Geologische Rundschau 76/2 | 599–631 | Stuttgart 1987

Multifactorial control of sedimentation patterns in an ocean marginal basin: the Lusitanian Basin (Portugal) during the Kimmeridgian and Tithonian

By Reinhold R. Leinfelder, Mainz *)

With 6 figures and 2 tables

Zusammenfassung

Das mesozoische Lusitanische Becken entstand als Teil des nordatlantischen Riftsystems. Während des Oberjuras lebte die tektonische Aktivität erneut auf und führte zu einer intensiven Differenzierung der faziellen Entwicklung. Die Kalke und Siliziklastika des Kimmeridge und Unteren Tithons, deren lithostratigraphische Beziehungen dargestellt werden, repräsentieren Becken-, Hang- und Flachwasserablagerungen sowie terrestrische Sedimente.

Synsedimentäre Tektonik bestimmte überwiegend die Ausbildung, Verteilung und Mächtigkeiten der Sedimente. Halokinese, eustatische Meeresspiegelschwankungen, exogene und biogene Faktoren kontrollierten die Faziesverteilung zusätzlich. Ein Vergleich der bathymetrischen Entwicklung aller größeren Beckenprofile und Standardisierung auf eine gemeinsame Zeitachse erlauben, die Überlagerung der einzelnen Kontrollfaktoren zu entschlüsseln sowie den Wert einiger biostratigraphischer Bezugshorizonte zu testen.

Während des Kimmeridge wurde die Paläogeographie vor allem durch starke Subsidenz des Beckenzentrums sowie durch eine große lineare Hebungszone im Norden und durch hohe klastische Zufuhr bestimmt. Am Ende des Kimmeridge und während des Tithons verlangsamte sich die Subsidenz. Anhebung innerer Beckenteile und schwankende Zufuhr von Klastika bewirkten die episodische Ausbreitung von Flachwasserkarbonaten und die verbreitete Entwicklung gemischt kalkig-klastischer Serien. Abschirmungs- und Abfangeffekte erlaubten kleinräumige Faziesvariationen. Zur Kreide hin verlandete das Becken durch klastische Zufüllung aus nördlicher, östlicher und vor allem nordwestlicher Richtung.

Abstract

The Mesozoic Lusitanian Basin developed as a part of the North Atlantic rift system. Tectonic rifting activity was rejuvenated during the Upper Jurassic, leading to intensive differentiation of facies development. Kimmeridgian and Lower Tithonian calcareous and siliciclastic sediments represent basinal and slope, shallow marine, and terrestrial environments. The lithostratigraphic arrangement of facies units is demonstrated. Sediment character, distribution and thicknesses are mainly controlled by synsedimentary faulting, with a partial overprint by uprise of salt diapirs. Eustatic sea level fluctuations, exogenic and biogenic factors resulted in additional control on facies development.

Comparing bathymetric development of major basin sections and simplified plotting on a common time scale is a simple tool to unravel the multifactorial control of sedimentation and to test the validity of some biostratigraphic markers.

During the Kimmeridgian, paleogeography was mainly determined by intensive subsidence of the basin center, by a large linear uplift zone in the north, and by a high amount of clastic influx. At the end of the stage and during the Tithonian, overall subsidence slowed down and inner basin uplifts arose further south. Degree of clastic input was variable. Thus shallow water carbonates were episodically widespread throughout the basin and mixed calcareous-clastic sequences were common. Sheltering and trapping effects resulted in local facies variations. Towards the Cretaceous the basin sanded up from northern, eastern and, particularly, northwestern directions.

Resumo

Durante o Mesozóico, a Bacia Lusitânica desenvolveu-se como parte do sistema »rift« do Atlântico do Norte. A actividade tectónica, tipo »rifting« renasceu durante o Jurásico Superior, causando uma diferenciação intensa no desenvolvimento de fácies. Os sedimentos calcários e siliciclásticos do Kimeridgiano e Titoniano inferior representam ambientes do mar mais ou menos profundo (fundo de bacia, declive, lagoa, delta) e ambientes continentais. O quadro litoestratigráfico das unidades de fácies é elaborado.

O carácter, a distribuição e a espessura dos sedimentos são sobretudo controlados pela actividade tectónica sinsedimentária. Movimentos halokinéticos, fluctuações eustáticas do nível do mar, e factores exogénicos e biológicos resultaram num controle adicional do desenvolvimento de fácies.

Uma medida simples para destrinçar os factores diferentes da sedimentação e para testar o valor de alguns »markers« bioestratigráficos é comparar o desenvolvimento batimêtrico dos cortes principais da bacia e estandardizá-los num comum eixo temporal.

*) Author's address: Dr. R. R. Leinfelder, Institut für Geowissenschaften, Johannes Gutenberg-Universität, Saarstr. 21, D-6500 Mainz.

Durante o Kimeridgiano, a paleogeografia foi dominada pela subsidência intensa do centro da bacia, por um grande levantamento estreito no Norte, e por um grau elevado de introdução de clásticos. No fim do estágio e durante o Titoniano, a subsidência geral diminuiu-se e novos levantamentos surgiram mais no Sul. O grau de introdução de clásticos foi variável. Por consequência, calcários de agua pouco profunda alargaram-se por vezes sobre grandes partes da bacia e sequências mistas de calcários e clásticos foram comum.

No fim do Jurásico o mar desapareceu por causa de enchimento da bacia por clásticos de proveniência norte, este, e partialmente noroeste.

Краткое содержание

Мезозойский бассейн Лузитании образовался как часть северо-атлантической системы рифтов. В верхней юре тектоническая активность оживилась и привела к дифференциации фациального развития. Отложения на дне бассейна, склонах, в мелководье, а также материковые седименты представлены известняками и кремнисто-обломочными составляющими киммериджского века и нижнего титона; описываются их литостратиграфические взаимоотношения.

Образование, распределение и мощность седиментов предопределялись синседиментными тектоническими процессами. Кроме того, соляная тектоника, эвстатические колебания уровня моря, экзогенные и биогенные факторы регулировали распределение фаций. На основании батиметрических сравнений профилей больших бассейнов и стандартизации их на общей временной оси удалось расшифровать наложения отдельных факторов и проверить значение некоторых биостратиграфических опорных горизонтов.

Палеогеография во время киммериджского века определялась прежде всего сильным опусканием центра бассейна, а также большим линейным поднятием на севере и приносом кластического материала. В конце киммериджского и в начале титонского веков проседание замедлилось. Поднятие внутренних частей бассейнов и колебания приноса кластического материала привели к временному отложению мелководных карбонатов, а также расширило развитие смешанных известково-кластических серий. В результате эффекта «защиты» и «перехвата» появились небольшие изменения фаций локального значения. В связи с усилением приноса кластического материала с севера, востока, и, прежде всего, с северо-запада, бассейн обмелел к началу мелового периода и превратился в сушу.

1. Introduction

The West Portuguese hill-land is mainly formed by the Mesozoic-Cenozoic sedimentary filling of the so-called Lusitanian Basin which developed along the western margin of the Iberian Hercynian Massiv.

Being of interest in terms of faunal and sedimentary development, stratigraphy, geotectonic setting and petroleum exploration, it attracted scientists already since last century (e.g., Sharpe 1850, for further references see below). In 1975 its development was, for the first time, seen under the light of the Atlantic opening (R. C. L. Wilson 1975a). Particularly the highly differentiated and partly fauna-rich Upper Jurassic sediments were since the last decade both matter of detailed case studies (e.g., Fürsich & Schmidt-Kittler 1980, Fürsich 1981, Fürsich & Werner 1986, Werner 1986) and more generalizing paleogeographic reconstruction attempts (Mouterde et al. 1979, for the entire Jurassic; R. C. L. Wilson 1979, for the Upper Jurassic). Complete basin analysis is, however, hampered by stratigraphic complications, caused mainly by the rarity of biostratigraphic markers and intense regional differences in facies development as a result of sediment pattern differentiation in an active ocean marginal basin.

The scope of this paper is to give a three-dimensional overview of the sedimentary and paleogeographic development during the Kimmeridgian and Tithonian, and to discuss control of sedimentation patterns by biologic and exogenic parameters, by local and regional tectonics, and by global changes of the eustatic sea-level.

2. Present-day state of the Lusitanian Basin

Not considering the Neogene cover, the basin is nowadays a fault-bounded half graben to graben structure. Direct sedimentary contact between the basement and the Mesozoic sequence along a major unconformity is exposed at several places on the eastern border of the basin. This border roughly follows the direction of basement elements (e.g., Teixeira & Gonçalves 1980). Rests of a western border are represented by the basement rocks of the Berlengas and Farilhoes Islands. The main part of the western limit is, though, only evident by local uplifts or by a diminished and outwedging sediment sequence in the offshore region (e.g., Vanney & Mougenot 1981).

3. Survey of basin development

The Lusitanian Basin developed as an assymetric graben basin along reactivated late Hercynian faults (e.g., R. C. L. Wilson 1975a). Dimension, assymetry and direction of the basin, and development

Fig. 1. Location of major sites mentioned in the text. Abbr.: A Alenquer, Arr Serra da Arrábida; Col Columbeira, Ms Monsanto, R Ramalhal, TV/M Torres Vedras/Matacães, V Vimeiro

phases allow classification as a tensional rifted margin basin *sensu* WILSON & WILLIAMS (1979). All three evolutionary model stages of an orthogonal divergent continental margin basin *sensu* MIALL (1984, tab. 9.3) may be recognized. Volcanism is, however, not exclusively related to phase 1. Certain differences arise also in the last stage of basin development, since tectonic activity reappears again:

The first stage, the rifted arch and rim basin phase, is characterized by terrestrial sediments, the lower part of the Gres de Silves of Upper Triassic age (PALAIN 1976), in which the first post-Hercynian volcanics are intercalated (cf. RIBEIRO et al. 1979).

In contrast with the characteristics of the model stage 1, volcanism in the Lusitanian Basin plays only a minor role at that time. An explanation might be that rifting in Portugal is eventually related to the dominance of red sea type vertical movements rather than extensional stress (cf. HUTCHINSON & ENGELS 1972; see also SCHMIDT 1984). Considering the fact that the Lusitanian rift was until the Cretaceous not accompanied by a westward lying spreading zone (DEWEY 1973, BIJU-DUVAL et al. 1977), it becomes clear that the basin belonged to a narrow but intensively subsiding rift system. Rifting possibly propagated from the Seabight Basin, west of Ireland, southwards and splitted off at the Biscay rift, to form an arm between Iberia and the Grand Banks (GRAU et al. 1973, R. C. L. WILSON 1975a, L. M. WILSON 1980, ZIEGLER 1982).

During the Triassic and Lower Jurassic broad-scale regional extension and wrenching in the North Atlantic and Tethyan domain of Pangaea was accompanied by only a moderate level of thermic activity (P. A. ZIEGLER, pers. commun.). As such, the Lusitanian rift cannot be related to local mantle-plume activity (»hot spots«) west and southwest of the Iberian peninsula as postulated by R. C. L. WILSON (1975a) and KRISTOFFERSON (1978), but rather to the geothermal development of deepseated mantle convectional systems (P. A. ZIEGLER, pers. comm.).

MIALL's second evolutionary stage is realized by a Hettangian fine-grained clastic through marly to dolomitic, lagoonal sequence containing thick evaporites, i.e. the upper part of the Gres de Silves and the Dagorda beds (FISCHER & PALAIN 1979, RIBEIRO et al. 1979, JANSA et al. 1980). Superimposed is the Sinemurian–Aalenian sequence which is differentiated into dysaerobic, radiolarian-bearing shales and periplatform oozes in the basin center, and marine siliciclastics, carbonate fans and dolomites along the basin margins (HALLAM 1971a, WRIGHT & WILSON 1984). Fans and rapid transition from shallow water to deep water facies evidence still active synsedimentary faulting.

Synsedimentary tectonic basin margin differentiation fades during the Middle Jurassic and thus opens the third evolutionary stage *sensu* MIALL (1984). Consequently, silicilastic input and differentiation into deeper and shallower parts diminishes. During the Bathonian and Callovian vast shallow-water carbonate deposits developed on costs of shrinking and shallowing basinal areas with marl/limestone sedimentation (e.g., RUGET-PERROT 1961, RIBEIRO et al. 1979). Off Portugal, however, pelagic sedimentation persisted (P. A. ZIEGLER, pers. comm.).

Inner basin tectonics together with the start of halokinetic movements led, though, to local Callovian non-deposition, erosion and angular bedding unconfirmities (e.g., ZBYSZEWSKI & MOITINHO 1960, RUGET-PERROT 1961, SEIFERT 1963) and associated differentiation into smaller subbasins. The medium to upper part of the Callovian is clearly regressive; the final Callovian is generally lacking (RUGET-PERROT 1961, MOUTERDE et al. 1979).

Subaerial exposure was widespread during the early Oxfordian, not only in large parts of the Lusitanian Basin (for ref. see below) but also in Northern Spain (BENKE 1981) and Western France (ENAY et al. 1982).

According to BOILLOT et al. (1975, 1979), BALDY et al. (1977) and MAUFRET et al. (1978), the Atlantic Ocean opened in two cycles, the first from the Hettangian to the Callovian, the second from the Upper Jurassic to Neogene. Each cycle started with a phase of intensive rifting, followed by a transgressive period due to the stop of thermic activity and related sinking of the rift margins. The subaerial exposure of parts of Western Europe lies at the beginning of the second rifting phase which durated from the Upper Jurassic to the Cenomanian (op.cit). Rifting pulses at the base of the Upper Jurassic are also obvious in the Biscay rift and the North sea rifts (ZIEGLER 1982) and are most probably related to the start of crustal separation in the Central Atlantic (P. A. ZIEGLER pers. comm.).

It should be noted that this presumed rifting phase did not lead to regressions at the beginning of the Upper Jurassic in other North Atlantic marginal basins (cf. HALLAM 1971b). On the contrary, a major transgression (»Atlantic transgression«) reflects the maturation of the southern North Atlantic into a real ocean (LANCELOT & WINTERER 1980). However, in the Grand Banks, most neighboured to the Lusitanian Basin during the Mesozoic, Oxfordian sediments are partially missing or exhibit brackish character (GRADSTEIN 1979, fig. 2).

R. C. L. WILSON (1979) also considers the Lower and Middle Jurassic situation in the Lusitanian Basin as an intracratonic calm phase following a Triassic-lowermost Liassic rift phase and being itself followed by an inversion phase during the Oxfordian (R. C. L. WILSON 1975a) and the main rift phase during the Kimmeridgian. The latter is characterized by active faulting, shifting of basin center, increased uprise of salt diapirs and subsequent facies differentiation into basinal shales, shallow water carbonate platforms and deltaic to terrestrial clastic sediments (cf. R. C. L. WILSON 1979 and below). Motor for this strong tectonic activity was possibly rapid spreading in the Central Atlantic and related rotation of Africa, resulting in now northward propagation of the West Iberian rift (P. A. ZIEGLER, pers. comm.). This coincides with general shallowing of the Lusitanian Basin towards north (see fig. 6) which was not the case during the Lower and Middle Jurassic (cf. MOUTERDE et al. 1979, figs. 1–9).

During the Cretaceous a mixed clastic-carbonate development persisted, but the Lusitanian Basin was continuously loosing its individuality due to the gradual disappearance of the accentuated western border (cf. REY 1979, BERTHOU & LAUVERJAT 1979), thus turning into a simple passive margin onshore province.

4. The Upper Jurassic sequence

4.1 The Upper Oxfordian

The mentioned Lower to Middle Oxfordian hiatus is commonly marked by erosive features in many parts of the Lusitanian Basin, including karstification, calcrete formation, development of pisolitic ironstones, reworking and local angular bedding unconfirmities (OERTEL 1956, RUGET-PERROT 1961, R. C. L. WILSON 1979, FELBER et al. 1982, LEINFELDER 1983). The Upper Oxfordian sediments overlying this unconformity are extremely variable, ranging from lignitic freshwater marls through algal carbonates to coral-bearing limestones (for review and references see R. C. L. WILSON 1979). In the extreme southeast and in the northwest of the basin red siliciclastics appear locally (op.cit.; LEINFELDER 1983).

In terms of facies aspects better studied show cases are e.g., the algal marsh to lagoonal sequence at Cabo Mondego (R. C. L. WILSON 1975b, 1979, WRIGHT 1985) and the Montejunto area, characterized by a rapid transition from deep water through allochtonous carbonates to shallow water limestones (R. C. L. WILSON group, in prep.).

4.2 The Kimmeridgian–Tithonian

Facies differentiation at local and regional scale is even more pronounced during the Kimmeridgian and Lower Tithonian (thus Lower Portlandian *sensu gallico* (fig. 2). Depositional environments range from terrestrial and deltaic to shallow bank, lagoonal, slope and even basinal environments. The late Tithonian, on the other hand, shows a more uniform development.

A brief account of the main sedimentary units is given below. Hierarchic classification (groups, for-

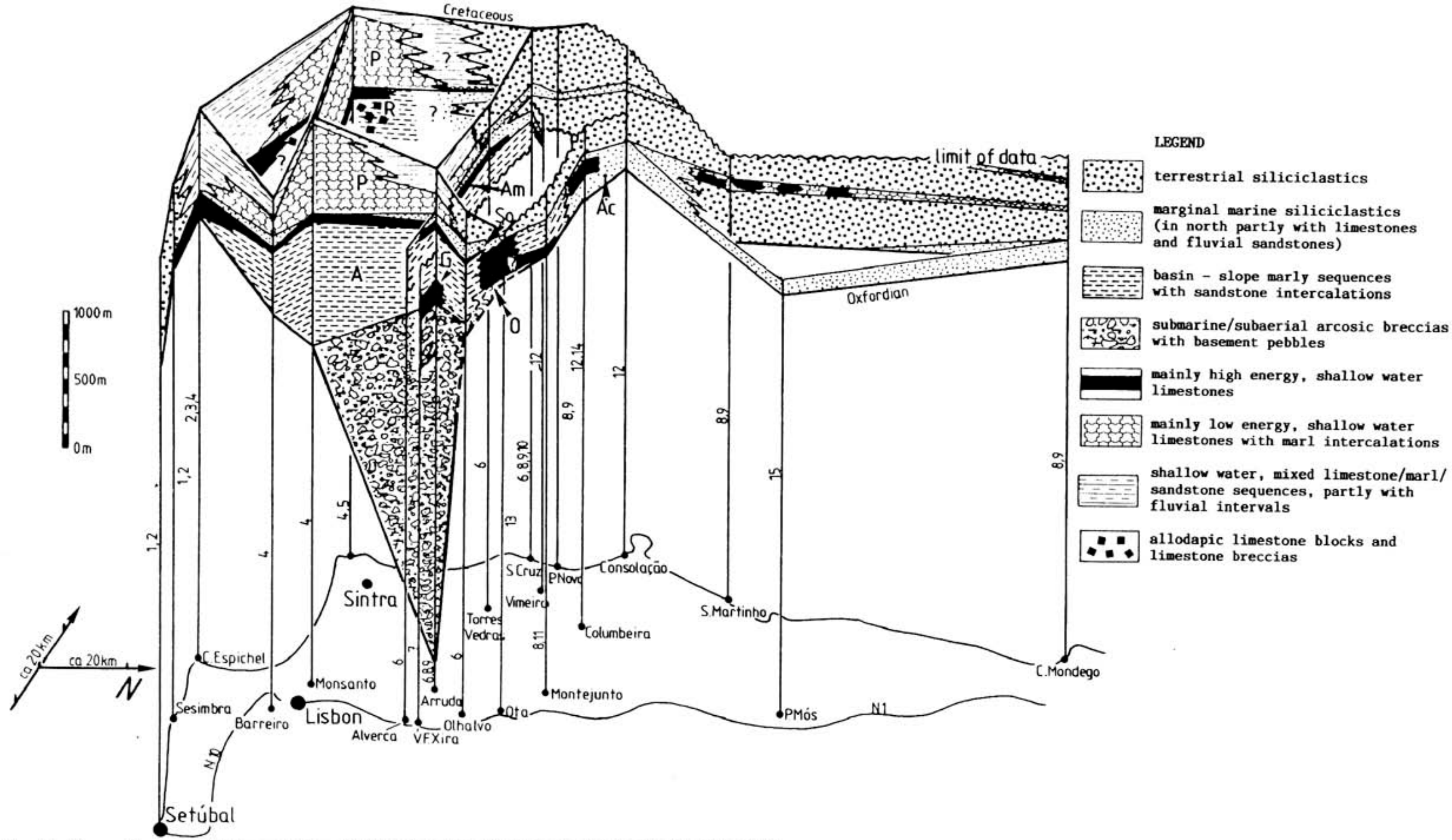


Fig. 2. Fence diagram of Kimmeridgian/Tithonian main facies units in the Lusitanian Basin.

A Abadia marls, Ac Alcobaça beds, Am Amaral limestone, G Monte Gordo limestone, O Ota limestone, P »Pteroceriano« beds, R Ramalhão shales, So Sobral beds.

Compilation based on evaluation of Leinfelder 1983 (1), Felber et al. 1982 (2), Fürsich & Schmidt-Kittler 1980 (3), Ramalho 1971 (4), Ellis 1984 (5), Leinfelder 1986 (6), Zbyszewski & Torre de Assunçao 1965 (7), R. C. L. Wilson, pers. commun. (8), R. C. L. Wilson 1979 (9), Fürsich 1981 (10), Zbyszewski et al. 1966 (11), Werner 1986 (12), Leinfelder et al. in prep. (13), Ruget-Perrot 1961 (14), Schmidt, 1984, 1985 (15), and further personal observations.

Horizontal correlation level is base of »Pterioceriano« beds and equivalent strata.

mations, members, etc.) and stratigraphic local names are mostly avoided except for well established items, since they are not used in the same way by all authors and are not essential for the purpose of this study.

Ota limestone complex

The Ota limestone complex, situated in the eastern part of the basin, undergoes intensive facies and biostratigraphic treatment at the moment (LEINFELDER, ERBENICH & RAMALHO in prep.). The complex consists of a wide variety of very pure shallow water limestones, including coral boundstones, intraclastic grainstones, loferites, black pebble packstones and various lagoonal low energy micritic types. Its isolated existence, tectonic limitation and lack of ammonites prohibited up to now any age determination. Recent finds of suited microfauna enable, however, to place large parts of the limestones into the Upper Kimmeridgian, although there are hints that Lower Kimmeridgian and even Upper Oxfordian might locally be included as well. Astonishing are facies similarities with the shallow water development of the Upper Oxfordian both in the Montejunto area and in the eastern Arrábida region (cf. LEINFELDER 1983).

Monte Gordo limestone

A small, isolated, reefal to lithoclastic limestone lentil, containing black pebbles and even detrital feldspar grains, exists closeby Vila Franca de Xira, further south, in about the same level as supposed for the Ota limestone. Not only did the sediment body grow on top of arcosic siliciclastics containing reworked basement pebbles, but it also passes laterally into sandy and marly deposits (ANDRADE 1934, ZBYSZEWSKI & TORRE DE ASSUNÇÃO 1965).

Amaral coral limestone

A thin sheet of coral limestones and oolites (10–50 m) surrounds the valley of Arruda dos Vinhos and parallels the Montejunto elevation. The Amaral limestone was described by DÖHLER (1984) and may be interpreted as lateral spreading of shallow water carbonates topping a marine siliciclastic shallowing sequence (see below).

Limestone development on and along salt diapirs

Marine shallow water, mostly high energy, rarely paralic, coal-bearing limestones developed during the Kimmeridgian on the crest of and along diapiric zones, particularly in the region of the Caldas da Rainha diapir (R. C. L. WILSON 1979, FÜRSICH & WERNER in prep.) and the Porto de Mós diapir (SCHMIDT 1984, 1985). Most common facies types are oolites, oncolites, biointrasparites and reefoid types containing corals, stromatoporoids and algae. Sediments may exhibit a certain degree of siliciclastic contamination.

Limestones, limestone/marl and marl/shale sequences around Sintra

RAMALHO (1971) gave the first detailed descriptions of the Kimmeridgian–Tithonian coastal sequences closeby Sintra. The Ramalhão shales are overlain by the marly Mem Martins limestones which are themselves superimposed by coral and oncolite-bearing calcareous deposits and by nodular limestones (fig. 3).

The sequence was reconsidered by R. C. L. WILSON (1979) and ELLIS (1984). It is interpreted by the latter author as a ramp development, characterized by local patch reefs and reefs, and by sporadic appearance of debris flows and turbidite clouds which were eventually caused by earthquakes.

Limestone/marl sequences in the northern basin center

A highly intercalatory sequence of marls, marly limestones, nodular limestones and coral biostromes around Arruda dos Vinhos, known as »Pteroceriano« beds (Lower Tithonian pp), is extensively treated in LEINFELDER (1986). The notion »Pteroceriano«, although still currently used in the regional literature is about to be abolished in a general stratigraphic reclassification because of incorrect mixing-up of lithostratigraphic and biostratigraphic meanings.

The lower part represents tranquil lagoonal shallow water environments below wave base, characterized by partly decreased salinity values and by occasional perturbances due to storms. Coral banks developed sporadically in more exposed areas and extended across the entire lagoon during the upper part of the »Pteroceriano« sedimentation (fig. 3).

Limestone/marl/sandstone intercalatory sequences

The late Lower to Upper Tithonian Freixial beds, a restricted lagoonal sequence of micritic and sparitic shallow water limestones, marls, marginal marine and terrestrial siliciclastics superimposes the »Pteroceriano« beds south of Arruda dos Vinhos (RAMALHO 1971, LEINFELDER 1986).

In the western part of the Serra da Arrábida, the Kimmeridgian and Tithonian consists of intraclastic, fenestral, bioclastic, coral-bearing and nodular shallow water limestones in which marls and siliciclastics soon become intercalated (RAMALHO 1971, FÜRSICH & SCHMIDT-KITTLER 1980, FELBER et al. 1982, LEINFELDER 1983) (fig. 3).

In the north of the basin, the Kimmeridgian basinal Abadia beds (see below) are partly substituted by the Alcobaça beds, a shallow marine, fauna-rich, sandy–marly sequence, containing intercalations of shallow water carbonates (e.g., RUGET-PERROT 1961, R. C. L. WILSON 1979, WERNER 1986, FÜRSICH & WERNER in prep.).

Marine marl/sandstone/breccia sequences between Sta. Cruz and Vila Franca de Xira

The most widespread facies in the center of the Lusitanian Basin during the Lower and Middle Kimmeridgian was a marly, ammonite-bearing, basinal sequence with interbedded, partly turbiditic limestones, the Abadia marls (e.g., R. C. L. WILSON 1979). Slump scar development and submarine canyon fillings consisting of basement breccias suggest also prograding slope environments, an interpretation which is also backed by seismic methods (R. C. L. WILSON, pers. commun.). In the uppermost part, however, shallow water benthic bivalve, echinoderm and coral fauna is locally abundant.

This sequence is underlain and towards the east partly substituted by up to 2500 m of arcosic basement conglomerates and breccias, containing locally large blocks of tumbled coral boundstones and thin layers of coral-bearing wackestones (cf. ANDRADE 1934, MEMPEL 1955, ZBYSZEWSKI & TORRE DE ASSUNÇÃO 1965, R. C. L. WILSON 1979; pers. observ.).

Marginal marine to terrestrial sandstone and conglomerate sequences

The Amaral limestone in the central basin area is superimposed by the marly–sandy Sobral sequence of presumed Upper Kimmeridgian age, representing environments of an estuarine delta type (LEINFELDER 1986).

The prevalent sediment types in the north and west of the Lusitanian Basin during the Kimmeridgian and Tithonian were terrestrial fine to very coarse-grained, often red, siliciclastic deposits. At the base or in an intercalated position of these so-called gres superiores marine clastics might appear (R. C. L. WILSON 1979, LEINFELDER 1986). In the surroundings of Alenquer a thin sheet of freshwater limestones, the Alenquer oncolite, is embedded within the clastics (LEINFELDER 1985).

The sequence along the coast of Sta. Cruz, for instance, consists of six superimposed units (sandy braided fluvial, meandering fluvial with pronounced point bar development, deltaic to fluvial, cyclic gravelly braided fluvial, meandering fluvial) (R. C. L. WILSON 1979 and pers. commun.; cf. also FÜRSICH 1981) (fig. 3).

Another important clastic, mainly terrestrial section further north at Consolação covers mainly the entire Kimmeridgian. At its base it exhibits a 80 m thick horizon containing coral meadows (WERNER 1986) (fig. 3).

In the eastern part of the Serra da Arrábida, a terrestrial red siliciclastic sequence substitutes the marine limestone development further west (FELBER et al. 1982, LEINFELDER 1983).

5. Stratigraphic correlation

Hitherto, there does not exist a generally accepted stratigraphic classification scheme for the West Portuguese Upper Jurassic. Classification is hampered by rapid lateral facies changes, rarity or lack of biostratigraphic markers and by a flood of often confusingly used stratigraphic items which on top commonly mix up both lithostratigraphy and biostratigraphy. For stratigraphic and nomenclatoric classification attempts see, e.g., MOUTERDE et al. (1972, 1979), RIBEIRO et al. (1979), R. C. L. WILSON (1979), FÜRSICH & WERNER (1986) and LEINFELDER (1986). The advanced stage of basin analysis allows a general lithostratigraphic reclassification which at the moment is under cooperative preparation among the different working groups of the Lusitanian Basin. As mentioned earlier, only a selection of commonly used stratigraphic items is therefore used in this study.

Biostratigraphic markers

In the lower part of the sediment package few ammonites were found in basinal marls (Ramalhão and Abadia beds) between Lisbon and Torres Vedras, giving a Lower to late Middle Kimmeridgian age (sensu gallico) (e.g., RUGET-PERROT 1961, MOUTERDE et al. 1972, 1979). Further north, ammonites were only very rarely detected in the lowest part of the respective series, eventually indicating

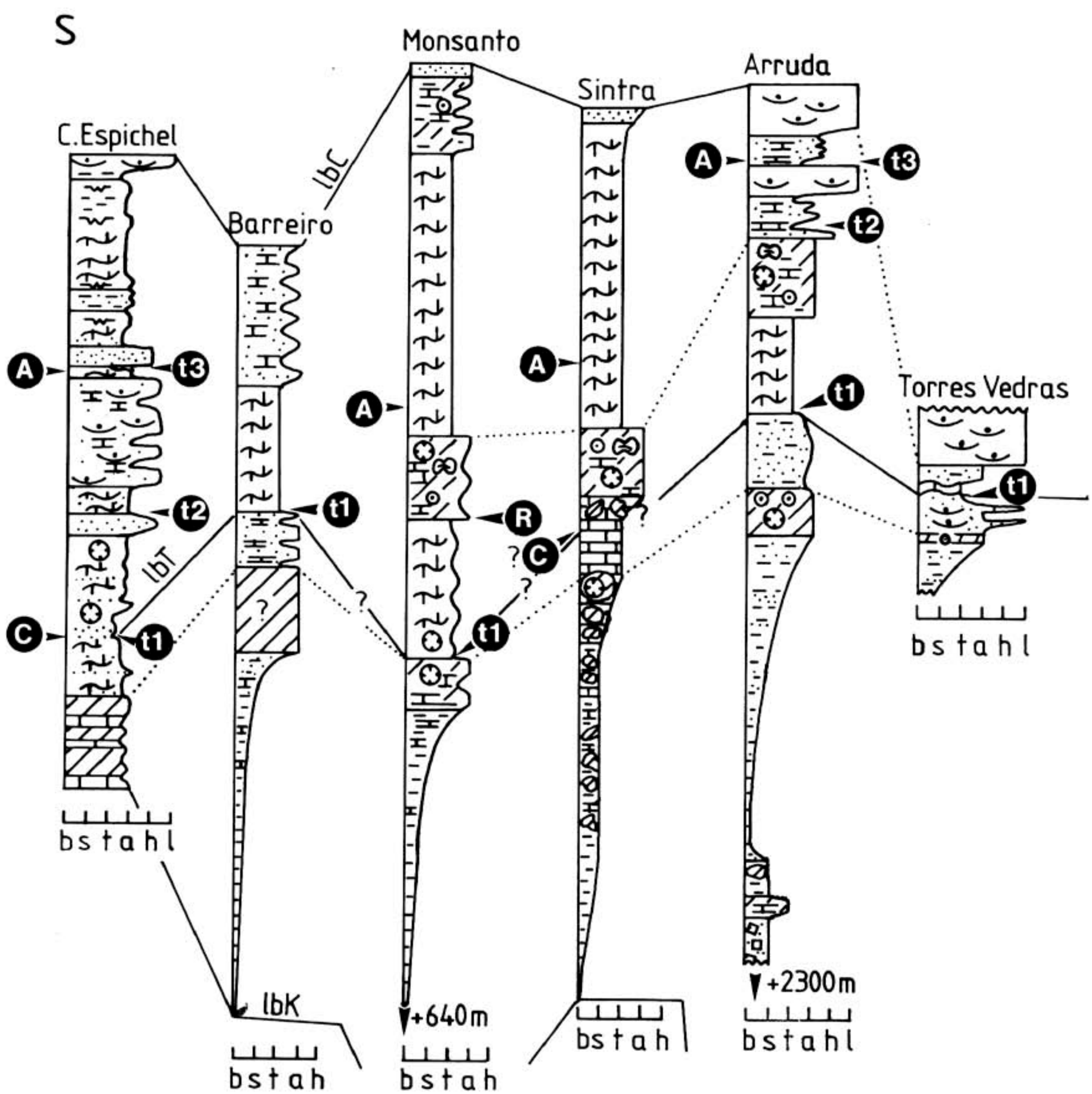


Fig. 3. Simplified principal Kimmeridgian/Tithonian sections of the Lusitanian Basin (after different authors, see references of fig. 2), and their relative bathymetric interpretation.

Abbrev.: b basin, s deeper slope, t tranquil shallow water (below fair weather wave base), a commonly agitated shallow water, h fluvial lowlands with high sinuosity rivers, l fluvial systems with low sinuosity rivers, lbK lower boundary of Kimmeridgian, lbT presumed lower boundary of Tithonian, lbC lower boundary of Cretaceous.

A: onset of *Anchispirocyclina lusitanica,* C: disappearance of *Clypeina jurassica,* R: Tithonian base of Monsanto log after RAMALHO (1971), t1, t2, t3: transgressive peaks, possibly related to VAIL et al. (1984) global eustatic highstands. t2, t3 may, however, also reflect local tectonic basin deepening.

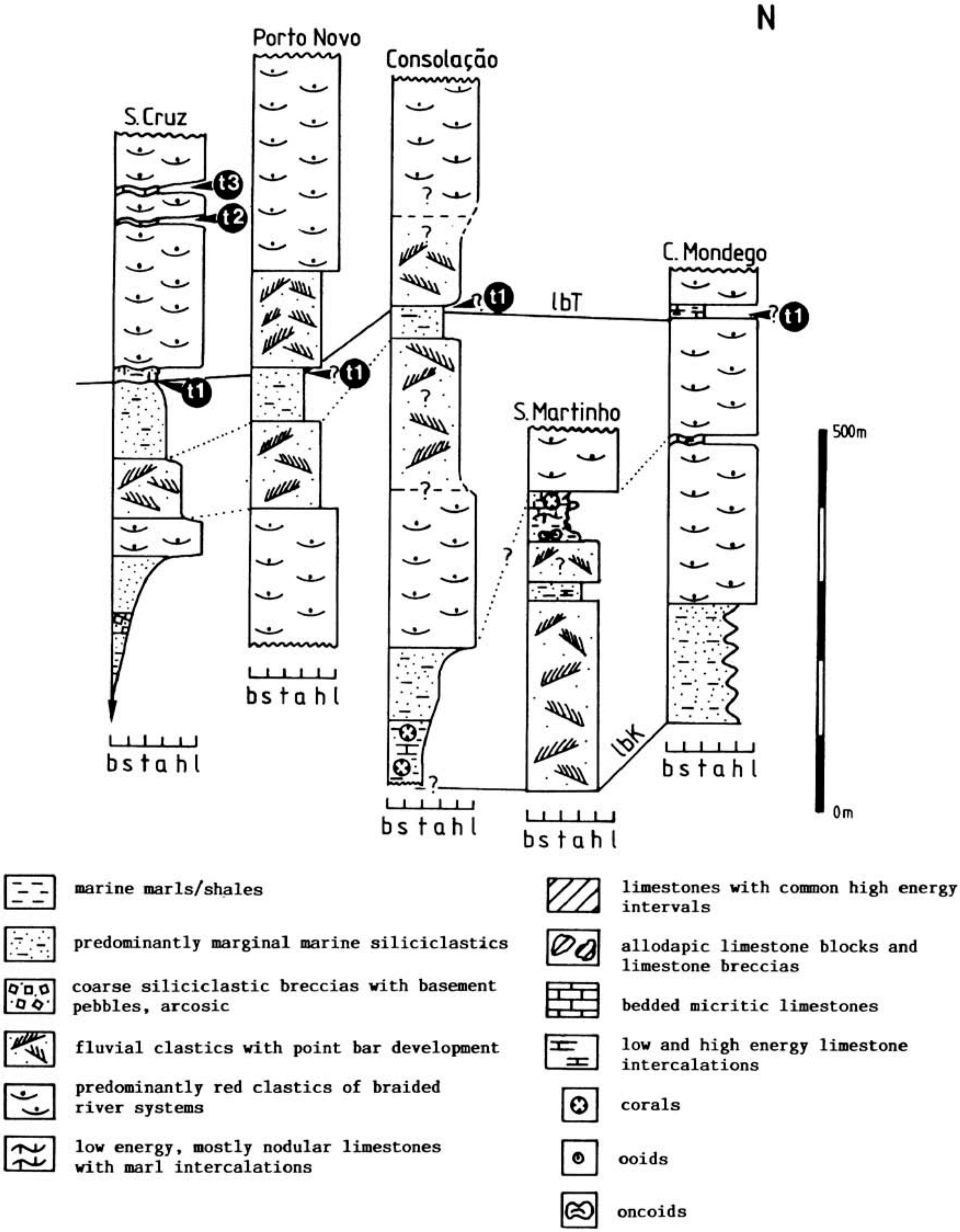
Dotted lines: facies correlation.

Nodular character of limestones hypothetic for Barreiro and Monsanto boreholes.

Location of sites, see figs. 1, 2.

Correlation after LEINFELDER (1986) and this paper.

equivalent ages (cf. LEINFELDER 1986, WERNER 1986). Elsewhere, a general Kimmeridgian age may be proved by foraminifers which in some regions may be further subdivided by use of diagnostic algae or ostracods (e.g., RAMALHO 1971, 1981, FÜRSICH & SCHMIDT-KITTLER 1980, FÜRSICH 1981, FELBER et



al. 1982, LEINFELDER 1986, WERNER 1986, LEINFELDER et al. in prep.).

Stratigraphic separation from underlying Upper Oxfordian sediments is rarely clear. Normally, the start of siliciclastic sedimentation superimposing the calcareous sediments is taken for the base of the

Kimmeridgian. In the basin center this is partly proved by ammonites (MOUTERDE et al. 1972).

Biostratigraphic markers indicating the Tithonian are rarer still. Ammonites giving a lowermost Tithonian age are, to the present authors knowledge, up to now only recorded from the basin center around Sintra and from the western Serra da Arrábida, further south (cf. RAMALHO 1971, MOUTERDE et al. 1972, 1979). By micropaleontological means, only the late Tithonian may be separated in the marine parts of the basin (RAMALHO 1971, FÜRSICH & SCHMIDT-KITTLER 1980, FELBER et al. 1982), although there exist preliminary results for a further differentiation of the lower part of the Tithonian (cf. RAMALHO 1981, LEINFELDER 1986).

Some implications on biostratigraphy result from application of approximated geohistory analysis (see chap. 10).

General lithostratigraphic situation

The general lithostratigraphic framework of sedimentary units is shown in a fence diagram in order to illustrate the geometric arrangement of the Kimmeridgian–Tithonian part of basin filling (fig. 2).

Except for the calcareous development in the Serra da Arrábida, the lower part of the sequence exhibits siliciclastic facies, ranging from very coarse-grained arcosic conglomerates to very fine-grained shales and marls (Abadia– Alcobaça level). The development in the north is predominantly terrestrial, passing to marine sediments towards south.

In the central to southern part of the basin, partly reefal limestones are superimposed which around Arruda dos Vinhos form a thin, more or less connected band (Amaral coral limestone). Correlation further south seems to rectify a sheet-like development. Further north, however, the limestones exhibit very variable thicknesses, seem to be often disconnected and apparently appear in different levels. A thin horizon of nodular, coral-bearing, biomicritic limestones and cyanophyte–red algal oncolites at S. Martinho was correlated by R. C. L. WILSON (1979) to the same unit. Although locally dated as Upper Kimmeridgian by algae (RAMALHO 1971, DÖHLER 1984), the horizon is not suited to form a good correlation level. Limestone bodies are not always connected to each other and are not developed in the west of the basin (Sta. Cruz–Consolação). Moreover, limestone development was certainly not synchronous (see chap. 10).

Except for the basin center between Lisbon (log Monsanto) and Sintra, and the Alverca region (Tejo valley), this level is again overlain by an, often only thin, horizon of siliciclastics which in the Serra da Arrábida may be partly substituted by red fluvial limestone arenites (LEINFELDER 1986).

Superimposed is a sequence of nodular limestones and marls in the southern and central parts of the basin (»Pteroceriano« beds) whose base is developed very widespread, reaching northwards in calcareous facies until a line Alenquer–Torres Vedras–Sta. Cruz (op.cit.). North of this line the limestones are substituted by a thin marine, rarely brackish, marly-clastic unit, characterized particularly by bivalve shell beds.

Overlying is a very mixed sequence, dominated by terrestrial coarse siliciclastics in the north and by a marine limestone/marl series in the south with variable marl contents and sandstone intercalations which generally augment upwards (Freixial beds p.p.). Further internal correlation is possible only for limited parts of the basin (cf. LEINFELDER 1986).

The main stratigraphic marker level: base of the »Pteroceriano« beds and corresponding levels

A perfect stratigraphic marker level should fulfil the following requirements:

- coverage of the entire basin
- laterally connected level
- little facies variability
- easy to recognize in the field
- synchronous development

Although there is no doubt about their usefulness and need, mere biostratigraphic markers may, however, have the disadvantage to be only recognized by the specialist, to contain fauna too laterally scattered to form a continuous marker sheet, or to only cover smaller parts of an entire basin, which is all true for the here treated sequence.

Storm beds are shown by AIGNER (1985) to represent almost perfect markers for the Swabian Muschelkalk development, covering most parts of the basin in continuous, perfectly synchronous sheets. In distal parts, however, facies differences may make identification difficult. Moreover, different storm layers may be confused with each other when outcrops over the basins are scattered. There exist storm-induced shell beds in the Lusitanian Basin which, though, have only very local value for correlating neighboured sections (LEINFELDER 1986).

The best stratigraphic marker in the studied sequence is the base of the calcareous/marly »Pteroceriano« beds and its lateral substitutes, since it fulfils most of the above mentioned criterions: As already outlined, the base of the »Pteroceriano« beds forms a laterally connected level and covers large parts of the basin, even in parts where sedimentation is otherwise completely performed by clastics. At the moment the horizon is mapped in the Torres Vedras–Arruda dos Vinhos region (e.g., STACHEL 1986), since the calcareous »Pteroceriano« sedimentation is not differentiated in official maps. Further south towards Lisbon, Sintra and the Serra da Arábida the level can be mostly connected to borehole and outcrop descriptions given by RAMALHO (1971) (cf. LEINFELDER 1986). Only in the eastern part of the Serra da Arábida there is no corresponding lithostratigraphic horizon subdividing the entirely clastic Kimmeridgian-Tithonian series.

At the coast north of Sta. Cruz there do not exist limestone horizons in the clastic Kimmeridgian-Tithonian development. There appears, however, a brachyhaline to brackish, marly-clastic level at Consolação, which is over and underlain by thick series of terrestrial clastics (WERNER 1986). This level reappears further south, at Porto Novo, exhibiting more marine fauna and containing a limestone conglomerate which apparently not only consists of eroded caliche crusts but also of reworked marine limestones in »Pteroceriano« facies. This layer can be easily connected to the limestone-containing interval at Sta. Cruz, a few kilometers further south. Nodular limestones appear here in the top part of a ca. 80 m thick clastic sequence dominated by fauna-rich marine intervals (FÜRSICH 1981). A very thin level of marly limestones and marls, intercalated in a fluvial series at Cabo Mondego in the north (R. C. L. WILSON 1979) belongs most likely to the same level. Direct tracing is, however, not possible due to outcrop conditions and structural situation.

The base of the »Pteroceriano« beds, including the thin limestones vestiges mentioned, is easy to recognize in the field due to their characteristic lithology and fossil content (nodular or marly micritic limestones with characteristic bivalve and trace fossil fauna), due to its normally sharp lower boundary and due to a sudden change from sandstone to limestone deposits in most areas (cf. LEINFELDER 1986). The level only suffers facies differentiation between Sta. Cruz and Consolação where limestones disappear. There the marker horizon can be easily persecuted by using the top of the mentioned marine level intercalated in the terrestrial sequence. There does not appear a corresponding marine level in the upper part of the S. Martinho sequence because of structural reasons.

Most difficult to consider is whether or not the marker level developed synchronously. It is dated at Cabo Espichel and Sintra where the detected lowermost Tithonian ammonites were found at the base of the »Pteroceriano« beds. Since there do not exist further finds, it still has to be discussed whether this dating can be extrapolated to other sites:

The mentioned marine interval of the Sta. Cruz section and the corresponding level in the Consolação section could be roughly dated as Upper Kimmeridgian/Lower Tithonian by means of a diagnostic ostracod (WERNER pers. commun. in FÜRSICH 1981, WERNER 1986), thus being too inaccurate to aid in the consideration of diachronism.

The base of the »Pteroceriano« beds is clearly the result of a transgression which CHAUMEAU (1962) considers to have been rapid, since maximum extension of the beds is already realized at their base. The rather thin marine interval of the northern part exhibits roughly similar thickness, also indicating a fairly rapid transgression/regression for the respective layers which should show much more pronounced increase in thickness towards the basin center if the responsible transgression would have been very slow and thus diachronous. The intercalated limestone levels should reflect times where clastic contamination sank below a critical value, allowing rapid, synchronous extension over the flooded area. Thus the base of the »Pteroceriano« limestone interval may be considered as a predominantly synchronous event, particularly where it is overlying marine clastics. A further possible hint for synchronism is the occasional appearance of a gigantic heterodont bivalve which already died out soon after its first appearance (FÜRSICH, pers. commun.; cf. LEINFELDER 1986). Luckily, this correlation level corresponds roughly to the Kimmeridgian/Tithonian boundary, thus increasing its importance even more.

6. Eustatic control of sedimentation

Despite general lateral facies transitions, the considered sequence is often characterized by crude cyclic or periodic facies repetitions.

For instance, three marine horizons characterized by marginal marine clastics, marls and, partly, limestones are separated by terrestrial clastics at Cabo Mondego. At Consolação, terrestrial clastics are subdivided by two marine levels. Around Arruda the basal marine clastics are overlain by a level of reefal and oolitic limestones which is again superimposed by marine clastics and marly limestone beds. The latter unit itself shows cyclic patterns considering the repetitive limestone/marl interbeddings. In the same region as well as further south the upper part of the

Tithonian is characterized by repetitive limestone/marl/sandstone sequences.

Such cyclic sedimentation patterns or facies repetitions can be often interpreted as the result of eustatic sea level fluctuations (see DUFF et al. 1967 and many articles in EINSELE & SEILACHER 1982, BAYER & SEILACHER 1985). Particularly periodic bathymetric changes of the sedimentation base are matter of interpretation through global sea level changes.

There is no doubt that strong sea level changes took place during the Upper Jurassic (e.g., HALLAM 1984, VAIL et al. 1977, VAIL & TODD 1981, VAIL et al. 1984), although many sequences are too easily interpreted as being only a result of such fluctuations. Moreover, the methodology of establishing sea level curves solely from seismic analysis lately suffers thorough critizism (e.g., PITMAN & GOLOCHENKO 1983; for the Upper Jurassic BROWN 1984, HALLAM 1984, MIALL 1986).

Pre-Quarternary eustatic sea level fluctuations are mainly considered to result from mid-oceanic ridge activity. There exists, however, no agreement on whether different spreading rates and related sinking speed of the lithosphere accounts directly for changes in the sea level due to enlargement or diminishing of ocean volume (e.g., VAIL et al. 1977, PITMAN 1978), or whether spreading speed is only an expression for mantle flow-induced global tectonics which is then reflected in sea level changes (e.g., SCHWAN 1980). Unconformities, one key tool to identify sea level fluctuations (cf. VAIL et al. 1977) may, however, also express regional tectonic pulses, particularly in passive margin provinces (LEPICHON 1980). More widespread unconformities may even be the product of globally synchronous episodes of cratonic uplift and not connected to sea floor spreading-induced eustatic changes (SLOSS 1984).

There is also no agreement on the general behaviour of the global sea level during the Kimmeridgian and Tithonian: HALLAM (1975, 1978) assumed a pronounced slow drop of the sea level, extending from the Jurassic highstand in the late Oxfordian until the end of the system. VAIL et al. (1977) gave two different curves: The curve of relative changes in global sea level, based on changes in coastal onlap, indicates a general rise during the entire Kimmeridgian, a sudden drop at the Kimmeridgian/Tithonian boundary, and continuing rise during the Tithonian and Berriasian, exceeding in extend the Callovian-Oxfordian transgression. The other estimate (op.cit., p. 91), based on mathematically calculated sea level changes, indicates that sea level was rising only during the Kimmeridgian, whereas in the Tithonian a slow drop was initiated. VAIL & TODD (1981) introduced a rapid, heavy drop around the mid-Kimmeridgian and three short-termed transgressive/regressive peaks, two of which lie within the Upper Kimmeridgian and one at the base of the Tithonian. The authors concluded an eustatic rise for the rest of the Tithonian.

For the Middle and Upper Jurassic HALLAM (1984) partly followed the calculated curve of VAIL et al. (1977). Except for the upper part of the Tithonian he gave a generally transgressive curve which is only modified by small-scale, short-termed drops. HALLAM also considered the Tithonian part of the Jurassic eustatic curve as very difficult to establish due to biostratigraphic problems, but reckoned a transgressive eustatic pulse at the base of this stage. Evidence would come from the Andes and from northwest Europe and western Siberia.

VAIL et al. (1984) published a revised global sea level curve, based on new biostratigraphic data which led to new age determinations for many Jurassic unconformities. They also transformed their sawtooth onlap curves, formerly understood as directly mirroring respective sea level changes, into a modified curve of global eustatic fluctuations. The resulting curve exhibits a slow drop during the entire Kimmeridgian, three transgressive/regressive peaks during the lower and middle Tithonian and a general transgressive character for the rest of the Tithonian.

The peaks seem to correspond to the Upper Kimmeridgian/Lower Tithonian peaks of VAIL & TODD (1981) which HALLAM (1984) considered as possible northwest European tectonic phenomenons, and thus not valid for the global curve. MITCHUM & ULIANA (1985), however, see the uppermost Kimmeridgian and the three intra-Tithonian unconformities realized in a shale–carbonate sequence in the Neuquén Basin of Argentina. This should point out the general importance of these unconfirmities which according to the above authors should correspond to global sea level lowstands.

GABILLY et al. (1985) dated the Jurassic unconformities of France by means of ammonite biostratigraphy and also interpreted the major ones as a result of global eustatic sea level changes. They report seven major intra-Kimmeridgian unconformities and one at the base of the Tithonian (below the *gigas* zone). The latter might correspond to a type 1 unconformity of VAIL et al. (1984), whereas the French intra-Kimmeridgian unconformities do not have global counterparts and might thus more likely correspond to West European tectonics (cf. chap. 3). Astonishingly, the outlined two further peaks of eventually global scale during the Tithonian did not find any depositional response in the French sedimentary record. This might be due to overcompensation of eustatic effects through high sedimentation rates.

In the Lusitanian Basin, a regressive Kimmeridgian sequence, as postulated by VAIL et al. (1984), can be recognized in many sections (fig. 3). Lowstands at the end of this stage are expressed by the shallow water carbonates of the Amaral coral limestone or by the regressive lower part of the partly superimposed Sobral sandstones (cf. LEINFELDER 1986). Fluvially deposited limestone arenites and lime conglomerates with a reddish, soil-produced matrix in the Serra da Arrábida (cf. FELBER et al. 1982, LEINFELDER 1983) could indicate the corresponding subaerial unconformity.

The following transgression, resulting in widespread deposition of »Pterioceriano« limestones at the base of the Tithonian could correspond to the lowest peak in the Tithonian part of the revised VAIL team curve (VAIL et al. 1984). This peak is situated 1 Mio years above the Tithonian base (corresponding to 1.3 Mio years above the base using KENT & GRADSTEIN's, 1985, revised geochronologic tabulation). The transgression would also coincide with the basal Tithonian transgressive pulse reckoned by HALLAM (1984). Detection of the two remaining intra-Tithonian eustatic peaks of VAIL et al. (1984) throughout all sections is not possible, what is eventually due to overcompensation by sedimentation. Effects should, though, be notable in areas with a general low sedimentation rate. As a matter of fact, in the Cabo Espichel, Arruda and Sta. Cruz sections, all characterized by a relatively low averadge bulk sedimentation rate for the Tithonian (cf. to chap. 10), two major transgressive peaks can be detected in the Tithonian part of the sequence (fig. 3). Correlation to the global sea level chart is tempting, yet causes problems:

The three »VAIL« peaks are respectively situated 1 MA, 2 MA, 3 MA above the Tithonian base (based on the VAN HINTE 1976 chronostratigraphic chart). To plot these ages in the geohistory diagrams given in fig. 5 values should transform to respectively 0, 1.6 MA, 3.2 MA above the Tithonian base[1]. This is in contrast with the actual position of the peaks. A possible solution of this problem could be that sedimentation rates were relatively higher in the Lower than in the Upper Tithonian. The indicated peaks could, however, also correspond to local tectonics (see chap. 10).

Cyclicity of repetitive limestone/marl/sandstone sequences at a smaller scale in the higher Tithonian is, however, very crude. At the Cabo Espichel, there exist at least 26 intercalations of clastic sediments which exhibit extremely variable thicknesses. This is interpreted as being the result of periodic tectonic hinterland activity rather than sea level fluctuations by FÜRSICH & SCHMIDT-KITTLER (1980). This point of view can be confirmed with the work of FELBER et al. (1982) who were able to trace the main horizons towards the hinterland, demonstrating that limestone levels thin out and disappear over a few tens of kilometers. This also excludes a third genetic possibility, a lateral shift of coastal environments.

An exclusive control of limestone/marl rhythms through oscillating sea level fluctuations can also be excluded for the studied sequence, since rhythms do not exhibit constant or continuously in/decreasing thicknesses for either the marl or the limestone levels or both (cf. LEINFELDER 1986, appendix), what would be a crucial characteristic for an eustatic control of sedimentation (EINSELE 1982).

[1] These geohistory curves are based on the more recent KENT & GRADSTEIN (1985) chronostratigraphic chart. Furthermore the lowest Tithonian »VAIL« peak is as an approximation correlated to the Kimmeridgian/Tithonian boundary.

7. Biologic control of sedimentation

There are many good examples for obvious biologic control of sedimentation in the sequence under study. Naturally, marine limestone development is *per se* already an example of biologic control. This control is particularly obvious in reefal facies. R. C. L. WILSON (1979) interpreted parts of the Praia do Guincho section closeby Sintra as a reef body and related talus sediments, thus clearly demonstrating direct biologic control of depositional environments. A similar situation exists in the Ota limestone, where a stripe of biogenic coral–algal limestones resulted in the establishing of a leeward low energy, lagoonal zone (LEINFELDER et al., in prep.).

Formation of nodular shallow water limestones at the Cabo Espichel and in the basin center was induced by burrowing activity of crustaceans (FÜRSICH & SCHMIDT-KITTLER 1980, LEINFELDER 1986).

The change from marly to calcareous sedimentation in the »Pteroceriano« beds is often favoured by pioneer fauna, such as oyster beds or highly specialized corals, or, more rarely, by the development of blue-green algal oncolites.

Widespread algal laminites and fenestral limestones in the Ota limestone or in the Cabo Espichel sequence are also biosedimentary structures in the original sense.

Very common, epibiontic miliolids in component-rich late Tithonian limestones indicate most likely intensive growth of sea grass and thus stabilisation of sediments (op.cit.).

8. Exogenic control of sedimentation

Algal laminites and fenestral limestones indicate tidal influences in a range of 2–4 m (R. C. L. WILSON 1975b, for Oxfordian carbonates). LEINFELDER (1986) interpreted the sedimentation patterns of the Sobral clastics as strongly influenced by tides, resulting in an estuarine delta type.

Common, up to two meter thick caliche horizons within terrestrial clastics on one hand, and frequent lignitic wood debris, partly with faint annual growth rings, on the other (op.cit.), indicate that climate was subtropically humid with long arid dry seasons. Thus it may be expected that degree of clastic input into the basin was seasonally changing according to the amount of riverine water charge.

Examples of influence of salinity fluctuations on faunal distribution and composition are widespread (FÜRSICH & SCHMIDT-KITTLER 1980, FÜRSICH 1981, LEINFELDER 1986, FÜRSICH & WERNER 1984, 1986, WERNER 1986). Oyster patch reefs, bivalve meadows, coral coppices and coral biostromes are obvious cases, where biologic control of sedimentation is itself controlled by exogenic factors. Many other environment-controlled faunal associations do, however, not directly influence the sedimentary facies, but are on the opposite themselves additionally controlled by lithofacies characteristics (e.g., grain size, organic content).

Shelly storm deposits are a direct proof for episodical occurrence of cyclone weather. Changes from marl to limestone sedimentation in the »Pterocerianoc« sequence may also be partly due to storm events: Storm-induced lime mud clouds, possibly originating from reefal areas around Sintra (see chap. 4.2) may have episodically helped to accelerate the re-establishment of carbonate deposition after breakdown phases caused by elevated terrigenous input. Hints are adaptions in the growth form of the mussel *Arcomytilus morrisi,* resulting in a change from epibyssate to semi-endobyssate life habit, rarity of microbored bioclasts in many beds and prevalence of a decapode deposit feeding animal producing *Rhizocorallium irregulare* tracks (LEINFELDER 1986). Wind-induced currents may also have accounted for facies changes from high to low energy settings in the Ota region (LEINFELDER et al., in prep.).

9. Tectonic control of sedimentation

There can be no doubt that tectonism was the main factor influencing and controlling sedimentation patterns in the Lusitanian Basin during the Kimmeridgian and Tithonian. The outlined interfingering and repetitions of facies types, restriction of certain types to local areas, and variations in sediment thicknesses clearly evidence a strong interplay of mainly fault-bounded subsidence and uplifts. This tectonic activity is related to regional rifting events in the North Atlantic region (see chap. 3).

Table 1 gives an overview of the main effects caused by differentiated hinterland and inner basin uplift or subsidence. Some examples are briefly considered as follows.

Sedimentologic effects of the Vila Franca fault system

The Vila Franca fault system is still recently tectonically notable. It caused heavy seismic activity in 1909. The epicenter of the catastrophic earthquake in 1755, destroying major parts of Lisbon center, was situated 250 km offshore, yet tectonic activity in the Tejo valley at Vila Franca was higher than in the surroundings (ZBYSZEWSKI & TORRE DE ASSUNÇÃO 1966).

Most obvious in terms of faulting effects are faults along the western border of the Tejo valley. Several kilometers further west faulting in similar direction is obvious for the western and eastern limitation of the Ota block. On Landsat satellite image, a further important line can be seen east of the Tejo river crossing the young valley filling.

Effects of synsedimentary faulting already during the Upper Jurassic are very obvious (fig. 4). In the Kimmeridgian Arruda subbasin (see below), clastic sediments are intensively coarsening towards east, i.e. towards the fault system. This is accompanied by a strong admixture of reworked basement rocks and detrital feldspar (fig. 4a). R. C. L. WILSON (1979) interpreted this as a questionable submarine and/or subaerial fan situation, initiated by a fault scarp caused by the Vila Franca faults. Growing of reefal limestones on and within this fan, e.g., the thick Monte Gordo limestone or thin levels at Castanheira further north, indicate the at least partial or episodical submarine character of the fan and calm phases where no clastics were introduced. Large allodapic limestone blocks and meter-sized clay boulders evidence steep slopes and eventual seismic pulses causing break-off of limestones. Black pebbles both in the Monte Gordo and Ota limestones point out occasional subaerial exposure of limestone areas, possibly due to local-scale tectonic uplifts. Deposition of black pebbles in form of very coarse conglomerates and breccias with a very limited distribution indicate fault controlled deposition within the Ota limestone.

In-situ blackening, intraformational solution phenomenons and vadose cements within the Ota limestone (LEINFELDER et al., in prep.) are also most probably related to local tectonic uplifts, although effects of eustatic sea-level oscillations cannot be excluded completed.

Rapid transition from the pure Ota limestone to surrounding marly sediments indicate fault-related, steep by-pass margins of the Ota bank. Lateral discontinuity of limestone bank development indicates downfaulting of blocks roughly perpendicular to the main faulting direction.

The late Kimmeridgian Amaral coral limestone is best and most typically developed in a narrow, NNE–SSW trending zone in the direct southern continuation of the Ota block (between Cabreira, Serra Isabel and Serra do Amaral, and Trancoso) (fig. 4b). Here reefal limestones are well differentiated into

U P L I F T

hinterland uplift

increased clastic input

regression (Tithonian of Serra da Arrábida)	**prograding of deltas into marly or calcareous areas, homopycnal freshwater outflow due to flattening of slopes: change in distribution of brackish sediments** (Sobral delta)	**change in inclination of basin margin slopes: turbidites** (Abadia beds) **slope break bioherms** (Monte Gordo limestone)	**shallowing of marine clastic sequences** (Abadia beds)	**breakdown of many carbonate-producing organisms**
		complete breakdown of carbonate production (marl levels in lower "Pteroceriano" beds)	**favouring adapted forms, e.g. baffling corals: biostromes** →	**inner basin morphology: differentiation into high and low energy facies** (Upper "Pteroceriano" beds)

inner basin uplift
(salt and non-salt induced)

local shallow water facies (Ota limestone, limestones at Caldas diapir)	**outthinning and outwetching sequences** (around most salt diapirs)	**subaerial exposure: karstification** (Ota limestone)
development of current systems controlling facies distribution (Ota limestone?)	**creation of bordered sub-basins: local low energy facies, facies boundaries clastics/limestones, channeling of sandstones** (lower Tithonian N Lisbon)	**chemical source for surrounding freshwater limestones** (Alenquer Oncolite)

S U B S I D E N C E

hinterland subsidence		**inner basin subsidence** (salt and non-salt induced)		
decreasing clastic input				
transgression (deepening at top of Sobral delta)	**establishing of carbonate producers and carbonate deposition** ("Pteroceriano" limestones p.p.)	**steepenig of local slopes: mass flow deposition** (arcosic gravel in Arruda borehole	**clastic traps** (lower "Pteroceriano" beds) ↔	**local high sedimentation due to sediment gathering: overcompensation of subsidence** (terrestrial clastics at S.Martinho)

Table 1. Effects of uplift and subsidence on the sedimentation pattern of Kimmeridgian/Tithonian deposits in the Lusitanian Basin. Most examples are treated in the text

subtypes and the associated oolites are also perfectly developed. Contamination by clastics is low in this region what contrasts the development further east and west. North of Arruda dos Vinhos, development of reefoid facies is almost exclusively restricted to this zone. Towards west and east only local marly coral bafflestones appear, whereas the main part of the Amaral limestone is represented by sandy oolites. Most characteristic for the entire area is the very sudden lateral disappearance of calcareous reefoid facies towards the Tejo valley.

The distribution of the Sobral delta is strongly influenced by the above indicated zone. Interestingly it thins out and disappears towards Alverca (7 km S Vila Franca), although further south similar sandstones reappear in the Lisbon area (borehole Barreiro 2, cf. fig. 3) which are clearly derived from the east (LEINFELDER 1986). This shows that the mentioned zone was not subsiding during times of the Sobral sedimentation and thus could not accept sediments.

The lower part of the »Pteroceriano« beds displays very low energy characteristics in the main region of its development south of Arruda dos Vinhos (nodular limestones and common marls). On the mentioned block, however, clastic contamination plays a much smaller role. Thicknesses are apparently less. Commonly intercalated in the sequence are high-energy limestones, partly with corals, as well as oncolites which indicate low sedimentation rates. This evidences the higher structural position of the block also during the early Lower Tithonian. High thicknesses reappear in a narrow zone closeby Vila Franca, indicating local wrench block subsidence in the close vicinity of the main Vila Franca fault (fig. 4c).

During sedimentation of the upper part of the »Pteroceriano« beds, characterized by a general shallowing, this zone underwent most likely subaerial exposure. There is no direct evidence since recent erosion truncated down to the lower »Pteroceriano« sediments, but perfect shelter from east-derived coarse clastics of presumed same age around Vila Franca is obvious (op.cit.) (fig. 4d).

Due to tectonic uplift, the Ota limestone complex suffered from karstification from the late Kimmeridgian onwards, serving as a chemical source for the intensive development of fresh water oncolites intercalated into surrounding red clastics (LEINFELDER 1985).

The mid? Kimmeridgian Monte Gordo limestone closeby also apparently underwent subaerial exposure prior to the Cretaceous, resulting in the removal of eventually overlying sediments so that requienid bivalves could encrust an Upper Jurassic surface (ELLIS 1983).

Apparently not the main fault crossing the town Vila Franca was acting as the eastern limitation of the basin. The main border must have been situated further east, eventually corresponding to the mentioned satellite image lineation. This can be concluded from the purity of Kimmeridgian shallow water limestone banks, indicating a certain distance from the hinterland, and from the facies development of the lower Tithonian in the Vila Franca area which also does not exhibit contamination by coarse clastics.

Halokinetic effects

Jurassic salt movements, known from Portugal since CHOFFAT (1882), are for a long time considered to have had considerable effects on the sedimentary record of the Lusitanian Basin (e.g., OERTEL 1956). Many authors guess that uprise of salt diapirs was fully or partly responsible for structural effects (e.g., GUÉRY 1984, for halokinetic control of mid-Liassic to Upper Dogger subsidence patterns; R. C. L. WILSON 1979, for additional diapiric effects on general Lower Oxfordian uplift and for partial halokinetic control of Kimmeridgian depocenters). Already ZBYSZEWSKI (1959) demanded that lateral forces are needed to initiate uprise, a point of view which was recently emphasized by SCHMIDT (1984) and LEINFELDER (1986) who consider that extensional and wrench faulting due to rifting triggered and enabled salt diapir uprise.

Actually, almost all portuguese salt diapirs are lined up along mainly straight, N or NNE trending fault zones. SCHMIDT (1984) calls attention on the common association of diapiric structures with basaltic rocks what proves deep reaching fault zones. He guesses that the respective faults were Mesozoic rejuvenations of Hercynian basement structures.

Upper Jurassic halokinetic movements within the Lusitanian Basin may be concluded from various hints:

The position of salt diapir structures often corresponds to Upper Jurassic paleogeographic highs, as postulated from facies analysis. There existed, however, Upper Jurassic shallow water zones along the diapir-occupied faults, like for instance in the Montejunto anticline (see chap. 4.1) which are not yet proved to be associated with salt structures.

Decrease of sediment thicknesses towards the structural apex and development of marginal troughs away from the axis (e.g., Sesimbra, LEINFELDER 1983; Columbeira, MONTENAT & GUÉRY 1984; Matacães, LEINFELDER 1986) and differences in facies development from either side of the topographic highs (Vimeiro–Caldas da Rainha, R. C. L. WILSON

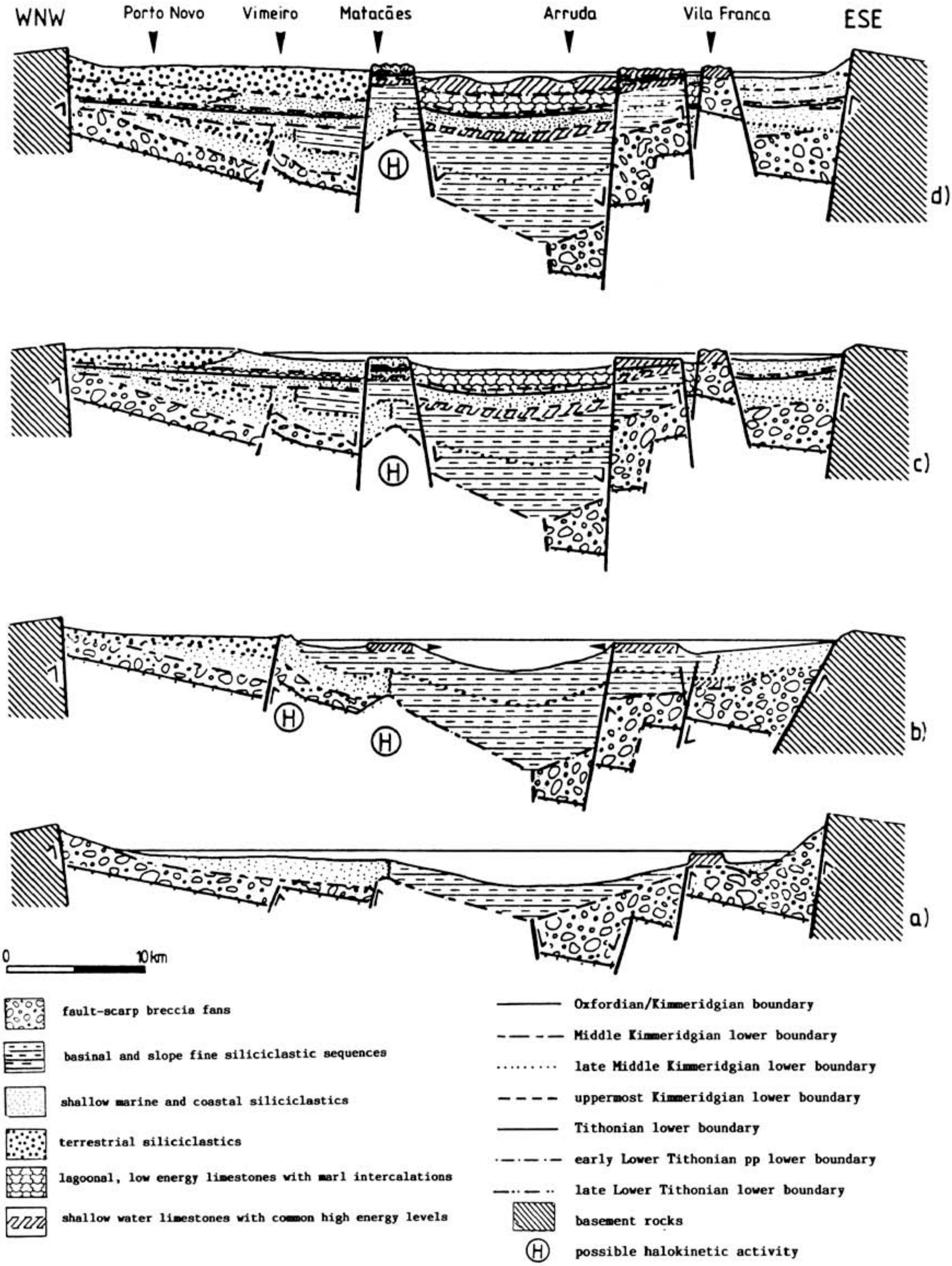


Fig. 4. Sketch of structural development for the Kimmeridgian and Lower Tithonian of the Lusitanian Basin, as evidenced by facies analysis. West of Porto Novo and east of Vila Franca hypothetic. Thicknesses not a scale. For true thicknesses see fig. 2.

1979, WERNER 1986) are typical features of active salt diapirs but could all as well result from block faulting and tilting. Ringlike or bipolar facies patterns, typically developed at Porto de Mós (SCHMIDT 1984) and to a lesser extent at Sesimbra (FELBER et al. 1982, LEINFELDER 1983) could theoretically also result from tilting and uplifting of isolated blocks. Even migration of marginal troughs, as occuring in the Columbeira area (MONTENAT & GUÉRY 1984) could result from continuous broadening of half grabens due to migration of downfaulting along en-echelon faults. Salt diapirs might have invaded such unstable fracture zones at a much later time than the Upper Jurassic. The existence of broad, so-called typhonic valleys (CHOFFAT 1882) in the center of the structures, created by the piercing and subsequent leaching of salt diapirs does not completely exclude this theoretic possibility for the mentioned examples. Tectonic uplift might cause migration of salts into the structure due to a new isostatic inequilibrium. The Formosinho anticline in the eastern Serra da Arrábida might be an example where salt occupied the core of a pre-existing flexure (cf. SEIFERT 1963, fig. 7). Salt might also pierce later through such structures to result in a typhonic valley.

Thus, caution is necessary for interpreting the origin of paleohighs, especially when they are not connected to diapiric structures. Examples are the zone between Vimeiro and Columbeira or the prolongation of the Torres Vedras–Matacães high towards north, which actually might be due only to Jurassic block faulting or upwarping.

It is assumed that the development of Upper Triassic salt deposits was restricted to the morphologically lowest parts of half graben structures which were created during the first rifting phase of the Lusitanian Basin (see chap. 3). Due to the differential loading of such structures a slowly moving salt rise may have started early (cf. JACKSON & TALBOT 1986). Wrench tectonics of the second Oxfordian-Kimmeridgian rifting phase resulted in differentiated strike-slip movements along the reactivated Upper Triassic faults. This triggered uprise of salt domes but also local tectonic uplift of former half graben flanks next to diapiric structures or without such. Diapirism eventually only started after uplifting in certain cases due to loading inequilibrium. Thus, the mentioned straight paleotopographic highs are due to a mixed block tectonic – salt tectonic history.

Halokinesis eventually started already in the Liassic (GUÉRY 1984) and was particularly active during the Callovian and Oxfordian (R. C. L. WILSON 1979). During the Kimmeridgian the Caldas diapiric zone was a remarkable facies boundary between clastics and carbonates (MONTENAT & GUÉRY 1984, WERNER 1986). R. C. L. WILSON (1979) guesses that the Vimeiro diapir further south possibly had influence on sediment thicknesses and facies differences of terrestrial clastics on both sides of the structure. The Porto de Mós diapiric structure exhibits strongly diminished sediment thicknesses during the Lower Kimmeridgian (SCHMIDT 1984). Jurassic activity in most structures stopped before the Tithonian (GUÉRY 1984). The Matacães structure and its northward prolongation was, however, obvious in terms of facies control mainly during the Lower Tithonian, where it largely bounded the calcareous/marly »Pteroceriano« development to the east (fig. 4c, d).

Examples of block subsidence effects and hinterland activity

R. C. L. WILSON (1979) identified two Kimmeridgian depocenters (»subbasins«), the Bombarral and the Arruda subbasins. Borehole Ramalhal 1, penetrating the Bombarral subbasin, displays from depth 1024 m to 394 m a predominantly marly-sandy marine shoaling upward sequence, exhibiting charophyte-bearing marls at the top (ZBYSZEWSKI et al. 1966). The sequence is termed »Pteroceriano« by the cited authors what is confusing in respect of facies development. Occurrence of the foraminifer *Alveosepta* aff. *jaccardi* at 950 m may indicate, like in the Consolação section (cf. WERNER 1986), a Lower to Middle Kimmeridgian age. The sequence is better interpreted as facies transition from the Abadia beds to the Alcobaça beds (eventually reaching the Sobral level). It is overlain by terrestrial clastics, representing most probably the late Kimmeridgian and Tithonian.

Hence, subsidence was, in connection with a slow sea level drop, progressively overcompensated by accelerated clastic input, resulting from increasing tectonic uplift of basement blocks, most likely at the western basin margin.

The Arruda subbasin, interpreted as wrench basin by R. C. L. WILSON (1979), was rapidly filled up with fan clastics during the Lower Kimmeridgian (see chap. 10). The rest of the sequence, displaying marly Abadia facies, also exhibits a general shoaling upward trend (fig. 3). Shallowing normally culminates in the overlying reefal and oolitic Amaral coral limestone. From field evidence alone it is difficult to decide, whether the change to calcareous facies in the Arruda subbasin is caused by aggradational shoaling accompanied with a sudden decrease of clastic hinterland input (thus decrease of tectonic activity) or whether it is due to a diachronous development resulting from progradation of a high energy slope break position. Occasional occurrence of low-energy micritic limestones within the Amaral limestone (e.g., at section Mata, LEINFELDER 1986) may indicate a general stop of terrigenous input. Main development of reefal limestones along a narrow zone east of Arruda dos Vinhos and sudden disappearance further east rather evidence an elevated narrow zone from which the slope break limestone may have prograded diachronously towards west and south. Hence, both processes apparently played a role what can also be concluded from the following chapter.

The Kimmeridgian was the time of greatest terrigenous influx due to pronounced uplift of hinterland by intensive block faulting in the second rifting phase (cf. R. C. L. WILSON 1979). This is not only true for the Lusitanian Basin but also for the entire North Atlantic where terrigenous influx displays a sharp peak, reaching mean accumulation rates of up to $2500\,g \times cm^{-2} \times MA^{-1}$ (EHRMANN & THIEDE 1986). In the Lusitanian Basin tectonic hinterland activity was, generalizing, apparently very high at the base of the Kimmeridgian, resulting in fault scarp-induced fan deposition. For the rest of the Kimmeridgian activity was at lower rate. This picture is modified by short-termed local or regional stops of hinterland activity to allow limestone formation, and by local development of high-energy limestones on uplifted blocks which were not influenced by surrounding clastic sedimentation.

During the Lower Tithonian the rapid nordward transition from »Pteroceriano« limestone/marl sedimentation to marginal marine and terrestrial clastics evidences active clastic traps. Since there are no hints for elevated barriers towards this direction and since the basal »Pteroceriano« transgression was for a short time reaching far towards north and northwest (see chap. 11), trapping of clastics was apparently maintained through rapid block subsidence, eventually supported by lateral salt creeping. Subsidence was compensated (but not overcompensated) by clastic input, resulting in a broad, low energy coastal plain where silt, sand and gravel-sized clastics were trapped because of loss of transport energy. Fine-grained clay fraction was, however, introduced into the marine basin to form, in interplay with small tectonic pulses, the typical limestone/marl sequences of the »Pteroceriano« formation. During sedimentation of the upper »Pteroceriano« beds, fluctuating clay input, and hence tectonic pulses, were strongly controlling biogenic reefal carbonate sedimentation (LEINFELDER 1986).

During the late Tithonian, the former clastic traps were mostly overcompensated by sand input, so that terrestrial clastics migrated further south and imaged the step-wise tectonic hinterland uplifts with rapid lateral progradings and subsequent withdrawals.

Further evidence of tectonic control on sedimentation results from approximated geohistory analysis, as given in the following chapter.

10. Approximation of geohistory analysis

Due to enormous differences in thicknesses and facies development it remains difficult to correlate individual Kimmeridgian/Tithonian sections even when considering their respective transgressive-regressive patterns (fig. 3). It remains furthermore almost impossible to find out even the approximate degree of diachronism for most levels, or to analyse whether regressive tendencies are a result of sedimentary overcompensation or are more likely tectonic or eustatic effects.

Important characteristics of basin development may be read from geohistory analysis curves (VAN HINTE 1978). Since biostratigraphic markers are mostly lacking, geohistory can only very approximately be applied on the studied sequence. Yet, many qualitative results can be derived from such an approach, and certain implications reveal the need for further studies.

The following data were used for the purpose:

Paleobathymetric interpretation, as given in fig. 3, is for the studied sequence converted into the following values

basin	−200 m
slope	−50 – −200 m, mean −125 m
tranquil shallow water	−20 – −80 m, mean −50 m
agitated shallow water	−50 – 0 m, mean −25 m
meandering rivers	0 – 50 m, mean 25 m
braided rivers, distal all. fans	0 – 100 m, mean 50 m

Absolute values are of minor importance, since it is only attempted to depict changes in basin topography. Hence, uncertainities are negligible.

Marker levels are the base of the Kimmeridgian, the Kimmeridgian/Tithonian boundary, as expressed by an easily recognizable transgressive peak (see chap. 5), and the top of the Tithonian. Duration for the Kimmeridgian is 4 MA, for the Tithonian 8 MA (after KENT & GRADSTEIN 1985).

As an approximation average bulk sedimentation rates have to be taken as constant for the entire Kimmeridgian and entire Tithonian respectively. An exception is section Arruda where the Kimmeridgian is divided into an upper marly part (450 m) with a presumed sedimentation rate similar to neighboured marly section Monsanto, what results in a different sedimentation rate for the fan breccias in the lower part of the Arruda log. If sections do not reach marker levels, averaged sedimentation rates from all sections are used. Sections situated on structural highs are not considered for the diagrams (fig. 5). Sediment thicknesses and corresponding sedimentation rates are given in tab. 2.

The following conclusions can be drawn from the diagrams (fig. 5):

* Bulk sedimentation rates were generally high during the Kimmeridgian. They, varied, however, considerably from site to site (approximately between 105 cm × $1000a^{-1}$ and 11 cm × $1000a^{-1}$, on the average 15–16 cm × $1000a^{-1}$). An exception is the Cabo Espichel with only 3 cm × $1000a^{-1}$, pointing out that clastics were gathered in the strongly subsiding parts of the basin. Basin center was situated between Lisbon (Monsanto) and Arruda. The enormous thickness around Arruda is due to a local wrench fault basin (see before), filled up during the Lower Kimmeridgian.

* In the north, marine incursions appear commonly in very different vertical position. The marine phase at Consolação and C. Mondego, however, ended at the same time. There was fairly strong basin subsidence in Porto Novo, Consolação, S. Martinho and C. Mondego (which was as or even more intensive than at C. Espichel, Barreiro and Sintra in the south), yet subsidence was normally overcompensated by coarse clastic input. Marine intervals should express lower rates of tectonic hinterland activity, so that due to starving input marine facies could rapidly spread. Strongest input and thus strong local hinterland activity already at the base of the Kimmeridgian was at Porto Novo and especially at S. Martinho.

* In the south, the sequence is generally shoaling up (except for the Serra da Arrábida). Calcareous facies appears earlier at the basin margins (Barreiro, Sintra, partly with allochtonous limestones) than in the basin

Kimmeridgian C.Espichel Barreiro 2 Monsanto Sintra Sta.Cruz Pto.Novo Consolaçao S.Martinho C.Mondego

	C.Espichel	Barreiro 2	Monsanto	Sintra	Arruda	Sta.Cruz	Pto.Novo	Consolaçao	S.Martinho	C.Mondego
Kimmeridgian										
thickness (m)	ca 120	540	1000	430	2750	>450	>360	ca 630	470	520
sed.rate (cm/1000 a)	3.0	13.5	25.0	10.8	68.8 (105/25)*	(15.6)	(15.6)	15.8	11.8	13.0
Tithonian										
thickness (m)	540	285	630	500	360	>300	>400	>270	>90	>50
sed.rate (cm/1000 a)	6.8	3.6	7.9	6.3	4.5	(5.8)	(5.8)	(5.8)	(5.8)	(5.8)

sedimentation rate Kimmeridgian: min. 3.0 max. 68.8 : mean 15.6 cm/1000 a (without considering extreme values)
sedimentation rate Tithonian: min. 3.6 max. 7.9 : mean 5.8 cm/1000 a

Table 2. Thicknesses and deduced bulk sedimentation rates for the Kimmeridgian and Tithonian part of major sections. For definition of Tithonian base refer to chap. 5. Time intervals after KENT & GRADSTEIN (1985). Asterisk marks splitting of Kimmeridgian bulk rate (see text). Further values in brackets represent averaged sedimentation rates.

Fig. 5. Simplified Geohistory Analysis: plot of bathymetric interpretation and cumulative thicknesses versus time for all sections.

x-axis: Kimmeridgian-Tithonian time scale (after KENT & GRADSTEIN 1985). Left base of Kimmeridgian, right top of Tithonian; large arrow Kimmeridgian/Tithonian boundary.

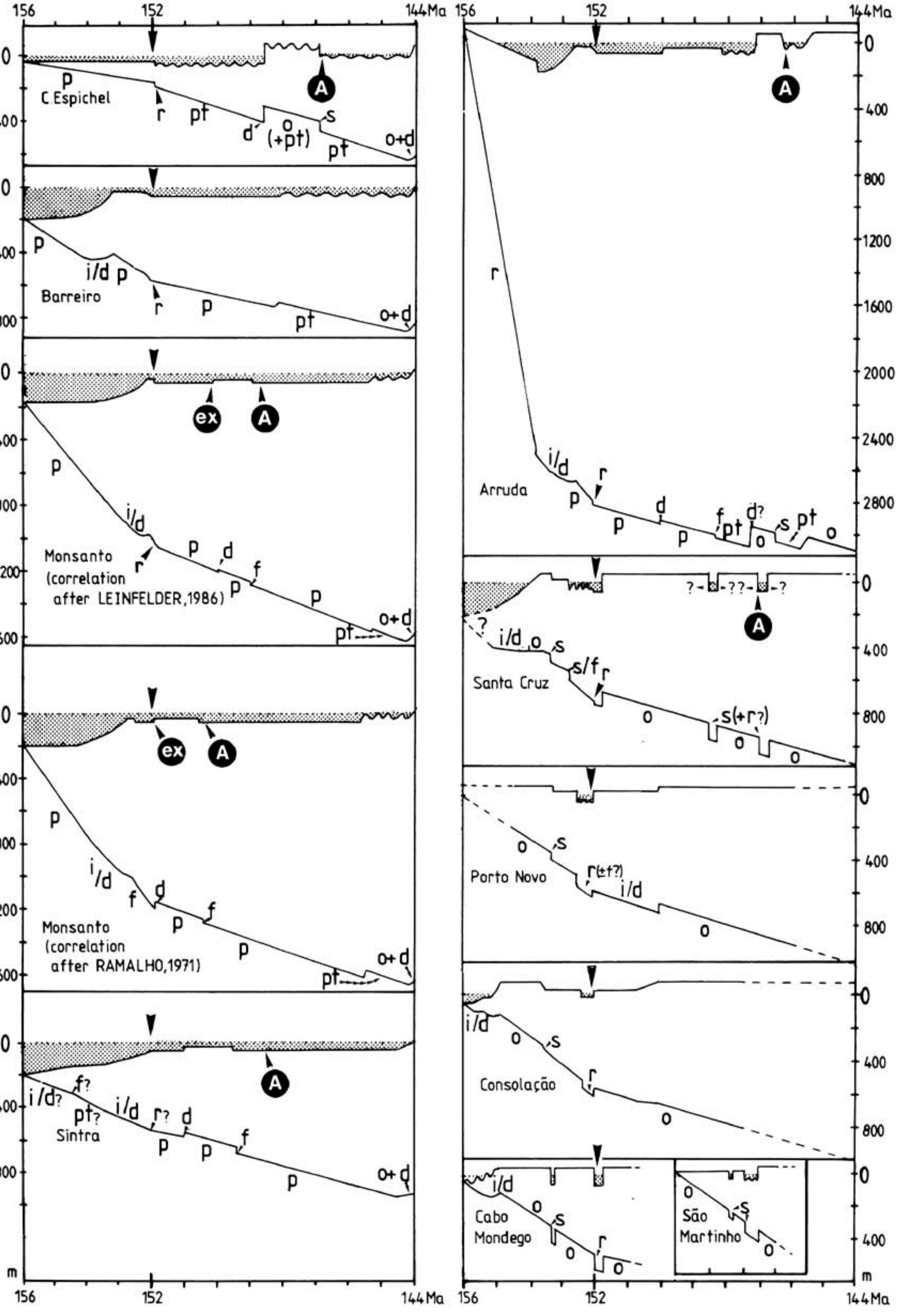
y-axis: thicknesses in meters

Abbrev.: A first appearance of *Anchispirocyclina lusitanica,* ex last occurrence of *Clypeina jurassica* in Monsanto log;

D tectonically or eustatically induced drop of sea level
F local basin deepening due to downfaulting
I shallowing due to increased clastic influx
O strong overcompensation of subsidence by sedimentation, due to tectonic hinterland activity
P sedimentation keeping pace with subsidence
PT sedimentation keeping pace with subsidence, yet mirroring tectonic pulses
R peak of sudden eustatic sea level rise
S deepening due to stop of depositional overcompensation.

Compare with fig. 3; for further explanation see text.

▶



156
152
144Ma
C.Espichel
Barreiro
Monsanto
(correlation
after LEINFELDER,1986)
Monsanto
(correlation
after RAMALHO,1971)
Sintra
Arruda
Santa Cruz
Porto Novo
Consolação
Cabo
Mondego
São
Martinho
A
ex
p
pt
r
d
o
s
f
i/d
o+d
(+pt)
s/f
s(+r?)
r(±f?)
m
0
400
800
1200
1600
2000
2400
2800

center (Monsanto, Arruda). This excludes a more aggradational shallowing where limestones should appear at the same time in the entire basin. Shallowing might be due to lateral sediment progradation or to a slow eustatic sea level fall. In the latter case, however, the western Arrábida sequence should also show a regressive pattern which is not fully the case (see below).

The marine, limestone-bearing level at S. Martinho, interpreted as equivalent to the Amaral limestone by R. C. L. WILSON (1979) and WERNER (1986), as well as the marine, marly-calcareous level at Cabo Mondego appear much earlier. This proves the postulated diachronous character of the interval (see chap. 5). If this level would correspond to the »Pteroceriano« transgression, what also cannot be excluded through paleontological facts, Kimmeridgian sedimentation rates for S. Martinho and Cabo Mondego would only amount to 7 cm $\times$ 1000a^{-1} and 9 cm $\times$ 1000a^{-1} respectively, what is in contrast to the overall picture.

* During the Tithonian, degree of basin deepening was much lower than during the Kimmeridgian, evidencing the slow-down of tectonic activity. Including compaction of Kimmeridgian sediments, whereby especially the marly-shaley sequence of the basin center might have undergone substantial compaction, sedimentation rates were 2.5 to 4 times lower than during the Kimmeridgian (Arruda 15 $\times$). Considering compactional effects, tectonic basin subsidence slowed down even more. Highest (uncorrected) subsidence rates appear in Monsanto (8 cm $\times$ 1000a^{-1}) and at the Cabo Espichel. In the latter case subsidence was higher by factor 2 than during the Kimmeridgian. A minimum was at the eastern margin of the basin (Barreiro: 3.6 cm $\times$ 1000a^{-1}). At the former western basin margin around Sintra, subsidence during the Tithonian did not slow down as much as in other regions, evidencing a shift in basin axis towards southwest.

* Tithonian basin deepening in the northwest was not lower than in marine areas further south. It was certainly differentiated from site to site, what, unfortunately cannot be depicted by the available means.

* In the Tithonian start of clastic input to the marine basin coincides roughly in the east (Arruda, Barreiro) and corresponds to a sudden increase in clastic input in the south (C. Espichel). From this one may conclude a regional uplift of the eastern hinterland, starting with a sudden peak. This resulted in penecontemporaneous fast introduction of clastics which was restricted to the eastern parts of the basin. Only much later clastics reached slightly further west (Monsanto).

Shallowing of marine facies is earlier in Sintra than in Arruda. The Arruda shallowing was, however, coeval with a change in mainland morphology in Porto Novo. Accepted that sedimentation rates were stable, this could eventually indicate a slow sea level drop, accentuated by tectonic pulses.

* Thick, oscillating transgressive/regressive cyclic limestone/marl/clastic sequences appear only in slowly subsiding areas (Cabo Espichel, Barreiro and Arruda). A general balance between slow subsidence and sedimentation kept the water level close to zero. Minor tectonic hinterland pulses resulted in short-time overcompensation of subsidence by terrestrial clastics. The two marine levels at Sta. Cruz could be interpreted as stops of hinterland activity in an otherwise slightly sediment-overcompensated area. To what extent sea level fluctuations may have played a role was discussed earlier.

* Kimmeridgian/Tithonian transition

It should be recalled that the base of the »Pteroceriano« limestones or the top of an corresponding marine clastic interval was taken as the stage boundary due to biostratigraphic investigations, field evidence and model considerations (see chap. 5). Taking this as a fact, it is obvious that lowering of the sedimentation base occured already since Middle to Upper Kimmeridgian at Sta. Cruz, Porto Novo and, possibly, Consolação. At Cabo Espichel a certain deepening is eventually already indicated during the Upper Kimmeridgian (decreasing miliolids, appearance of sponge spicules, cf. RAMALHO 1971). At the base of the Tithonian a short-termed rapid deepening is obvious by the appearance of abundant sponge spicules, occurrence of ammonites, ironhydroxide impregnations (cf. op.cit.) and by a positive strong peak in foraminiferal diversity (cf. FÜRSICH & SCHMIDT-KITTLER 1980). Arruda exhibits rapid deepening at the top of the Sobral sandstones, what possibly is transferable to Barreiro log. Deepening in Sintra is not very obvious. Limestones consisting mainly of allochtonous reef blocks and breccias grade into a low energy limestone/marl sequence. Eventual deepening is apparently overprinted by a flattening of the depositional slope.

Deepening occurred thus in many places, even in sediment-overcompensated terrestrial areas, what backs the interpretation of an eustatic sea level highstand. Transgression was additionally favoured by a decrease of tectonically induced clastic influx which lowered the terrestrial sedimentation base already in Middle to Upper Kimmeridgian times.

* Biostratigraphic implications

Dating by *Clypeina jurassica*

Lithostratigraphic correlation of log Monsanto may be performed by interpreting the transition from a 74 m thick micritic limestone interval characterized partly by reefal microfacies (zone 4 and 5 of section Monsanto after RAMALHO 1971, 96) to a 130 m thick dasycladacean-dominated, apparently lagoonal sequence (»calcaires cryptocristallins graveleux«, zone 6 of respective section, op.cit.) as the transgressive correlation level. Deepening may be hardly reflected, yet the overlying 109 m of partly oolitic and oncolitic limestones (zones 7, 8 of cited section) would correspond to a similar shallowing level at Sintra. Facies similarities in this aspect were also mentioned by RAMALHO (1971). Additionally both shallowing levels would plot in a comparable height at about 2 MA above the Tithonian base.

However, the mentioned »gravelly limestones« (zone 6), which according to this plotting should be of lowermost Tithonian age, exhibit abundant specimens of the dasycladacean alga *Clypeina jurassica,* on grounds of which RAMALHO (1971) dated these sediments as uppermost Kimmeridgian.

If this is so, consequences in the geohistory curve are considerable (fig. 5). Sedimentation rate for the Kimmeridgian would be very high (29.8 cm $\times$ 1000a^{-1}) and decrease of rate for the Tithonian would be by factor 5.7 which is high compared to lithologically similar closeby sections. This section would moreover display a transgression – regression rhythm different to the rest of the basin. The Kimmeridgian/Tithonian boundary would correspond to the change from the mentioned dasycladacean-dominated lagoonal facies (zone 6) to lagoonal facies characterized by the appearance of oolitic and oncolitic beds (zones 7, 8 of cited section). This should rather reflect a regression although related differences in depositional depth are certainly small. The foraminifer *Anchispirocyclina* would appear already 1.7 MA above the Tithonian base, whereas in the rest of the sequence its earliest appearance is around 3 MA above the base (see below).

As yet there is no solution to these discrepancies. Good regional biostratigraphic results were achieved by considering *Clypeina jurassica* as typical for the Upper Kimmeridgian of Portugal (cf. RAMALHO 1971, 1981). The basal Tithonian transgression could eventually be overcompensated by sedimentation, although appearance of oolites and oncolites stands against high sedimentation rates. Another explanation could be local tectonic uplift.

If one judges that the match of the transgressive-regressive pattern has to be correct, then *Clypeina jurassica* must have appeared still during the Lower Tithonian (if reworking of the alga from underlying sediments is excluded). As a matter of fact, *Clypeina jurassica* is known in most parts of Central and Southern Europe from the (late?) Kimmeridgian throughout the Tithonian to the lowermost Cretaceous (e.g., FARINACCI & RADOICIC 1974, DRAGASTAN 1975, BENEST et al. 1975, AZEMA et al. 1977). This is also true for neighboured Spain (GARCÍA-HERNANDEZ & LÓPEZ-GARRIDO 1979). Even in Portugal it was locally detected in the Tithonian, although in much lower frequency (RAMALHO 1971, tabl. H; 1981).

Hitherto, none of the two possible cases can be definitely ruled out, until reconsiderations of the Monsanto log under facies aspects and further biostratigraphic data are available.

Dating by *Anchispirocyclina lusitanica*

Anchispirocyclina lusitanica is the guide fossil for the *Lusitanica* Biozone which is typical for the Upper Tithonian in the Mediterranean realm (e.g., BENEST et al. 1975, AZEMA et al. 1977, GARCÍA-HERNANDEZ & LÓPEZ-GARRIDO 1979). Regardless of the above mentioned complications (i.e. neglecting case b of interpretation of section Monsanto), and based on the applied methods the form appears early in Monsanto (ca. 3.3 MA above Tithonian base, case a) and Sintra (ca. 3.4 MA above base) but late in Cabo Espichel (5.1 MA) and Arruda (5.7 MA). From Barreiro log the form is not reported; its theoretical appearance might lie within the clastic-influenced interval above 3.8 MA over base.

If one considers the first appearance as a synchronous event, explanation must be that sedimentation rates have varied considerably and that consequently the given diagrams based on constant bulk rates are not a good approximation to the actual situation. Sedimentation rates at Arruda and Cabo Espichel must then have been very rapid for the lower part of the Tithonian, whereas rates in the basin center must have been considerably lower for similar sediments. This is unlikely, particularly when taking facies dependance of the form's appearance into consideration. Lack of the form in Barreiro logs might be due to the ecological stress of mixed clastic–carbonate sedimentation. Below the form's first appearance in the Arruda section there occur terrestrial sediments (plotted 4.7–5.7 MA above Tithonian base) and a mixed clastic carbonate sequence (3.7–4.7 MA above base); below the first appearance at Cabo Espichel there is a predominantly clastic sequence (plotted 3.2–5.1 MA above Tithonian base). Sediments representing better ecologic conditions below these sequences were possibly deposited before the first possible appearance of the foraminifer. It is not quite clear under this aspect why marine intervals within the mentioned mixed sequences around 3.5–5.0 MA above Tithonian base do not exhibit occasional specimens of the form. This might be due to the much lower population density at the base of the taxon range zone (evaluation of RAMALHO 1971 and own observations) and to insufficient sampling density.

Considering an error of 1 MA due to approximations and generalizations through the applied method, a boundary between the early Tithonian (Portland A *sensu* RAMALHO 1971, 1981) and late Tithonian (Portland B *sensu* RAMALHO) could be drawn at around 3–4 MA above base, subdividing the 8 MA duration of the Tithonian (after KENT & GRADSTEIN 1985) into a lower part lasting 3.5±0.5 MA and an upper part durating 4.5±0.5 MA.

It should be, however, noted that according to SEPTFONTAINE (1981) the *Lusitanica* zone starts already during the Kimmeridgian (if including *Anchispirocyclina praelusitanica* which is difficult to distinguish in unoriented thin-sections). This would even allow interpretation b for the Monsanto section where *Anchispirocyclina* would already appear 1.7 MA above the Tithonian base.

As a conclusion it may be stated that the first appearance of *Anchispirocyclina lusitanica* is not a good marker level. A better correlation horizon would be the bloom of the form (acme subzone) in the higher part of the Tithonian which in the given examples plots around 5–6 MA above the Tithonian base.

11. Paleogeography of the Lusitanian Basin during the Kimmeridgian and Tithonian

Generalized paleogeographic sketch maps for the Kimmeridgian and Tithonian of the Lusitanian Basin were first given by R. C. L. WILSON (1975a) and later by MOUTERDE et al. (1979) and in more detail by R. C. L. WILSON (1979). FELBER et al. (1982), and recently WERNER (1986) and LEINFELDER (1986) added maps for selected parts of the basin, which refine and partly contrast the given image.

Based on the cited authors and on the results of this paper on stratigraphic correlation and basin development, fig. 6 gives a set of rearranged and refined maps. Due to uncertainities in correlation caused by lack of biostratigraphic markers and due to lack of further sedimentological data, they still exhibit model character. Together with the added structural sketches (fig. 4) they are, however, thought to sufficiently illustrate the paleogeographic development.

During the Lower Kimmeridgian facies differentiation already reached a highpoint due to intensive block faulting and heavy input of terrestrial clastics. In the south, limestone development exhibited supratidal character around the Sesimbra diapir. East and particularly southeast-derived siliciclastics were only of subordinate importance.

The basin center showed differentiation into parts with and without carbonate development, indicating that between Cabo Espichel and north of Sintra no western uplifted basement block was existing. Fault scarps produced coarse-grained subaerial/submarine fans, eventually filling wrench fault subbasins. It remains unclear whether Upper Oxfordian local shallow water carbonate facies persisted to the Lower Kimmeridgian and whether the shallow water Ota bank existed already.

Further north, a narrow zone was limited towards east by a tectonic uplift which was strongly modified by uprise of salt diapirs. This zone exhibited clastic-dominated, coral-bearing, lagoonal and prodelta facies around Peniche, and delta cycles further north (Cabo Mondego). In between, basin deepening was overcompensated by clastic input to result in the development of a high sinuosity fluvial plain. East of the uplift region paralic to marginal marine facies, influenced by the Porto de Mós diapir, was deepening southwards towards the basin center.

The Middle to early Upper Kimmeridgian displays similar trends in the south with some modifications. In the Serra da Arrábida, reworked calciclastics were increasingly introduced from the east. Around Sintra, an accentuated slope break and reefs, developing most likely west of this zone, are indicated by allochtonous limestone sequences. Further north, fan breccias disappeared, instead clastic slope and basin sedimentation was widespread. The Ota limestone bank started to develop, as did carbonate platform sedimentation on top and east of the Vimeiro–Caldas da Rainha–Alcobaça uplift. Connection of the two carbonate areas is unlikely since east-derived clastics must have entered the basin in between. Along the present-day coast, marine, partly calcareous sediments occurred after a time of terrestrial sedimentation. High clastic input from the Berlengas area prevented establishment of lime sedimentation around Peniche.

In the course of the Upper Kimmeridgian the entire northern part of the basin was filled up by clastics what resulted in the development of vast fluvial areas. Towards south, clastics turned gradually into marine facies. Due to mainly progradational filling of the marin basin and due to a certain stop of hinterland activity, a thin sheet of reefoid limestones spread diachronously across the basin. The Ota limestone already underwent karstification due to local tectonic uplift. Limestone development also took place between Sintra and Lisbon. It is, though, not clear whether the entire center became occupied by limestone deposits. In the Serra da Arrábida terrestrial siliciclastics were gradually prograding westwards.

At the top of the Kimmeridgian, terrestrial clastics were advancing from the north, resulting in the progradation of an estuarine delta towards south. Clastics were outwedging towards southeast due to local uplifts controlling sediment distribution. East-derived clastics appear at Barreiro, closeby Lisbon. In the Serra da Arrábida terrestrial clastics, underlain by a sheet of calciclastics, prograded markedly towards west. The Sesimbra diapir stopped rising, thus having no further control on sedimentation. Towards the Kimmeridgian upper boundary, a slight transgression is obvious, being the prelude of a general sea level rise in the early Tithonian.

In the course of this sea level rise marine facies reappeared in regions formerly abandoned by the sea. Possibly due to a certain stop of terrigenous input from the western borderlands, the sea occupied a narrow stripe along the recent coast between Sta. Cruz and Cabo Mondego. At the peak of the transgression, close at the base of the Tithonian, limestones appeared at both localities. Between these sites, higher clastic input prevented limestone formation. Freshwater admixture was high at Peniche, resulting in brackish facies. Further south, the basin was occupied by a low-energy sequence of micritic limestones and marls. Local high-energy facies points out uplifts in the Mataçães–Montejunto area and SE Arruda dos Vinhos. Local high thicknesses along the recent Tejo valley were certainly caused by wrench tectonics. In the Serra da Arrábida the sea only remained in the western part, producing a mixed carbonate/clastic sequence reflecting coast line fluctuations caused by short-termed tectonic hinterland uplifts.

Still in the very early Tithonian the sea disappeared again from the newly flooded areas, withdrawing even more than before. Concomitantly tectonic activity, particularly along the western basin margin, started again. It caused an accentuated pushing back of the sea due to sedimentary overcompensation. Marly limestone sedimentation was restricted to the basin center and to a lagoonal protrusion towards north which was confined and sheltered from clastics by the Mataçães–Montejunto uplift and by the uplift zone east of Arruda. Transition towards shallow marine clastics further north occurred along a stable line demonstrating equilibrium of clastic deposition with subsidence. During the mid Lower Tithonian shallowing in the marine parts of the basin resulted in widespread development of higher energetic sediments.

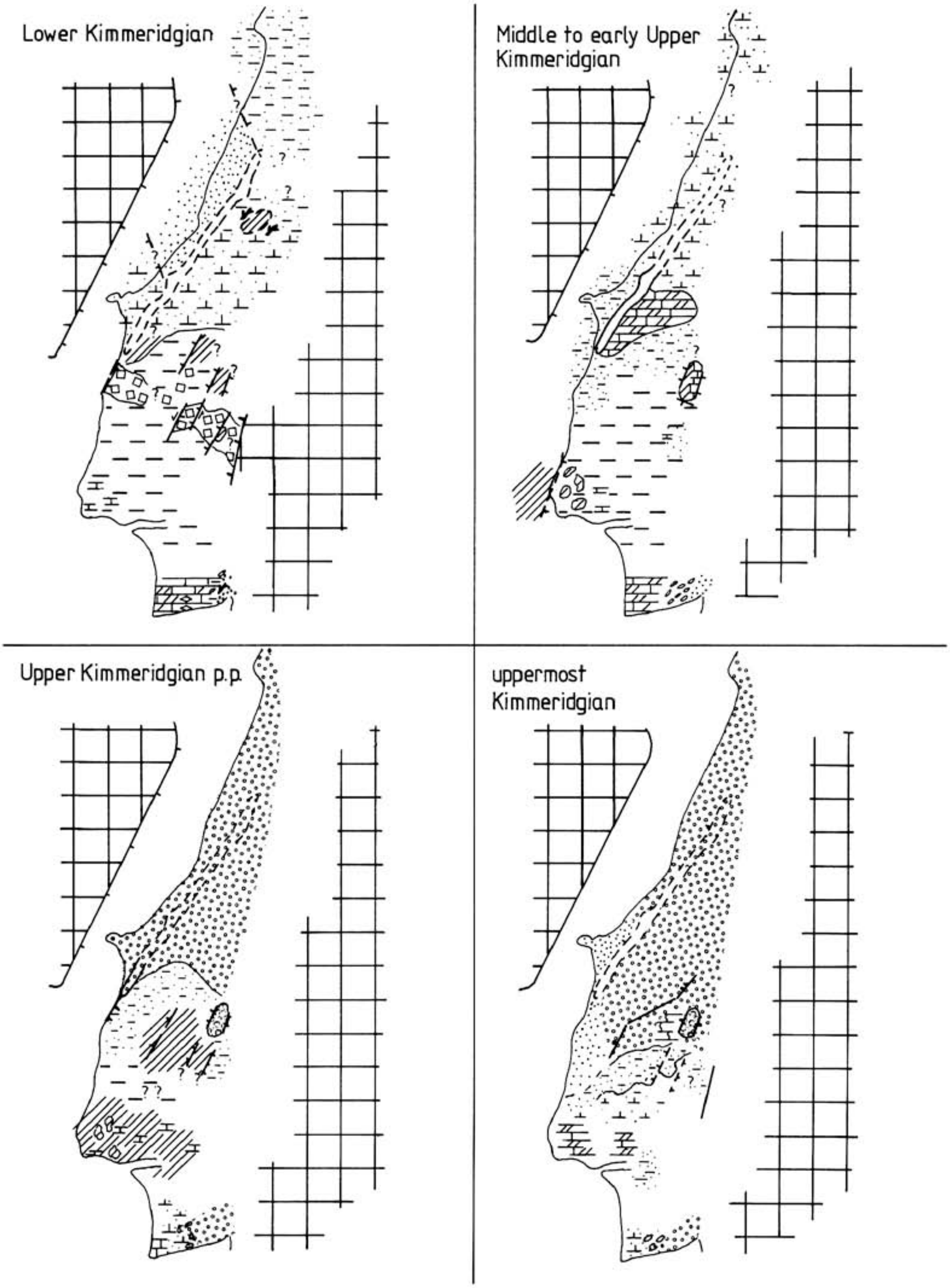
Basin filling by introduced clastics continued during the late Lower and Upper Tithonian, resulting in a further continuous push-back of marine regions. Tectonic and halokinetic intrabasin differentiation ceased, so that a simple facies pattern was produced: The constantly shrinking marine area was characterized by a mixed carbonate/marl/clastic, lagoonal sedimentation directly reflecting hinterland uplifts and eventually further global sea level oscillations. Clastics prograded diachronously from the north, northwest and east. They reached Sintra only at the Cretaceous boundary.

Continuous marginal marine, Purbeck-type sedimentation into the Lower Cretaceous eventually persisted only in the Lisbon and Sintra area (Ramalho 1971, Rey 1979), whereas the rest of the basin was lifted up. This resulted in the widespread development of an erosional subaerial unconformity, the »Neocomian Surface« (Vanney & Mougenot 1981).

Conclusions

1) Kimmeridgian and Tithonian sediments in the Lusitanian Basin were deposited in very variable settings, including basinal areas, slope environments, slope break buildups, low and high-energy lagoonal settings, delta systems, lakes and alluvial areas.

2) Basinal, slope, lagoonal and lake environments may likewise display siliciclastic and carbonate facies. The respective development is controlled by biologic and ecologic factors, and particularly by the interaction between energy regime and amount of clastic influx which is a function of tectonic activity: Calcareous lake facies, for instance, only developed during tectonically calm phases of ceasing clastic hinterland supply. Carbonate precipitation was, however, dependant on an uplifted closeby lime source area and on the biologic activity of algae. Marine carbonate buildups were related to tectonic or halokinetic uplifts and suitable environmental conditions (e.g., normal salinity) resulting in strong biologic activity. Allodapic carbonates originating from slope break buildups contaminated siliciclastically dominated slope settings when slopes were tectonically steepened. High-energy carbonates were also created in clastic shoaling-up zones, either during intervals of low sediment input or due to progradational migra-


Lower Kimmeridgian
Middle to early Upper Kimmeridgian
Upper Kimmeridgian p.p.
uppermost Kimmeridgian

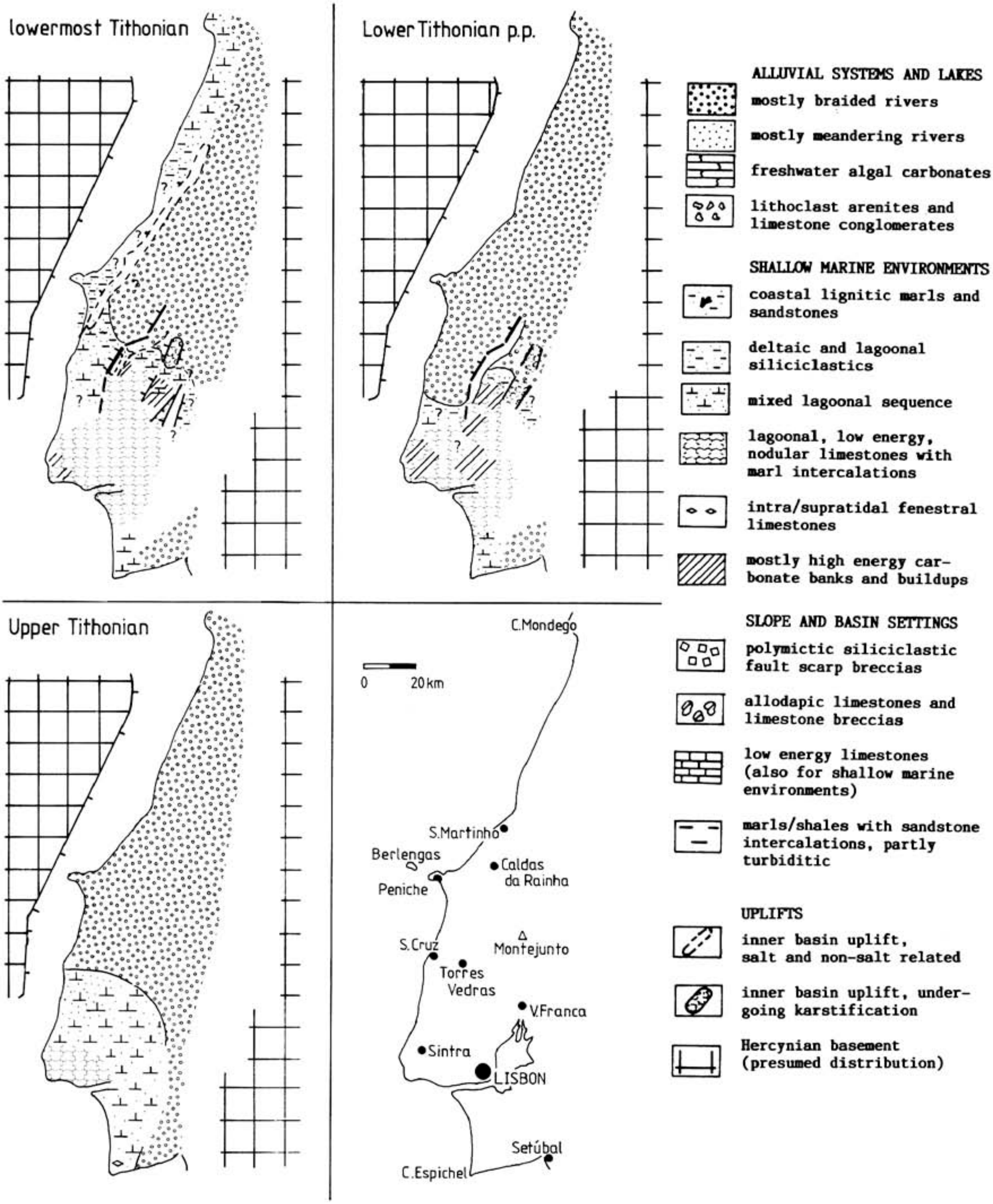


Fig. 6. Paleogeography and facies distribution of the Lusitanian Basin during the Kimmeridgian and Tithonian (partly after R. C. L. Wilson 1979 and Werner 1986).

tion of inner basin high-energy settings. Limestone development in low-energy, basinal or lagoonal environments represents either times of very low clastic influx, or was due to a certain distance from clastic sourcelands, or was enabled by effective, subsiding clastic traps and uplifted clastic fences.

Examples of tectonic control on clastic sedimentation are turbiditic and debris flow deposits, and changes from high sinuosity to low sinuosity fluvial systems, reflecting uplift of source areas. Very important as source area for clastics was an uplifted basement block west of the present-day Portuguese coast. During the Kimmeridgian, most clastics entered between Consolação and S. Martinho, resulting in complete overcompensation of basin subsidence during the Lower Kimmeridgian and in local brackish water characteristics of the marine transgression at the Kimmeridgian/Tithonian boundary. Further south, between Sintra and Cabo Espichel, no influence of west-derived clastics is obvious. A westward shallowing is, however, notable between Lisbon and Sintra.

3) Effects of halokinesis on facies distribution and thicknesses were widespread during the Kimmeridgian and, to a lesser extent, Tithonian. Uprise of salt diapirs was, however, only accompanying or posterior to rift tectonics which reactivated late Hercynian and Triassic faults. Occasionally, salt might have invaded uplift structures at a much later time. Block faulting and tilting without diapiric influence also result in similar facies pattern.

4) Influence of eustatic sea level changes on sedimentation is difficult to decipher. Tectonic control commonly led to a mosaic-like facies pattern and, together with common overcompensation of bathymetric deepening by clastic sediment supply, frequently overprinted eventual eustatic effects. An exception is the base of the »Pteroceriano« beds which is clearly transgressive north of Lisbon. Sediment characteristics evoke a rapid, mainly synchronous spreading of limestone facies over major parts of the basin at the peak of the transgression. In the basin center this transgression is less obvious, to be more distinct again in the southernmost part of the basin. There, this main correlation level (MCL) can be biostratigraphically dated as Kimmeridgian/Tithonian boundary. Thus, the MCL might probably correspond to a VAIL et al. (1984) peak in the global sea level curve.

Approximation of geohistory methods reveals that two intercalations of marine limestones in Tithonian terrestrial clastics south of Sta. Cruz may correspond to two transgressive peaks south of Arruda. They might either represent two further intra-Tithonian global eustatic peaks of the VAIL curve or are, more likely, due to local block faulting.

5) Geohistory curves are normally worked out after a refined biostratigraphic framework is established and calibrated by absolute age values. Simplified application of geohistory methods on the studied sequence provides, however, an useful tool to unravel the mutual influence of basin subsidence, hinterland activity and eustacy. For this purpose bathymetric log interpretation and cumulative thicknesses were plotted versus time, by using three synchronous and datable MCL (respective base of Kimmeridgian, Tithonian, Cretaceous) and averaging bulk accumulation rates for the respective intervals. This results in many time-correlatable events such as the simultaneous onset of *Anchispirocyclina* in settings where ecologic conditions were favourable (2 of 5 sections with *Anchispirocyclina*), and the coinciding maximum frequency of the form (all 5 sections with *Anchispirocyclina*), or the matching of downfaulting intervals. Thus, simplifications seem acceptable and permit a good degree of confidence on the validity of the resulting pattern. The method supports and emphasizes the overall dominance of tectonic activity on the sedimentary record (fig. 5):

After rapid deepening of the sedimentation base at the Oxfordian/Kimmeridgian boundary, a general shallowing started sooner or later in all marine parts of the basin, reflecting increased influx of clastic material and most likely a slight drop of eustatic sea level. Tectonic effects are well visible in Sintra (allodapic limestones), Arruda (wrench fault basin subsidence) and by the rapid and complete filling of marine basin parts with clastics in the north. A certain stop of tectonic activity led at the same time (±2 MA above Kimmeridgian base) to deepening of the sedimentation base in Arruda (fault scarp breccia to basinal marls), Sta. Cruz, Porto Novo, Consolação (braided rivers to meandering rivers), S. Martinho and Cabo Mondego (short-termed marine ingression). Further stop of tectonic uplift and related stop of clastic input at the end of the Kimmeridgian resulted in shallow water carbonate sedimentation, keeping pace with subsidence (Barreiro, Monsanto?), and start of a marine transgression in the north, accompanied by a general sea level rise. The western Arrábida exhibited a different development. Due to low basin subsidence and lacking terrigenous influx, carbonate sedimentation was continuously keeping pace with the water level.

During the Tithonian, sedimentation was generally in equilibrium with subsidence in the south and center of the basin, partly mirroring small tectonic pulses. In the north, complete overcompensation of subsidence resulted in overall terrestrial facies. Shallowing starts simultaneously in Monsanto (interpretation a), Arruda, Porto Novo and Consolação, but earlier in Sintra. Shallowing was possibly caused by regional uplift, with Sintra being a local prelude. Further shallowing/deepening phases occur then only locally. Tectonic deepening happened at the same time in Monsanto (case a) and Sintra; simultaneously shallowing is also evident for the Cabo Espichel and neighboured section Barreiro. Coeval are furthermore eventually two deepening events in Arruda and Sta. Cruz (see above). Increased clastic influx and further shallowing appears diachronously in all marine parts of the basin towards the Cretaceous boundary, indicating a general uplift of the basin and its margins which was eventually accompanied by a slow eustatic drop.

6) Geohistory methods have also some implications on Portuguese biostratigraphy. Thus, in the Lusitanian Basin, degree of synchronism is much higher when correlating the maximum frequency of *Anchispirocyclina lusitanica* in the Upper Tithonian, rather than the first occurrence of the foraminifer.

Discrepancies arise when dating the Upper Kimmeridgian by the alga *Clypeina jurassica*. Further investigations are needed to find out whether the form also may frequently appear in the Tithonian or whether presuppositions of the applied geohistory methods are locally not valid.

7) The development of the Lusitanian Basin is integrated into the general development of the North Atlantic rift system. The Upper Jurassic sequence of the basin reflects a second major rifting phase, related to spreading activity in the southern North Atlantic. During the Kimmeridgian extensional and wrenching forces caused rapid subsidence. Subsidence slowed down considerably during the Tithonian and uplifts occurred more widespread. This evidences thermal doming of the axial part of the North Atlantic rift what led to partial deep truncation of sediments down to Triassic deposits during the final Jurassic and early Cretaceous (P. A. ZIEGLER, pers. comun.).

Other major North Atlantic marginal basins commonly display similar development. Thus, the Northwest African basins and the U.S. eastern basins also subsided along major Hercynian fault zones; they exhibit an Upper Triassic/Lower Liassic initial rift sedimentation and subsequent carbonate platform or mixed clastic/carbonate development (cf. LANCELOT & WINTERER 1980, JANSA & WIEDMANN 1982). A major unconformity at the base of the Cretaceous is very obvious in the East Coast basins (McWHAE 1981). No Lower Oxfordian regression or hiatus is, however, evident in other basins except for some Grand Banks basins (cf. GRADSTEIN 1979) and Upper Jurassic carbonate development is commonly more widespread (JANSA 1981, JANSA & WIEDMANN 1982). Basin shape is often more assymetric than in the Lusitanian Basin, resulting in unidirectional shelf-shelfbreak-slope patterns. The Upper Jurassic of the Nova Scotian Basin is a well studied example, consisting from NW to SE of the terrestrial to marginal marine Mic Mac sandstones, the Baccaro ramp to platform carbonates of the Abenaki Formation, and the deep water shales of the Verril Canyon Formation (ELIUK 1978, McWHAE 1981). Commonly, these facies patterns become overriden by a final Jurassic clastic wedge, prograding towards the Atlantic axis (cf. McWHAE 1981, JANSA & WIEDMANN 1982). Carbonate platforms may also sharply break off to form carbonate gravity flow deposits interfingering with pelagic limestones, as it is true for the Upper Jurassic of the Moroccan Mazagan Platform (JANSA et al. 1984, STEIGER & JANSA 1984).

Neighboured European basins such as the Northern Spain, Aquitaine and Algarve Basins, but also the Moroccan Atlas Basin developed perpendicular or oblique to the Atlantic main rift axis along major transform directions. They commonly exhibit a fairly different development, strongly influenced by the local setting. Thus, the Upper Jurassic of the Aquitanian Basin, distant from large basement uplifts, is dominated by carbonate and marl facies (DELFAUD & DELVALLE 1980, ENAY et al. 1982). The Northern Spain Basin is characterized by marginal marine to brackish sandstones or a hiatus during the entire Oxfordian, and by a mixed clastic/carbonate development with differentiation into terrestrial, marginal marine and deeper marine settings during the late Upper Jurassic (R. C. L. WILSON 1975a, ALVARRADO 1980). Major clastic input occurred, however, only during the Lower Cretaceous (R. C. L. WILSON 1975a). The Algarve and Atlas Basins partly connected Atlantic rifting zones and Tethyan pelagic regions. Hence, the Upper Jurassic sequences are constituted of intercalating shallow marine, partly brackish, carbonates, talus sediments and pelagic deposits (cf. ROCHA & MARQUES 1979, MOUTERDE et al. 1979, RAMALHO 1981, STETS & WURSTER 1982, HÜSSNER 1985).

Acknowledgements

Chris Wilson, Graham Hill (Milton Keynes) and Rick Hiscott (St. John) introduced me to the coastal sections at Porto Novo and S. Martinho, and to the Vila Franca area. Franz Fürsich, Winfried Werner (Munich) and Norbert Schmidt-Kittler (Mainz) made me understand the Cabo

Espichel, Santa Cruz, Consolação and Columbeira sections. Miguel Ramalho (Lisbon) accompanied me to the eastern Serra da Arrábida and the Arruda region. For this and for the accompanying informations and fruitful discussions my special thanks are due to them all.

I am greatly thankful to Peter Ziegler (Den Haag) for critically reading chapter 3 and for his useful comments.

The »Serviços Geológicos de Portugal«, Lisbon, kindly granted working permission.

References

AIGNER, T. (1985): Storm depositional systems. Dynamic stratigraphy in modern and ancient shallow-marine sequences. – Lecture Notes in Earth Science, **3**, 174 pp.

ALVARADO, M. (1980): Introducción a la geología general de Espanha. – Bol. Geol. Mineralog., **91**, 1–65.

ANDRADE, C. F. (1934): Um recife corálico no Jurássico de Vila Franca de Xira. – Bol. Mus. Miner. Geol. Univ. Lisboa, **3**, 1–11.

AZÉMA, J., CHABRIER, G., FOURCADE, E. & JAFFREZO, M. (1977): Nouvelles données micropaléontologiques, stratigraphiques et paléogéographiques sur le Portlandien et le Néocomien de Sardaigne. – Rev. Micropaleont., **20**, 125–176.

BALDY, P. BOILLOT, G., DUPEUBLE, P. A., MALOD, J., MOITA, I. & MOUGENOT, D. (1977): Carte géologique du plateau continental sud-portugais et sud-espagnol (Golfe de Cadix). – Bull. Soc. Géol. France, (7), **19**, 703–724.

BAYER, U. & SEILACHER, A. (eds) (1985): Sedimentary and evolutionary cycles. – Lecture notes in Earth Sciences, **1**, 465 pp., Berlin (Springer).

BENEST, M.-C., COIFFAT, P., FOURCADE, E. & JAFFREZO, M. (1975): Essai de déterminaticn de la limite Jurassique – Cretacé par l'étude des microfaciès dans les séries de la plate-forme du domaine méditerranéen occidental. – Colloque sur la limite Jurassique – Crétacé, Lyon, Neuchatel, 1973 – Mém. Bull. Rech. Géol. Min., **86**, 169–181.

BENKE, K. (1981): Die Dogger/Malm-Wende in den NW-Keltiberischen Ketten (Spanien) und angrenzenden Gebieten – Sedimentologie, Stratigraphie und Paläogeographie. – Facies, **4**, 95–164.

BERTHOU, P. Y. & LAUVERJAT, J. (1979): Essai de synthèse paléogéographique et paléobiostratigraphique du bassin occidental portugais au cours du Crétacé Supérieur. – Ciências da Terra (UNL), **5**, 121–144.

BIJU-DUVAL, B., DERCOURT, J. & LEPICHON, X. (1977): From the Tethys Ocean to the Mediterranean Seas. A plate tectonic model of the evolution of the western Alpine System. – Int. Sym. Struct. Hist. Mediterr. Basins, Split 1976, 143–164, Paris (ed. Technip).

BOILLOT, G., DUPEUBLE, P.-A. & MUSELLEC, P. (1975): Carte géologique du plateau continental nord-portugais. – Bull. Soc. Géol. France, (7), **17**, 462–480.

–, MALOD, J. A. & MOUGENOT, D. (1979): Evolution géologique de la marge ouest-iberique. – Ciências da Terra, **5**, 215–222.

BROWN, S. (1984): Jurassic. – In K. W. GLENNIE (ed), Introduction to the petroleum geology of the North Sea, 103–131, Oxford (Blackwell).

CHAUMEAU, J. (1962): Etude géologique de l'anticlinorium d'Arruda et de ses abords. – C.P.P. and Compagnie des Pétroles France Afrique, Lisbon (unpubl. report.).

CHOFFAT, P. (1882): Note préliminaire sur les vallées tiphoniques et les éruptions d'ophite et de teschenite en Portugal. – Bull. Soc. Géol. France, (3), **10**, 267–295.

DELFAUD, J. & DELVALLE, J. (1980): La cuenca de Aquitania y su margen Pirenaica. – Bol. Geol. Miner., **91**, 107–142.

DEWEY, J. F., PITMAN, W. C., III, RYAN, W. B. F. & BONNIN, J. (1973): Plate tectonics and the evolution of the Alpine System. – Bull. Geol. Soc. Amer., **84**, 3137–3180.

DÖHLER, M. (1984): Die regressive Flachwassersedimentation im höheren Kimmeridge der mittleren Estremadura (Portugal) zwischen Torres Vedras und Alenquer. – DFG-Programm Biofaziesanalyse im Mesozoikum der Estremadura (Portugal), Schm. 452/5, 39 pp, University of Munich (unpublished report).

DRAGASTAN, O. (1975): Upper Jurassic and Lower Cretaceous microfacies from the Bicaz Valley Basin (East Carpathian). – Mem. Inst. Geol. Geophys., **21**, 87 pp.

DUFF, P. M. D., HALLAM, A. & WALTON, E. K. (1967): Cyclic sedimentation. – Developm. Sedimentol., **10**, 280 pp.

EHRMANN, W. U. & THIEDE, J. (1986): Correlation of terrigenous and biogenic sediment fluxes in the North Atlantic Ocean during the past 150 my. – Geol. Rdsch., **75**, 43–55.

EINSELE, G. (1982): Limestone – marl cycles (periodites): diagenesis, significance, causes – a review. In: G. EINSELE & A. SEILACHER (eds); Cyclic and event stratification, 9–53, Berlin (Springer).

– & SEILACHER, A. (eds.) (1982): Cyclic and event stratification. – 536 pp., Berlin (Springer).

ELIUK, L. S. (1978): The Abenaki Formation, Nova Scotia Shelf, Canada – A depositional and diagenetic model for a Mesozoic carbonate platform. – Bull. Canad. Soc. Petrol. Geol., **26**, 424–514.

ELLIS, P. M. (1983): A Cretaceous encrusted surface on the Upper Jurassic limestones of Monte Gordo, Vila Franca de Xira, Portugal. – Comun. Serv. Geol. Portugal, **69**, 103–106.

– (1984): Upper Jurassic carbonates from the Lusitanian Basin, Portugal, and their subsurface counterparts in the Nova Scotian Shelf. – 193 pp, unpublished phD-thesis, Open University, Walton Hall.

ENAY, R., MANGOLD, C., CARIOU, E., CONTINI, D., DEBRAND-PASSARD, S., DONZE, P., GABILLY, J., Lefavrais-Raymond, A., Mouterde, R. & Thierry, J. (1980): Synthèse paléogéographique du Jurassique Français. – Docum. Lab. Géol. Lyon, **5**, 210 pp.

Farinacci, A. & Radoicic, R. (1964): Correlazione fra serie gioresi e cretacee dell'Apennino Centrale e delle Dinaridi esterne. – Rendeconti, (2), **7**, 269–300.

Felber, J., Hoffmann, R., Leinfelder, R., Oschmann, W. & Ziegler, R. (1982): Biofaziesanalyse randmariner Ablagerungsbereiche im Oberen Jura Portugals. II. Paläogeographie und Faziesverteilung in der Serra da Arrábida. – N. Jb. Geol. Paläont. Abh., **163**, 301–330.

Fischer, J. C. & Palain, C. (1971): Nouvelles observations sédimentologiques et paléobiologiques sur l'Hettangien du Portugal. – Comun. Serv. Geol. Portugal, **13**, 65–83.

Fürsich, F. T. (1981): Salinity-controlled benthic associations from the Upper Jurassic of Portugal. – Lethaia, **14**, 203–223.

– & Schmidt-Kittler, N. (1980): Biofacies analysis of Upper Jurassic marginally marine environments of Portugal I. The carbonate-dominated facies at Cabo Espichel. With a contribution of M. Ramalho. – Geol. Rdsch., **69**, 943–981.

– & Werner, W. (1984): Salinity zonation of benthic associations in the Upper Jurassic of the Lusitanian Basin (Portugal). – Géobios, Mém. sp., **8**, 85–92.

– & – (1986): Benthic association and their environmental significance in the Lusitanian Basin (Upper Jurassic, Portugal). – N. Jb. Geol. Paläont. Abh., **172**, 271–329.

Gabilly, J., Cariou, E. & Hantzpergue, P. (1985): Les grandes discontinuités stratigraphiques au Jurassique: témoins d'événements eustatiques, biologiques et sédimentaires. – Bull. Soc. Géol. France, (8), **1**, 391–401.

García-Hernandes, M. & López-Garrido, C. (1979): El tránsito Jurásico-Cretácico en la zona Prebética. – Cuad. Geolog., **10**, 535–544.

Gradstein, F. M. (1979): Jurassic micropaleontology of the Grand Banks. – Ciências da Terra (UNL), **5**, 85–96.

Grau, G., Montadert, L., Delteil, R. & Winnock, E. (1973): Structure of the european continental margin between Portugal and Ireland from seismic data. – Tectonophys., **20**, 319–339.

Guéry, F. (1984): Evolution sédimentaire et dynamique du bassin marginal Ouest-Portugais au Jurassique (Province de Estremadura, secteur de Caldas da Rainha – Montejunto). (Abstract). – Unpubl. thesis, N° 84–01, Université Claude Bernard, Lyon.

Hallam, A. (1971a): Facies analysis of the Lias of West Central Portugal. – N. Jb. Geol. Paläont. Abh., **139**, 226–265.

– (1971b): Mesozoic geology and the opening of the North Atlantic. – J. Geol., **79**, 129–157.

– (1975): Jurassic environments. – 269 pp, Cambridge (Cambr. Univ. Press).

– (1978): Eustatic cycles in the Jurassic. – Paleogeogr., Palaeoclimat., Palaeoecol., **23**, 1–32.

– (1984): Pre-Quarternary sea-level changes. – Ann. Rev. Earth Planet. Sci., **12**, 205–243.

Hüssner, H. (1985): Jurassische Karbonate des westlichen Hohen Atlas (Marrokko): Mikrofaziesanalyse und plattentektonischer Rahmen. – Facies, **12**, 141–218.

Hutchinson, R. W. & Engels, G. G. (1972): Tectonic evolution in the southern Red Sea and its possible significance to older rifted continental margins. – Bull. Geol. Soc. Amer., **83**, 2989–3001.

Jackson, M. P. A. & Talbot, C. J. (1986): External shapes, strain rates, and dynamics of salt structures. – Bull. Geol. Soc. Amer., **97**, 305–323.

Jansa, L. (1981): Mesozoic carbonate platforms and banks of the eastern North American margin. – Marine Geol., **44**, 97–117.

–, Bujak, J. P. & Williams, G. L. (1980): Upper Triassic salt deposits of the western North Atlantic. – Canad. J. Earth. Sci., **17**, 547–559.

–, Steiger, T. H. & Bradshaw, M. (1984): Mesozoic carbonate deposition on the outer continental margin off Morocco. – Init. Rep. DSDP, **79**, 857–891.

– & Wiedmann, J. (1982): Mesozoic-Cenozoic development of the eastern North American and Northwest African continental margins: a comparison. – In: U. v. Rad, K. Hinz, M. Sarntheim & E. Seibold (eds), Geology of the Northwest African continental margin, 215–269, Berlin (Springer).

Kent, D. & Gradstein, F. M. (1985): A Cretaceous and Jurassic Geochronology. – Bull. Geol. Soc. Amer., **96**, 1419–1427.

Kristofferson, Y. (1978): Sea floor spreading and the early opening of the North Atlantic. – Earth Planet. Sci. Letters, **38**, 273–290.

Lancelot, Y. & Winterer, E. L. (1980): Evolution of the Moroccan oceanic basin and adjacent continental margin – a synthesis. – Init. Report DSDP, **50**, 801–821.

Leinfelder, R. R. (1983): New mapping results on sheet Setúbal (Sesimbra to Portinho da Arrábida, Serra da Arrábida, Portugal). – Comun. Serv. Geol. Portugal, **69**, 295–324.

– (1985): Cyanophyte calcification morphotypes and depositional environments (Alenquer Oncolite, Upper Kimmeridgian?, Portugal). – Facies, **12**, 253–274.

– (1986): Facies, stratigraphy and paleogeographic analysis of Upper? Kimmeridgian to Upper Portlandian sediments in the environs of Arruda dos Vinhos, Estremadura, Portugal. – Münch. Geowiss. Abh. (A), **7**, 215 pp.

Le Pichon, X. (1980): Subsidence, transgressions, régressions: notions élémentaires sur les mécanismes physiques. – In: Stratigraphie et paléogéographie: principes et méthodes, 135–147, Paris (Doin Edit.).

Maufret, A., Boillot, G., Auxietre, J. L. & Dunant, F. P. (1978): Evolution structurale de la marge continentale au Nord-Ouest de la péninsule Ibérique. – Bull. Soc. Géol. France, **20**, 375–388.

McWhae, J. R. H. (1981): Structure and spreading history of the northwestern Atlantic region from the Scotian Shelf to Baffin Bay. – In: J. W. Kerr & A. J. Fergusson (eds), Geology of the North Atlantic borderlands, Canad. Soc. Petrol. Geol., Mem., **7**, 299–332.

Mempel, G. (1955): Zur Paläogeographie des Oberen Jura in Mittel-Portugal. – Zeitschr. Deutsch. Geol. Gesellsch., **105**, 106–126.

Miall, A. D. (1984): Principles of sedimentary basin analysis. – 490 pp., New York (Springer).

– (1986): Eustatic sea level changes interpreted from seismic stratigraphy: a critique of the methodology with particular reference to the North Sea Jurassic record. – Bull. Amer. Ass. Petrol. Geol., **70**, 131–137.

Mitchum, R. M. Jr. & Uliana, M. A. (1985): Seismic stratigraphy of carbonate depositional sequences, Upper Jurassic – Lower Cretaceous, Neuquén Basin, Argentina. – Amer. Ass. Petrol. Geol., Mem., **39**, 255–274.

Montenat, C. & Guéry, F. (1984): L'intrusion diapirique de Caldas da Rainha et l'halocinèse Jurassique sur la marge Portugaise. – C. R. Acad. Sci. Paris, **298** (ser. 2, 20), 901–906.

Mougenot, D., Monteiro, J. H., Dupeuble, P. A. & Malod, J. A. (1979): La marge continentale Sud-Portugaise: évolution structurale et sédimentaire. – Ciências da Terra (UNL), **5**, 223–245.

Mouterde, R., Ramalho, M., Rocha, R. B., Ruget, C. & Tintant, H. (1972): Le Jurassique du Portugal. – Esquisse stratigraphique et zonale. – Bol. Soc. Geol. Portugal, **18**, 73–104.

–, Rocha, B., Ruget, C. & Tintant, H. (1979): Faciès, biostratigraphie et paléogéographie du Jurassique Portugais. – Ciências da Terra (UNL), **5**, 29–52.

Oertel, G. (1956): Transgressionen im Malm der portugiesischen Estremadura. – Geol. Rdsch., **45**, 304–314.

Palain, C. (1976): Une série détritique terrigène, les »Grès de Silves«: Trias et Lias Inférieur du Portugal. – Serv. Geol. Portugal, Mem., **25**, 377 pp.

Pitman, W. C. III. (1978): Relationship between eustacy and stratigraphic sequences of passive margins. – Bull. Geol. Soc. Amer., **89**, 1389–1403.

– & Golovchenko, X. (1983): The effect of sea level change on the shelfedge and slope of passive margins. – Soc. Econ. Paleont. Mineral., Sp. P., **33**, 41–58.

Ramalho, M. M. (1971): Contribution à l'étude micropaléontologique et stratigraphique du Jurassic Supérieur et du Crétacé Inférieur des environs de Lisbonne (Portugal). – Serv. Geol. Portugal, Mem., **19**, 212 pp.

– (1981): Note préliminaire sur le microfacies du Jurassique Supérieur Portugais. – Comun. Serv. Geol. Portugal., **67**, 41–45.

Rey, J. (1979): Le Crétacé Inférieur de la marge Atlantique Portugaise: biostratigraphie, organisation sequentielle, évolution paléogéographique. – Ciências da Terra (UNL), **5**, 97–120.

Ribeiro, A., Antunes, M. T., Ferreira, M. P., Rocha, R. B., Soares, A. F., Zbyszweski, G., Moitino de Almeida, F., Carvalho, D. de & Monteiro, J. H. (1979): Introduction à la géologie générale du Portugal. – Publ. Serv. Geol. Portugal, 114 pp.

Rocha, R. B. & Marques, B. (1979): Le Jurassique de l'Algarve (Portugal): Esquisse stratigraphique et évolution paléogéographique. – Cuad. Geolog., **10**, 35–43.

Ruget-Perrot, C. (1961): Etude stratigraphique sur le Dogger et le Malm Inférieur du Portugal au Nord du Tage – Bajocien, Bathonien, Callovien, Lusitanien. – Serv. Geol. Portugal, Mem., **7**, 197 pp.

Schmidt, D. (1984): Biofazielle und sedimentpetrographische Untersuchung des Diapirs von Porto de Mós (Mittelportugal). – 193 pp, unpubl. thesis, Freie Universität, Berlin.

– (1985): Faziesausbildung und Diapirismus im Oberjura von Mittelportugal am Beispiel des Diapirs von Porto de Mós. – Berliner geowiss. Abh., (A), **60**, 49–89.

Schwan, W. (1980): Geodynamik peaks in Alpinotype orogenies and changes in ocean-floor spreading during late Jurassic – late Tertiary time. – Bull. Amer. Ass. Petrol. Geol., **64**, 359–373.

Sloss, L. L. (1984): Comparative anatomy of cratonic unconformities. Amer. Ass. Petrol. Geol. Mem, **36**, 1–6.

Stachel, T. (1986): Bericht zur zweimonatigen Kartierung im Raum SW Sobral de Monte Agraço (westliches Portugal). – unpubl. report, Universität Mainz.

Steiger, T. & Jansa, L. F. (1984): Jurassic limestones of the seaward edge of the Mazagan carbonate platform, Northwest African continental margin, Morocco. – Init. rep. DSDP, **29**, 449–491.

Stets, J. & Wurster, P. (1982): Atlas and Atlantic – Structural relations. – In: U. v. Rad, K. Hinz, M. Sarntheim & E. Seibold (eds.), Geology of the Northwest African continental margin, 67–85, Berlin (Springer).

Teixeira, C. & Gonçalves, C. (1980): Introduçao à geologia de Portugal. – 477 pp., Braga (Livraria Cruz).

Vail, P. R., Hardenbol, J. & Todd, R. G. (1984): Jurassic unconformities, chronostratigraphy and sea-level changes from seismic stratigraphy and biostratigraphy. – Amer. Ass. Petrol. Geol., Mem., **36**, 129–144.

–, Mitchum, R. M. Jr., Todd, R. G., Widmaier, J. M., Thompson, S. III, Sangree, J. B., Bubb, J. N. & Hatlelid, W. G. (1977): Seismic stratigraphy and global changes of sea level. – Amer. Ass. Petrol. Geol. Mem., **26**, 49–212.

– & Todd, R. G. (1981): Northern North Sea Jurassic unconformities, chronostratigraphy and sea-level changes from seismic stratigraphy. – In: L. V. Illing & G. D. Hobson (eds.), Petroleum geology of the continental shelf of North-West Europe, 216–235, London (Heyden Inst. Petrol).

Van Hinte, J. E. (1976): A Jurassic time scale. – Bull Amer. Ass. Petrol. Geol., **60**, 498–516.

– (1978): Geohistory analysis – application of micro paleontology in exploration geology. – Bull. Amer Ass. Petrol. Geol., **62**, 201–222.

Vanney, J.-R. & Mougenot, D. (1981): La plate-forme continentale du Portugal et les provinces adjacentes analyse geomorphologique. – Serv. Geol. Portugal Mem., **28**, 86 pp.

Werner, W. (1986): Palökologische und biofazielle Analyse des Kimmeridge (Oberjura) von Consolação, Mittel portugal. – Zitteliana, **13**, 109 pp.

WILSON, L. M. (1981): Circum-North Atlantic tectonostratigraphy reconstruction. – Canad. Soc. Petrol. Geol., Mem., **7**, 167–184.

WILSON, R. C. L. (1975a): Atlantic opening and Mesozoic continental margin of Iberia. – Earth Planet. Sci. Letters, **25**, 33–43.

– (1975b): Some examples of shoaling deposits from the Upper Jurassic of Portugal. – In: R. N. GINSBURG (ed.), Tidal deposits, 363–371, Berlin (Springer).

– (1979): A reconnaissance study of Upper Jurassic sediments of the Lusitanian Basin. – Ciências da Terra (UNL), **5**, 53–84.

– & WILLIAMS, C. A. (1979): Oceanic transform structures and the development of Atlantic continental margin sedimentary basins – a review. – J. Geol. Soc. London, **136**, 311–320.

WRIGHT, V. P. (1985): Algal marsh deposits from the Upper Jurassic of Portugal. – In: D. F. TOOMEY & M. H. NITECKI (eds), Paleoalgology, 330–341, Berlin (Springer).

– & WILSON, R. C. L. (1984): A carbonate submarine fan from the Jurassic of Portugal. – J. Sedim. Petrol., **54**, 394–412.

ZBYSZEWEKSI, G. (1959): Etude structurale de l'aire tiphonique de Caldas de Rainha. – Serv. Geol. Portugal, Mem., **3**, 182 pp.

– & MOITINHO DE ALMEIDA, F. (1960): Notícia explicativa da carta geológica de Portugal, 1:50000, folha 26-D, Caldas da Rainha. – Serv. Geol. Portugal, Lisbon.

– & TORRE DE ASSUNÇÃO, C. (1965): Notícia explicativa da carta geológica de Portugal, 1:50000, folha 30-D, Alenquer. – Serv. Geol. Portugal, Lisbon.

–, VEIGA FERREIRA, O. DA, MANUPPELLA, G. & TORRE DE ASSUNÇÃO, C. (1966): Notícia explicativa da carta geológica de Portugal, 1:50000, folha 30-B, Bombarral. – Serv. Geol. Portugal, Lisbon.

ZIEGLER, P. A. (1982): Geological atlas of Western and Central Europe. – Vol. **1, 2**, 130 pp & 109 pp, Shell Internat. Petrol. Maatschappij BV, Amsterdam (Elsevier).