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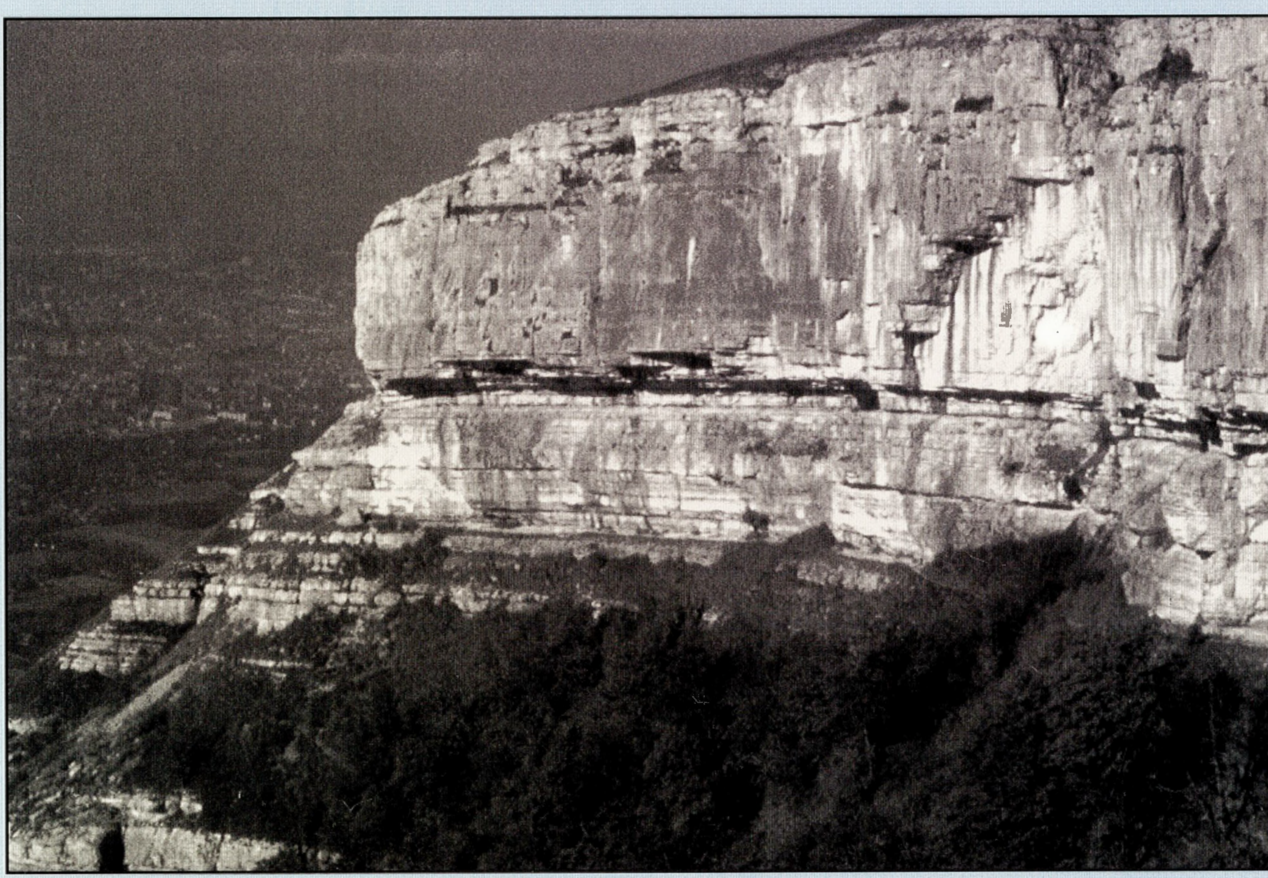

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The evolution of the French Jura platform during the Late Berriasian to Early Valanginian: controlling factors and timing

Sedimentology, sequence analysis, and correlation with successions in the Atlantic Atlas (Morocco) and the Vocontian Trough (SE France)

THÈSE

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"The things of the universe are not sliced off from one another with a hatchet, neither the hot from the cold nor the cold from the hot."

Anaxagoras

ABSTRACT

The objective of the present study is the understanding of the principal factors influencing the sedimentation of carbonate-dominated systems from the Late Berriasian to the Early Valanginian of the French and Swiss Jura (Pierre-Châtel, Vions, and Chambotte Formations). Climate, tectonic and eustatic changes are studied through analysis of facies, discontinuities, sequence-stratigraphy and cyclostratigraphy. The comparison with other paleogeographic settings, the hemipelagic to pelagic Vocontian Trough (France) and a mixed carbonate-siliciclastic ramp system in the Atlantic Atlas (Morocco), furnishes additional evidence on the interaction of factors controlling sedimentation and their temporal and spatial relevance.

Sedimentological analysis of 8 reference sections from the western Jura Mountains in France and 5 complementary sections from the central Swiss Jura, the French Vocontian basin and Moroccan Essaouira basin, permits the elaboration of depositional models of the Northern-Tethys and Atlantic margins in the earliest Cretaceous. The Jura platform was subdivided into a continental domain, coastal/tidal area, internal lagoon, open lagoon, barrier, and external lagoon. A ramp and slope/basin domain predominantly occur on the Atlantic margin and in the Vocontian Trough.

The importance of short-term breaks in sedimentation, which manifest the most rapid and substantial environmental changes in sedimentary systems, is taken into account by a detailed discontinuity analysis. On the basis of 8 criteria (geometry, lateral extent, morphology, biological activity, mineralization, facies contrast, diagenetic contrast, and biostratigraphy), the environmental relevance of discontinuity surfaces is assessed. Four groups of discontinuity can be distinguished, which are related to environmental changes indicating subaerial exposure, subaqueous omission, subaqueous erosion, and changes in texture and facies.

Quasi-periodic environmental changes are expressed in the stratigraphic record through repetitive variations of sedimentological and geochemical characteristics. Three hierarchies of such depositional sequences are evidenced and can be attributed to the effect of relative sea-level changes. Deposits and discontinuities on all scales correspond to sea-level lowstand, transgression, maximum flooding, sea-level highstand, and regression. Elementary, small-scale, and medium-scale sequences can therefore be described in terms of sequence- and cyclostratigraphic concepts. Superimposition of different frequencies of sea-level change leads to the multiplication of characteristic discontinuities. Superimposed sea-level falls cause repeated formation of sequence boundaries (SB). Superimposed initial- and maximum floodings favor the formation of transgressive (TS) and maximum-flooding (MF) surfaces, respectively. On all scales, depositional sequences bounded by these discontinuities can be differentiated. SB-sequences are characterized by subaerial exposure surfaces and deepening-shallowing trends, whereas TS- and MF-sequences are delimited by discontinuities or intervals indicating initial flooding and deepening-shallowing trend, and maximum flooding and shallowing-deepening trend, respectively. The stacking pattern of the different types of depositional sequences reflects relative sea-level changes on a larger scale.

Integration of biostratigraphic data allows correlation of depositional sequences not only across the platform, but also with basinal sections and sections on the Atlantic margin. It suggests that medium-scale and probably also small-scale sequences reflect environmental changes of intercontinental extent. Typical arrangements of sequences in 5:1/4:1 relationships point to sea-level and/or climatic changes that were in phase with insolation variations in the Milankovitch frequency band. Elementary sequences, interpreted to correspond to the precession cycle (20 ky), predominantly occur in the basin where the sedimentary system is less prone to the formation of autocycles, as compared to the highly dynamic platform environments. Small-scale sequences corresponding to the first eccentricity cycle (100 ky) are clearly evidenced on the Jura platform and in the Vocontian Trough, whereas on the Atlantic margin depositional sequences of a similar scale cannot be attributed to an external forcing factor. However, medium-scale sequences related to the second eccentricity cycle (400 ky) can unambiguously be correlated between Atlantic and Tethyan domains. Cyclostratigraphic analysis suggests a duration of 6 My for the studied interval (approx. 1.6 My for the Pierre-Châtel Fm., 2 My for the Vions Fm., 2.4 My for the Chambotte Fm.), which corresponds reasonably well to the radiometrically deduced duration of 5.5 My ± 4.8 My (Gradstein et al. 1995, Hardenbol et al. 1999*).

The correlation within a narrow framework of timelines gives evidence for several intervals of differential subsidence, as indicated by contemporaneous deposition and condensation and/or subaerial exposure. Tectonic activity expressed as accelerated subsidence and blockfaulting is probably related to changes in stress patterns due to accelerated rifting in the North Atlantic, changes in motion between the African and European plates and initial rotation of the Iberian block. Tectonic activity in the *Picteti/Alpillensis* and *Otopeta/Pertransiens* ammonite zones is correlatable between Atlantic and Tethyan domains and coincides with major sea-level drops postulated on the "eustatic" sea-level curve of Haq et al. (1987). There is no evidence for major sea-level drops of comparable amplitude in either domain suggesting that the "eustatic" events indicated by these authors probably contain an important tectonic component. Sequence boundaries

* released in 1999, but backdated to 1998

identified in the sequence-stratigraphic framework of Hardenbol et al. (1999) can be recognized, but they generally correspond to well-expressed boundaries of medium-scale sequences, whereas larger-scale, presumably 3rd-order relative sea-level trends, cannot be correlated between the studied domains.

Factors giving evidence for the general climatic evolution include subaerial exposures and associated diagenetic patterns underlined by stable isotope composition, siliciclastics (clay mineral composition, detrital quartz), presence and abundance of organic matter, and faunal assemblages. They indicate semi-arid conditions in the Purbeckian, a transitional climate for the Pierre-Châtel Formation, and more humid and seasonal conditions through the Vions and the Upper Chambotte Formations with a short (800 ky), more arid interlude in the Lower Chambotte Formation.

The evolution of the Jura platform as result of the interaction of eustatic sea-level changes, climate changes and tectonic activity can be summarized as follows:

1. After widespread progradation the Jura platform attained a flat-topped morphology at the end of the *Subalpina* zone (top Purbeckian). Tectonic activity initiated differential subsidence and/or local uplift during a lowstand in sea-level, which led to partial platform exposure (SB Be4). Resuming subsidence and a beginning sea-level rise on the 2nd order led to high-energy conditions in proximal platform positions on a now distally-steepened ramp morphology. However, effective carbonate production in a transitional, semi-arid climate and well-oxygenated environments allowed the platform to catch up with rising relative sea level and to prograde rapidly (early *Paramimounum* zone, Pierre-Châtel Formation).

2. Slowing relative sea-level rise, tectonic activity, and high-frequency, low-amplitude sea-level falls in the middle of the *Paramimounum* zone caused subaerial exposure with local karstification (SB Be5). Associated morphological changes in the hinterland, together with a more humid climate, explain the abrupt arrival of detrital quartz and organic material. For the duration of approximately 1 My the platform stayed in a state of keep-up as a result of continuing slow sea-level rise and low subsidence rates. The flat-topped platform morphology, attained through aggradation and progradation, recorded small-scale eustatic sea-level changes with widespread, repetitive exposures. The majority of siliciclastics and organic matter were trapped in shallow lagoons on the platform. This caused localized mesotrophic conditions, and the main area of carbonate production was thus restricted to the platform rim.

3. The sedimentary system began to change with elevated differential subsidence in the *Pictetí/Alpillensis* zones. Tidal influence and strong currents became predominant on the highly structured platform that was marked by isolated barrier islands. A relative sea-level rise is recorded and is due to generally elevated subsidence, accelerated sea-level rise on the 2nd order, and low aggradation potential of the platform, in environments that were stressed by continuing detrital input.

4. The detrital input then gradually disappeared in the upper part of the *Alpillensis* zone as a response to a more arid climate and effective winnowing in predominantly high-energy environments. The backstepping of depositional environments and continuing differential subsidence led to a distally-steepened ramp morphology and absence of barrier systems in the lowest Chambotte Formation. However, high carbonate production in the now oligotrophic environments again induced platform progradation and shoaling at the top of the *Otopeta* zone.

5. An abrupt tilt of the platform with significantly enhanced differential subsidence is evidenced by deep truncation and subaerial exposure in proximal platform positions and deep-ramp conditions with renewed siliciclastic input in more distal positions. A transitional climate with elevated seasonality is evidenced by high activity of storms and input of siliciclastics and organic matter. However, carbonate production was sufficient to compensate again for the subsidence pattern and induce local progradation (Upper Chambotte Fm.), only to be interrupted again by a phase of intensified differential subsidence and a platform crisis at the top of the *Campylotoxus* zone.

Short-term climatic and sea-level variations are responsible for bathymetric variations and facies distribution in platform and basin environments on a local to regional scale. However, depending on the general configuration and sensitivity of the sedimentary system, patterns of environmental change can be correlated over large distances. Climatic changes and episodic tectonic activity on the scale of 1 to 2 My mainly influence platform morphology, and presence and distribution of siliciclastics. This in turn influences to an important extent the efficiency of carbonate production and patterns of platform progradation, retrogradation and aggradation on a regional scale. Internal feedback mechanisms between sea-level change, plate-tectonic activity and carbon burial rates affecting the global carbon cycle act on the long term (several My) and control general platform evolution and the change between climate modes (e.g., Weissert & Mohr 1996). However, at least on a regional scale, such feedback mechanisms probably also led to higher-frequency fluctuations between climate modes (humid / seasonal / temperate-warm vs. semi-arid / balanced / hot) and to a highly irregular long-term trend with episodic set-backs leading towards the late Early Cretaceous greenhouse climax.

Zusammenfassung

Die vorliegende Arbeit soll zum Verständnis der Umweltfaktoren beitragen, die die Sedimentation in den karbonat-dominierten Systemen des Späten Berriasian und des Frühen Valanginian im Französischen und Schweizer Jura (Pierre-Châtel-, Vions- und Chambotte Formationen) beeinflussten. Veränderungen des Klimas, des relativen und eustatischen Meerespiegels und der tektonischen Aktivität werden anhand der Analyse von Fazies und Diskontinuitäten, sowie mit Hilfe der Anwendung von Sequenzstratigraphie und Zyklostratigraphie untersucht. Der Vergleich mit paläogeographisch unterschiedlich gelegenen Regionen, dem hemipelagischen bis pelagischen Vocontischen Trog (Süd-Frankreich) und dem gemischt sliziklastisch-karbonatischen Rampensystem der Atlantischen Plattform (Marokko) liefert zusätzliche Informationen über Wechselwirkungen zwischen verschiedenen Umweltfaktoren und der zeitlichen und räumlichen Ausdehnung ihrer Veränderungen.

Die sedimentologische Analyse von acht Referenzprofilen im westlichen Jura (Frankreich) und fünf komplementären Profilen im zentralen Schweizer Jura, dem Vocontischen Trog und dem Essaouira Becken (Marokko) erlaubt eine detaillierte Rekonstruktion der Ablagerungssysteme in der nördlichen Tethys und am Atlantischen Kontinentalrand in der untersten Kreide. Die Jura Plattform kann in kontinentale, küstennah-tidale, inner-lagunäre, offen-lagunäre, barriereartige und extern-lagunäre Ablagerungsräume unterteilt werden. Rampen sind vorwiegend auf der Atlantischen Plattform präsent, und ausgesprochene Tiefwasserbereiche treten nur im Vocontischen Trog auf.

Der Wichtigkeit von kurzzeitigen Unterbrechungen in der Sedimentation, die als sedimentäre Zeugnisse der relativ schnellsten und substantiellsten Umweltveränderungen verstanden werden können, wird mit einer detaillierten "Diskontinuitätsanalyse" Rechnung getragen. Anhand von acht Kriterien (Geometrie, laterale Ausdehnung, Morphologie, biologische Aktivität, Mineralisation, Fazies-Kontrast, diagenetischer Kontrast und Biostratigraphie) wird die Aussage der Diskontinuitäten über Umweltveränderungen evaluiert. Dabei können vier Gruppen von Diskontinuitäten unterschieden werden. Sie zeigen Veränderungen des sedimentären Systems in Form von Emersion, submariner Omission, submariner Erosion und Wechsel in Textur und Fazies an.

Quasi-periodische Umweltveränderungen sind als sich wiederholende Änderungen in sedimentologischen und geochemischen Charakteristika und einem spezifischen Muster ihrer vertikalen Abfolge in der stratigraphischen Abfolge ausgedrückt. Drei Hierarchien solcher Ablagerungssequenzen können unterschieden und als Resultat relativer Meerespiegelschwankungen verschiedener Frequenzen identifiziert werden. Ablagerungen und Diskontinuitäten auf allen Hierarchien können Meeresspiegeltiefstand, Transgression, schnellster Überflutung, Meeresspiegelhochstand und Regression zugeordnet werden. Aus diesem Grund können alle elementaren, kleinen und mittelgrossen Sequenzen auch mit Konzepten der Sequenz- und Zyklostratigraphie beschrieben und erklärt werden. Überlagerung verschiedener Frequenzen von Meeresspiegelveränderung führt zur Vervielfältigung charakteristischer Diskontinuitäten. Ein überlagerter Meeresspiegelfall führt zur wiederholten Bildung von Sequenzgrenzen ("Sequence **B**oundaries" im Sinn der klassischen Sequenzstratigraphie). Überlagerte "initiale Überflutung" und "schnellste Überflutung" favorisieren die wiederholte Bildung von Transgressions- und Überflutungsdiskontinuitäten ("**T**ransgressive **S**urfaces" und "**M**aximum-**F**looding surfaces"). Bei allen Hierarchien können Ablagerungs-sequenzen, die von solchen Diskontinuitäten begrenzt sind, beobachtet werden. SB-Sequenzen zeichnen sich durch Emersionsdiskontinuitäten und einen "tiefer-flacher" bathymetrischen Trend aus. TS- und MF-Sequenzen sind von Diskontinuitäten oder Zonen begrenzt, die initiale oder schnellste Überflutung anzeigen. Erstere sind durch einen "tiefer-flacher" bathymetrischen Trend charakterisiert, wohingegen MF-Sequenzen einen "flacher-tiefer" Trend bezeugen. Die vertikale Anordnung der verschiedenen Sequenztypen reflektiert einen langfristigen Trend in der Veränderung des relativen Meeresspiegels.

Die Integration von biostratigraphischen Daten erlaubt die Korrelation der Sequenzen nicht nur innerhalb der Plattform, sondern auch mit den Beckenprofilen und den Profilen auf dem atlantischen Schelf. Die gute Übereinstimmung lässt vermuten, dass kleine und mittlere Sequenzen überregionale Umweltveränderungen anzeigen. Die typische Anordnung der Sequenzen in einem 5:1/4:1 Verhältnis deutet auf einen Zusammenhang von Meeresspiegelschwankungen und Klimawechseln mit Insolationsveränderungen im Frequenzbereich der Milankovitch Zyklen hin. Elementare Sequenzen, die als Ausdruck des Präzessionszyklus (Dauer etwa 20.000 Jahre) interpretiert werden, können vor allem in Beckenbereichen identifiziert werden. Diese sind im allgemeinen weniger empfindlich für autozyklische Prozesse als die dynamischen Plattformbereiche. Kleine Sequenzen, die dem ersten Exzentrizitätszyklus (100 ka) zugeordnet werden, sind deutlich auf der Juraplattform und im Vocontischen Trog zu erkennen. Auf dem Atlantischen Schelf hingegen können Sequenzen einer vergleichbaren Grössenordnung nicht eindeutig einem äusseren Einfluss zugeordnet werden. Mittelgrosse Sequenzen, die den zweiten Exzentrizitätszyklus reflektieren, können jedoch eindeutig zwischen Atlantik und Tethys korreliert werden. Die zyklostratigraphische Analyse ergibt eine Dauer von 6 Mio. Jahren für den untersuchten Zeitabschnitt (ca. 1.6 Mio. Jahre für die Pierre-Châtel Fm., 2 Mio. Jahre für die Vions Fm., 2.4 Mio. Jahre für die Chambotte Fm.).

Dieses Ergebnis stimmt mit der mittleren Dauer von 5.5 ± 4.8 Mio. Jahren überein, die von radiometrischen Daten abgeleitet wurde (Gradstein et. al. 1995, Hardenbol et al. 1999).

Die Korrelation innerhalb eines engen Rasters von Zeitlinien weist auf wiederholte Zeitintervalle mit erhöhter differenzieller Subsidenz hin, die sich durch kontemporäre Ablagerungen und Kondensationen und/oder Emersionen in verschiedenen Plattformbereichen auszeichnen. Solch tektonische Aktivität mit erhöhten Subsidenzraten und Bruchschollenverschiebungen ist wahrscheinlich auf veränderte Stressfelder in der Kruste in Folge eines beschleunigten Riftings im Nord-Atlantik und Bewegungsänderungen zwischen der Afrikanischen und Europäischen Kontinentalplatte zurückzuführen. Erhöhte tektonische Aktivität in den *Picteti/Alpillensis* und *Otopeta/Pertransiens* Ammonitenzonen kann zwischen den verschiedene Untersuchungsbereichen korreliert werden. Sie fällt zusammen mit postulierten signifikanten Meeresspiegelabsenkungen auf der "eustatischen" Meerespiegelkurve von Haq et al. (1987). In den Untersuchungsgebieten können jedoch keine Anzeichen für Meeresspiegelabsenkungen mit vergleichbaren Amplituden nachgewiesen werden. Das legt die Vermutung nahe, dass diese als eustatisch gedeuteten Ereignisse wahrscheinlich eine wichtige tektonische Komponente aufweisen. Sequenzgrenzen, die auf dem globalen sequenzstratigraphischen Schema von Hardenbol et al. (1999) verzeichnet sind, können in allen Untersuchungsgebieten nachgewiesen werden. Sie entsprechen allerdings gut entwickelten Sequenzgrenzen der mittelgrossen Sequenzen. Grössere Trends in der Entwicklung des relativen Meeresspiegels, die etwa denen der 3. Ordnung (Vail et al. 1977) entsprechen, sind hingegen nicht korrelierbar.

Faktoren, die eine Interpretation der generellen klimatischen Entwicklung zulassen, beinhalten Emersionen und assoziierte frühdiagenetische und isotopengeochemische Veränderungen, Siliziklastika (quantitativ und qualitativ), Vorhandensein und Menge organischer Substanz, sowie paläoökologische Indizien. Sie deuten ein semiarides Klima im Purbeck, ein Übergangsklima in der Pierre-Châtel Formation, und humide, saisonale Bedingungen in der Vions Formation und oberen Chambotte Formation an. Kurzzeitig (ca. 800.000 Jahre) aridere Bedingungen herrschten während der Ablagerung der unteren Chambotte Formation vor.

Die detaillierte Entwicklung der Jura Plattform als Resultat der Wechselwirkungen von eustatischen Meeresspiegelschwankungen, Klimawechsel und tektonischer Aktivität kann folgendermassen zusammengefasst werden:

1. Infolge einer ausgedehnten Progradation erlangte die Jura Plattform am Ende der *Subalpina* Zone eine grosse Ausdehnung mit einer flachen Morphologie. Eine erste Phase tektonischer Aktivität führte zu erhöhter differenzieller Subsidenz und lokalen Hebungsbewegungen, die im Zusammenwirken mit einem generellen Meeresspiegeltiefstand zu partiellem Auftauchen der Plattform führten (SB Be4). Anhaltende differenzielle Subsidenz und ein beginnender Meeresspiegelanstieg 2. Ordnung führten anschliessend zu einer Überflutung der Plattform und hoch-energetischen Bedingungen in proximalen Plattformbereichen. In dieser Phase zeigte die Plattform die Morphologie einer distal steilen Rampe ("distally-steepened ramp"). Effiziente Karbonatproduktion in einem warmen Übergangsklima (von arid zu humid) und in einem oligotrophen Milieu führte dazu, dass die Sedimentationsrate den relativen Meeresspiegelanstieg einholte und die Plattform erneut progradierte (untere *Paramimounum* Zone, Pierre-Châtel Formation).

2. Eine Verlangsamung des relativen Meeresspiegelanstiegs, erneute tektonische Aktivität und Meeresspiegelabsenkungen mit hoher Frequenz aber geringer Amplitude führten zum Auftauchen der Plattform mit regionaler Karstbildung in der mittleren *Paramimounum* Zone (SB Be5). Veränderungen in der Morphologie des Hinterlandes in Verbindung mit einem nun deutlich humideren Klima erklären das abrupte Auftreten von detritischem Quarz und organischer Materie. Für den Zeitraum von etwa 1 Mio. Jahre konnte die Plattform gut dem relativen Meeresspiegelanstieg folgen. Dies ist auf einen nur langsam ansteigenden eustatischen Meeresspiegel und vor allem auf geringe Subsidenzraten zurückzuführen. Die erneute Verflachung der Plattformmorphologie, eine Folge der kontinuierlichen Aggradation und Progradation, führte zu einer ausgesprochenen Sensibilität für hoch-frequente Meeresspiegelschwankungen, was durch wiederholtes und lateral aushaltendes Auftauchen der Plattform in diesem Zeitraum belegt ist. Der Grossteil des kontinentalen Detritus und des organischen Materials wurde in den flachen Lagunenbereichen auf der Plattform zurückgehalten. Dies führte lokal zu mesotrophen Bedingungen, und als Konsequenz war die grösste Karbonatproduktion auf Plattformrandbereiche beschränkt.

3. Das Ablagerungssystem begann sich mit erneut erhöhter differenzieller Subsidenz und wahrscheinlicher Bruchschollentektonik in den *Picteti/Alpillensis* Zonen zu verändern. Deutlich ausgeprägte tidale Strömungen begannen auf der nun morphologisch strukturierten Plattform mit vorgelagerten Inseln zu dominieren. Ein relativer Meeresspiegelanstieg zeichnete sich ab, was durch allgemein erhöhte Subsidenzraten, Beschleunigung des eustatischen Meerespiegelanstiegs 2. Ordnung, und vor allem durch vermindertes Potential von Karbonatproduktion durch anhaltenden detritischen Einfluss zu erklären ist.

4. Im oberen Teil der *Alpillensis* Zone begann sich der detritischer Einfluss zu verringern. Dies kann als Folge von aridere Klima und effizientem Auswaschen des Detritus in den nun vorwiegend hochenergetischen Ablagerungssystemen angesehen werden. Eine Retrogradation der Plattform und eine sich fortsetzende differenzielle Subsidenz führten

zu einer Änderung in der Plattformmorphologie zu einer distal steilen Rampe mit einhergehender Überflutung der vorgelagerten Inseln an der Basis der Chambotte Formation. Diese Veränderungen führten erneut zu oligotrophen Bedingungen und hoher Karbonatproduktion auf der gesamten Plattform. Die Folge davon, Aggradation und Progradation der Plattform, erlaubten das Einholen des ansteigenden relativen Meeresspiegels und eine Verflachung der Ablagerungsbereiche zum Ende der *Otopeta* Zone.

5. Ein abruptes Kippen der Plattform, verursacht durch erheblich ansteigende differenzielle Subsidenzraten, kann mit einer tiefgreifenden Erosion in proximalen Bereichen und offen-mariner, tiefer Rampenfazies in distalen Bereichen der Plattform nachgewiesen werden. Ein Klimawechsel zu humideren und deutlich saisonaleren Bedingungen wird durch erneuten Eintrag von Siliziklastika und organischem Material sowie durch Sturmablagerungen bezeugt. Die Karbonatproduktion blieb jedoch ausreichend hoch, um zumindestens lokal die Subsidenz zu kompensieren und geringe Progradation einzuleiten. Die Erholung der Plattform wurde jedoch am Top der *Campylotoxus* Zone jäh von einer erneuten tektonischen Phase und kontinuierlichem eustatischen Meeresspiegelanstieg unterbrochen, und eine schwere Krise wurde ausgelöst.

Hochfrequente Klima- und Meeresspiegelschwankungen mit geringer Amplitude sind für lokale und regionale bathymetrische Veränderungen und die Faziesverteilung in Plattform- und Beckenbereichen verantwortlich. Muster dieser Umweltveränderungen können jedoch in Abhängigkeit von der Sensibilität der entsprechenden Ablagerungsysteme über weite Strecken korreliert werden. Klimawechsel und episodische tektonische Aktivität mit einer Frequenz von etwa 1 bis 2 Mio. Jahren beeinflussen direkt die Plattformmorphologie und das Auftreten von Siliziklastika. Dies wiederum beeinflusst in einem regionalen Rahmen die Effizienz der Karbonatproduktion und damit das Muster von Progradation, Retrogradation und Aggradation der Plattform. Bei den Wechselwirkungen zwischen Eustasie, Plattentektonik (Vulkanismus) und den C_{carb}-C_{org} Begrabungsraten, die den globalen Kohlenstoffzyklus beinflussen, handelt es sich um langfristige Prozesse (mehrere Mio. Jahre). Sie nehmen Einfluss auf den generellen Modus der Plattformentwicklung und den Wechsel zwischen verschiedenen klimatischen Modi (vgl. Weissert & Mohr 1996). Zumindest auf einer regionalen Basis führten wahrscheinlich ähnliche Wechselwirkungen im Zusammenwirken aller Faktoren zu viel kurzfristigeren Schwankungen zwischen humidem, saisonalem, warm-temperiertem und semiaridem, ausgeglichenem, warmem Klima. Die Enwicklung hin zum Treibhausklima der Unterkreide scheint demnach deutlich wechselhaft mit kurzfristigen, episodischen Schwankungen verlaufen zu sein.

RESUMÉ

L'objectif de cette étude est de comprendre les principaux facteurs influençant la sédimentation des systèmes à dominante carbonatée du Berriasien supérieur au Valanginien inférieur dans le Jura suisse et français (Formations de Pierre-Châtel, de Vions et de la Chambotte). Les changements climatiques, tectoniques et eustatiques ont été étudiés à travers les analyses de facies, de discontinuités, de stratigraphie séquentielle et de cyclostratigraphie. La comparaison avec d'autres contextes paléogéographiques comme le domaine hémipélagique à pélagique du bassin Vocontien (France) et le système de rampe mixte siliciclastique-carbonatée dans l'Atlas Atlantique (Maroc), fournit des indications additionnelles sur l'interaction des facteurs contrôlant la sédimentation et sur leur importance temporelle et spatiale.

Huit coupes de référence se situent dans la partie ouest du Jura français et quatre coupes complémentaires ont été levées dans le Jura Central suisse, le bassin Vocontien français et le bassin marocain d'Essaouira. Leur analyse sédimentologique permet l'élaboration de modèles de dépôt pour la Téthys et la marge atlantique dans le Crétacé inférieur. La plate-forme a été ainsi subdivisée en domaine continental, en zone côtière/tidale, en lagon interne et ouvert, en barrière, et en lagon externe. Domaines de rampe et de talus/basin prédominent sur la marge Atlantique et dans le Bassin Vocontien.

Une analyse détaillée des discontinuités met en évidence l'importance des arrêts de sédimentation de courte durée, issus de changements environnementaux très rapides et substantiels dans les systèmes sédimentaires. L'importance environnementale des surfaces de discontinuités a été étudiée sur la base de huit critères (géométrie, extension latérale, morphologie, activité biologique, minéralisation, contraste de faciès et de diagenèse, biostratigraphie). Quatre groupes de discontinuités peuvent ainsi être distingués. Ils sont reliés à des changements environnementaux indiquant une exposition subaérienne, une absence de sédimentation, une érosion sous-marine, ou un changement dans la texture et le faciès.

Des changements environnementaux quasi-périodiques sont exprimés dans l'enregistrement stratigraphique par des variations répétitives dans les caractéristiques sédimentologiques et géochimiques, ainsi que dans l'empilement séquentiel ("stacking pattern"). Trois niveaux de hiérarchie ont été observés dans les séquences de dépôt et attribués à l'effet des changements du niveau marin relatif. Les dépôts et les discontinuités au niveau de chaque échelle de séquence traduisent un bas niveau marin, une transgression, un maximum d'inondation, un haut niveau marin et une régression. Des séquences élémentaires, des petites séquences et des séquences moyennes peuvent ainsi être décrites en terme de concepts de stratigraphie séquentielle et de cyclostratigraphie. La superposition de différentes fréquences du niveau marin mène à la multiplication des discontinuités caractéristiques. Des chutes de niveau marin superposées ont comme conséquence la formation répétée de limites de séquence. Les inondations initiales et maximales superposées favorisent la formation respectivement de surfaces de transgression et de maximum d'inondation. A toutes ces différentes échelles, les séquences de dépôt, délimitées par ces discontinuités, peuvent être différentiées. Les "séquences SB" (contexte de chute du niveau marin) sont caractérisées par des surfaces d'émersion et par une évolution en "deepening-shallowing". Les séquences TS (contexte de transgression) sont délimitées par des discontinuités ou des intervalles indiquant une inondation initiale et une évolution en "deepening shallowing". Les séquences MF (contexte de "maximum flooding") sont elles décrites par un maximum d'inondation et une évolution en "shallowing-deepening". L'empilement séquentiel ("stacking pattern") des différents types de séquences de dépôt reflète les changements du niveau marin relatif à plus grande échelle.

L'intégration des données biostratigraphiques permet la corrélation des séquences de dépôt non seulement à travers la plate-forme, mais aussi avec les coupes de bassin et celles de la marge Atlantique. Il est suggéré que des changements environnementaux d'extension intercontinentale peuvent être reflétés par les séquences moyennes et probablement aussi par les petites séquences. L'arrangement typique des séquences avec un rapport de 5:1/4:1 indique des changements de niveau marin et/ou climatiques qui sont en phase avec des variations d'insolation dans la bande de fréquence de Milankovitch. Les séquences élémentaires sont interprétées comme correspondant au cycle de précession (20 ka). Elles sont prédominantes dans le bassin où le système sédimentaire est moins propice à la formation d'autocycles en comparaison avec les environnements hautement dynamiques de la plate-forme. Les petites séquences correspondent au premier cycle de l'excentricité (100 ka). Elles sont bien visibles sur la plate-forme jurassienne et dans le Bassin Vocontien, au contraire de la marge Atlantique où les séquences de dépôt de même échelle ne peuvent être attribuées à un "forcing" externe. Cependant, les séquences moyennes, qui sont reliées au second cycle de l'excentricité (400 ka), peuvent être explicitement corrélée entre les domaines Atlantique et Téthysien. L'analyse cyclostratigraphique suggère une durée de 6 Ma pour l'intervalle étudié (estimation par formation: Pierre-Châtel: 1.6 Ma; Vions: 2 Ma; Chambotte: 2.4 Ma), ce qui correspond bien à la durée de 5.5 Ma ± 4.8 Ma déduite des âges radiométriques (Gradstein et al. 1995, Hardenbol et al. 1999).

La corrélation effectuée dans un cadre temporel précis indique plusieurs intervalles possédant une subsidence différentielle, indiqués par des dépôts contemporains, des niveaux condensés et/ou des émersions. L'activité tectonique, exprimée par une subsidence accélérée et des blocs basculés, est probablement reliée à de multiples changements dans les contraintes de stress, comme un rifting accéléré dans l'Atlantique Nord, des changements dans le mouvement entre les plaques africaine et européenne, ainsi qu'une rotation initiale du bloc ibérique. L'activité tectonique comprise dans l'espace de temps correspondant aux zones d'ammonites *Picteti/Alpillensis* et *Otopeta/Pertransiens* est corrélable entre le domaine Atlantique et Téthysien. Elle coïncide avec les chutes majeures du niveau "eustatique" postulées par Haq et al. (1987). Aucune chute majeure de niveau marin ayant une amplitude comparable n'a été observée dans les domaines étudiés, ce qui suggérerait que les événements "eustatiques" indiqués par ces auteurs contiennent une importante composante tectonique. Les limites de séquences, qui ont été indiquées sur la charte de stratigraphie séquentielle de Hardenbol et al. (1999), peuvent être identifiées, mais elles correspondent généralement aux limites bien marquées des séquences moyennes, tandis que les séquences de plus grande échelle, présumées appartenir aux variations de 3ème ordre du niveau marin relatif, ne peuvent pas être corrélées entre les différents domaines étudiés.

Les facteurs attestant de l'évolution climatique générale englobent les emersions et les évolutions diagénétiques associées. Elles sont soulignées par la composition des isotopes stables, les siliciclastiques (composition des minéraux argileux, quartz détritique), la présence et l'abondance de la matière organique et les assemblages faunistiques. Ils indiquent des conditions semi-arides dans le Purbeckien, un climat de transition pour la Formation de Pierre-Châtel, des conditions plus humides et saisonnières dans la Formation de Vions et de la Chambotte supérieure avec un court interlude plus aride dans la partie inférieure de la Chambotte.

Au vue de l'interaction des changements de niveau eustatique, du climat et de l'activité tectonique, l'évolution de la plate-forme jurassienne peut être résumée de la façon suivante:

1. Après une importante progradation, la plate-forme jurassienne atteint une morphologie plate à la fin de la zone à *subalpina* (sommet du Purbeckien). L'activité tectonique initie une subsidence différentielle et/ou un "uplift" durant un bas niveau marin, conduisant à une exposition partielle de la plate-forme (SB Be4). Une reprise de la subsidence associée à un début d'augmentation du niveau marin sur le 2ème ordre amène des conditions de haute énergie en position de plate-forme proximale sur ce qui est désormais une rampe plus accentuée en position distale ("distally steepened ramp"). Cependant, la production carbonatée effective dans un climat de transition semi-aride et des environnements bien oxygénés permet à la plate-forme de rattrapper la montée du niveau marin relatif ("catch-up") et de prograder rapidement durant l'espace de temps équivalent à la zone à *Paramimounum* inférieur (Formation de Pierre-Châtel).

2. Un ralentissement dans la montée ou une légère chute du niveau marin relatif sur le long terme, associé à l'activité tectonique et à des chutes du niveau marin de haute fréquence et de basse amplitude dans le milieu de la zone à *Paramimounum*, cause une émersion avec une karstification locale (SB Be5). Des changements morphologiques dans l'arrière-pays et un climat plus humide expliquent l'arrivée abrupte de quartz détritique. Pendant une durée approximative de 1 Ma, la plate-forme reste en état d'équilibre ("keep up") résultant de l'augmentation lente du niveau marin et des faibles taux de subsidence. La plate-forme acquiert une morphologie plate par aggradation et progradation, ce qui induit un bon enregistrement des petits changements eustatiques avec des émersions répétitives très étendues. La majorité des siliciclastiques et de la matière organique se trouve piégée dans les lagons peu profonds sur la plate-forme. Ce phénomène engendre des conditions mésotrophiques locales, ayant comme conséquence la restriction de la production carbonatée à la bordure de la plate-forme.

3. Le système sédimentaire commence à changer avec une subsidence différentielle élevée dans la zone à *Picteti/ Alpillensis*. L'influence tidale et les courants forts deviennent prédominants sur la plate-forme hautement structurée, marquée par des îles de barrière isolées. Une augmentation du niveau marin relatif est observée. Elle est due à une subsidence générale élevée, une augmentation du niveau marin sur le 2ème ordre et un faible potentiel d'aggradation sur la plate-forme dans des environnements stressés par un apport détritique continu.

4. L'apport détritique diminue graduellement dans la partie supérieure de la zone à *Alpillensis*, résultant d'un climat plus aride et d'un vannage ("winnowing") dans des environnements à prédominance de haute énergie. Le "backstepping" des environnements de dépôt et la subsidence différentielle continue mène à une morphologie de "distally steepened ramp" et à une absence de système de barrière dans la partie inférieure de la Formation de la Chambotte. Cependant, la grande production carbonatée dans des environnements désormais oligotrophiques induit à nouveau une progradation de la plate-forme et a une diminution de la profondeur au sommet de la zone à *Otopeta*.

5. Un brusque basculement de la plate-forme, accompagné d'une amplification de la subsidence différentielle est mis en évidence par une troncation profonde et une émersion de la plate-forme proximale et des conditions de rampe profonde associées à un renouvellement des apports siliciclastiques en position distale. Une grande activité de tempête, combinée à des apports en siliciclastique et en matière organique, atteste d'un climat de transition avec une saisonnalité élevée. Cependant, la production carbonatée est suffisante pour compenser la subsidence et induire une progradation locale (Formation de la Chambotte supérieure). Ce processus est interrompu à nouveau par une phase de subsidence différentielle intensifiée au sommet de la zone à *Campylotoxus*.

Les variations climatiques et de niveau marin à court terme sont responsable des variations bathymétriques et de la distribution des faciès dans les environnements de plate-forme et de bassin sur une échelle locale à régionale. Cependant, grâce à la configuration générale et à la sensibilité du système sédimentaire les tendences des changements environnementaux peuvent être corrélé sur de longues distances. Les changements climatiques et les activités tectoniques épisodiques sur une échelle de 1 à 2 Ma contrôlent principalement la morphologie de la plate-forme ainsi que la présence et la distribution des siliciclastiques. Ceci influence à tour de rôle et en grande partie l'efficacité de la production carbonatée et la progradation, la rétrogradation et l'aggradation de la plate-forme sur une échelle régionale. Les mécanismes internes de "feed-back" entre le changement de niveau marin, l'activité des plaques tectoniques et le taux d'enfouissement du carbone affectant le cycle global du carbone jouent sur le long terme (plusieurs Ma) et contrôlent l'évolution générale de la plate-forme ainsi que le changement entre les modes climatiques (e.g., Weissert & Mohr 1996). Régionalement, de tels mécanismes de rétroaction conduisent probablement à des fluctuations de période bien plus courte entre les différents types de climat (humide / saisonnier / tempéré vs. semi-aride / equilibré / chaud), induisant un développement irrégulier du long terme, vers la période chaude du Crétacé inférieur.

CONTENTS

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1 - Introduction

Sedimentary environments are complex, dynamic systems. Their evolution through time, represented by the stratigraphic record, has all the characteristics of non-linear dynamics, as it is the output of many interacting and coupled parameters (Smith 1994). New and revived stratigraphic concepts such as sequence stratigraphy (Vail et al. 1977), genetic stratigraphy (Galloway 1989), and cyclostratigraphy (e.g., Fischer 1986, Strasser 1994) have established a way of thinking which appreciates this perception. In particular they brought up a greater awareness of the dynamic behavior of sedimentary systems and demonstrated the necessity for models that do not simply use cause and effect relationships, but rather look for interdependencies and feedback mechanisms between many variables.

Sequence- and cyclostratigraphic concepts allow to establish high-resolution correlations of stratigraphic sections in a bio- and/or chronostratigraphic framework. With such correlations the sedimentary record permits qualitative and quantitative statements about the relative importance and interactions of tectonics, climate change and sea-level change, not only for the long term (millions of years), but also for time periods in the range of tens- to hundreds- of thousands of years. This allows to analyze lateral facies variations and depositional geometries with much higher precision and becomes even relevant for today's environmental problems.

In times when the term "global change" is in the news almost every day, and the anthropogenic influence on Earth's climate is evident and source of much debate, the sedimentologist working in the Cretaceous cannot give answers to current questions. His contribution, however, is to demonstrate the sensitivity of the ecosphere (bio- and geosphere) to environmental changes and illustrate the result of complex reactions when its dynamic equilibrium was disturbed.

In this respect the Lower Cretaceous of the Swiss and French Jura makes a particularly good example. A sound biostratigraphic framework is established on the basis of ammonites (Le Hégarat & Remane 1968, Le Hégarat 1971, Bulot 1995), benthic foraminifers (Clavel et al. 1986, Blanc 1996), calpionellids (Remane 1985, Blanc 1996), and charophyte-ostracod assemblages (Detraz & Mojon 1989). Numerous correlations on the northern Tethyan margin have been carried out (e.g., Detraz & Mojon 1989, Waehry 1989, Pasquier 1995, Blanc 1996) and cyclostratigraphic studies have demonstrated that sedimentary systems during the Lower Cretaceous did record orbital parameters, and allow to work within a narrow framework of timelines (Strasser 1988, 1994, Pasquier 1995, Pasquier & Strasser 1997, Strasser & Hillgärtner 1998).

The sedimentary systems in the Upper Berriasian and Lower Valanginian are dominated by carbonates, but the recurrent and abrupt influx of siliciclastics, if only for that reason, suggests important environmental changes. The sediments recording these changes provide important information about the local, regional and global conditions that governed the depositional systems at these times. Comparison of platform sediments with deposits in other paleogeographic settings (South-East of France, West Morocco) contributes to discriminate factors which control sedimentation and underlines local and regional differences.

1.1. Objectives

Sedimentary research has long gone beyond being a purely descriptive science but detailed observation and description will always be the basis of any theoretical interpretation and will survive scientific trends and models "en vogue". It is the aim of this study, therefore, to give before all a detailed description and a comprehensive sedimentological analysis of the reference sections. The main objective then is to establish a high-resolution correlation of the sections by means of sequence- and cyclostratigraphy. Given this framework, the relative roles of tectonics, sea-level change and climate change are elaborated. Comparison with other paleogeographic settings furthermore sheds light on interactions of these factors and their spatial relevance. For a synthesis, a detailed model of the evolution of the Jura platform during the Late Berriasian and Early Valanginian is proposed.

1.2. Geographic situation

The main study area is located in the southern Jura in France to the west and south of Geneva (Fig. 1.1). Eight sections were logged and studied in detail in this region. One investigated section, originally logged by Pasquier (1995), is located about 80 km to the northeast in the Swiss Jura (Fig. 1.1). Two basinal sections are located in southeastern France in the region around Sisteron (Fig. 1.2). Two additional sections are studied in the western Atlantic Atlas of Morocco, south of the coastal town of Essaouira (Fig. 1.3).

1.3. Paleogeographic situation

During the Late Jurassic and earliest Cretaceous the western Tethys was already subject to sinistral translative movements between Africa and Laurasia induced by northward propagation of the Atlantic spreading axis. As a consequence of these movements compressional stresses developed in the eastern parts of the Tethys, whereas in the western parts sea-floor spreading diminished (Ziegler 1988). This and accelerated rifting in the North Atlantic led to a general thermal doming and tectono-eustatic sea-level fall, which caused widespread emergence of the Armorican, Central, and Rheno-Bohemian Massifs during Late Jurassic times (Late Kimmerian unconformity).

The continental regions were the source for important siliciclastic influx into the Paris basin and onto the northern Tethyan shelf. This is recorded in the Lower Berriasian Wealden Facies in southwestern England and the Paris basin (Ziegler 1988), and the Purbeckian in the Jura region. The Jura domain represents part of the northern margin of the oceanic Ligurian Tethys (Fig. 1.4) and makes up the central part of the North-Tethys platform. Sedimentation in the Middle to Late Berriasian and Early Valanginian is characterized by shallow-marine carbonates with intermittent siliciclastic influence during a general sea-level rise. However, the platform was never very far from exposure with maximum water depths of a few tens of meters only.

The Vocontian through represents a branch of the Ligurian Tethys which was bordered to the north and northwest by the Jura platform and to the south and southwest

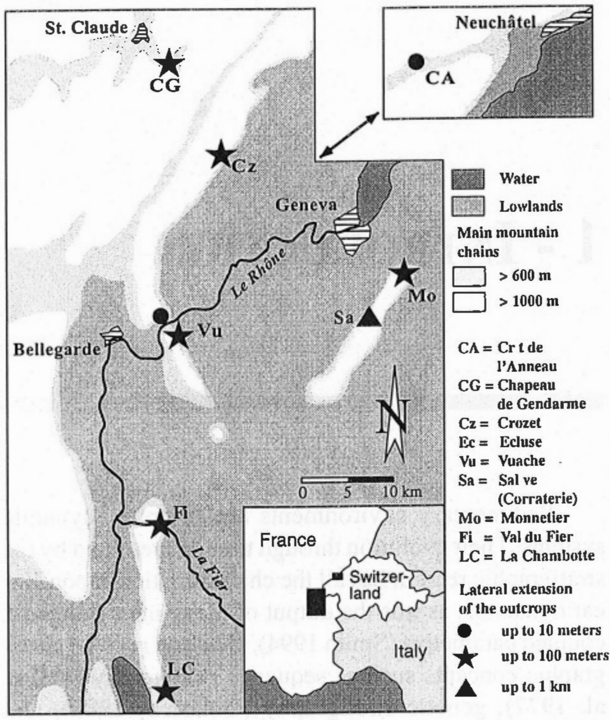


Fig. 1.1. Location of sections in the French and Swiss Jura.

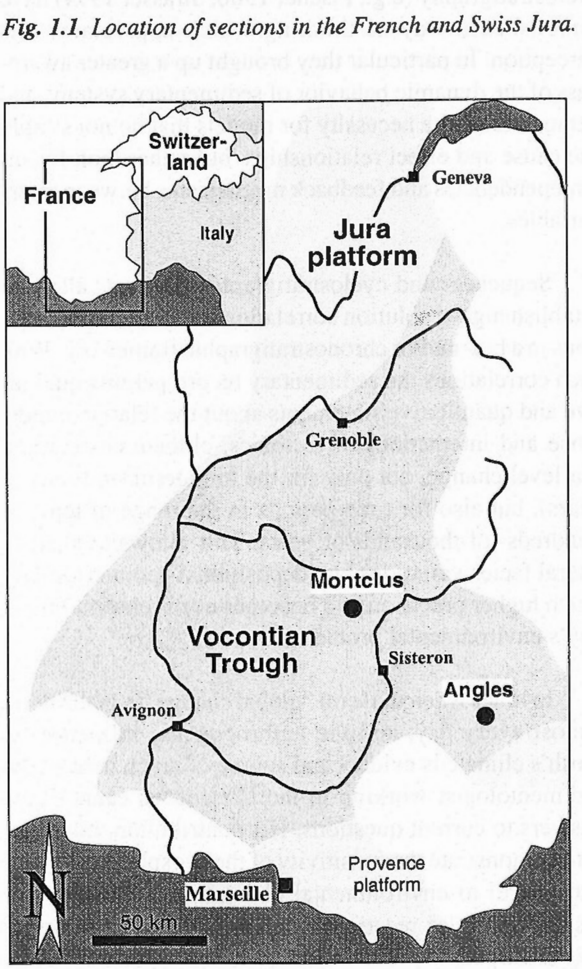


Fig. 1.2. Geographic position of Montclus and Angles sections and paleogeographic domains between Jura and Provence platforms (modified after Pasquier 1995).

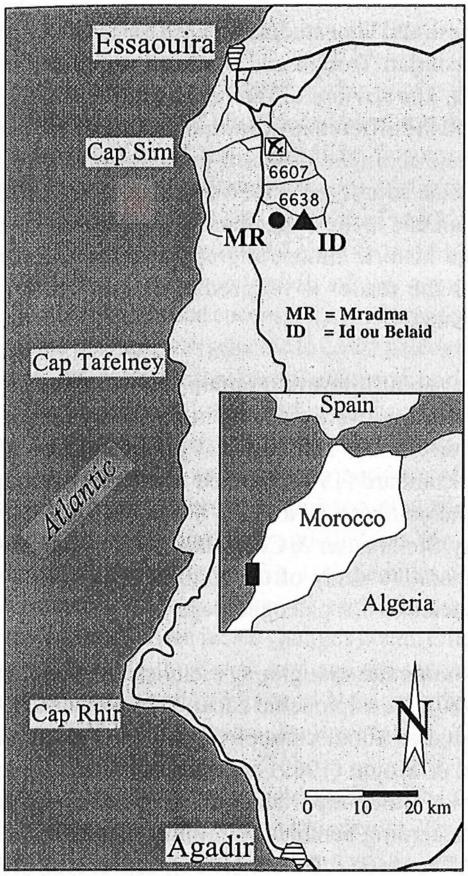


Fig. 1.3. Geographic location of the Morrocan sections.

by the Marseille (Provence) platform and the Corsica-Sardinia high (Figs. 1.2, 1.4, Ziegler 1988). Throughout all of the Early Cretaceous this domain was characterized by hemipelagic to pelagic sedimentation with intermittent biodetrital influx from the surrounding platforms.

The Essaouira basin, located in the Atlantic Atlas, formed part of the northwestern shelf of the Sahara platform (Fig. 1.4). This region was marked by the extensional tectonic regime of the Central Atlantic and transform movements along the Azores fracture zone causing episodic phases of elevated subsidence rates. Sedimentation was calcareous with a strong siliciclastic influence. The studied area was located on a distal, open-marine platform with water depths of around 20 to 50 meters, which never attained emergence during the Early Cretaceous (Taj-Eddine 1992).

The paleolatitude of the Jura platform, according to paleogeographic reconstructions (Smith et al. 1994, Fourcade et al. 1993, Funnell 1990, Ziegler 1988, Dercourt et al. 1985, Barron et al. 1981), was approximately 26° to 33° N and, as such, corresponds to the subtropical zone. In the same reconstructions the Vocontian Trough was located 2° to 3° farther to the south and the Essaouira region was located 10° to 12° to the south. The northern margin of the Sahara platform therefore had a paleogeographic position between 14° to 23° N that corresponds to today's tropical climate belt.

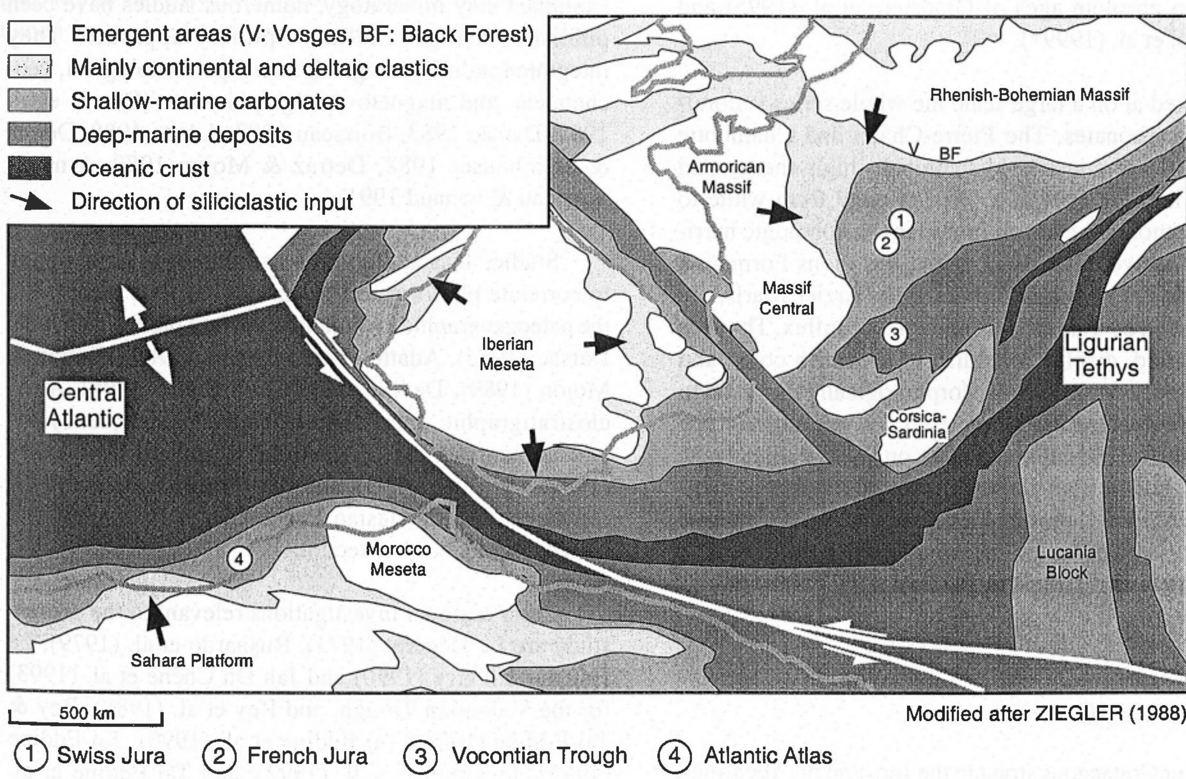


Fig. 1.4. Paleogeographic reconstruction of the western Tethys domain in the earliest Cretaceous.

1.4. Stratigraphy

The nomenclature used in this study is based on formal lithostratigraphic subdivisions for the southern Jura established by Steinhauser and Lombard (1969) and modified by Clavel et al. (1986). The large lithologic variety along the northern Tethyan shelf led to many different stratigraphic subdivisions for regions that are not very far apart. Figure 1.5 summarizes and compares Berriasian and Lower Valanginian lithostratigraphic units from eastern Switzerland throughout the Jura chain towards the southwest and the Essaouira Basin in the Atlantic Atlas. The very monotonous nature of the sediments in the Vocontian Trough during this time interval did not stimulate a lithostratigraphic subdivision.

The strata studied in the Jura includes the Pierre-Châtel, Vions, and Chambotte Formations, the latter subdivided in Lower Chambotte, Guiers Member, Marnes d'Arzier, and Upper Chambotte. The Calcaire Roux Formation is only encountered in a few sections. The Formations correspond to the ammonite zones *Occitanica*, *Boisseri*, and *Roubaudiana* (Clavel et al. 1986). Ammonite subzones and calpionellid zones are illustrated in Figure 1.5.

On the basis of correlations and biostratigraphy established and refined by Le Hégarat (1971, 1980), Remane (1985), and Clavel (1986), the lithostratigraphic interval studied encompasses approximately 5.5 My ± 4.8 My according to absolute ages of Gradstein et al. (1995) and Hardenbol et al. (1999*).

Looked at on a large scale the whole series is dominated by carbonates. The Pierre-Châtel and Chambotte Formations are constituted mainly of high-energy and lagoonal limestones with a color ranging from white to gray. Commonly they form important morphologic barriers and massive cliffs. In contrast, the Vions Formation and the Guiers Member, including the Arzier Marls, are characterized by a significant siliciclastic influx. They are well stratified, display a reddish to brownish color, and can be slightly dolomitized. Morphologically they tend to form lows preferentially covered by vegetation. The Calcaire Roux Formation indicates only a low siliciclastic influence, but the high content in crinoid fragments, common silicification (late diagenetic nodules) and the intense reddish-brown color of the weathered rock give it a characteristic appearance in the field.

1.5. Historic

Lower Cretaceous strata in the Jura and the Vocontian Trough were subject of intense scientific examination, already in the 19th century (e.g., Desor 1854, Coquand 1871). These and later studies focused on basinal sections in the Vocontian Trough and platform sediments in the Swiss Jura. The obvious difference in lithology of the two realms, and the difference in ammonite abundance caused much debate about the platform-to-basin correlations for the Berriasian stratotype in the Vocontian Trough, and the Valanginian one in the Swiss Jura. For a detailed description of the historic lithostratigraphic terms and further references the reader is referred to Darsac (1983) and Adatte (1988).

More recent lithostratigraphic studies of the central and southern Jura were carried out by Mouty (1966) and Steinhauser (1969). The classical publication by Steinhauser & Lombard (1969) defined new lithostratigraphic units in the southern Jura using a nomenclature current until today. Steinhauser & Charollais (1971) published the first comparative study of the southern and central Jura with implications for paleogeography.

The ammonite stratigraphy established by Le Hégarat (1971, 1980), the calpionellid zonation by Remane (1985), and the studies about charophyte-ostracod assemblages by Detraz & Mojon (1989) are essential biostratigraphic frameworks. Other important biostratigraphic constraints mainly concerning benthic foraminifers have been added by Arnaud-Vanneau & Darsac (1984), Clavel et al. (1986), Charollais & Wernli (1995), and Blanc (1996).

Beginning with Persoz & Remane (1976), who examined clay mineralogy, numerous studies have been published using a multidisciplinary approach. They integrated sedimentological, micro-paleontological, geochemical, and magneto-stratigraphic data (Darsac et al. 1982, Darsac 1983, Boisseau 1987, Adatte 1988, Detraz & Steinhauser 1988, Detraz & Mojon 1989, Arnaud-Vanneau & Arnaud 1991).

Studies that additionally use sequence stratigraphy to correlate platform and basin sections, and reconstruct the paleogeographic evolution of the Jura platform include: Darsac (1983), Adatte (1988), Detraz (1989), Detraz & Mojon (1989), Deville (1991), and Blanc (1996). A cyclostratigraphic approach utilizing high-frequency sequence analysis for correlations is applied by Waehry (1989), Pasquier (1995), and Pasquier & Strasser (1997) for the Lower Berriasian Pierre-Châtel Formation and corresponding basinal sections.

Other regional investigations relevant to the present study are Le Hégarat (1971), Busnardo et al. (1979), Le Hégarat & Ferry (1990) and Jan Du Chêne et al. (1993) for the Vocontian Trough, and Rey et al. (1988), Rey & Taj-Eddine (1989), Taj-Eddine et al. (1990), Taj-Eddine (1992), Taj-Eddine et al. (1992), and Taj-Eddine et al. (1993) for the Atlantic Atlas.

* released in 1999, but backdated to 1998

1.6. Methodology

Taking the main objective of this study, that is the examination of type and timing of environmental changes and the distinction of the controlling factors involved, detailed investigation of well-exposed sections is indispensable. Eight reference sections in the French Jura were chosen according to continuous outcrop conditions to allow a detailed field examination of the complete and tectonically undisturbed successions on a cm-scale. Their geographic position was selected to cover paleogeographic platform cross-sections perpendicular and horizontal to the platform strike. In order to judge lateral facies variability and lateral extent of discontinuity surfaces, distances between single sections vary between 400 meters to a maximum of 60 kilometers. Outcrops that allow physical tracing of individual beds and surfaces for a significant distance were preferred (Fig. 1.1). Samples were taken in the immediate surroundings of each observable bedding surface and where facies changes occur. One supplementary section in the Swiss Jura and the sections in the Vocontian Trough and in Morocco were examined in detail in the field but fewer samples were taken.

Rock samples were cut, and etched slabs were prepared to be examined under the binocular. Thin-section analysis of 1200 samples was then carried out to determine depositional facies and early-diagenetic alteration. Cathodoluminescence and scanning-electron microscope were used to obtain more accurate data for selected samples. Marls and clays were washed and sieved and the obtained grain-size fractions analyzed under the binocular.

Stable-isotope analysis of 45 samples of the Salève sectio was carried out by Dr. M. Joachimski (Erlàngen / Germany). Rock powder was extracted with a dentist drill from polished rock samples, whereas large bioclasts and cement veins were avoided. The analyses of this "modified whole rock" has shown to be representative for the overall isotopic signature (excluding biological fractionation and late diagenetic signatures, Plunkett 1997). The samples have subsequently been prepared and analyzed by Dr. M. Joachimski (Univ. Erlangen, Germany). Carbonate powders were reacted with 100% phosphoric acid (density >1.9) at 75°C in an on-line carbonate preparation line (Carbo-Kiel - single sample acid bath) connected to a Finnigan Mat 252 mass-spectrometer. All values are reported in per mil relative to VPDB by assigning a $d^{13}C$ value of +1.951 and a $d^{18}O$ value of -2.201 to NBS19. Analytical precision was controlled by replicate measurements of IAEA Standard CO-1. The standard deviation ranges between 0.01-0.03‰ for $d^{13}C$ and 0.03-0.08‰ for $d^{18}O$. For isotope data and their interpretation refer to chapters 5.3 and 8.

All samples are numbered by section; a specimen with the number Sa113 represents sample number 113 in the Salève section. All materials concerning this study (thin sections, selected rock slabs, sieved fractions) are deposited at the Institute of Geology at the University of Fribourg in Switzerland.

2 - FACIES ANALYSIS

2.1. INTRODUCTION

2.1.1. Definitions

Microfacies

Microfacies describes an association of all paleontological and sediment-petrographic criteria which can be defined at a small scale, e.g., in thin section or on a polished slab (Flügel 1982, 1992). Wilson (1975) and Flügel (1982) defined a suite of standard microfacies types which are associated with specific facies zones (see below).

Facies

Facies is the sum of all organic and inorganic characteristics of a sediment including color, composition, texture and sedimentary structures (Flügel 1992, Tucker & Wright 1990).

Facies zone

A Facies zone corresponds to an area that can be differentiated by an association of lithology, sedimentary structures, and organisms and, thus is reflecting specific environmental conditions in a depositional system. This term is used synonymously to "facies belts" of Wilson (1975) and Flügel (1982).

Facies association

Facies associations of Tucker & Wright (1990) which are used to describe "groups of facies occurring together" and are reflecting "the same general environment" have a similar notion as facies zone. However, this term is used also to refer to a vertical succession of related facies.

2.1.2. General Remarks

Although the Upper Berriasian and Lower Valanginian of the French and Swiss Jura appear as a relatively homogeneous, dominantly calcareous succession, they bear an enormous lithological variability when looked at it in detail. This reflects the complex depositional mechanisms of shallow-marine carbonate and mixed carbonate-siliciclastic systems fringing continental areas. Carbonates are, in contrast to siliciclastics, essentially produced biogenically and are of autochthonous or semi-autochthonous origin. Therefore the producing organisms or the "carbonate factory" reflect ecological conditions influenced by water temperature, salinity, turbidity, bathymetry, and currents. Microfacies analysis - the study of composition and texture (Tucker & Wright 1990) - is therefore an essential tool for reconstructing carbonate depositional environments. In the following chapters all encountered criteria for microfacies and facies analysis and their environmental significance are discussed.

2.2. CONSTITUENTS

Rock-composing constituents give substantial information about depositional environments and ecological conditions. This chapter briefly describes the main elements occurring in Upper Berriasian and Lower Valanginian formations of the French Jura and reviews their paleoecological importance.

2.2.1. Non-skeletal grains

Peloids

Peloids are rounded to well rounded, spherical to elliptical grains formed of crypto- or microcrystalline material without an internal structure. Commonly they have sizes smaller than 1 mm and may be rich in organic material. Their origin often remains uncertain since several modes of formation are possible (Fig. 2.1, Tucker & Wright 1990).

In the present study 10 to 40% of the sediment is made up of peloids which corresponds to quantities of 30% observed in modern platform environments (Tucker & Wright 1990). Approximately 95% of the observed peloids range in size from 10 to 60 μm, are well rounded and subspherical and occur preferentially in a lagoonal context. They are interpreted as mainly of fecal origin since micritization of grains usually is partial and identification of their origin is possible. Only micritic ooids that undergo secondary micritization can be difficult to distinguish. The other, bigger, commonly less well-sorted and rounded peloids are interpreted as micritized grains and/or reworked mudclasts. Grains interpreted as "true" fecal pellets on the basis of typical internal structures are extremely rare.

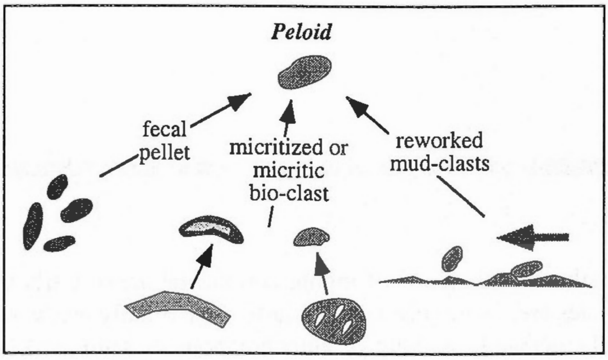


Fig. 2.1. Possible origin of peloids (after Tucker & Wright 1990).

Ooids

Ooids are spherical to elliptical coated grains with concentric calcareous cortices around a nucleus that can be of variable composition (Flügel 1982, Tucker & Wright 1990). They can form in different environments (marine, lacustrine, fluviatile, continental) and can show various microfabrics and mineralogies (Richter 1983).

Marine ooids commonly are formed by chemical or biochemical precipitation of calcium carbonate. The detailed role of biogenic activity in this process is not yet clear. However, it is known that the presence of algae, bacteria and/or organic matter influences their formation (Flügel 1992). Other environmental factors which are important for ooid formation include: water temperature of more than 18-20 °C, $CaCO_3$ super-saturation of the sea water, shallow water depths (1-5 m) with at least occa-

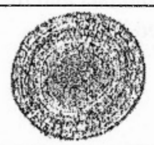
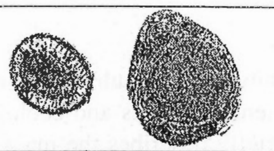
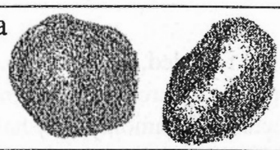
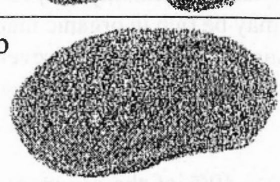
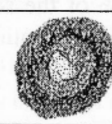
Ooid-type		Description	Accessory components	Depositional environment	Formations
1	*radial fibrous*				
a		> 5-10 thin (<30 m) laminae high sphericity patchy micritisation rare micritic laminae 0.2 -0.6 mm fl often reworked, abrasion of individual laminae Fe-stained	intraclasts reworked bioclasts echinoderms bryozoa thick bivalves	- high-energy , external ooid bars or reworking from these environments into bioclastic high-energy bars - conditions favorable for ooid-formation remained constant for a longer time interval - high iron contents in sea-water and phases of non-deposition are indicated	Guiers Member, (upper Chambotte), Calcaire Roux
b		< 5-10 thick (>20 m) laminae moderate to high sphericity patchy micritisation alternating, irregular micritic laminae 0.3 -1 mm fl often reworked, broken and abraded laminae (Fe-stained)	intraclasts reworked bioclasts high energy lagoonal context	- moderate to high-energy ooid bars and/or reworking from these environments into bioclastic bars - conditions favorable for ooid-formation remained constant for a shorter time interval - high iron contents in sea-water and phases of non-deposition are indicated	Pierre Ch tel , (Vions), Guiers Member, (upper Chambotte), Calcaire Roux
c		superficial 1-2 thin (<30 m) laminae moderate sphericity patchy micritisation 0.3 -1 mm fl	lagoonal context	- bioclastic/ooid bars and/or reworking from these environments into lagoons - conditions favorable for ooid-formation prevailed only for a short time	Vions, lower Chambotte
2	*micritic*				
a		micritic or micritized laminae occasionally in alternance with few radial fibrous laminae moderate sphericity 0.3 -0.6 mm fl	lagoonal context intraclasts, quartz oncoids moderate to high diversity of lagoonal benthic organisms	- lagoonal environments with intermittent high energy, tidal ooid/bioclastic bars in lagoons - conditions for intense microbial activity are commonly favorable	Vions, lower Chambotte
b		micritic or micritized laminae occasionally in alternance with few radial fibrous laminae low sphericity 0.6 -2 mm fl	lagoonal context oncoids, lumps, composite grains high diversity of lagoonal benthic organisms	- lagoonal environments with intermittent high energy - conditions for intense microbial activity are favorable - fuzzy transitions between ooids and oncoids	(Vions), lower Chambotte
c		micritic or micritized laminae rarely in alternance with few radial fibrous laminae moderate to high sphericity 0.3 -0.5 mm fl often reworked and intensively Fe-stained	lagoonal context intraclasts, black facies, quartz reworked bioclasts lagoonal benthic organisms	- lagoonal environments with intermittent high energy, tidal ooid/bioclastic bars in lagoons - conditions for microbial activity are commonly favorable - iron/pyrite staining indicates high content of iron in sea water /anoxic conditions	Vions

Fig. 2.2. Ooid characteristics and classification.

sional high-energy conditions, slightly elevated salinity, supply of potential nuclei, few other organisms that extract carbonate from the sea water, and environmental conditions that remain stable for a relatively long period of time (Flügel 1992). The diagnostic value of ooids alone is diminished by the fact that these conditions occur in several marine environments and that ooids are easily transported and reworked. Consequently, taphonomic factors such as hydrodynamic sorting play an important role in their distribution.

Ooids in the Lower Cretaceous of the French Jura are supposedly of a marine origin since they occur only together with other marine organisms. In 40 to 50% of the samples ooids appear simultaneously with intraclasts and reworked bioclasts indicating an semi-autochthonous to allochthonous origin. This is especially the case in proximal platform positions. Figure 2.2 summarizes the different occurring types and the associated environments of formation/deposition. Whereas type-1 ooids occur preferentially in high-energy, open-marine environments, type-2 ooids commonly occur in a more lagoonal context. Both types of ooids can show an intense staining by iron-minerals, especially in the Vions Formation and Guiers Member/Upper Chambotte Formation. High iron influx due to lateritic weathering in the hinterland (Gygi 1981) and prolonged non-deposition and reworking (Burkhalter 1995) were favorable for the formation of iron-ooids in the Jurassic. Similar environmental conditions may have prevailed during certain time intervals in the Early Cretaceous.

Oncoids

Oncoids are irregularly formed calcareous components with non-concentric, micritic laminae that are due to biogenic accretion around a nucleus. Principally, cyanobacteria are involved in the formation of oncoids, but other encrusting organisms such as filamentous algae, foraminifers, serpulids and bryozoans can contribute to a large part in oncoid growth.

Oncoids occur in lacustrine, fluviatile and marine environments. Their size, their shape and their surface texture generally depend on sedimentation rates and energy conditions in the depositional environment (Dahanayake 1978, Ciarapica & Passeri 1983). The size becomes larger with lower sedimentation rates, sphericity and smoothness of the surface increase with energy.

In the studied sediments oncoids are common and attain sizes between 2 mm and 3 cm. The smaller oncoids resemble micritic ooids of type 2b in internal structure but are less spherical and larger. They occur in shallow lagoons intermittently exposed to higher current velocities, as indicated by reworking. Larger oncoids frequently show only faint or no lamination at all, but low-energy lagoonal context and local encrusting by foraminifers and filamentous algae are criteria for their distinction. Large, rounded aggregates of *Bacinella-Lithocodium* associations and *Lithocodium* together with microbial encrusting are also classified as oncoids (s.l.). They display a comparable encrusting habit but no distinct lamination. Their presence is almost exclusively limited to intervals with low or no siliciclastic influence and commonly to shallow lagoons with intermittent, moderate- to high-energy conditions.

Pisoids

Coated grains with laminae rich in iron-minerals and resembling ooids in size and texture and commonly observed in association with paleosols. Other definitions, such as large ooids do exist in literature (Tucker & Wright 1990). Here, they are interpreted as pisoids that formed in a vadose environment during lateritic soil development (Fig. 2.3).

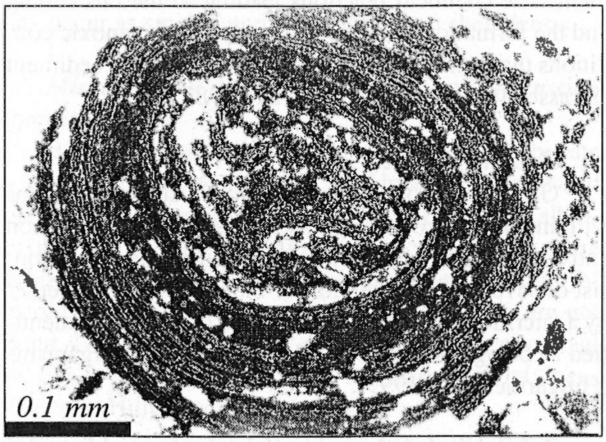


Fig. 2.3. Pisoid occurring in a paleosol. Laminae are made up of iron-rich minerals (probably goethite and/or chamosite) and incorporate detrital quartz grains.

Lithoclasts

Two types of lithoclast are distinguished: intraclasts that are components of slightly consolidated sediment and derived from the same general environment as the hosting rock by syn-sedimentary reworking, and extraclasts that are consolidated and eroded fragments of sedimentary and/or crystalline rock eroded in a foreign environment to that of the host sediment.

Intraclasts are very abundant in the studied Lower Cretaceous sediments. They are mainly represented by reworked and slightly rounded mudclasts or fragments of soft sediment in tidal environments and high-energy bioclastic bars. Alternating low-energy conditions and high current velocities in addition to lateral migration and associated erosion of channel banks are the main sources for these allochems. Bioclasts indicating a prolonged reworking and a filling with different sediment have the same significance. Such intraclasts occur preferentially in open-marine, high-energy environments.

Only few extraclasts are identified since it is often difficult to distinguish them from intraclasts. Single angular components, indicating (early) lithification with an unusual facies, occur in proximal platform positions and in shallow water and during times of elevated siliciclastic input.

Black pebbles are a special form of lithoclasts observed in the sediments. In most cases they have all characteristics of intraclasts being predominantly micritic, deformed and rounded, but they show an intensive black staining and a high content of pyrite. They always occur together with blackened fossils and black ooids. The blackening is caused by high contents of pyrite, commonly present as microscopic framboids. The presence of pyrite in such quantities indicates considerable quantities of bacterially decomposed organic matter and iron in the sediment. The chemical reactions leading to the blackening and the formation of pyrite generally require anoxic conditions that can occur in bottom waters or in the sediment (Strasser 1984, Strasser & Davaud 1983).

Grain aggregates

Grapestones occasionally occur in shallow lagoons with intense microbial activity indicated by micritization and encrusting of components. The composite grains consist of several ooids, peloids, or bioclasts bound together by a micritic envelope, which in some cases, can be identified as being composed of filamentous micro-organisms (Hillgärtner et al. 1999).

2.2.2. Skeletal grains (bioclasts and fossils)

Macrofauna

The platform sediments of the Lower Cretaceous are macroscopically very unexciting with respect to fossils. Only few specimens of bivalves (*Pholadomya*), echinoids, gastropods and brachiopods were collected. None of them has a biostratigraphic significance and, thus they are not specifically determined.

Benthic foraminifers

Benthic foraminifers can be looked at as the most significant microfossils in the sediments of the Upper Berriasian and Lower Valanginian of the northern Tethyan platform. They occur abundantly in most environments and have an important biostratigraphic value. Numerous studies have evaluated their occurrence in time and their paleoecology (Steinhauser & Lombard 1969, Darsac 1983, Salvini-Bonnard et al. 1984, Arnaud-Vanneau & Darsac 1984, Clavel et al. 1986, Boisseau 1987, Arnaud-Vanneau et al. 1988, Zaninetti et al. 1988, Adatte 1988, Blanc 1996). The effect of variations of environmental factors such as salinity, nutrient supply, physical disturbance and illumination on the occurrence of specific foraminifers and their diversity have been studied by Arnaud-Vanneau (1994). The determinations in this study are carried out to the genus-level and in obvious cases to the species level on the basis of published illustrations. Foraminifer diversity is estimated in relative values (low, moderate, high) according to the number of occurring species/groups.

Four associations of benthic foraminifers typical for different paleogeographic settings on the platform have been distinguished by Darsac (1983) and Arnaud-Vanneau & Darsac (1984). The observations in this study are in large parts consistent with their results.

<u>Ramp (proximal to distal):</u> Benthic foraminifers which are generally of small size (± 200 µm) are mainly represented by simple agglutinating forms (*Glomospira, Erlandia, Textularia, Haphlophragmoides, Valvulina*), some complex agglutinating forms (*Pseudotextulariella*), and nodosariids (*Lenticulina*). Small miliolids occur less commonly.

<u>Platform barrier / open lagoon :</u> A great abundance of benthic foraminifers is attained in this paleogeographic setting. Generally they are of large size (±1 mm) and are dominated by complex agglutinating forms (*Pseudocyclammina, Everticyclammina, Pfenderina, Pseudotextulariella*). Simple agglutinating forms (*Nautiloculina, Charentia*) and miliolids are also abundant. Trocholins and nubeculariids are present but of small size.

<u>Open lagoon (internal / external platform):</u> Here the foraminifer diversity achieves its maximum. Simple agglutinating foraminifers (*Textularia, Verneuilina, Arenobulimia, Nautiloculina*) and numerous trocholins attain their largest sizes (up to 1 mm). Many forms with calcareous porcelaneous tests occur (*Pseudotriloculina, Triloculina, Quinqueloculina, Sigmoilina, Pavlovecina allobrogensis*) and complex agglutinating forms (*Pseudotextulariella courtionensis, Pseudocyclammina*) may be present. Nubeculariids and *Tubiphytes* (classified as foraminifer-algae association, Schmid 1995) attain their highest abundance and size. Lenticulins occur abundantly as reworked constituents in tidally influenced lagoons.

<u>Restricted lagoons / coastal environment:</u> Abundance and diversity of benthic foraminifers is significantly reduced in these environments. Only few small miliolids and reworked lenticulins in tidally influenced settings are found.

Algae

Two groups of algae are encountered on the Jura platform: green algae and red algae, whereas the latter are rare and occur only in high-energy platform-barrier positions. Chlorophytes are represented by a variety of dasycladales (*Clypeina, Lagenoporella*, Conrad 1977) which occur in open-marine lagoons with low to moderate energy, and codiaceans that appear under open marine, moderate- to high-energy conditions. Charophytes are present in proximal platform positions and in a restricted marine/brackish context in the Vions Formation. Charophyte-ostracod assemblages have been studied in detail by Detraz & Mojon (1989).

Bivalves

These molluscs are omnipresent in small quantities. However, in more restricted lagoons they occur in abundance. Specimens of the genus *Pinnidae* are found in open lagoonal position and large endobenthic bivalves (*Pholadomya*) are discovered exclusively in tidal domains. A few, thin bioconstructions of rudists are observed in shallow, open-marine lagoons with reduced sedimentation rates. Lithophags and oysters frequently colonize firm- and hardgrounds.

Gastropods

Like bivalves they occur in all environments. In muddy to sandy restricted lagoons they occur in greater abundance.

Ostracods

In all environments with moderate- to low-energy conditions ostracods are present to abundant. In tidally dominated and/or restricted, marginal-marine lagoons they form an important constituent (Detraz & Mojon 1989).

Brachiopods

These stenohaline sessile organisms prefer a firm substrate in open-marine lagoons. They occur in open to external platform settings with well agitated waters.

Echinoderms

Echinoderm fragments are present in all types of sediment in variable quantities. Two different groups can be distinguished: crinoids and echinoids and/or asteroids (plates and spines). Their habitat is the open-marine realm with normal salinity. In external platform and ramp settings crinoid fragments are the most abundant fraction (Plate 3.5). Echinoids occur preferentially in open lagoonal environments. However, their fragments are easily transported and displaced by currents, which explains their great abundance in tidally-influenced, internal lagoons of the Vions Formation.

Bryozoans

Bryozoans are suspension feeding organisms that prefer stenohaline habitats. The most abundant and diverse fauna occurs together with crinoids indicating their preference for firmgrounds in open-marine, external platform positions (Plate 3.5, Walter 1997). Reworked fragments frequently appear in tidally-influenced, internal platform settings. Encrusting forms can be observed in open lagoons were they are associated with other encrusters (Plate 4.3).

Corals and sponges

Both types of organisms are very rare in the studied strata. No in-situ bioconstructions are observed and corals and calcareous sponges only occur as reworked clasts in some high-energy beach environments and lagoons that presumably were bordered by small patch reefs (Plates 1.5, 4.5). One exceptional abundance of calcareous sponges, together with crinoids and bryozoans in a marly interval at the base of the Guiers Member, is interpreted as deep ramp setting (Plate 1.3). Only few encrusting sponges occasionally occur in open lagoons.

Serpulids

Serpulids are suspension feeders that have the potential to create significant bioconstructions in lagoonal environments, but only few, slightly reworked serpulid constructions, approximately 5 cm in diameter, are observed in protected lagoonal environments (Plate 1.6). Commonly they occur as small, encrusting variety on shell debris.

Microbes and associated binding and encrusting organisms

Binding of sediment and encrusting of hard substrate is a common phenomenon in the Upper Berriasian and Lower Valanginian of the French Jura. It occurs in all intertidal to lagoonal settings and contributes to the stabilization of mobile sediment and the formation of oncoids (s.l.). Larger bioconstructions, however, are not found. The initial binding of carbonate sand by microbial activity is subject of a separate investigation and documented in Hillgärtner et al. (1999).

A variety of bacteria with photo- and heterotrophic metabolisms is to a large part involved in this process. Their activity is witnessed by micritic laminar, rarely columnar, stromatolitic crusts in the intertidal regime (Plate 2.6). A cloudy unstructured growth, commonly found in protected lagoons is interpreted as thrombolite. It is associated with an abundance of oncoids that generally are formed by cyanobacterial encrusting. A variety of other organisms, some with an uncertain status, contribute to the sediment binding and encrusting activity. They occur in ecologically similar conditions as oncoids (s.str.). *Cayeuxia*, formerly interpreted as a green algae, is now attributed to cyanobacteria (Leinfelder et al. 1993). *Tubiphytes* is ascribed to an association of foraminifers and algae (Schmid 1995), whereas *Thaumatoporella* is looked at as a green algae (Leinfelder et al. 1993). *Lithocodium* and *Bacinella* are associations of a foraminifer with cyanobacteria and algae, respectively (Leinfelder et al. 1993, Leinfelder & Schmid 1996). They are very common in open-marine lagoons and form rounded aggregates up to several centimeters in size. They indicate shallow open-marine waters that are well agitated and oxygenated, and hold a low turbidity (Dupraz 1998, pers. comm.).

Crustaceans and Vertebrates

Some fossilized bone fragments of turtles (Berger, Plint pers. comm.) and pincers of crustaceans were found in the Vions Formation. Blondel (1984) and Blondel et al. (1989) described cheloniens and decapods (*Mecochiridae, Axiidae*) in the same formation. These discoveries and the intense bioturbation (*Thalassinoides*) indicate a significant activity of crustaceans in a shallow, brackish-lagoonal to open-lagoonal environment.

2.2.3. Siliciclastics

Siliciclastic detritus commonly is derived from continental areas and is transported onto the platform by rivers, currents and wind. For the Berriasian, Steinhauser & Charollais (1971) have postulated transport of siliciclastics by coastal currents from the northeast to the southwest on the Jura platform, implying a connection with the Wealden facies in the Paris Basin.

The most abundant terrigenous particles are quartz grains with diameters ranging from 50 µm to 200 µm and a mean of 100 µm (Plate 9). Quantities of up to 35% can be reached, but 10% quartz content are rarely exceeded. Mean values are around 4% (refer also to chapter 8.3.1 and Fig. 8.3). In association with quartz some heavy minerals are observed in thin section (zircon, turmaline, pyroxene). This composition indicates a mature siliciclastic fraction. In marly intervals of the Vions Formation up to 30% quartz silt and up to 50% clay are detected.

2.2.4. Organic matter

In intervals rich in siliciclastics small fragments of charred organic matter are common. Paleosols are characterized by charred roots and occasionally associated with lenticular coal horizons with thickness up to 5 cm. Most organic matter was presumably present in a colloidal and dissipated form (refer to chapter 2.4.3).

2.3. Sedimentary structures

Primary sedimentary structures can be created by physical, biological, and chemical agents. They include all structures formed at the time of deposition or shortly thereafter and before the consolidation of the sediments. Specific structures form in erosional, depositional, and omissional regimes and bear an important potential for the interpretation of paleoenvironments. Detailed descriptions and discussions are given for example in Walker & James (1992), Tucker & Wright (1990), Allen (1984), or Scholle et al. (1983).

2.3.1. Structures related to the action of currents

Cross-stratification is observed at cm- to meter-scales and related to the transport and deposition of commonly sand-sized material under uni- to multidirectional currents with relatively high velocities. Depending on the angle and extent of foresets and vertical succession of different types of cross-bedding bathymetric trends from shoreface to foreshore/beach can be deduced (e.g., Tucker & Wright 1990).

Cm-sized current ripples represent a preserved bedform. They are found predominantly in shallow water depths in intertidal environments (Plate 4.6).

Cm-sized wave ripples evidence wave action in a very shallow-marine to intertidal context.

Flaser-bedding is typical for the alternation of current activity and quiescence in tidally influenced environments. It is very common in proximal platform positions in the Vions Formation (Plate 5.5).

Parallel lamination with a slight inclination points to high-energy conditions commonly encountered in beach environments.

Hummocky cross-stratifications are large, meter-scale, convex bedforms and characteristic for water depths between fair-weather wave-base and storm wave-base in a mid-ramp setting (Burchette & Wright 1992) (Plate 5.3). HCS is interpreted as being related to storm events (Duke 1985).

Erosion of semi-consolidated sediment commonly is indicated by truncated bedforms and sedimentary structures. It is encountered in all environments where sudden changes in energy conditions occur. This may be related to processes such as storms, waves, tides, gravity flows, switching channels, or changes in base level (Plate 4.7).

Grading of sediment in a normal or inverse sense is due to changing hydrodynamic properties of the transporting current. This generally is observed in high-energy settings.

2.3.2. Structures due to rapid sedimentation

Loading structures such as ball and pillow or flame structures are caused by rapid sedimentation of sand-sized sediment on water-saturated marl or shale intervals and the resulting equilibration of pressure gradients in the interstitial waters.

2.3.3. Structures creating primary porosity

Circum-granular cracks are created by contraction of a muddy matrix around solid constituents due to strong evaporation and drying-out of the sediment. These structures commonly accompany paleosol formation (Plates 3.1 and 5.2).

Birdseyes are interpreted as porosity created by active algal/microbial growth and trapping of gases released by decomposing organic matter and algal metabolism (Plate 2.6). This structure characteristically occurs in intertidal to supratidal environments (Shinn 1983).

Keystone vugs is a variety of fenestrae that is attributed to trapping of air bubbles in well-sorted sands, commonly in the swash and back-wash zone of beaches (Plate 4.1).

2.3.4. Structures related to biogenic activity

Characteristic associations of trace fossils are called ichnofacies. They encompass all in-situ traces left behind in the sediment by living animals during moving, breeding, and feeding activity. Three groups can be distinguished. In association with omission surfaces the three observed ichnofacies occur as pre-omission suite, omission suite, and post-omission suite, respectively (Fürsich 1979, Bromley 1975).

Soft-sediment ichnofacies

It is characterized by a diffuse bioturbation giving a mottled aspect to the sediment and obliterating sediment textures and structures. No specific trace fossils of this ichnofacies are identified.

Ichnofacies in semi-consolidated substrates

The corresponding trace fossils belong to the *Psiloichnus*, *Glossifungites*, and *Skolithos-Cruziana* ichnofacies (Pemberton et al. 1992) and typically are characterized by penetrative bioturbation. The burrows commonly display sharp boundaries and are filled with a different sediment. Well-preserved trace fossils of this type can be observed especially in the Vions Formation.

Arenicolites and *Psiloichnus* are trace fossils characterized by J- to U-shaped burrows with lengths of up to 60 cm's (Simpson 1975) and more than a meter (Frey et al. 1984), respectively. Burrows with these characteristics preferentially are found in lagoons with tidal influence, relatively high sedimentation rates, and anoxic conditions in the sediment, as indicated by abundant pyrite. *Arenicolites* and *Psiloichnus* are both typical for marginal-marine environments with tidal influence. The trace possibly belongs to crabs or a suspension- and/or deposit-feeding worm in its resemblance to traces of recent echiurans (*echiurus echiurus*) and enteropneusts (*balanoglossus clavigerus*) (Bromley 1990).

Thalassinoides is a burrow system consisting of vertical shafts, which connect largely horizontal tunnel networks. They display variable burrow diameters (2-10 cm) and typical Y-shaped branching patterns (Kennedy 1975). These burrows are attributed to crustaceans and decapods and are common in lagoons that undergo episodic omission phases and consolidation of the sediment. Hence, they are often associated with firm- and hardground formation (Bromley 1975, Fürsich 1979).

Ichnofacies associated with hard substrates

Characteristic for this ichnofacies (*Typranites*) is a bioperforation of the substrate. It typically displays destructive solution or scratching by boring organisms. This can be differentiated from penetrative bioturbation by a truncation of the entire sediment texture. In this study, only lithophags and other boring bivalves are encountered in association with hardgrounds (Plate 6).

Rhizoturbation

Roots that penetrate the sediment can also cause a disturbance of the internal sediment texture and stratification. This rhizoturbation that causes a mottled aspect of the sediment, and pebbly, irregular surfaces commonly is associated with cementation and concretions in vadose environments.

2.4. Early diagenesis

The early diagenetic history of a sediment, that encompasses all processes taking effect after deposition, essentially is marked by compaction (geometric rearrangement of constituents), cementation, and alteration (recrystallization, dissolution).

Involved processes are influenced by changing external factors in the depositional environment, and their traces can eventually furnish details about the regional post-depositional history (e.g., Plunkett 1997, Tucker 1993). In the present study early diagenesis was only looked at with the objective to obtain further indications of sea-level and climate change. The analytical methods used are thin-section petrography (plane and cross-polarized), scanning-electron-microscopy (SEM), and stable-isotope analysis.

2.4.1. Compaction

Compaction in unconsolidated sediment can lead to loading structures (see above) and nodular appearance of the limestone, especially when intercalation of heterogeneous sediments is present (differential compaction). An enhancement of this nodular bedding and formation of stylolites are related to burial compaction of the lithified sediment. Here dissolution occurs preferentially in places where rheological properties change. In pure limestones such changes are caused by facies changes, but for the most part, these structures are found in mixed calcareous-siliciclastic intervals. Dissolution can continue throughout the whole diagenetic pathway and is not restricted to early burial compaction (e.g., vertical stylolites due to tectonic stress).

2.4.2. Cementation

Cementation is the initial process leading to sediment lithification. Environment and composition of pore fluids cause precipitation of characteristic cement types (e.g., Bathurst 1975, Moore 1989, Tucker & Wright 1990). Subaerial exposure and influence of meteoric waters can lead to rapid cementation of carbonates (Wright 1994, Tucker & Wright 1990). Special attention was given to cements with asymmetric crystal habits that can indicate precipitation in vadose environments (Moore 1989). These include stalactite cements and associated meniscus cements (Plate 6.7). Problems with sediment textures that commonly are looked at as indicators for specific diagenetic environments are described in detail in Hillgärtner et al (1999). Early cementation in marine-phreatic environments is commonly associated with omission phases, or phases of reduced accumulation rate and intensified pumping of pore fluids through the sediment. It characteristically displays fibrous and micritic crystal-fringes. Such syn-sedimentary cementation can lead to the formation of firm- and hardgrounds (e.g., Shinn 1969, Dravis 1979) which consecutively are subject to biogenic alteration (bioperforation, encrusting, etc., refer also to chapter 3). Differences in the chemical environment commonly are reflected by the cement mineralogy. Pseudomorphs can indicate crystallization of characteristic minerals such as gypsum or anhydrite. Together with small euhedral dolomite in the matrix they are attributed to elevated ion concentrations of sulfate and/or magnesium in hypersaline pore fluids probably because of intensified evaporation (Tucker & Wright 1990).

Microbial activity in stabilization and early cementation of sandy carbonates

Characteristic fabrics such as micrite envelopes, calcified filaments and micritic grain-to-grain bridges are observed along discontinuity surfaces in this study and in Jurassic sediments. They compare well with textures observed in a modern firmground on an ooid sandwave near Wood Cay (Schooner Cays, Bahamas, Hillgärtner et al. 1999). Their similarity with microbial fabrics described in grapestones and in intertidal to continental-vadose environments suggests that microbial activity played an important role in the initial stabilization and cementation of carbonate sands. Problems with the environmental significance of such fabrics are discussed in Hillgärtner et al. (1999).

2.4.3. Alteration

Recrystallization and dissolution of a primarily lithified substrate are the main processes of early diagenetic alteration. Minerals that are metastable with respect to the pore fluids are transformed or dissolved and possibly replaced. Their new chemical composition then reflects the changing environmental conditions.

Such alteration is especially effective under the influence of meteoric waters and even more so when sediments are subaerially exposed. Small dissolution cavities in association with paleosol formation are interpreted as microkarst due to enrichment of CO_2 and humic acids in the pore fluids. Such processes commonly are associated with secondary infilling of the created porosity. The internal sediment can be vadose crystal silt (Tucker & Wright 1990), or lime-mud derived from sediments of a subsequent marine incursion. Commonly, such infills display geopetal structures.

Biocorrosion occurs when rootlets and roots penetrate lithified substrate (Plate 3.6). In many cases the zone around the root mould is micritized and/or calcified, and residual organic matter is preserved in the cavity. Structures such as anastomosing pedotubules, alveolar septal fabric, clotted fabric, incipient calcretization, and mottled micrite are interpreted as results of paleosol alteration (Wright 1994, Klappa 1980, Bain & Foos 1993, Mack et al. 1993). However, the alteration never penetrates more than a few tens of centimeters (max. 60 cm) in the studied sections, suggesting short-lived pedogenesis and/or partial erosion.

Carbonate precipitation and recrystallization under the influence of soil-gas and evaporation also have a significant effect on the isotopic composition of the limestone (Videtich & Matthews 1980). This is discussed in more detail in chapter 5.2.1.

Patchy dolomitization of matrix and components is interpreted as alteration during early burial diagenesis in the mixing zone (Badiozamani 1973, Tucker & Wright 1990). This process principally is related to elevated Mg^{2+}/Ca^{2+} ratios in zones where meteoric and marine waters interfere, and is influenced by the presence of organic matter (Soussi & M'Rabet 1994) and sulfates (Whitaker et al. 1994). Dolomitization in the studied sections occurs almost exclusively in conjunction with siliciclastics and organic matter which point to fresh-water influence.

Black facies

A very common early diagenetic alteration in the Lower Cretaceous is the blackening of limestone, termed "black facies" in this study (Plate 4.2). This blackening occurs independently of the environmental context (high energy, open platform, internal lagoons), but is always related to the presence of siliciclastics. In sedimentary environments this blackening essentially is the result of a bacterial decomposition of organic matter and mineralization of pyrite (framboids) which occurs, when iron oxides are present, and reducing conditions prevail (Jørgensen 1983, Berner 1989). The terrigenous influence on the blackening is not only restricted to the influx of iron oxides (attached to clays and silt/sand), but to the influx of terrestrial organic matter and nutrients (phosphates, nitrates). Elevated nutrient levels lead to an increased primary productivity which in turn causes higher consumption of oxygen by benthic organisms and aerobic bacteria (e.g., Hallock & Schlager 1986). Eventually, this leads to eutrophication of the environment, anaerobic degradation of organic matter by sulfate reducing bacteria, and consequently formation of pyrite (Pittet 1993, Jørgensen 1983, Siesser & Rogers 1976).

Quartz neoformation

Another common alteration in the observed limestones is the formation of authigenic quartz. It makes up approximately 2 to 20 % of the total quartz. These percentages correlate positively with the total abundance of quartz and consequently have no influence on the relative abundances of detrital quartz. Neoformation commonly

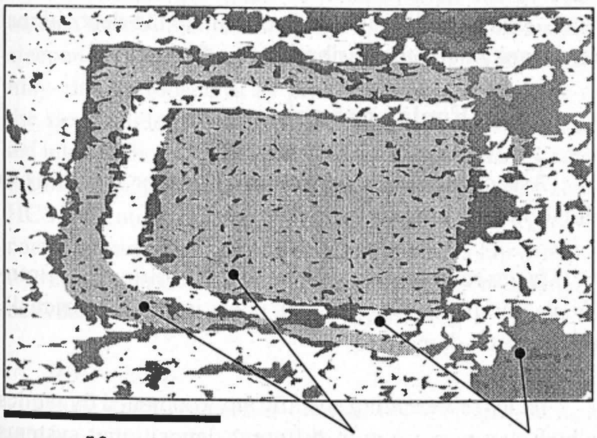


Fig. 2.4. Quartz grain with growth-related zonation in sample LC 121.1. Note slightly rhombohedral habit of the crystal suggesting pseudomorphic after dolomite.

is indicated by pseudomorphosis after dolomite resulting in clearly rhomboedric crystal habits (Plates 9.4, 9.5). Occasionally, zonations in crystals imply different phases of silicification (Fig. 2.4). Overgrowth on detrital quartz grains can only in rare cases be unambiguously identified. The distribution and type of quartz neoformation are coherent with observations of Pittet (1996) in the Oxfordian of the Swiss Jura and seem to be common phenomena in comparable early-diagenetic environments.

Early-diagenetic dolomitization commonly coincides with presence and high abundance of detrital quartz in the Upper Berriasian and is not pervasive except at the base of La Chambotte section. This positive correlation suggests that influx of fresh water, that carries the detrital load and can favor dolomitization in the mixing zone, is an important controlling factor. Maturation of organic matter (abundant in the Vions Formation) can lead to the generation of CO_2-rich fluids (Tucker 1993) and may have caused leaching of early-diagenetic dolomite. Silicification is favored in diagenetic environments rich in Mg and SiO_2 which could explain the association of dedolomitization and silicification (cf. Pittet 1996). The principal source of the SiO_2 cannot be assumed to be biologic, since sponges, diatoms, radiolarians, or other organisms with siliceous skeletons are not observed. Consequently, silica used in the authigenesis of quartz has to be of detrital origin itself. Alteration of feldspars and detrital quartz under a humid climate and transformation of clay minerals (smectite to illite) can liberate silica in the form of hydroxides (Hancock & Taylor 1978, Pye & Mazullo 1994). The precipitation of the quartz precursors (quartzine, opal-CT) from silica-rich solutions and their subsequent transformation into quartz is stimulated by the presence of magnesium-hydroxides and detrital minerals. (Williams & Crerar 1985, Hendry & Trewin 1995, Pittet 1996). This may explain the observed quantitative relationships.

2.5. Facies

Given the great abundance of different microfacies that occur in the Lower Cretaceous of the Jura platform and the fact that one microfacies type may occur in various environments (Pasquier 1995), they are not individually specified here but are directly integrated into a facies classification.

Figure 2.5a-c shows the facies types identified in the sections of the French Jura and Vocontian Trough. They are grouped into facies associations reflecting generalized depositional environments. Synthetic cross-sections of the platform are used to summarize the regional distribution of significant constituents and facies associations and regroup them into facies zones across the shelf (Fig. 2.6). This follows the approach of Wilson (1975) and Flügel (1982). Facies zone 1 and 2 are not represented in the French Jura but are encountered in sections in the Vocontian Trough and in Morocco.

2.5.1. Criteria for a facies distinction

Ecology and abundance of constituents

The applied facies classification takes into account all organic and inorganic constituents in consideration of their relative abundance, and in conjunction with their paleoecological significance indicated above. The abundance is estimated on the basis of thin sections and with the help of polished slabs (for larger constituents). This procedure is carried out differently according to particle size in adapting the microscope/binocular magnifications (Pittet 1996).

Magnifications between 2.5x and 40x are used for larger components like molluscs, oncoids, *Lithocodium-Bacinella* associations, etc.A magnification of 40x is used for the majority of particles like peloids, ooids, small oncoids, small molluscs, echinoderms, agglutinating foraminifers, algae, etc. A magnification of 100x is used for small foraminifers and ostracods.

The abundance is estimated in five categories that are compared to approximate percentages (in brackets):

- present, when a few components occur in the thin section/slab (0.1-2%);

- rare, when the specific component is easily found in the thin section/slab (2-5%);

- common, when a few components occur in the field of observation (5-15%);

- abundant, when components are abundant in the field of observation (15-70%);

- dominant, the component is a principal rock constituent (rarely the case) (70-100%).

The abundance of pyrite, dolomite, and glauconite is evaluated in the same way, whereas for quartz the percentage is estimated with the help of frequency diagrams (Bacelle & Bosellini, in Flügel 1982).

Texture

The limestone texture is defined according to Folk (1962) and Dunham (1962). It is a purely descriptive classification but holds certain implications about energy conditions in depositional environments regarding the presence and absence of lime-mud (Folk 1962, Dunham 1962, Plumley et al. 1962, Flügel 1982).

The term "Boundstone" (B) is used to describe all actively constructed facies including microbial mats and binding by *Lithocodium-Bacinella* associations. The term "Rudstone" (R) is used to describe facies with more than 10% of grains larger than 2 mm, following the classification of Dunham (1962). However, the term "Floatstone" is not applied. To be able to distinguish mud- and grain-supported textures for large grain-sizes the terms "Grainstone"(G) and "Wacke-/ Packstone"(W/P) are used in association with (R), instead.

Sedimentary structures

Sedimentary structures are incorporated in the facies classification according to their environmental significance indicated above.

Early diagenesis

This comprises the early diagenetic imprint on the rock, as far as it is characteristic for a specific depositional environment.

Facies context

Facies cannot be described by one sample only. It has to be kept in mind that one specimen, and even more so one thin section, represents only a fraction of the possible variability in a given environment. Therefore the adjacent lithologies (vertically and laterally) are incorporated in the concept of facies to allow a more comprehensive understanding of facies variability and facies evolution through time.

In cases where vertical facies transitions are dominantly gradual, or the facies change across small-scale discontinuity surfaces is modest, facies evolution can reflect lateral facies associations (according to Walther's law). This as well helps to discriminate the depositional environment and the attribution to a facies zone.

2.5.2. Facies groups

Associated facies are grouped together according to the following criteria:

- facies representing generalized depositional environments: ramp, lagoon (external, open, internal, restricted), tidal/coastal domain;

- facies representing specific environmental dynamics which can occur within different depositional systems (dune, bar, beach, bioconstruction, tempestite, marls);

- facies displaying environmentally relevant early diagenetic imprints and alteration. This "facies" is used as an additional suffix.

Selected examples of micro- and macrofacies are figured in Plates 1 to 5.

2.5.3. Facies zones

Eight facies zones are differentiated in this study and are numbered from distal to proximal positions (Fig. 2.5a-c).

FZ1 - slope / basin

This zone is not represented in the reference sections studied in detail in the Jura, but is typical for the sections in the Vocontian realm. Sedimentation is of a hemipelagic to pelagic type with alternating marl and limestone beds. Constituents are dominated by a pelagic-planctonic association of radiolarians, foraminifers, calpionellids, sponge spicules, calcisphaerulids, ammonites, and planctonic echinoderms. Locally, mass flow deposits such as debris flows and turbidites intercalate and may import platform detritus.

FZ2 - ramp

The ramp is an environment encountered rarely in the studied sediments on the Jura platform. It corresponds to the mid-ramp to outer-ramp setting sensu Burchette & Wright (1992) and is characterized by a deposition well below fair-weather wave-base and probably most of the time below storm wave-base (no tempestites). Consequently it is dominated by bio-micrites displaying intense bioturbation. It is the only facies-zone where sponges and sponge spicules are abundant in association with crinoids and bryozoans.

FZ3 - external lagoon

External lagoons are dominantly high-energy open-marine environments commonly encountered on distally-steepened ramps. They lack the protection of a barrier and may also be looked at as proximal ramp, depending on the slope angle (Read 1982, Schlager 1997). In the studied interval they are characterized by an open-marine fauna (crinoids, bryozoans, lenticulins) and commonly show HCS indicating regular storm influence. Local hydrodynamic subtidal bars indicate agitation above fair-weather wave-base; the bathymetric range therefore is estimated between 5 and 30 meters water depth.

FZ4 - barrier

The platform-barrier domain on the Jura platform essentially consists of high-energy subtidal bars with occasional indications for emergence. They are composed of open-marine carbonates dominated by ooids, peloids, large agglutinating foraminifers, coarse coral and other bioclastic rubble, echinoderms, and *Lithocodium-Bacinella* "oncoids". Typical is the reworked habit of the constituents. Shoaling may be indicated by beaches or vadose diagenesis. Barriers may also be formed by isolated islands where prolonged subaerial exposure is witnessed by paleosol formation. Biogenically constructed reefs are not observed in this study, but have been described during deposition of the Vions Formation by Detraz & Mojon (1989), Detraz (1989), Steinhauser (1969) and Darsac (1983), further to the east (Bauges Massif).

FZ5 - open lagoon (high-energy, protected)

This facies zone is the most common in the Berriasian/Valanginian of the Swiss and French Jura platform. It is characterized by open marine conditions with normal salinity and relatively low turbidity. Depositional environments range from high-energy lagoons (exposed to waves and/or currents) to lagoons that are protected, but not restricted. Limestone texture varies between grain- and mudstone, accordingly. Fauna and flora generally are diverse, especially foraminifers, which are represented by trocholins, large textulariids, large miliolids, and cyclamminids. All types of oncoids are abundant, their size and form being dependent on energy conditions. Patchy growth and sediment binding by *Lithocodium-Bacinella* associations and *Tubiphytes* occur when sedimentation rates are low. In open, protected environments dasycladacean mud- to wackestones are common. In contrast, higher-energy settings that are influenced by tidal currents and siliciclastic input show reworking, dominance of echinoderm fragments, peloids and ooids. During certain time intervals (Vions Formation) firm- and hardground formation can be characteristic in this facies zone.

FZ6 - internal lagoon (semi-restricted, restricted, tidally influenced)

Internal lagoons occupy the entire protected to restricted realm of the platform. Diversity of fauna and flora commonly is reduced, salinity may be increased (evaporation) or reduced (fresh-water influence), and living conditions can be adverse due to high intensity of tidal currents or anoxia in bottom waters. An increased terrigeneous influence is manifested in this environment by contents of quartz and marls, as well as particulate organic matter. In general, the facies displays intense bioturbation by few types of organisms and euryhaline fauna such as bivalves, gastropods, ostracods, serpulids, (few) charophytes, and small miliolids. The taphonomic influence of currents is the reason for the abundance of echinoderm fragments (usually stenohaline organisms) in internal lagoons strongly affected by tides. Here, flaser-bedding is a characteristic sedimentary structure.

FZ7 - coastal / tidal

The coastal and tidal domains include all environments with intertidal to supratidal characteristics. This can be tidal flats featuring laminated microbial mats, stromatolitic and/or microbial encrusting/binding with desiccation features or, more rarely, beaches with characteristic laminations. Tidal marshes are distinguished by mud to sand flats with coal seams and high contents of particulate organic matter.

Facies	Description	Dunham	Sedimentary Structures	Allochems	Fauna and Flora	Diagnostic Features	Interpretation of Depositional Environment	Facies-zone
RAMP								
R 1	Biomicrite	W	intense bioturbation		sponges, bryozoans, corals, incr. *Lithocodium*, echinoderms (partly crinoids), spicules, small textulariids and miliolids	generally black facies, quartz and pyrite	ramp (proximal), below storm-wavebase, detritic influence	FZ2 FZ3
LAGOON								***external***
L 1	Biopelmicrite (-sparite) with abundant echinoderms	W-P(G)	bioturbation, (lamination, x-bedding, ripples)	(well-sorted, abundant small) peloids and intraclasts, rare ooids (1a)	small, abundant, low diversity foraminifers, lenticulins, up to 35% echinoderm fragments (crinoids), brachiopods, bryozoans, sponges	quartz always present, < 0.1mm, black facies	open-marine, external high-energy lagoon, influence of tidal currents ?, (subtidal dunes or shoreface, transition to B3)	FZ3
L 1b	Biopelmicrite	P		peloids, ooids (1b), intraclasts	lenticulins, trocholins, brachiopods, bryozoans, crinoids		open-marine, external lagoon	FZ3
L 2	Bio (onco/pel) -micrite to sparite	P-G	HCS	peloids, oncoids	echinoderms, oncoids, *Lithocodium*, diverse small foraminifers		open-marine, external lagoon, storm influence	FZ3
								open
L 3	Biomicrite	M-W			foraminifers, some echinoderms, spicules		open-marine, low-energy lagoon, (protected)	FZ5
L 4	Bio (onco/oo/pel) -micrite, (-sparite)	W-G	bioturbation (sometimes intense, A4)	peloids, oncoids, ooids (1b,1c)	abundant and large cyclaminids and trocholins, diverse fauna, *Lithocodium*, few echinoderms		open-marine "carbonate lagoon", varying energy, below - above fairweather-wavebase (fwb)	FZ5
L 5	Bio (onco/oo/pel) -micrite with corals	W-P (G) R-Fl		peloids, oncoids, ooids (1b,1c)	abundant coral debris		open-marine lagoon surrounding reef or with patchreefs	FZ4 FZ5
L 6	Bio (onco/pel) -micrite, (-sparite)	P-G, R-Fl	coarse grained	peloids, oncoids, (intraclasts)	abundant *Lithocodium*, (some corals), echinoderms		open-marine, high-energy lagoon, slightly reduced sedimentation rate, (reworking)	FZ4 FZ5
L 7	Onco (bio/pel) -micrite to *Bacinella-Lithocodium* (algal) Boundstone	W-(P) B		oncoids, peloids	cloudy growth of *Bacinella -Lithocodium* associations or *Bacinella* framework between components of a grainstone, agglutinating foraminifers, commonly diverse fauna	quartz sometimes present	open-marine lagoon, stabilization of sediment, (by algal filaments?, *Bacinella*), reduced sedimentation rate, condensation	FZ5
L 8	Bio (pel/intra/oo) -micrite with abundant echinoderms	W-P	intense bioturbation, sometimes channel erosion	intraclasts, peloids, ooids some oncoids	10 - 20 % echinoderm debris (commonly urchins), ostracods, diverse foraminifers, lenticulins	abundant pyrite and quartz ≤ 0.1 mm, reworking	shallow, open lagoon, detrital influence, influenced by tidal currents, (condensation)	FZ5 FZ6
L 9	Onco (bio/oo/pel) -micrite	W-P	intense bioturbation	abundant oncoids, ooids, peloids	*Lithocodium*, echinoderms, molluscs cyclaminids, diverse fauna	quartz - 10%	open-marine lagoon, slightly reduced sedimentation rate, more or less protected	FZ5
L 10	Bio (onco/pel) -micrite with serpulids	W-(P)		oncoids, peloids	abundant serpulids, diverse fauna		open-marine lagoon, reduced sedimentation rate	FZ5
L 11	Bio (onco/pel) -micrite with dasycladaceae	M-W	bioturbation	oncoids, peloids	abundant green algae (dasycladaceans), bivalves ostracods, gastropods, echinoderms, some foraminifers		open-marine lagoon, protected, low-energy conditions	FZ5
								internal
L 12	Oo (bio/onco/pel) -micrite	W-P(G)		ooids, either (2a,2b or 1a-c), oncoids, intraclasts	echinoderms, molluscs, ostracods, foraminifers, (cyclamminids, trocholins, rarely brachipods and bryzoans)	qtz, dolomite, pyrite	internal to open lagoon, varying energy	FZ5 FZ6
L 13	Biopel (onco/oo) -micrite	W-P(G)	well sorted, bioturbation	peloids (abundant small), intraclasts, reworked ooids, oncoids	echinoderm debris (commonly urchins), (low diversity) foraminifers, ostracods, gastropods, rare dasycladaceans	quartz always present, ≤ 0.1mm often pyrite-rich, organic matter	shallow lagoon, internal to open, (distal) detrital influence, varying energy	FZ5 FZ6
L 14	Bio (pel/intra/oo) -micrite with abundant echinoderms	(W)P-G	ripples and flasers, intense bioturbation, sometimes channel erosion	peloids, intraclasts, (black pebbles), few broken ooids (1b)	5 - 15 % echinoderm debris (commonly urchins), abundant ostracods, low diversity foraminifers, lenticulins, molluscs	abundant pyrite and quartz 0.1 - 0.2 mm, some phosphate, reworking	shallow lagoon, proximal detrital influence, strong influence of tidal currents, condensation	FZ6
L 15	Bio(onco/pel) -micrite	M-W	intense bioturbation (A4)	peloids, oncoids	low diversity fauna, ostracods, few echinoderms, rare dasycladaceans and *Lithocodium*, molluscs	qtz	protected (internal) lagoon, reduced sedimentation rate, (rare storm influence)	FZ6
								restricted
L 16	Biopelmicrite	M-W	bioturbation		ostracods, some small miliolids, sometimes charophytes	often qtz	restricted lagoon or tidal pool (variable salinity - brackish lake?)	FZ6
L 17	Micrite (dolomitized)	M (W)			rare ostracods		restricted lagoon, transition to tidal mudflat	FZ6

Fig. 2.5 a. Principal facies occurring in the reference sections. First part.

Facies	Description	Dunham	Sedimentary Structures	Allochems	Fauna and Flora	Diagnostic Features	Interpretation of Depositional Environment	Facies-zone
TIDAL / COASTAL DOMAIN								
T 1	stromatolitic microbial mats, with silty, peloidal laminae	M/B	lamination (mm), birdseyes	peloids			microbial tidal flat	FZ7
T 2	homogeneous micrite to biomicrite	M/B	desiccation cracks, circumgranular cracks, birdseyes bioturbation	peloids	rare fauna: gastropods, bivalves, and ostracods	dolomitisation, qtz	mud flat	FZ7
T 2a	thrombolitic						microbial mud flat	
T 3	Bio (pel/onco/oo) -micrite	M-W	bioturbation, (birdeyes)	peloids, ooids (2a-c), oncoids	rare fauna: gastropods, bivalves, ostracods foraminifers	dolomitisation, qtz	intertidal lagoon to tidal flat	FZ7
T 4	Biopelmicrite with abundant quartz	W-P	ripples, lamination, bioturbation	well sorted peloids, intraclasts	ostracods, some foraminifers and other bioclasts	-30 % quartz	sandy mud flat to sand flat	FZ7
T 5	Bio (pel/onco) -micrite to -sparite	P-G	normal grading, erosional contact at base of bed	intraclasts, peloids, oncoids		(meteoric diagenesis)	tidal channel	FZ6 FZ7
T 6	Bio (intra/pel) -micrite	W		intraclasts (laminated M), peloids	bivalves, ostracods, gastropods, (charophytes)		tidal flat to semi-restricted intertidal lagoon	FZ6 FZ7
T 7	Biopelmicrite	M		peloids	ostracods, gasteropods, charophytes	"chalky" appearance	restricted (brackish) lagoon, coastal lake	FZ7 FZ8
DUNE								
D 1	Bio(onco/oo/pel) -micrite	P-G	sometimes foresets, lamination, normal grading	peloids, ooids (1b,1c,2a), oncoids		dm - beds	submerged hydrodynamic dunes (lagoonal)	FZ3-FZ6
D 2	Bio (oo/pel) sparite	G	very well sorted	peloids, ooids (1c,2a)		meteoric diagenesis	aeolian dunes?	FZ8
BAR								
B 1	Oo (bio/pel) sparite	G	well sorted, (big foresets), lamination	ooids (1a-c,2a), peloids		(dm) m - beds	open-marine submerged ooid bars	FZ3-FZ5
B 2	Bio (oo/onco/pel) -sparite	G	well sorted, (big foresets), lamination		few echinoderms	(dm) m - beds	open-marine submerged bioclastic bars	FZ5
B 3	Bio (oo/onco/pel) - Oo (bio/onco/pel) -sparite	(P-)G	well-med. sorted, foresets, normal grading, lamination, reworking	broken ooids (1a,1b, 2a), peloids, intraclasts	abundant ooids and / or echinoderms, bryozoans, oysters	pyrite	open-marine oolitic to bioclastic bars, tidal- or storm influence	FZ3-FZ5
B 4	Oo (bio/pel) micrite	W-P		ooids (1a-c, 2a), peloids, intraclasts		dm - beds, in bar context, same composition as bar	abandoned bar, interbar lagoon	FZ5
B 5	Bio intrasparite / -micrite	(P-)G	foresets, normal grading, lamination, reworking, x - bedding	peloids, broken ooids (1a), some intraclasts	abundant and big echinoderm-(crinoid) fragments, abundant and big bryozoans, bivalves	silicification, pyrite, Fe, glauconite	open-marine external bioclastic bars, proximal ramp (Calcaire Roux)	FZ2 FZ3
B 6	Oo (bio/pel) -micrite / sparite	P-G	emersion featues A1/2	ooids (1c, 2a), intraclasts	echinoderms, diverse bioclasts, corals		short-lived , high-energy emersive (ooid) bars (shoaling), may be tidally influenced, (internal) lagoon	FZ5 FZ6
BEACH								
Bch 1	Bio (onco/oo/pel) -sparite	G	lamination (cm), gradation (normal and inverse), keystone vugs	ooids (1c, 2a), peloids, oncoids, intraclasts	echinoderms, diverse bioclasts, corals	possible meteoric diagenesis	beach	FZ4 FZ7
Bch 2	Bio (onco/oo/pel) -sparite	G-R	coarse grained, x - bedding, planar lamination		echinoderms, diverse bioclasts, corals	beach context	lower beach / shoreface	FZ4
BIOCONSTRUCTION								
Bh1	Rudist framestone	Fr			in situ rudists bio (onko,oo,pel) micrite as matrix		rudist bioherm in shallow lagoonal, moderate to high-energy positions	FZ5
TEMPESTITE								
Tt 1	Bio (oo/intra/pel) -sparite	(P)-G	lamination (cm), gradation, ripples, erosional base	ooids,peloids, intraclasts	composition somewhat foreign to surrounding facies	cm - beds	tempestite - washover into or onto restricted / tidal facies	FZ6 FZ7
MARLS								
M 1	sterile marls (argillaceous)	m	lamination (mm)		ostracods?	dark grey to brown	restricted, protected lagoons, tidal pools with varying siliciclastic influence	FZ6 FZ7
M 2	silty to sandy marl	m			ostracods, foraminifers		lagoonal / tidal, open to semi-restricted, siliciclastic influence	FZ6 FZ7
M 3	bioclastic marls (argillaceous)	m	nodular appearance		sponges, echinoderms, foraminifers, bryozoans	grey	open marine, below storm wave-base	FZ2 FZ3

Fig. 2.5 b. Principal facies occurring in the reference sections. Second part.

Facies	Description	Dunham	Sedimentary Structures	Allochems	Fauna and Flora	Diagnostic Features	Interpretation of Depositional Environment	Facies-zone
BASIN								***pelagic***
Bp 1	Biomicrite	M-W and Marls	even bedding, lamination, bioturbation	peloids	radiolaria, calpionellids, cadosines, stomiosphaeres, foraminifera, echinoderms sponge spicules, ostracods bivalves, belemnites, bryozoa	limestone-marl alternations, org.-matter, phosphate, pyrite, quartz	normal pelagic, basinal sedimentation, low detritic influence	FZ1
Bp 2	Biomicrite	M-P (Marls)	erosive base, sole marks, (lamination), grading, bioturbation	intraclasts, peloids	radiolaria, calpionellids, cadosines, stomiosphaeres, foraminifera, echinoderms sponge spicules, ostracods bivalves, belemnites, bryozoa	org.-matter, phosphate, pyrite, quartz	pelagic turbidites and small debris flows sometimes with terrigeneous influence	FZ1
Bp 3	Biomicrite	M-W and Marls	erosive base, lenticular beds, beds with irregular base and top, low bioturbation	intraclasts, peloids	radiolaria, calpionellids, cadosines, stomiosphaeres, foraminifera, echinoderms sponge spicules, ostracods bivalves, belemnites, bryozoa	org.-matter, phosphate, pyrite, quartz	slumped pelagic deposits	FZ1
Bp 4	Biomicrite	M-W	lamination cross-bedding	peloids	radiolaria, calpionellids, cadosines, stomiosphaeres, foraminifera, echinoderms sponge spicules, ostracods bivalves, belemnites, bryozoa	org.-matter, phosphate, pyrite, quartz	pelagic turbidites and/or contourites	FZ1

Sub-Facies	Occurring in:	Dunham	Sedimentary Structures	Diagnostic Features	Interpretation of Diagenetic Environment	Facies-zone
ALTERATION / EARLY DIAGENESIS						
A 1	in lagoonal - tidal facies	M-P	desiccation cracks circumgranular cracks	dolomitization possible	desiccation, subaerial exposure	
A 2	in lagoonal - tidal facies	M-G	roots, root casts, brecciation (rhizo-turbation)	dolomitization, dissolution, meteoric diagenesis	paleosol, subaerial exposure	
A 3	in lagoonal - tidal facies	M-G	boring (lithophags), incrusting, multiphase bioturbation	Fe - staining	hardground / condensation	
A 4	in lagoonal - tidal facies	M-G	intense bioturbation, commonly with sharp edges and filled with different sediment	Fe - staining	firmground / condensation	
A 5	in lagoonal high-energy facies and bars	P-G	internal sediment (infil-tration), geopetal struc-tures, sometimes circ-umgranular cracks	meteoric diagenesis possible	abandonement of high-energy environments, stab-ilisation of mobile sedi-ment, possible emersion	

Fig. 2.5 c. Principal facies and early-diagenetic alteration occurring in the reference sections. Third part.

FZ8 - continental

Continental depositional environments are only suspected. Two observations of chalky carbonates with "rare" to "common" charophytes are interpreted as coastal lakes. In one locality (Lower Chambotte Fm., Salève section) facies of very well-sorted oo-pel-sparite is perceived as aeolian dune in a near-shore position (Kindler et al. 1997). All other indications of continental environments are restricted to reworking of continentally derived material and alteration of limestones. They are commonly manifested as black pebbles and paleosols, karstification and/or other indications of vadose diagenesis, respectively. The extent of limestone alteration suggests, however, that intervals where continental conditions prevailed were of "short" duration.

2.6. Facies model

To facilitate the understanding of the lateral and vertical variations and architecture of large-scale depositional systems and the environmental conditions by which they were governed, theoretical sedimentological models are developed. These generalized, interpretative facies models are based on the study of associations of litho-, micro-, bio-, ichno-, and diagenetic facies of sedimentary rocks which, in an uniformitarian approach, are compared with facies associations in modern sedimentary environments (e.g., Wilson 1975, Flügel 1982, Scholle 1983, Tucker & Wright 1990, Tucker et al. 1990, Walker 1992).

Two types of facies model commonly are used, one of which looks at the combined features of many local examples of a specific depositional environment, whereas the other represents a summary of a local/regional situation (Walker 1990). The latter type, resembling the "Wilson model" (Wilson 1975, Flügel 1982), is used in this study. Being static and in most cases two-dimensional it has to be kept in mind that such models are very simplified and rarely reflect the real depositional diversity.

However, to compensate somewhat for the static nature of such regional models two cross-sections of the Berriasian-Valanginian Jura platform are presented taking into account different platform morphologßies that occurred through time. They represent two end-members of various transitional platform configurations ranging from that of a flat-topped platform to that of a distally steepened ramp with local barrier islands (Figs. 2.6a,b). The models point out the lateral distribution and extent (variation) of facies zones (see below) and summarize the occurrence of characteristic constituents. A three-dimensional block-diagram (Fig. 2.7) is used to illustrate better the lateral coexistence of different depositional environments on the shelf.

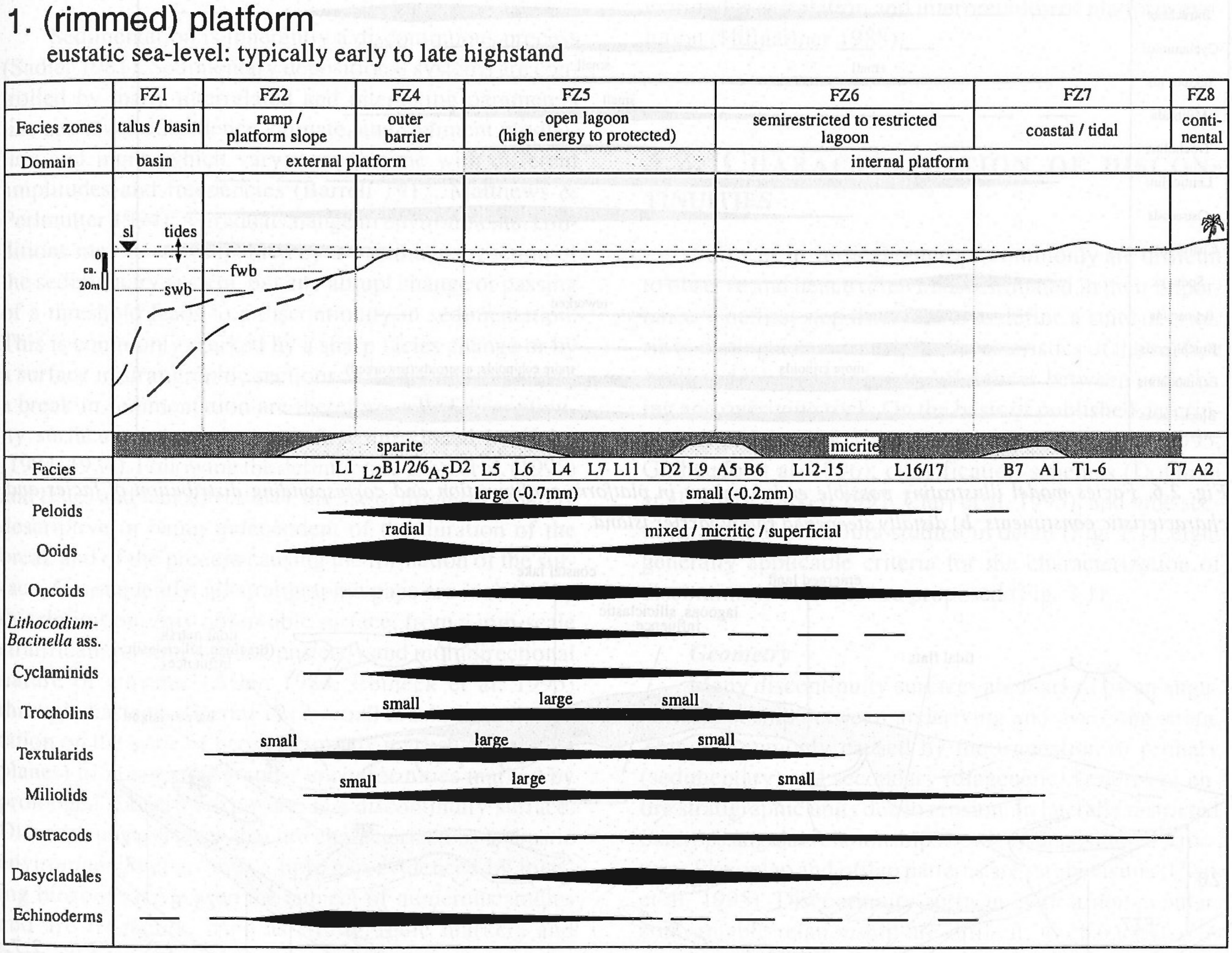


Fig. 2.6. Facies model illustrating possible end-members in platform configuration and corresponding distribution of facies and characteristic constituents; a) (rimmed) platform.

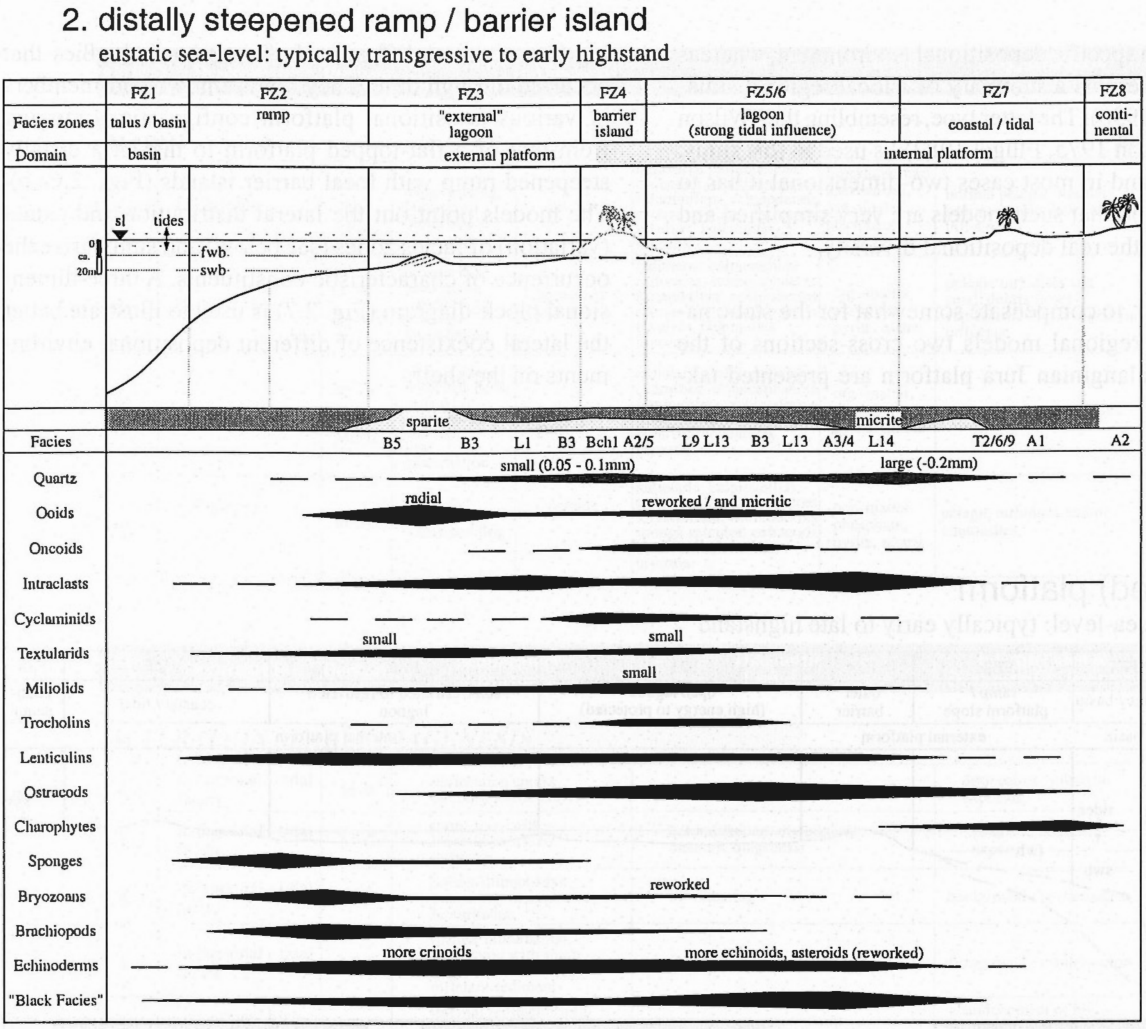


*Fig. 2.6. Facies model illustrating possible end-members in platform configuration and corresponding distribution of facies and characteristic constituents; **b**) distally steepened ramp/barrier island.*

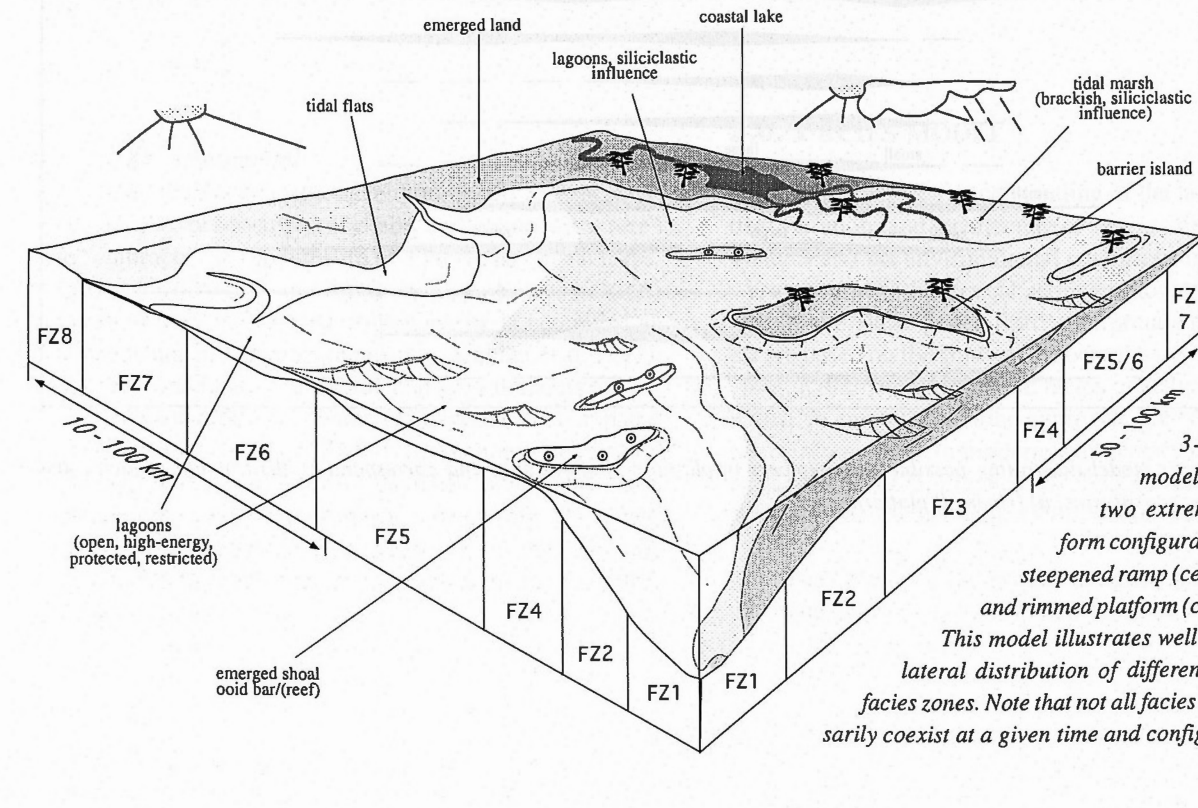


Fig. 2.7. 3-D facies model integrating two extremes of platform configuration: distally steepened ramp (center to right) and rimmed platform (center to left). This model illustrates well the possible lateral distribution of different facies and facies zones. Note that not all facies zones necessarily coexist at a given time and configuration.

3 - Discontinuity Analysis

3.1. Introduction

Sedimentation is inherently a discontinuous process (Sadler 1981). Sedimentary depositional systems are controlled by many interrelated and interacting parameters like sea level, subsidence, climate, and sediment production and input, which vary through time with different amplitudes and frequencies (Barrell 1917, Matthews & Perlmutter 1994). A gradual change in environmental conditions may be accompanied by a continuous reaction of the sedimentary system, but any abrupt change or passing of a threshold leads to a discontinuity in sedimentation. This is commonly marked by a sharp facies change or by a surface in stratigraphic sections. All surfaces indicating a break in sedimentation are therefore called discontinuity surfaces, a term that was first introduced by Heim (1924, 1934). Following the definition of Clari et al. (1995) and Bates and Jackson (1987), the use of this term is purely descriptive in being independent of the duration of the break and of the process causing the formation of the surface. Consequently, all stratigraphic gaps are included in this definition. Any observable surface, from a mm-scale stratification formed by the unsteady and multidirectional nature of currents (Allen 1984, Reineck et al. 1995) through diastems (Barrell 1917: small breaks in sedimentation on the scale of beds commonly marked as bedding planes) to major stratigraphic unconformities marked by prolonged subaerial exposure, is a discontinuity surface. Discontinuity surfaces that are characterized by a drastic environmental change or a time gap evidenced by missing biozones have been the subject of numerous studies and are frequently used as stratigraphic markers and boundaries of lithostratigraphic units, or are interpreted as sequence-stratigraphic bounding surfaces. However, subtle discontinuities on the scale of a bed and below biostratigraphic resolution are the rule in stratigraphic successions. Here, a process-oriented study of all surfaces in a stratigraphic section is necessary to determine their relative importance for the interpretation of the evolution of the sedimentary system. A systematic characterization and classification of small-scale discontinuities occurring on the shallow-marine Jura platform is proposed. Distribution patterns of such surfaces furnish additional information for correlation and interpretation of platform evolution (Hillgärtner 1988).

3.2. Characterization of Discontinuities

Surfaces in vertical outcrops commonly are difficult to observe and hence often underestimated in their importance. The first step therefore is to define a suite of criteria to distinguish even subtle characteristics of the surface itself and take into account differences between underlying and overlying rock. On the basis of published descriptions (Fürsich 1979, Bain & Foos 1993, Clari et al. 1995, Ghibaudo et al. 1996), classification schemes (Doglioni et al. 1990, Ricken 1991, Clari et al. 1995), and nine sections in the French Jura studied in detail (Fig. 1.1), eight generally applicable criteria for the characterization of discontinuity surfaces are proposed (Fig. 3.1).

Geometry

Many discontinuity surfaces are marked by an angular relationship between underlying and overlying strata. This is commonly caused by the truncation of primary (sedimentary) and secondary (diagenetic) features or entire stratigraphic units due to erosion. In laterally restricted outcrops angular relationships between depositional structures like onlap and offlap patterns are rarely visible (Clari et al. 1995). Discontinuity surfaces with a non-angular, conformable relationship are difficult to characterize in the field, and petrographic, geochemical and/or biostratigraphic evidence is needed for their interpretation.

Lateral extent

Observation of the lateral extent of discontinuity surfaces is a criterion strongly dependent on outcrop conditions. Single surfaces with wide lateral continuity tend to indicate an environmental change of at least regional importance (Walker & Eyles 1991, Meckel & Galloway

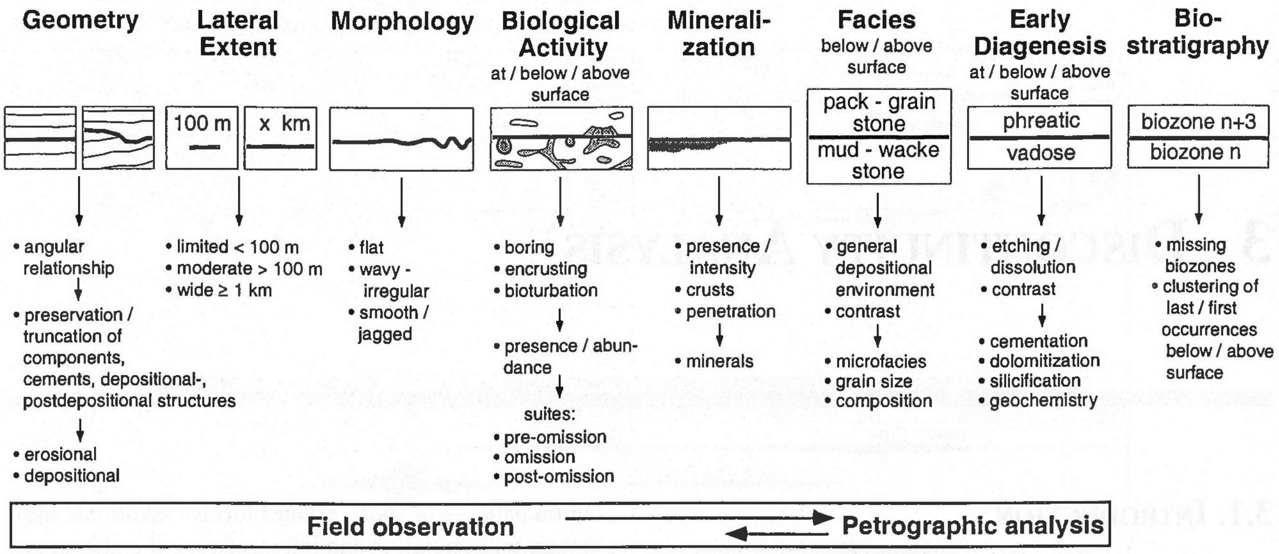


Fig. 3.1. Characterization criteria for discontinuity surfaces. Criteria are applied to the surface itself and the rock immediately above and below (Hillgärtner 1998).

1996). Stratification with low lateral continuity (several meters only) is commonly related to a locally restricted depositional process, such as cross-bedding in a subtidal bar. The environmental significance of such surfaces is included in the facies interpretation. In most studied outcrops, surfaces can be traced laterally over 100 meters to a maximum of 1 km (Fig. 1.1). Physical correlation of single surfaces between different outcrops (500 m-3 km) is possible only in cases where prominent surfaces are marked by unequivocal features.

Surface Morphology

The morphology of a surface can also be a criterion for the characterization of discontinuities (Jaanusson 1961, Read & Grover 1977, Fürsich 1979), although one of secondary relevance only. An irregular, wavy habit may indicate minor erosion in an otherwise homogeneous sediment without visible sedimentary structures. Biological activity can have a constructive or destructive influence on the surface morphology. Microbial crusts and bioherms create an irregular positive relief, whereas bioturbation and bioerosion cause rough surfaces by destructive modification. On the other hand, very flat, sharply cut surfaces, especially those in a shallow-marine environment, are commonly caused by high-energy abrasion of a homogeneously lithified substrate. Differential compaction may also lead to irregular surfaces, indicating an often subtle facies change. In other cases, stylolitization superimposed on a pre-existing discontinuity surface may lead to an irregular, commonly jagged surface morphology.

Biological activity

The importance of trace fossils for the genetic interpretation of discontinuity surfaces has already been pointed out by Bromley (1975) and Fürsich (1979). Intensity and type of bioturbation below and above the surface, as well as signs of encrusting and boring organisms at the surface are used as characterization criteria. A classification into pre-omission, omission, and post-omission suite of trace fossils is used to indicate their relationship with the hiatus (Bromley 1975). Lithification of the discontinuity surface is indicated by an omission suite including boring organisms that cut sharply through the fabric of the underlying rock (Purser 1969, Shinn 1970, Ghibaudo et al. 1996).

Mineralization

Sediments which are mineralized by iron and manganese oxides, phosphates, or authigenic minerals such as glauconite are associated with condensation (Föllmi et al. 1991, Carson & Crowley 1993, Gomez & Fernandez-Lopez 1994, Burkhalter 1995). Consequently, discontinuity surfaces showing *in situ* crusts of such a mineral paragenesis indicate a considerable break in sedimentation, commonly in subtidal environments. In contrast, crusts of aluminum-iron oxides and pyrite are found to be the result of paleosol formation (terra rossa) and alteration during subsequent marine flooding (Wright 1994). Penetrative staining of the underlying rock by iron oxides can also point to oxidation due to karstification and paleosol formation.

Facies contrast

Any sharp change in facies across a surface underlines the discontinuous nature of sedimentation. However, when the facies change takes place in the same depositional system, as for example by superposition of lagoonal mudstone onto lagoonal packstone, the relevance of the discontinuity surface is often difficult to assess and an incomplete sedimentary record can not automatically be inferred. An important break in sedimentation and a drastic environmental change can only be inferred from a superposition of facies contradicting Walther's law (Clari et al. 1995). Special attention is given to grain-size changes being basically a measure for energy variations, changes in composition (siliciclastics vs. carbonates), and changes in carbonate microfacies. Surfaces marking a facies contrast are often enhanced by late diagenetic processes such as pressure dissolution and stylolitization, which preferentially occur along a contact of rocks with different lithologic properties.

Early diagenetic contrast

Diagenetic evidence for a discontinuity is given where the underlying rock shows vadose cementation or early diagenetic alteration due to evaporation, meteoric waters, and/or paleosol formation, all indicating subaerial exposure (Videtich & Matthews 1980, Bain & Foos 1993, Wright 1994, Beach 1995). Where erosion has removed other sedimentological evidence, relative enrichment in ^{18}O may point to evaporation during exposure, and ^{13}C depletion can indicate the influence of soil gas (Videtich & Matthews 1980, Joachimski 1994). Different compactional features in underlying and overlying rock can point to different phases of early cementation (Clari et al. 1995). This can also be evidenced by cement stratigraphy, stable isotopes, and trace element analysis (Goldstein et al. 1991, Plunkett 1997).

Biostratigraphy

Bio- and chronostratigraphic data are the only means for time assessment of a hiatus. They can even be the only way to identify a discontinuity when there is a lack of any other diagnostic features, such as may be the case in basinal settings with rather monotonous sedimentation. The main limitation, however, is time-resolution, which commonly is too low for the majority of discontinuity surfaces occurring in the sedimentary record. In a more general sense and with less precision than bio- and chronostratigraphy other methods can furnish information about the time gap represented by a discontinuity. These include the degree of clustering of first and last occurrences of taxa around discontinuity surfaces and taphonomic characteristics of bioclastic concentrations along a hiatus (Holland 1995, Kidwell 1993). No biostratigraphic evidence indicating prolonged time gaps in the succession was obtained in the present study, which means that all observed discontinuities are below biostratigraphic resolution.

3.3. Observed surfaces and their interpretation

The nine sections studied were initially chosen according to continuous outcrop conditions to allow a study of the complete and tectonically undisturbed succession. This was an elementary condition, since the study was to be based on a detailed field examination on a cm-scale. Samples were taken in the immediate surroundings of each observable bedding surface and where possible of the surface itself. On the basis of field and petrographic evidence obtained during facies analysis, all surfaces were then characterized according to the criteria defined above. Nine major groups of surfaces can be observed in the studied succession, distinguished by their common features and environment of formation. Their characteristics are summarized in Figure 3.2 and their detailed interpretation is given below.

Subtidal firmground to incipient hardground

Low accumulation rate in a subtidal lagoonal environment favors consolidation and incipient cementation of the sediment at and directly below the water-sediment interface. Intense bioturbation by *Thalassinoides*-producing organisms accompanies this process. Burrows of the omission suite are commonly filled with the same facies found in the overlying rock (Plate 6.1). In rare cases the burrows even preserve a facies not recorded otherwise, indicating recurrent phases of deposition and erosion. Such filling of open burrows can also occur during storms ("tubular tempestites") and therefore indicate episodic high-energy events (Wanless et al. 1988). When the protective layer of unconsolidated sediment above the layer of incipient cementation is stripped off, weak erosion commonly forms an irregular surface because of the inhomogeneous nature of the lithification.

Subtidal hardground

The characteristics and the setting of this surface group are very similar to the one above, the only distinction being a more advanced lithification of the substrate. This allows colonization of the surface by boring and encrusting organisms, which commonly include *Lithophaga*, other boring bivalves (Plate 6.2), sponges and oysters. These hardgrounds compare well to the ones described by Kennedy and Garrison (1975) and Fürsich (1979). However, the abundance and diversity of both borers and encrusters are always low and no superimposed borings are observed, all pointing to a single omission phase. No evidence of early meteoric or vadose cementation was found and, therefore, an entirely subtidal origin of the lithification can be assumed.

High-energy hardground

This hardground type also shows encrusting and boring by the same suite of organisms as the subtidal one (Plate 6.3). However, only few *Thalassinoides* of the pre-omission suite occur. Early cementation is usually restricted to porous subtidal to lower intertidal high-energy carbonates like bioclastic- or ooid shoals. The lateral extent of these hardgrounds is rather restricted, commonly reaching not more than a few tens of meters. Their environmental setting and lateral continuity compare well to a recent occurrence of hardgrounds in oolitic shoals on Eleuthera Bank, Bahamas (Dravis 1979). There, they occur in a setting of relative high sedimentation rates and

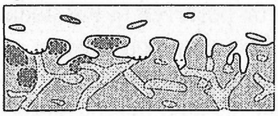
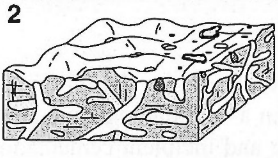
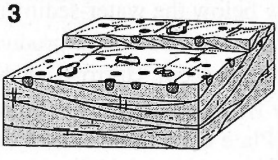
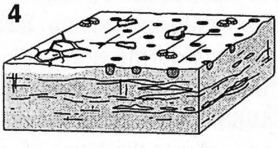
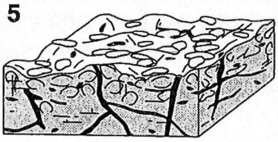
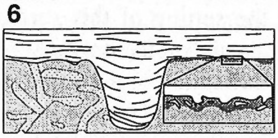
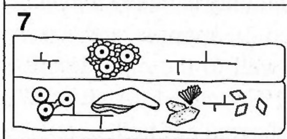
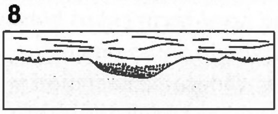
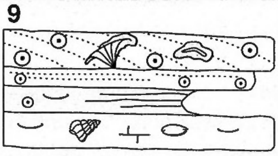
Surface Groups	Characterization						Interpretation (for detailed information see text)
	Geometry	Lateral Extent	Morphology	Biological Activity at / below / (above) surface	Minera-lization	Facies and Contrast across surface	Early Diagenesis below surface	
1 	erosional: cutting primary structures	100 m-1 km	irregular: relief often > 5 cm	bioturbation suites: intense preomission, moderate-intense omission: all *Thalassinoides*	weak FeOx	grain size composition microfacies	beginning marine phreatic cementation fibrous aragonite	**firmground to incipient hardground in subtidal environment**
2 	weakly erosional	> 100 m	flat to undulating	bioturbation suites: intense preomission, intense omission, weak-moderate post-omission: all *Thalassinoides* boring, encrusting: *Lithophaga*, Oysters	weak-strong FeOx MnOx possibly phosphates or sulfides	grain size composition microfacies	marine phreatic cementation fibrous aragonite microbial activity	**hardground in subtidal low-energy environment**
3 	weakly erosional	< 1 km	flat	bioturbation suites: occasionally weak omission: *Thalassinoides* boring, encrusting: *Lithophaga*, Oysters	weak FeOx	grain size	marine phreatic cementation fibrous aragonite microbial activity	**hardground in subtidal high-energy environment**
4 	weakly erosional	100 m-1 km	flat	bioturbation suites: weak preomission; (unidentified) boring, encrusting *Lithophaga*, Oysters	weak FeOx	composition	cementation microbial activity calcretization (desiccation) occasionaly relative depletion in ^{13}C and enrichment in ^{18}O	**hardground in inter- supratidal environment**
5 	weakly erosional	> 1 km	undulating often brecciated	bioturbation suites: weak-intense preomission: (unidentified) rhizoturbation: colonization by plants	strong FeOx	grain size composition microfacies	pedogenesis microbial activity dolomitization calcretization silicification relative depletion in ^{13}C and enrichment in ^{18}O	**paleosol commonly superimposed on subtidal facies**
6 	erosional: cutting primary and secondary structures	> 1 km	flat undulating jagged pitted etched cavities	bioturbation: weak-intense	strong FeOx penetrative staining	grain size composition microfacies	vadose cementation dissolution association with pedogenesis, gravitational cements and sediment infill in cavities	**microkarst / epikarst (karst)**
7 	depositional (erosinal)	< 1 km	flat	bioturbation: weak-moderate		microfacies grain size	cemetation contrast across the surface often vadose cem., dolomitization, silicification below discontinuity	**diagenetic discontinuity**
8 	erosional	< 1 km	undulating irregular	bioturbation: weak-intense		grain size composition microfacies	possible cementation contrast across the surface due to different rock properties	**erosion surface in sub- intertidal environment**
9 	depositional: preserving primary structures	< 1 km	flat undulating	bioturbation: weak-intense		grain size composition microfacies	possible cementation contrast across the surface due to different rock properties	**simple discontinuity surface in sub- intertidal environment**

Fig. 3.2. Groups of surfaces and their characteristic features observed in the Lower Cretaceous sections of the French and Swiss Jura (Hillgärtner 1998).

strong agitation by currents. The hydrodynamically active environment causes local winnowing, and the high rate of sea-water percolation and pumping through the porous sediment favors early cementation. Endolithic and chasmolithic algae (Dravis 1979) and other organisms inhabiting empty pore space (such as *Bacinella)* can contribute to the initial stabilization of the sediment (Hillgärtner et al. 1999).

Inter- to supratidal hardground

Shallow intertidal to supratidal environments also favor early lithification. Repeated drying, evaporation, and subsequent wetting lead to complex diagenetic processes. They may be evidenced by desiccation indicators (circum granular or desiccation cracks), precipitation of evaporites (gypsum), calcretization, and synsedimentary dolomitization. In some cases, the intertidal to supratidal environment of formation is indicated by stromatolitic microbial mats showing sheet cracks or mudstones with birdseyes. A sediment surface thus stabilized and lithified is prone to be rapidly colonized by boring and encrusting endofauna and epifauna during the following marine incursion. In many cases it will be difficult to distinguish from a subtidal hardground, except by facies analysis. Rarely, such surfaces display a very flat, knife-cut nature (Plate 6.4) where only the lower parts of the borings are preserved. They compare to the planar erosion surfaces of tidal origin (Read & Grover 1977) representing abrasion on a tidal, probably wave-cut platform.

Paleosol

Subaerial exposure of the sediment surface causes colonization by plants and development of a soil. Diagenetic processes associated with pedogenesis destroy primary structures and textures by strong micritization and recrystallization. Dissolution processes and rhizoturbation during exposure, as well as erosion and reworking of the soil during the subsequent marine flooding, cause the commonly brecciated to pebbly appearance of the surface. Roots, root molds, microscopic rhizoliths and other fabrics of soil development like soil pisoids and calcrete mottles (Bain & Foos 1993, Wright 1994) are evident (Plate 6.5). However, neither a vertical zonation of the soils nor a complete pedogenic transformation of the primary sediment were found, both of which would indicate an advanced stage of soil development (Martin-Chivelet & Gimenez 1992, Mack et al. 1993). This points to generally short-lived subaerial exposures with the formation of protosols (Mylroie & Carew 1995).

Microkarst, epikarst

Dissolution of carbonate by CO_2-enriched meteoric waters is indicated by sharp-cut and often intensively stained surfaces displaying a micro-relief with small cavities (Plate 6.6). Laterally, such a surface can show a relief of up to 15 centimeters with small paleokarst pits (Plate 5.1, Vanstone 1998). No caves or larger dissolutional features were observed, though. This can be attributed to a micro-karstification of lithified carbonates and formation of epikarst underneath a soil cover during seasonally humid climates (Wright 1994, Mylroie & Carew 1995), and to a limited duration of subaerial exposure (D'Argenio et al. 1997). Gravitational infill of cavities by stained and reworked lithoclasts and green, clayey calcisiltites in association with secondary micrite in form of crusts and diffuse patches (Beach 1995), and stalactite cements are also typical features of such karst development (D'Argenio et al. 1997).

Diagenetic discontinuity

A sharp change in the style of early diagenesis and cementation (not associated with any other change) commonly causes the accentuation of a surface by stylolitization due to differences in rheological properties of the bounding rock layers. All diagenetic discontinuities observed in the studied sections display a vadose diagenesis of the underlying rock contrasting marine phreatic diagenesis in the overlying rock (Plate 6.7). Such discontinuities are found in high-energy skeletal carbonates where sedimentary structures are absent or obliterated by bioturbation and cannot provide evidence for erosion. Vadose zones commonly extend downwards in the strata for a few tens of centimeters only, indicating no major lowering of the base-level.

Inter- to subtidal erosion surface

Erosion of underlying strata in inter- to subtidal settings may be encountered in a large variety of facies, but in all of them it is fundamentally related to an increase of energy in the depositional system. This may be event-related in the case of storms or gravity flows representing relatively short time spans in which erosion takes place. Long-lasting erosion and condensation can occur during lowering of base-level and wave-base, causing winnowing, sediment starvation, and erosion of subtidal sediments. The erosion surface itself is rarely indicative of the environment: facies and depositional environment are the keys for interpretation (Plate 6.8).

Simple discontinuity surface

Surfaces related to an abrupt facies change neither manifesting condensation nor erosion may include changes in texture, sorting, grain size, and mineralogy. Similarly to the intertidal to subtidal erosion surfaces, they rarely contain self-evident features which indicate an environment of formation. A variety of factors influencing a depositional system (see below) causes such discontinuities, and their environmental relevance can only be assessed by a detailed facies analysis of the host rock. Many of the bedding planes and stratification surfaces in the sedimentary record are of this type.

3.4. The significance of discontinuities

Commonly, discontinuities such as those described above are expressed as bedding planes and are responsible for the layering of strata in outcrop and subsurface. The various types of discontinuity surfaces commonly show different morphologic expressions in the field due to differential weathering, which can give a subjective impression about their relative importance. In any case, they reflect changes in the depositional environment at many different scales. The magnitude of the time gap they represent can vary considerably and may represent as much or more time as the sediments between them (Algeo & Wilkinson 1988, Walker & Eyles 1991).

The importance of environmental variations marked by discontinuities is generally evaluated by the time span represented by the corresponding stratigraphic gap and the lateral extent of the discontinuity surface (Salvador 1987, Nummedal & Swift 1987, Doglioni et al. 1990). Major time gaps on the order of several My commonly are related to environmental changes in response to global, long-term processes of tectono-eustatic origin (Sloss 1963). The stratigraphic record between such major unconformities commonly comprises various orders of sequences and stratigraphic units that are bounded by discontinuities representing shorter time-intervals and having laterally more restricted importance (Plotnick 1986, Ricken 1991). This hierarchical perception reflects the view of the stratigraphic record as a periodic accumulation of sediment in response to relative sea-level changes of varying amplitude and frequency (Vail et al. 1977, 1991, Goodwin & Anderson 1985, Posamentier et al. 1988, Goldhammer et al. 1990, 1991, Mitchum & Van Wagoner 1991, Osleger & Read 1991, D'Argenio et al. 1997).

However, a hierarchical perception of discontinuities based on duration of the gap (Ricken 1991) and lateral extent of the surface seems to be inappropriate, at least for shallow-marine platforms. Such environments are very sensitive to relative sea-level changes and react in different ways to variable superpositions and variations of amplitude and frequency of eustatic sea-level changes associated with different forcing mechanisms. High amplitude sea-level changes in icehouse worlds lead to well developed discontinuities due to longer duration of exposure and a major drop of the ground-water table causing extensive meteoric alteration of the sediments. Low amplitude sea-level changes in greenhouse worlds, in contrast, are typically characterized by poorly developed discontinuities reflecting short-lived emersions and sea-level falls not far below the platform surface (Read 1995). This means that hierarchical accommodation changes do not necessarily produce the same hierarchy of bounding discontinuities in the stratigraphic record. In addition, the conditions for the occurrence of laterally extensive single surfaces related to relative sea-level changes seem to apply only to a very restricted range of environments (Cartwright et al. 1993). Only low-angle platforms with a low morphology where changes in relative sea level lead to coeval environmental reactions over large areas may develop single, laterally extensive surfaces. Even in such environments factors such as amplitudes of relative sea-level change, lateral variability of depositional systems, and variations in sediment supply, accumulation and redistribution have a significant influence on the lateral extent of individual discontinuities.

In the studied sections of the Lower Cretaceous no hiatus exceeds the scale of biostratigraphic resolution, which is about 500 ky to 1 My, and single surfaces are rarely correlatable for more than a few kilometers. Therefore, other criteria than duration of the hiatus and lateral extent of the discontinuity surface have to be used to determine the significance of the observed discontinuities.

Environmental variables

On a shallow-marine carbonate platform, the principle variables controlling sedimentary processes are eustatic sea level, tectonic activity (including subsidence), and climate (Strasser 1991). These variables are interdependent in a complex way and have indirect global or at least regional effects. Variables which have a direct effect on the sedimentary system on a local to regional scale can be reduced to relative sea level, accumulation rate, the type of sediment available for sedimentation, and the energy regime (Fig. 3.3A). The variations in the type of sediment may be related to changes in autochthonous production or input from external sources, as for example siliciclastics washed in from the hinterland. Any significant and rapid change of any of these four variables causes a specific reaction in the sedimentary system. The reaction is manifested in form of subaqueous erosion, subaerial exposure, subaqueous omission, or texture and facies changes, which in most, if not all, cases produce a discontinuity surface. The relation of the observed surface-groups to these environmental reactions is illustrated in Figure 3.3A. "False discontinuities" or bedding planes caused only by late diagenetic processes (Bathurst 1991) were not encountered in this study.

3.5. Classification of Discontinuities

On the basis of these expressions of environmental changes in the stratigraphic record, a simple classification is proposed for the observed discontinuities (Fig. 3.3B). Four surface types have been distinguished taking into account the importance of allogenic forcing of environmental change. The first type includes all surfaces resulting from subaerial exposure, regardless of how they are manifested in the stratigraphic record. In most cases encountered in this study, however, exposures are indicated by an overprinting of subtidal facies. This is an unequivocal sign for a relative lowering of sea-level and cannot be caused by progradation or lateral migration of facies belts or changes in the sediment supply (Schlager 1993). It is therefore a marker for an allogenic forcing of the sedimentary system (Strasser 1991). The second type

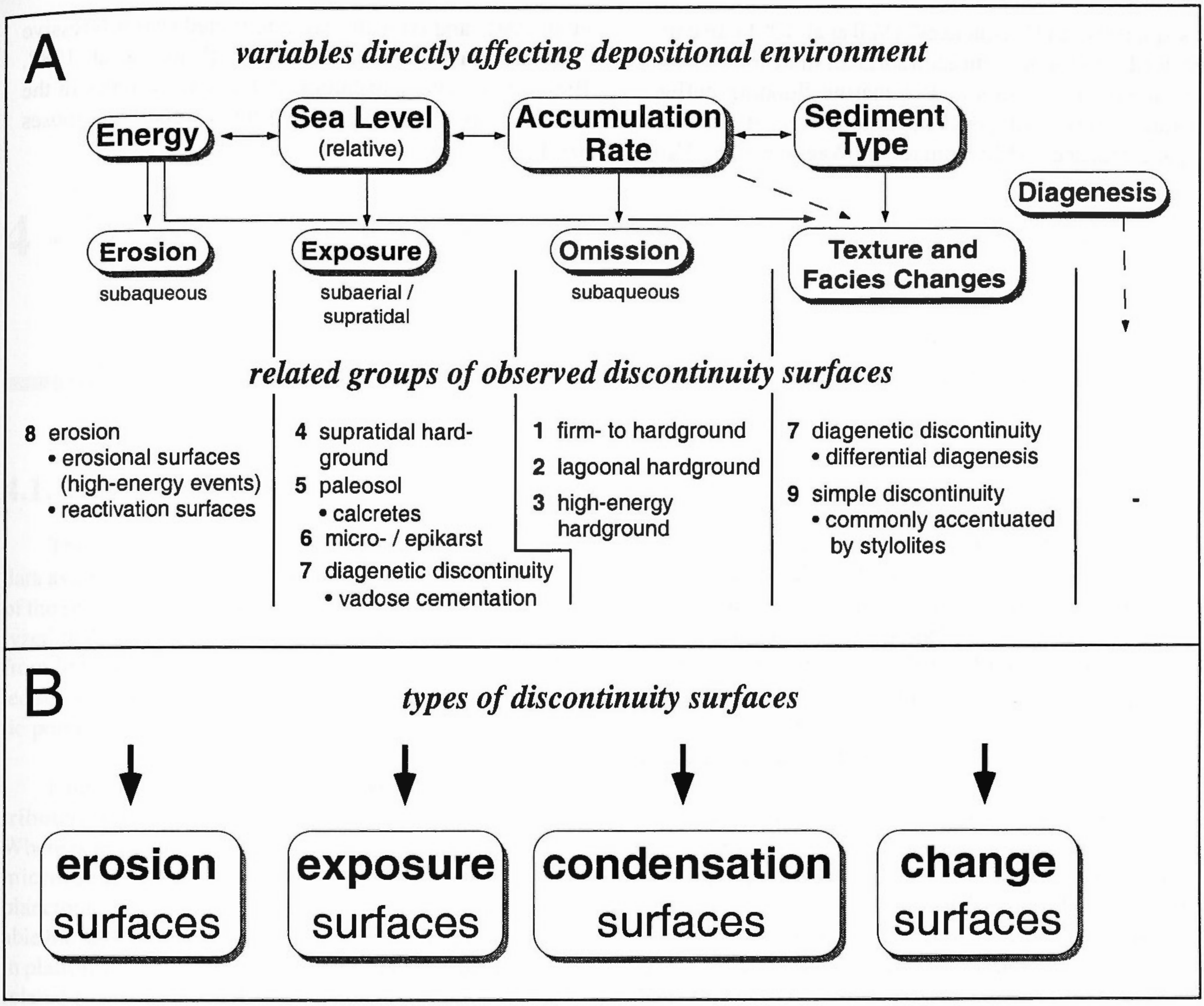


Fig. 3.3. Environmental variables directly controlling sedimentation on a shallow-marine carbonate platform. Influences such as climate change or tectonic activity are considered as indirect effects. A) Variables which for any given position on the platform lead to a specific reaction of the sedimentary system when they change rapidly or by a considerable amount. The observed discontinuity surfaces that are related to these reactions of the sedimentary system are listed below. B) Classification of discontinuity surfaces according to the predominant environmental change (Hillgärtner 1998).

comprises all discontinuities related to stratigraphic condensation in a subtidal environment. Stratigraphic condensation on shallow-marine platforms is very common and can be caused by local processes such as sediment bypassing or winnowing by locally restricted currents. It is, however, commonly related to relative sea-level changes (Galloway 1989, Kidwell 1993, Gomez & Fernandez-Lopez 1994, Burkhalter 1995). Omission can occur during initial-flooding or maximum-flooding phases causing sediment starvation, or during maximum-regression phases where exposure is not attained but lowered wave-base induces winnowing and submarine erosion (Osleger 1991). The third type includes all discontinuities that show evidence of subaqueous erosion. The fourth type describes all surfaces indicating facies and/or texture changes. In most cases, surfaces indicating small-scale erosion or facies changes are related to locally restricted depositional processes. However, allogenic forcing related to climatic and/or relative sea-level changes can for example lead to changes in wave base or energy conditions leading to erosion or to a sudden input of siliciclastics (Strasser & Hillgärtner 1998). Some surfaces show a combination of subtidal condensation and subsequent exposure. Hardgrounds that display a vadose overprinting or pedogenic alteration are considered as exposure surfaces because here the latter process is clearly related to an allogenic forcing and indicates an environmental change of at least regional importance (Strasser 1991).

This classification is basically consistent with cyclostratigraphic and sequence-stratigraphic analyses where surfaces indicating exposure or at least a shoaling-up facies evolution capped by supratidal sediments are used to delimit "peritidal cycles" (Goldhammer et al. 1993) or "simple sequences" and "sequences" (Vail et al. 1991). In contrast, well-constrained omission surfaces indicating small-scale submarine erosion and/or marine flooding define "subtidal cycles" (Osleger 1991, Goldhammer et al. 1993) or "parasequences" (Mitchum & Van Wagoner 1991, Vail et al. 1991) and typically are interpreted as transgressive and/or maximum-flooding surfaces (Loutit et al. 1988, Brett 1995). The distribution of the surface types in the studied sections and their value for correlation purposes are discussed in chapter 6.

4 - BIOSTRATIGRAPHY

4.1. INTRODUCTION

This chapter gives an overview of the biostratigraphic data available for the studied interval. Fortunately several of the sections studied here are well known and were analyzed in detail by others in this respect. All data retrieved from literature that could be precisely placed in the studied sections are integrated and used as chronostratigraphic tie points.

Fauna relevant for biostratigraphic analyses is distributed depending on the paleogeographic setting. Whereas in basinal sections abundance of ammonites and microfauna, such as calpionellids, dinoflagellates, planctonic foraminifers, and nannofossils provides valuable biostratigraphic datums, these organisms are very rare in platform facies. Here, benthic foraminifers that are correlated to ammonite and calpionellid zones are the main biostratigraphic tool. Occasionally, charophyte-ostracod assemblages in internal platform positions can be used as markers.

Fig. 4.1 illustrates the relevant biostratigraphic scales and their correlation for the Berriasian and Valanginian stages.

4.2. AMMONITES

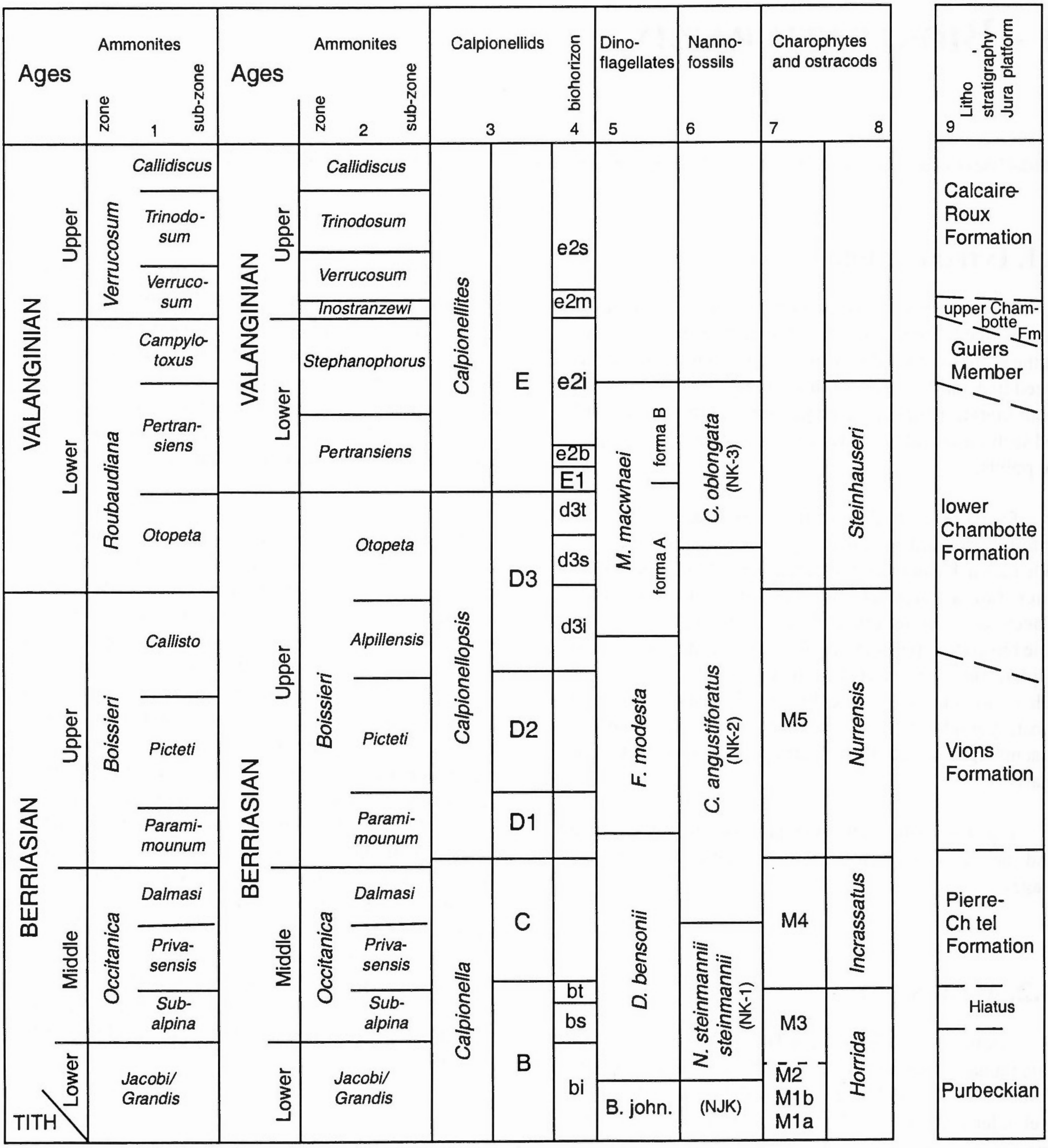
Sections in the Vocontian Trough are rich in ammonite fauna and are well studied in this respect (Le Hégarat & Remane 1968, Le Hégarat 1971, 1980, Atrops & Reboulet 1993, Bulot 1995, Bulot et al. 1997). The ammonite stratigraphy used here is based on ammonite zonations provided by Bulot (1995) and Le Hégarat (1980) for the Berriasian, whereas the zonation by Le Hégarat (1971, 1980) is used for the Early Valanginian. The reason for this combination of two ammonite zonations lies in the fact that relevant studies about benthic foraminifers and calpionellids on the platform are calibrated to the scales of Bulot and Le Hégarat (in Blanc 1996), whereas the majority of the Valanginian biostratigraphic data (basinal sections) are well calibrated to the scale of Le Hégarat (1971, 1980). The Berriasian-Valanginian boundary is placed at the base of calpionellid zone E, which almost exactly corresponds to the base of the *Pertransiens* ammonite zone, as it was recommended by the working group on the "2nd International Symposium on Cretaceous Stage Boundaries" (Bulot 1996).

Ammonites observed in basinal sections during this study were neither collected nor identified, since the biostratigraphic framework is already well established for these areas. No ammonites were found in platform sediments of the French and Swiss Jura during this field campaign. However, exceptional ammonite discoveries on the platform by Clavel (1986) and Waehry (1989) allowed the dating of the top of the Purbeckian as *Jacobi-Grandis* zone and the base of the Pierre-Châtel Formation at Mount Salève as *Privasensis* subzone. This suggests the existence of a hiatus during the *Subalpina* subzone on the platform (Deville 1991). The base of the Vions Formation coincides with the appearance of the benthic foraminifer *Pavlovecina allobrogensis* (see below) and could be attributed to the *Paramimounum* subzone at Mount Musiéges (situated between Vuache and Val du Fier sections). Localized ammonite occurrences in the Jura mountains imply Early to Late Valanginian ages (*Pertransiens*?, *Campylotoxus*, *Verrucosum*? subzones) for the characteristic Calcaire Roux-type facies (Detraz & Mojon 1989, Blanc 1996).

4.3. CALPIONELLIDS

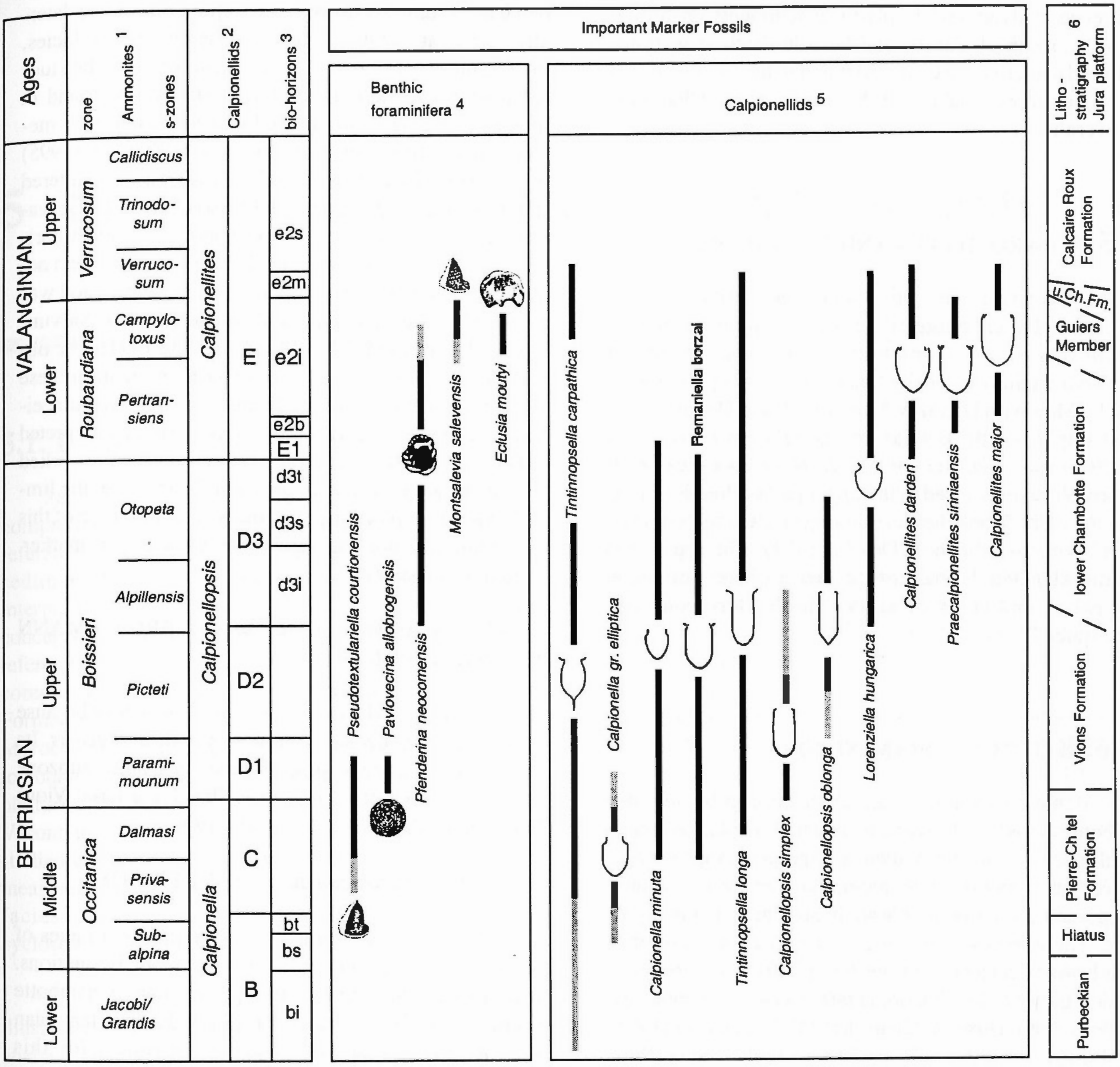
Calpionellids are most abundant in the basinal domains. They were intensively studied by Remane (1963, 1985), Le Hégarat & Remane (1968), Le Hégarat & Ferry (1990), and Blanc (1995) in the Vocontian Trough, and the established biozonation is well correlated with ammonite zones (Fig. 4.1). Important marker species observed in this study and mentioned in the sections are illustrated in Fig. 4.2.

On the Jura platform only few intervals contain scarce calpionellids. The few individual specimens encountered in this study were not determinable. However, Darsac (1983) could identify *Calpionellopsis* sp. at the base of



1 after Le H garat & Remane (1968), Le H garat (1971,1980)
2 after Le H garat (1980) & Bulot (1995), in Blanc (1996)
3 after Remane (1985)
4 after Blanc (1996)
5 after Monteil (1993)
6 after Bralower et al. (1989)
7 after Detraz & Mojon (1989)
8 after Schudack (1993)
9 this study, mainly based on Clavel et al. (1986)

Fig. 4.1. *Comparison of different biostratigraphic charts for the earliest Cretaceous applicable in the Tethyan domain.*



1 after Le H garat & Remane (1968), Le H garat (1971,1980), Bulot (1995)
2 after Remane(1985)
3 after Blanc (1996)
4 after Clavel et al. (1986), Blanc (1996), Charollais & Wernli (1995)
5 after Blanc (1996)
6 this study, mainly based on Clavel et al. (1986)

***Fig. 4.2.** Major marker fossils in platform and basin environments used in the present study.*

the Vions Formation in La Chambotte section and Zaninetti et al. (1988) described *Calpionellopsis oblonga* (zone D) in the middle part of the Vions Formation in the Monnetier section. This allows to attribute the Vions Formation to the Upper Berriasian. Associations of *Calpionellites darderi* and *Tintinopsella carpathica* found in the Salève section (Deville 1991) allow to place the Guiers Member in the *Pertransiens* and *Campylotoxus* ammonite subzones.

4.4. Other microfauna

Fauna such as dinoflagellates and nannofossils are well known and abundant in basinal sections and the Vocontian Trough (e.g., Bralower et al. 1989, Monteil 1992, 1993). They were not studied here and for more detailed analyses and their biostratigraphic arrangement specifically in the Angles section the reader is referred to Jan du Chêne et al. (1993) and Giraud et al. (1995). Occurrence of dynocysts in platform sediments on Mount Vuache, south of Vuache and Fort de l'Ecluse sections, allowed to ascribe the Guiers Member to the *Campylotoxus* subzone (unpubl. data by E. Monteil, cited in Charollais & Wernli 1995).

4.5. Charophytes and ostracods

Charophyte-ostracod assemblages, although only occasionally encountered in marly, marginal-marine sediments of the Vions Formation are an important biostratigraphic tool on the Jura platform. They were studied by Mojon (in Detraz & Mojon 1989 and Deville 1991) and imply a Late Berriasian age (*Paramimounum* to *Callisto* subzones) for the Vions. More specifically, an assemblage discovered in the upper part of the Vions Formation of the Monnetier section suggests a Late Berriasian age (*Callisto* subzone) (Deville 1991). The top of the Purbeckian and the base of the Pierre-Châtel Formation reveal assemblages of early Middle Berriasian age (Pasquier 1995).

4.6. Benthic foraminifers

Benthic foraminifers are abundant and broadly distributed across the Lower Cretaceous Jura platform. Several species with a limited biostratigraphic range serve as excellent chronostratigraphic markers of which the most important ones used in this study are illustrated in Figure 4.2. The extensive literature on benthic foraminifers in the Lower Cretaceous of the Jura platform is here represented only by a few important references: e.g., Steinhauser (1969), Steinhauser & Charollais (1971), Darsac (1983), Salvini-Bonnard et al. (1984), Clavel et al. (1986), Zaninetti et al. (1988), Adatte (1988), Martini & Zaninetti (1995), Charollais & Wernli (1995), and Blanc (1996).

Pavlovecina allobrogensis STEINHAUSER, BRÖNNIMANN & KOEHN ZANINETTI 1969

This characteristic, large foraminifer is attributed to the *Paramimounum* subzone (Clavel 1986). *Keramosphaera allobrogensis* STEINHAUSER, BRÖNNIMANN & KOEHN ZANINETTI, a name formerly used and well known in the Jura domain, is the type species of *Pavlovecina* LOEBLICH & TAPPAN, n.gen. (Loeblich & Tappan 1988). Its first occurrence generally coincides with the appearance of siliciclastically influenced carbonates, typically attributed to the Vions Formation, and it is dated as *Paramimounum* subzone (Clavel et al. 1986). The first appearance probably is facies dependent in some sections, since in pure limestones laterally correlatable with *Pavlovecina*-bearing mixed facies, the foraminifer is not observed. In almost all of the studied platform sections *P. allobrogensis* can be found in intervals ranging from one bed to a maximum of 9 meters. This confirms observations made by Pasquier (1995) in the eastern Jura where *P.allobrogensis* was encountered in intervals up to 5 meters thick. However, this observation opposes older descriptions, where one distinct horizon with *P. allobrogensis* was identified and a second occurrence about 10 meters higher up in the succession was proposed in the Monnetier and Salève sections (Salvini-Bonnard et al. 1984, Clavel 1986, Deville 1991). The observations of a second *P. allobrogensis* horizon in these sections cannot be confirmed here. It is suggested that either the extended interval of occurrence was misinterpreted as two distinct horizons, or that an occurrence of reworked specimens gave such an impression. In any case, the limited biostratigraphic range and the wide distribution of this foraminifer, making it an ideal chronostratigraphic marker, cannot be questioned.

Pseudotextulariella courtionensis BRÖNNIMANN & CONRAD 1966

This textularid foraminifer is mentioned here because of its characteristic association with *P. allobrogensis*. Its first occurrence can be placed in the *Privasensis* subzone and it typically occurs in Pierre-Châtel and basal Vions Formations (Pasquier 1995, Blanc 1996).

Pfenderina neocomensis PFENDER 1938

This species is common in open platform facies of the upper Vions and entire Lower Chambotte Formations. It disappears abruptly at the top the Lower Chambotte Formation implying a Late Berriasian to Early Valanginian age (*Alpillensis* to *Pertransiens* subzones) for this lithostratigraphic interval (Blanc 1996). Occurrences together with *Montsalevia salevensis* (*Campylotoxus* subzone, Blanc 1996) cannot be confirmed in the studied sections.

Montsalevia salevensis CHAROLLAIS, BRÖNNIMANN & ZANINETTI 1966 and ***Eclusia moutyi*** SEPTFONTAINE 1971

M. salevensis is exclusively found in open to external lagoonal facies interpreted as representing the Guiers Member. It occurs together with *E. moutyi* in the Fort de l'Ecluse section (Charollais & Wernli 1995) and *Calpionellites darderi* in the Salève section (Deville 1991), both indicating Lower Valanginian (*Campylotoxus* subzone) age.

5 - Sedimentological and Sequence Analysis

5.1. Introduction

Sequence analysis, as applied here, is based on a comprehensive sedimentological analysis of the eight reference sections in the French Jura. The encountered sedimentological patterns and phenomena are then interpreted in terms of sequence- and cyclostratigraphic concepts. The same criteria and methods tested for the reference sections are applied in the interpretation of the complementary section in the Jura, the Pierre-Châtel Formation in general, and the complementary sections in Morocco, which were sampled in less detail. This allows to obtain a wider spatial and temporal control of the Jura platform. The basinal domain covered by sections in the Vocontian Trough is ruled by different, but generally well studied environmental dynamics. The sections there are measured and described in detail, but the monotonous facies variations allow to apply interpretative cyclostratigraphic concepts without exhaustive sampling.

The first part of this chapter is concerned with the illustration and definition of the methodology of sequence analysis, and of the concepts and terms used. Secondly, all sections will be presented and interpreted individually, following the scheme:

- *Geographic, geologic and stratigraphic setting;*
- *Sedimentological analysis (general facies evolution and main events);*
- *Interpretation and sequence evolution.*

5.1.1. Stratigraphic sequences and cycles

It has been realized for a long time now that the discontinuous and cyclic nature of the sedimentary rock record reflects a variety of recurrent environmental changes through time (e.g., Suess 1885, Gilbert 1895, Barrell 1917). Concepts such as allostratigraphy, genetic stratigraphy and sequence stratigraphy (for related references see Posamentier & James 1993) aim to relate the cyclicity in the rock record to a time-stratigraphic framework. Their common basis is the division of the stratigraphic successions into chronostratigraphic "sequences" bounded by discontinuity surfaces, whereas significance and genesis of the discontinuities are valued differently. Common cyclostratigraphic concepts, in a different approach and on a different scale relate cyclicity in the rock record to orbital signals and attempt to establish an independent, high-resolution chronostratigraphic framework (Fischer 1995, DeBoer & Smith 1994, Kauffman et al. 1991).

Sequence stratigraphy, the most accepted stratigraphic concept today, was initially put forward by a research group of EXXON (Vail 1977). They presented a model that explains stratigraphic patterns and the evolution of sedimentary basins mainly as the effects of large-scale eustatic sea-level changes. The model that has been extensively discussed (e.g., Vail 1987, Posamentier et al. 1988, 1992 Posamentier & James 1992, Posamentier & Allen 1993, Hunt & Tucker 1992, 1995, Kolla et al. 1995, Miall 1986, 1991, 1997) was developed as a tool to predict the geometry of sediment bodies in the search for petroleum reservoirs. It was essentially based on seismic studies in siliciclastic-dominated systems at passive margins. One of the major problems was the low vertical resolution of the source data, which only resolve major lithological changes. With advancing seismic techniques and especially with detailed field studies, higher-resolution data of lithological variations are obtained nowadays, whereas the geometry of sediment bodies often is less evident (restricted outcrops). The resulting modifications and refinements of the initial concept suiting individual case studies led to a continuous redefinition of principles and terms, and an ever finer stratigraphic subdivision (e.g., Mitchum & Van Wagoner 1991). This has raised questions about the validity of global correlations of apparently eustatic events as presented in "global cycle charts" (Miall 1986, 1991, 1992, 1997). In addition, it has been realized that eustatic variations are not necessarily the dominating factor controlling the formation of stratigraphic sequences in tectonically unstable settings and carbonate systems (e.g., Schlager 1989, 1991, 1993, Hubbard 1988, Cloetingh 1988, Posamentier & Allen 1993).

Abundant high-resolution sedimentological studies relate small-scale cyclicity to astronomically defined insolation variations, that trigger changes in climate and oceanic circulation (Milankovitch 1941, Berger 1988, 1990). The understanding of the mechanisms involved and their sedimentary signatures is a main domain of research today (e.g., Fischer 1986, 1991, Goldhammer et al. 1993, 1994, Strasser 1994, Read et al. 1995, Pittet & Strasser 1998a, 1998b). Cyclical variations of sediment types and mineralogy, of palynofacies, and of geochemical, paleo-ecological, and various other markers are related to processes such as changes in oceanic productivity and redox potential on the sea floor (e.g., Einsele & Ricken 1991), shifting climate belts (e.g., Perlmutter & Matthews 1989, 1992, Hallam 1994, Pittet 1996), or short-term eustatic sea-level variations (e.g., Montañez & Osleger 1993, Pittet 1994, Balog et al. 1997). These studies demonstrate that orbital signals recorded in a variety of environments in many different ways are the result of complex interactions of atmosphere, hydrosphere, and lithosphere. The controlling orbital parameters (precession, obliquity, eccentricity) are well defined in terms of cycle duration back to Mesozoic times (Berger et al. 1989, Berger & Loutre 1994). The fact that they are so often recorded in sedimentary environments, regardless of intermediate feedback mechanisms, make the corresponding sedimentological signatures a particularly good and independent chronostratigraphic tool (Kauffman et al. 1991, Cotillon 1991, Fischer 1995). The obtained high-resolution time framework allows quantification of sedimentary and tectonic processes (Pittet 1994, 1996).

It has to be mentioned however, that some authors doubt the cyclical nature of the stratigraphic record (e.g. Drummond & Wilkinson 1993, 1996, Wilkinson et al. 1996, 1997). They base their critic essentially on the argument that statistical analysis of thickness variation of stratal elements and lithofacies variation in several published cyclostratigraphic analysis (Wilkinson et al. 1996) and in one section with a large number of peritidal cycles (Wilkinson et al. 1997) compare to data obtained from randomly created successions of peritidal cycles. Although their critic of an unreflected use of cyclostratigraphic principles is reasonable their work basically shows that it is essential to distinguish between external and intrinsic factors forcing a sedimentary system. This can only be attained with a detailed 3-dimensional analysis of the sedimentary system. An oversimplification of facies associations and an analysis of one section alone do not account for the complex evolution of sedimentary systems through time.

In terms of hierarchy, sequence- and cyclostratigraphy can be looked at as two end-members in the analysis of stratigraphic sequences. Sequence stratigraphy was initially concerned with large-scale stratigraphic sequences with thicknesses of tens to thousands of meters deposited in response to eustatic and tectono-eustatic cycles, and with durations of millions of years (Vail et al. 1977). Cyclostratigraphic studies, in contrast, initially described small, meter-scale depositional sequences (called cycles in these concepts) in basin settings with regular facies variations, and which formed in response to climatic and glacio-eustatic cycles with durations of 10-100 thousands of years (cited in Fischer et al. 1990, Weedon 1993).

More recently, the concepts of sequence stratigraphy are applied to smaller-scale systems (Mitchum & Van Wagoner 1991), and the terminology is proposed to be entirely independent of scale (Posamentier et al. 1992). Modern cyclostratigraphy seeks to reconcile large-scale, basin-wide processes with Milankovitch-type climate variations (Perlmutter & Matthews 1990, 1992) and the concepts are applied also to platform environments (e.g., Fischer 1964, Goldhammer et al. 1990). The major problem today is the unfeasibility to scale down the simplistic sequence-stratigraphic model and apply it to all kinds of environments without modification, for the reasons described above. On the other hand, no systematic and accepted terminology exists in cyclostratigraphic concepts for complex depositional sequences (sedimentary cycles) in platform environments, which genetically compare well to "sequences" sensu Vail et al. (1991). Although the fundamental controlling processes and the scale of time and space are not the same, sedimentary systems show a comparable logic in the response to accommodation changes on all scales.

For the time being, all concepts undergo continuous modification, and many studies apply customized terminologies suiting their specific case (e.g., Naish & Kamp 1997). The terminology used herein (defined below) evolved from abundant high-resolution analyses of platform carbonates (Fribourg working group) and attempts to combine sequence- and cyclostratigraphic aspects (Strasser et al. 1999, Strasser & Hillgärtner 1998, Hillgärtner 1998, Strasser 1998, Hillgärtner et al. 1998, Pittet & Strasser 1998a, 1998b, Pasquier & Strasser 1997, Pittet 1996, Pasquier 1995, Strasser 1994). It is partly based on Vail et al. (1991) and adapted to the specific study conditions where: 1) it is impossible to determine directly the geometry of sediment bodies; 2) discontinuity surfaces can be traced (physically) only for limited distances (1.5 km); 3) the reference sections supply exclusively data about environmental dynamics on the platform, and only complementary sections provide information on contemporaneous basinal dynamics. For practical reasons and in order to avoid the creation of new and awkward terminologies, the general principles of sequence stratigraphy (Vail et al. 1977, Vail et al. 1991, Posamentier et al. 1988, Strasser, 1994) are applied to all scales of stratigraphic patterns.

5.2. Depositional sequences

Definition

A depositional sequence is a succession of genetically related sediments whose facies evolution and/or stacking pattern is repetitive within a section. Commonly, depositional sequences are delimited by discontinuity surfaces or zones of well marked (facies) change indicating an inversion in the trend of environmental change. They are scale and time independent and are the stratigraphic expression of recurring environmental changes. For simplicity the terms "cycle" and "cyclic" will be used for such changes, although recurrences in natural systems are at the most quasi-periodic. In this study the term "sequence" is used synonymously with "depositional sequence" and therefore complies only partially with the original definition (Vail et al. 1977).

5.2.1. Criteria for sequence identification and interpretation

Criteria to identify initially and interpret subsequently depositional sequences are manifold. Sedimentological analysis provides various parameters such as general facies evolution, variations in ecological indicators, interpretation of beds and discontinuities, and vertical variations in bed thickness ("stacking pattern"). Specific analyses of diagenetic features, stable isotope signatures, and the study of clay-mineral composition and palynofacies can yield supporting evidence (Plunkett 1997, Pittet & Gorin 1997, Pasquier 1995, Joachimski 1994).

Facies evolution

The study of lateral facies variations and vertical facies evolution is the principal tool in sequence analysis. It allows four-dimensional interpretations of depositional environments. Particularly on platforms, depositional environments are intimately related to bathymetric conditions and their evolution evidences intervals of shallowing (shallowing-up) or deepening (deepening-up). With sufficient lateral control transgressive and regressive phases and variations in sedimentation rate can be implied. Facies evolution is the sedimentary expression of environmental changes, which can be the result of cycles intrinsic to the depositional system (autocycles), or follow cyclic or quasi-linear (mainly long-term cyclic) external forcing mechanisms (allocycles) (Strasser 1991).

Discontinuity surfaces

Discontinuity surfaces on shallow-marine carbonate platforms can display a wide variety of characteristic elements (refer to chapter 3). All surfaces (excluded are "false discontinuities" of purely late-diagenetic origin) reflect reactions of the sedimentary system to rapid and drastic environmental changes. Such surfaces actually record the most important times during platform evolution, namely times of the highest dynamics in environmental change. However, only a detailed and individual investigation of each surface can reveal information on what environmental change caused its formation. Reactions of sedimentary systems to changes in energy regime, relative sea level, accumulation rate, and sediment type are expressed by subaqueous erosion, exposure (including any erosion in a subaerial setting), sub-aqueous omission, and facies changes. A repetitive occurrence of important environmental changes marked by discontinuity surfaces can be used to delimit depositional sequences. Refer also to Chapter 3.5.

Bioturbation

Ichnofacies and the degree of bioturbation, in general, are closely related to specific environmental conditions. Hence, they can be an excellent tool to delimit depositional sequences (Knaust 1998, Ghibaudo 1996, Pemberton et al. 1992). In this study only *Typranites*, *Glossifungites*, and *Skolithos* ichnofacies are identified (refer to chapter 2). More attention is given to the degree of bioturbation which is a function of accumulation rate, oxygenation, substrate type, and nutrient content of the sediment (e.g., Wetzel 1991, Wignall 1993).

In platform environments oxygenation levels are rarely the limiting factors for the occurrence of bioturbation. Nutrient supply commonly is sufficient to sustain intense biological activity. Terrigeneous input, however, can significantly increase nutrient levels which may result in major overturns of benthic communities and consequently in a drop of carbonate production and accumulation rate (e.g. Hallock & Schlager 1986, Dupraz & Strasser 1998). In the studied sections the intensity of bioturbation is highest in lagoonal and tidal environments. Herein the intensity seems to be independent of substrate type (e.g., muddy vs. grainy), but commonly is closely associated with firm- and hardgrounds (chapter 3). This implies that sediment accumulation-rate is the most important controlling factor on bioturbation intensity in these environments.

Important variations in accumulation rate can be caused by relative sea-level changes: rapid initial rise causing a lag time before significant carbonate production starts (initial flooding) (Strasser 1991); fastest rise of sea-level and increase in bathymetry inducing changes in benthic communities, drop in carbonate production-rate and sediment starvation (maximum flooding); rapid sea-level fall with decrease in bathymetry and/or increase in energy causing sediment by-pass situations. Commonly associated with sea-level changes are variations in climatic conditions and water quality (e.g. salinity, turbidity) which additionally influence carbonate production.

Taphofacies

Bioclastic concentrations (e.g., shellbeds) commonly are associated with discontinuity surfaces (chapter 3) and, thus, can be significant for the identification and characterization of depositional sequences. They indicate low sediment accumulation rates, caused either by sediment starvation during sea-level rise, or erosion, reworking, and condensation during sea-level fall and lowstand (Kidwell 1993, Abott & Carter 1994, Naish & Kamp 1997). Other sedimentological and paleontological criteria, such as sedimentary structures and faunal assemblage are needed to determine their environmental significance (e.g., Kidwell 1989). Such bioclastic concentrations are rare on the Jura platform, but quite abundant on the Moroccan shelf. There ammonite accumulations and oyster shellbeds commonly are associated with submarine hardgrounds and erosional surfaces, pointing to condensation during rising and falling sea level, respectively.

Bedding and stacking pattern

In mixed carbonate-siliciclastic sedimentary systems the identification and interpretation of individual beds in terms of sequence analysis can be rather difficult. Single beds are always defined by discontinuity surfaces of various types, as mentioned above (except for late diagenetic bedding, Bathurst 1991). The presence of bedding therefore has diverse reasons, and its appearance in the field can be modified by the effects of weathering. In certain cases single beds can be described as depositional sequences using the criteria defined above. The stacking pattern and thickness variations of all depositional sequences (including beds) can point to environmental changes through time, such as variations of accommodation and accumulation rate. Thick and thin depositional sequences (in a relative sense) with an internal shallowing-up trend can indicate high and low accommodation, respectively. However, when they display an internal deepening trend accommodation was not the limiting factor. Here, varying thickness indicates changing accumulation rates, which can also be a function of accelerated rise in relative sea level (Fig. 5.1). The stacking pattern of depositional sequences can therefore be taken as a proxy for trends in relative sea-level changes, in consideration of eustasy, subsidence, sediment production, and sediment input. Yet, these principles cannot be taken as a rule and have to be applied with caution since they hold only true for depositional sequences representing an equal amount of time.

Clay content

Variation of clay content in platform carbonates commonly is expressed by differential weathering of the marly intervals in outcropping strata. This can accentuate apparent bedding and, thus, can help to discriminate bedding rhythms and depositional sequences.

Most clay minerals in marly sediments are of detritic origin and result from erosion of crystalline bedrock or reworking of paleosols. Surficial runoff and streams then transport the clays into the marine realm. In general, deposition of marls is related to low-energy conditions. Marly intervals are observed in shallow protected and restricted lagoons and in deeper, open-marine settings below the action of waves.

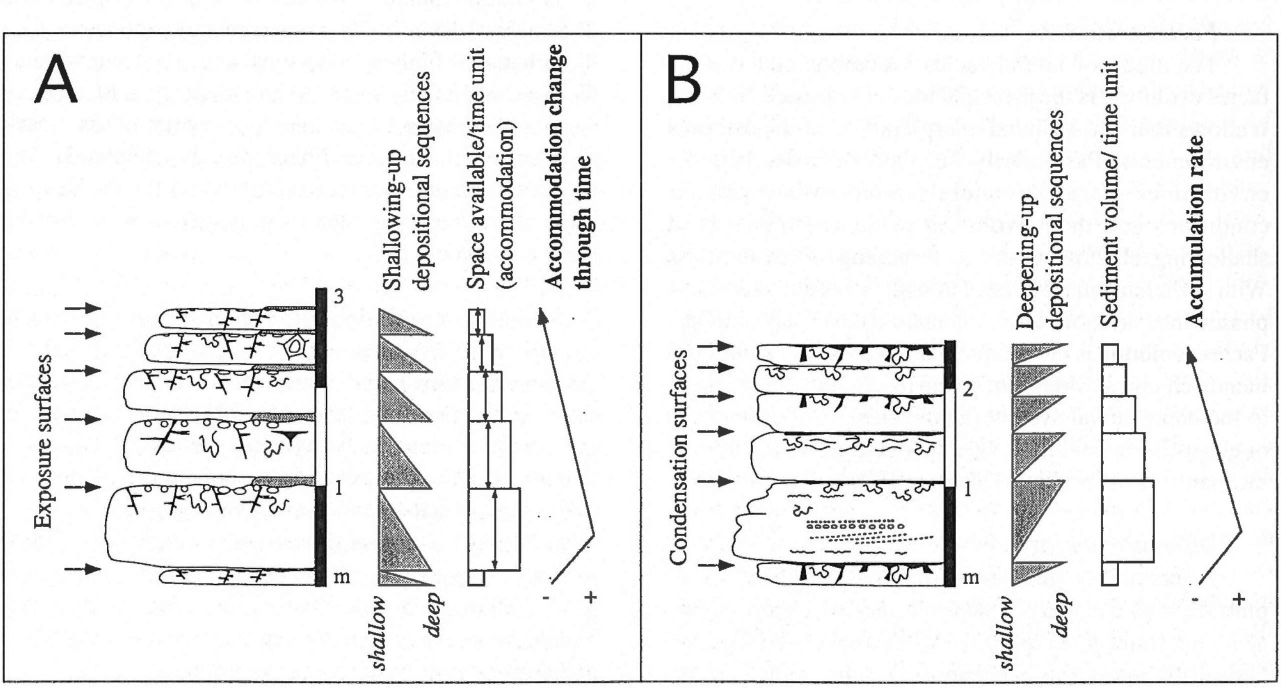


Fig. 5.1. Hypothetical stacking pattern of depositional sequences (here beds). A) Varying thickness of individual beds with internal shallowing-up facies evolution implies changes in accommodation through time. B) A similar stacking pattern indicates variation in sediment accumulation rate when sequences show a deepening-up facies trend (for a legend of sedimentary features refer to Fig. 5.9).

The relative abundance of clay minerals in carbonates is the result of two phenomena (Einsele & Ricken 1991): 1) increase/decrease of terrigeneous input with a constant rate of carbonate production (dilution cycle); 2) decrease/increase in carbonate production with constant terrigeneous influx (productivity cycle). On shallow platforms both phenomena are closely related because abundant terrigeneous material in the water column raises turbidity, which negatively effects phototrophic organisms and, thus, carbonate production (Wilson 1975). Excess nutrient levels related to the increase in terrigeneous input can lead to eutrophication also inhibiting the dominantly oligotrophic carbonate production (Hallock & Schlager 1996, Dupraz & Strasser 1998).

Occurrence of clay in platform carbonates can be related to four main types of environmental change (Fig 5.2, Strasser & Hillgärtner 1998):

1. Rapid fall of relative sea-level increases the erosion potential and, thus, clay input into the shallow-marine realm. Accordingly, the clays are associated with the shallowest and/or most restricted facies. Association with charophytes and fresh-water carbonates can point to local freshwater lakes on the shallow, partly emergent platform.

2. Low sea levels can lead to isolation of pools and shallow lagoons from the open marine influence. The resulting low-energy conditions result in the settling out of clays. In this case, the marly layers exhibit restricted fauna, increased density of bioturbation (lowered accumulation rates) or dysoxic facies due to reduced water circulation.

3. Low-energy conditions can also be created by rapidly rising sea level (initial and maximum flooding), which positions the sea-floor below wave base. These marly layers are characterized by a relatively open-marine fauna, facies is relatively deep, and bioturbation intensity may be high. A rapid rise in sea level can also favor the reworking of previously exposed paleosols and additionally introduce clay minerals.

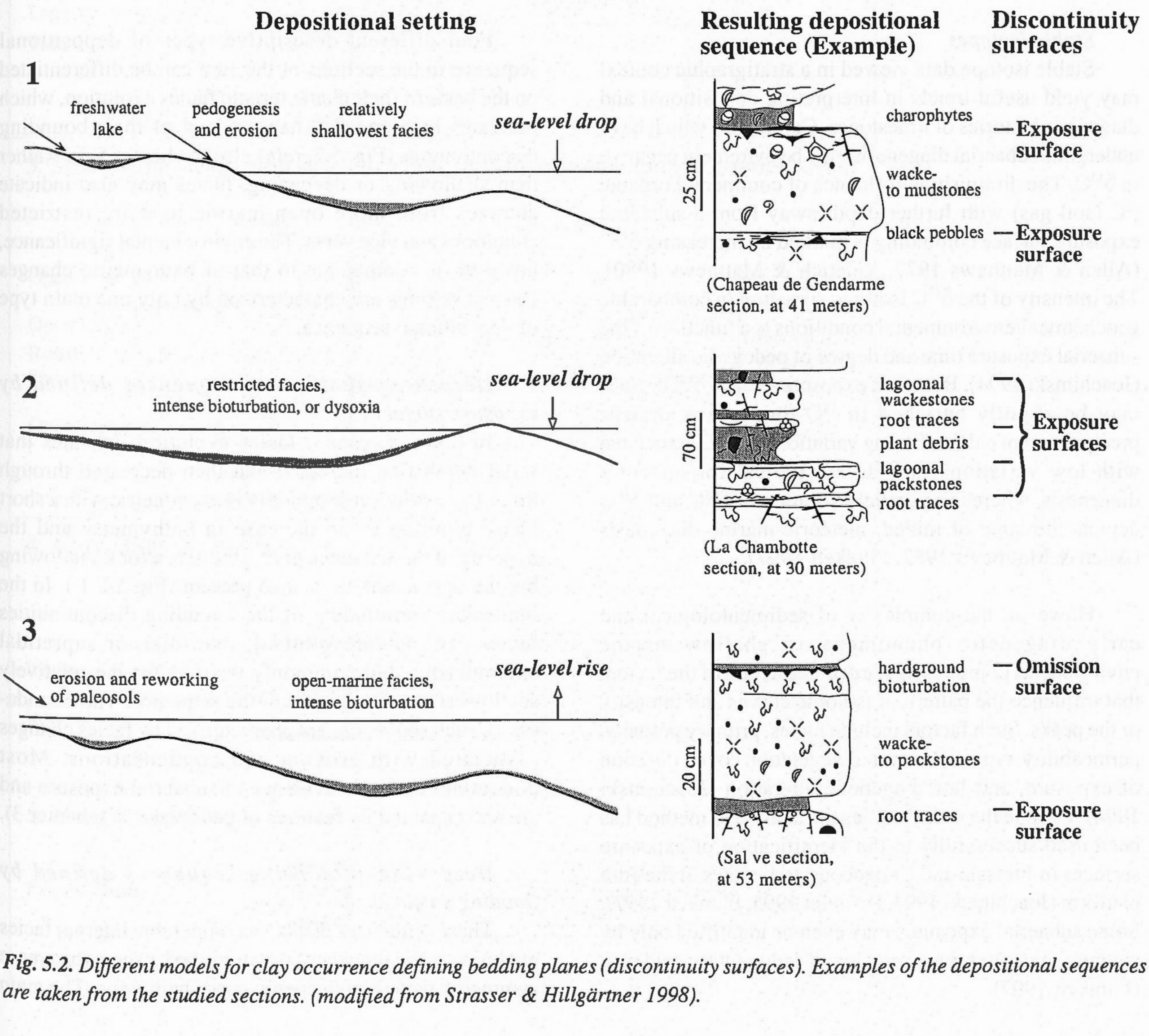


Fig. 5.2. Different models for clay occurrence defining bedding planes (discontinuity surfaces). Examples of the depositional sequences are taken from the studied sections. (modified from Strasser & Hillgärtner 1998).

4. Increased rainfall in the hinterland also raises clay input into the marine system. Such climate changes depend on paleolatitude and on atmospheric circulation patterns (Perlmutter & Matthews 1989, Kindler et al. 1997, Pittet & Strasser 1998a). In certain paleogeographic positions and paleoclimatic regimes, climate may be more humid at high sea-level stands, in other contexts more rain occurs at low sea-levels.

The composition of clay-mineral assemblages can furnish important information on provenance of the sediment, environmental conditions (climate), and may even show a relation to sequence stratigraphic patterns (Deconinck et al. 1985, Deconinck & Strasser 1987, Deconinck 1993). However, clay mineral assemblages are difficult to interpret in terms of sequence analysis. Many intermediate factors, such as current patterns on the platform, or diagenetic alteration of the sediments can influence their composition (Pasquier 1995, Pittet 1996). Refer to chapter 8 where available data on clay-mineral assemblages are used in the interpretation of the climatic evolution.

Stable Isotopes

Stable isotope data viewed in a stratigraphic context may yield useful trends in interpreting depositional and diagenetic histories of limestones. Carbonates which have undergone subaerial diagenesis may be extremely negative in $\delta^{13}C$. The diminishing influence of continental organic ^{12}C (soil gas) with further depth away from a subaerial exposure surface commonly is marked by increasing $\delta^{13}C$ (Allen & Matthews 1977, Videtich & Matthews 1980). The intensity of the $\delta^{13}C$ isotopic signature in comparable geochemical/environmental conditions is a function of the subaerial exposure time and degree of pedogenic alteration (Joachimski 1994). Below an exposure surface $\delta^{18}O$ values may be slightly enriched in ^{18}O due to evaporative precipitation of calcite. Strong variations in $\delta^{13}C$ associated with low variations in $\delta^{18}O$ are typical for meteoric diagenesis, whereas a parallel evolution of $\delta^{13}C$ and $\delta^{18}O$ depicts the zone of mixed, meteoric-marine diagenesis (Allen & Matthews 1982, Plunkett 1997).

However, the complexity of sedimentological and early diagenetic phenomena in shallow-marine environments complicates the differentiation of the factors that influence the pattern of isotopic curves and intensity of the peaks. Such factors include facies, primary porosity, permeability, type and extent of vegetation cover, duration of exposure, and late diagenetic alteration (Joachimski 1994). Despite these difficulties the analytical method has been used successfully in the identification of exposure surfaces in Jurassic and Cretaceous carbonates of the Jura platform (Joachimski 1994, Pasquier 1995, Plunkett 1997). Some subaerial exposures may even be identified only by isotopic signatures due to erosion of sedimentary evidence (Plunkett 1997).

Other criteria

For reasons of completeness some criteria useful to interpret the depositional and diagenetic evolution (and thus depositional sequences) are briefly mentioned. However, they are not applied in this study due to unfavorable lithologies (mainly limestone, rare marly intervals) and the laborious analyses involved.

In palynofacies analysis particulate organic matter is used to determine the ratio of terrestrial vs. marine influence (Steffen & Gorin 1993, Pittet & Gorin 1997).

Geochemical analysis of trace elements, such as Mg, Sr, Mn, and Fe can supply information on diagenetic pathways (Plunkett 1997) and on environmental conditions (e.g., Persoz & Remane 1976, Emmanuel & Renard 1993).

Cement stratigraphy can also be a powerful tool to identify and evaluate diagenetic events related to environmental and climatic changes (Plunkett 1997).

5.2.2. Types of depositional sequences

Four different descriptive types of depositional sequence in the sections of the Jura can be differentiated on the basis of their characteristic facies evolution, which indicates bathymetric changes, and of their bounding discontinuities (Fig. 5.3, refer also to chapter 5.3). Rather than shallowing or deepening, facies may also indicate changes from more open-marine to more restricted conditions and vice versa. The environmental significance, however, is comparable to that of bathymetric changes Basinal settings are characterized by only one main type of depositional sequence.

Deepening-shallowing sequences defined by exposure surfaces

In these sequences facies evolution indicates that water depth first increased and then decreased through time. The evolution commonly is asymmetric with a short phase pointing to an increase in bathymetry and the majority of the sequence giving evidence for a shallowing but the opposite trend is also present (Fig 5.3 1.). In the immediate surrounding of the bounding discontinuities facies can indicate subtidal, intertidal, or supratidal environments, but commonly testifies for the relatively shallowest conditions within the sequence. The boundaries of such sequences are characterized by facies changes associated with erosion and condensations. Most discontinuities show evidence of subaerial exposure and are accompanied by features of pedogenesis (chapter 3).

Deepening-shallowing sequences defined by flooding surfaces

These sequences display a comparable internal facies evolution as the sequences just described, except for a more symmetric pattern of deepening and shallowing. The main

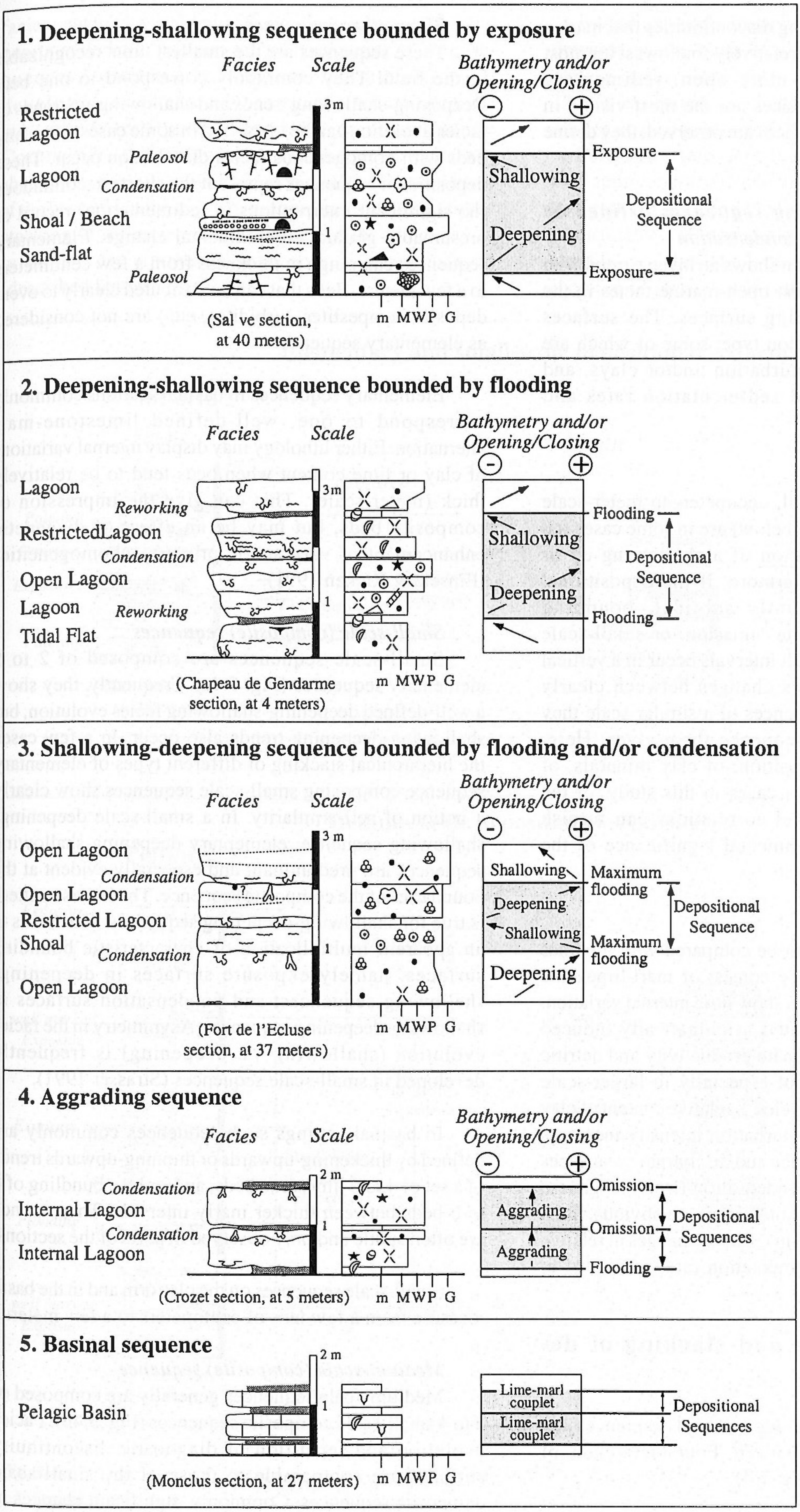


Fig. 5.3. Examples for different types of depositional sequences occurring in the studied sections. For a detailed explanation see text (legend in Fig. 5.9).

difference lies in the bounding discontinuities that mark a rapid facies change from the relatively shallowest (or most restricted) to deeper (or more open) sedimentary environments. As these surfaces are the most visible in the field and no exposure surfaces are observed, they define the depositional sequence.

Shallowing-deepening sequences defined by maximum flooding and/or condensation

Here, the facies evolution shows an inverse trend with well marked, deepest or most open-marine facies in the surrounding of the bounding surfaces. The surfaces commonly are of the omission type, some of which are underlined by intense bioturbation and/or clays, and suggest strongly reduced sedimentation rates and condensation.

Aggrading sequences

Facies changes in small, decimeter- to meter-scale depositional sequences (see below) are in some cases too subtle to allow identification of a shallowing-up or deepening-up trend. Furthermore, if the depositional environment was constantly subtidal, bounding discontinuities only indicate omission, or small-scale erosion. However, when such intervals occur in a vertical context of repetitive facies changes between clearly identified depositional sequences of a similar scale they can be distinguished as sequence themselves. Here, analysis of isotopic composition, of clay minerals, of palynofacies, or, as in most cases in this study, of the general context with lateral correlations can furnish information on the environmental significance of the sequence.

Basinal sequences

These sequences cannot be compared with the ones found on the platform. They consist of marl-limestone alternations where both facies show little internal variation. These alternations are related to climatically induced changes in planctonic carbonate productivity and detritic input of clays. In some, but especially in larger-scale depositional sequences (see below), relative content of clay versus limestone varies, bioturbation intensity increases in certain parts of the sequence and discontinuity surfaces that display evidence for condensation (iron/manganese crusts, ammonite concentrations) are more obvious. Such variability can be interpreted in terms of changes in relative sea level, climate and sedimentation rate (refer also to chapter 6.4).

5.2.3. Hierarchy and stacking of depositional sequences

In all studied sections the depositional sequences are stacked in a hierarchical pattern. Four hierarchies of sequence are recognized (Fig. 5.4):

Elementary sequences

These sequences are the smallest units recognizable in the field. They commonly correspond to one bed. Deepening-shallowing trends and shallowing-up trends of facies evolution can be identified in some cases, but many beds with homogeneous facies distribution occur. These depositional sequences represent the shortest, continuous (no significant interruptions in sedimentation) record of presumably gradual environmental change. Elementary sequences can range in thickness from a few centimeters to a few meters. Beds that can be attributed clearly to event deposits (tempestites, turbidites, etc.) are not considered as elementary sequence.

Elementary sequences in basinal sections commonly correspond to one, well-defined limestone-marl alternation. Either lithology may display internal variations of clay or lime content when beds tend to be relatively thick (meter-scale). This can give the impression of composite beds, but may be an effect of diagenetic enhancement of very subtle primary inhomogeneities (Einsele & Ricken 1991).

Small-scale (composite) sequences

Small-scale sequences are composed of 2 to 6 elementary sequences (Fig. 5.4a). Frequently, they show a well-defined deepening-shallowing facies evolution, but shallowing-deepening trends also occur. In a few cases the hierarchical stacking of different types of elementary sequence composing small-scale sequences show clearly a notion of self-similarity. In a small-scale deepening-shallowing sequence, elementary deepening-shallowing sequences are predominant and especially evident at the boundaries of the composite sequence. The inverse pattern is true for shallowing-deepening sequences. This leads to an apparent multiplication of characteristic bounding surfaces, namely exposure surfaces in deepening-shallowing sequences, and condensation surfaces in shallowing-deepening sequences. Asymmetry in the facies evolution (shallowing vs. deepening) is frequently developed in small-scale sequences (Strasser 1991).

In basinal settings such sequences commonly are defined by thickening-upwards or thinning-upwards trends of a set of 4 to 6 limestone beds, and/or by a bundling of 4 to 6 beds between thicker marly intervals. These trends are often subtle and only observed in parts of the sections.

Small-scale sequences on the platform and in the basin measure from a few tens of centimeters to a few meters.

Medium-scale (composite) sequence

Medium-scale sequences generally are composed of 3 to 4 small-scale composite sequences (Fig. 5.4b). Facies evolution and repetition of diagnostic discontinuity surfaces are comparable to those of the small-scale composite sequences. Commonly, significant changes in facies, such as a sudden influx of siliciclastics (detrital quartz and clays), happen across boundaries of these sequences. Additionally, they are delimited by well-developed discontinuity surfaces (which may be multiplied) testifying to exposure or condensation. Commonly, they display a symmetric deepening-shallowing evolution with thinner small-scale sequences at the base and at the top. Yet, in some cases a characteristic stacking of different sequence types, in the same way as in large-scale composite sequences, can be observed (described below).

Medium-scale sequences in the basin can only be detected on the basis of abrupt changes in the stacking pattern of the elementary and/or small-scale sequences. This is expressed by changes in thickness of limestone and/or marl layers, which mirror general changes in sedimentation rate and changes in the ratio of lime versus clay sedimentation. Sudden variations of this type can be underlined by small slumps or turbidites, and/or contourites, all indicating different environmental conditions at the boundaries of these sequences. A gradual increase of marls towards the center of the sequence can

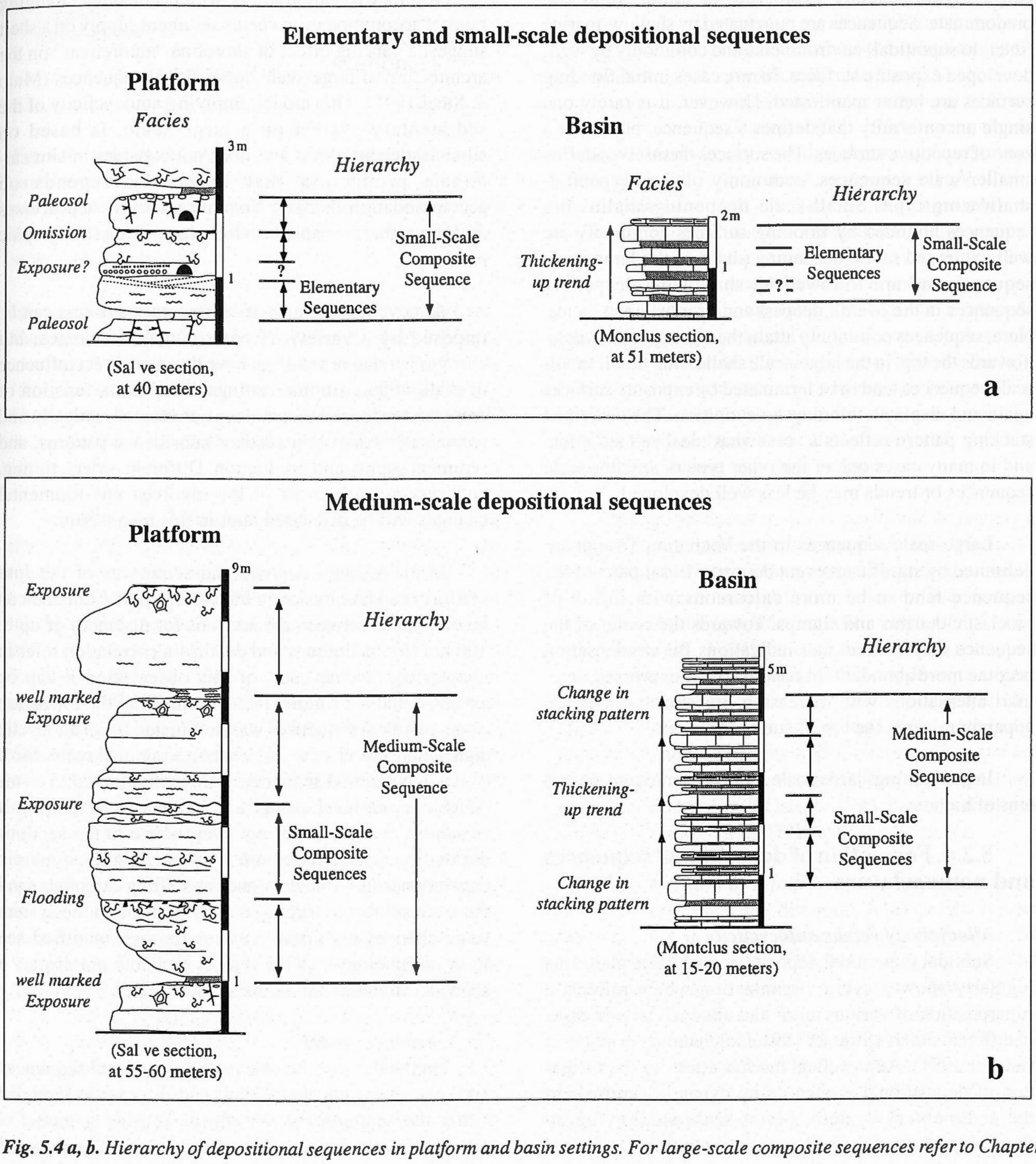


Fig. 5.4 a, b. Hierarchy of depositional sequences in platform and basin settings. For large-scale composite sequences refer to Chapter 5.3 (legend in Fig. 5.9).

be present. Sometimes, however, color changes that delimit sequences are the only evidence and point to changing contents of trace elements such as iron.

Medium-scale composite sequences on the platform and in the basin measure from a few meters to a few tens of meters.

Large-scale (composite) sequences

These sequences are characterized by a deepening-shallowing facies evolution, whereby the individual trends can be symmetrically expressed, or either trend can predominate. Sequences are punctuated by shallow-marine (inter- to supratidal) environments and commonly by well-developed exposure surfaces. In rare cases initial flooding surfaces are better manifested. However, it is rarely one single unconformity that defines a sequence, but rather a zone of repetitive surfaces. The surfaces themselves define smaller-scale sequences, commonly of the deepening-shallowing type. Small-scale deepening-shallowing sequences bounded by flooding surfaces commonly are well expressed in the deepening phase of the large-scale sequence and are followed by shallowing-deepening sequences in the overall deepest and/or most open facies. Here, sequences commonly attain their greatest thickness. Towards the top, in the large-scale shallowing trend, small-scale sequences tend to be terminated by exposure surfaces again and display a thinning-up evolution. The described stacking pattern reflects a somewhat idealized situation, and in many cases one or the other type of smaller-scale sequences or trends may be less well developed.

Large-scale sequences in the Vocontian Trough are delimited by significant event deposits. Basal parts of the sequence tend to be more calcareous with influx of bioclastic detritus and slumps. Towards the center of the sequence clay content and indications for condensation become more abundant. In contrast, well-expressed lime-marl alternations with increasing limestone content are apparent towards the top of such sequences.

In both settings large-scale sequences measure several tens of meters.

5.2.4. Formation of depositional sequences and nomenclature

Allocyclicity versus autocyclicity

Subtidal to peritidal deposits on carbonate platforms regularly show a cyclic organization which reflects a superposition of various auto- and allocyclical processes of different orders (Strasser 1991, Goldhammer et al. 1993, Osleger 1991). Autocyclical models attest for the formation of depositional sequences by dynamics intrinsic to the sedimentary system. Meter-scale shallowing-up sequences can be formed by combinations of vertical aggradation and progradation of tidal flats or lateral migration of tidal channels and shifting carbonate shoals (e.g., Pratt & James 1986, Strasser 1991, Pratt et al. 1992). These models, however, convince only in explaining laterally discontinuous shallowing-up sequences. Subtidal depositional sequences and sequences displaying vadose diagenetic overprinting and evidence for significant subaerial exposure are not accounted for (Osleger 1991, Strasser 1991). For the discussion of problems and limitations of the models the reader is referred to Pratt et al. (1992).

Volumetric calculations which consider changing rates of accommodation versus sediment supply on a shelf suggest a varying effect of shoreline "autoretreat" on the architecture of large-scale depositional sequences (Muto & Steel 1997). This model, implying autocyclicity of the sedimentary system on a large scale, is based on siliciclastic shorelines and does not consider in-situ carbonate production that intimately depends on accommodation changes. No other autocyclical processes are known that are capable to form large-scale stratigraphic patterns.

Allocyclical forcing of sedimentary systems can be imposed by a variety of environmental changes, but changes in relative sea-level have the most direct influence in shallow depositional settings. They are a function of complex feedback mechanisms between climatically and tectonically controlled eustasy, subsidence patterns, and sediment supply and production. Different orders, timing, and varying influences of the involved environmental changes will be discussed later in this manuscript.

In many cases depositional sequences of the Jura platform can be physically traced for up to 1 km and can be correlated between the sections for distances of up to 100 km (for techniques and details on correlation refer to chapter 6). On the basis of this observation it can be assumed that the forcing factor controlling the formation of depositional sequences was dominated by an allocyclic signal, and was effective at least on a regional scale. Most of the depositional sequences can be interpreted in terms of relative sea-level change and sequence stratigraphy. It is natural, however, that not everywhere in the sections depositional sequences are evident. Highly dynamic depositional environments, such as shifting carbonate sand bars, are subject to forcing factors intrinsic to the system (e.g., changes in current patterns through modified sea floor morphology), and will most probably not display a sequence architecture on the smaller scales.

Sequence model

On the basis of the observed depositional sequences and their hierarchical stacking, and in order to interpret better the sedimentary record, the following model is developed. It is assumed that the theoretical position of characteristic deposits and discontinuities on a relative sea-

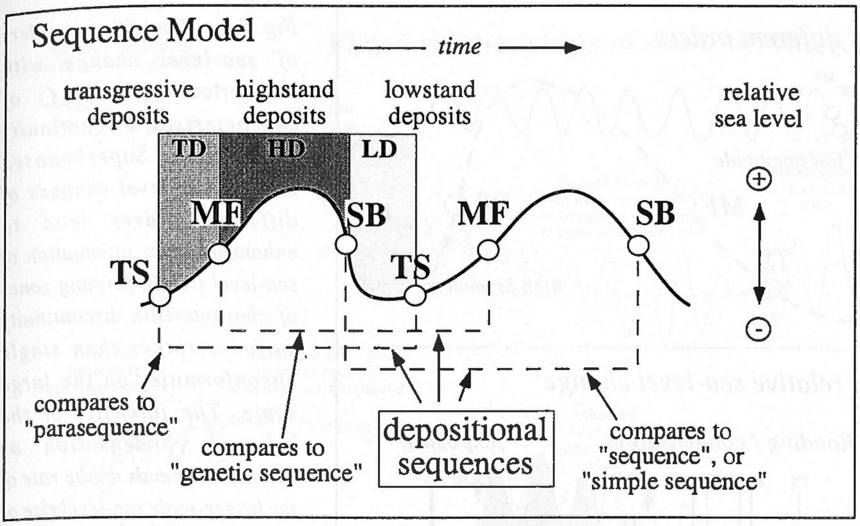


Fig. 5.5. Depositional sequences in terms of relative sea-level change and well marked environmental changes. The theoretical position of characteristic deposits and discontinuities on a relative sea-level curve is independent of the scale of the relative sea-level variation. For a detailed explanation of terms used and references of the established terms refer to text.

level curve is independent of the scale of the relative sea-level variation (Fig. 5.5). Model and terminology are based on sequence- and cyclostratigraphic models (Vail et al. 1977, Strasser 1991, 1994), but terms are slightly modified. This is done in order to avoid terms such as "systems tract" or "parasequence", which imply specific architectures and scales.

On all scales, deposits on shallow carbonate platforms that indicate relative deepening and/or opening of the system are called "transgressive deposits" (**TD**). They correspond to the phase where accommodation increases during relative sea-level rise, and after an initial flooding of a relatively shallower substrate. The discontinuity depicting initial flooding is termed the "transgressive surface" (**TS**). Just after the initial flooding carbonate production is low. Yet, with increasing bathymetry and the recovery of the carbonate factory (Kendall & Schlager 1981) water depth becomes ideal for maximum carbonate production (Tipper 1997). Consequently, the potential to fill up the available space becomes larger. This is when the thickest beds are created. Intensively bioturbated and/or marly intervals, or distinct omission surfaces testifying for the most open-marine conditions correspond to the fastest rise in relative sea level. They are denoted as "maximum flooding zones" (**MFZ**), or as "maximum-flooding surfaces" (**MFS**), respectively. The more general term used is "maximum-flooding" (**MF**). Deposits then thin upwards and indicate shallowing trends that correspond to slowing relative sea-level rise and initial fall. They are depicted as "highstand deposits" (**HD**). Commonly, they are terminated by a discontinuity surface, or an interval indicating subaerial exposure: the "sequence boundary" (**SB**), or the "sequence boundary zone" (**SBZ**).

Theoretically, this surface or interval corresponds to the time of fastest relative sea-level fall and, thus, to destruction of accommodation when the possibility to create subaerial exposure is the highest. The deposits that relate to the phase of a relative sea-level lowstand and theoretically are located between SB and TS, are called "lowstand deposits" (**LD**). In reality, however, most SBZs on the large scale reflect general times of relative sea-level fall and therefore, at least in parts, correspond to deposits traditionally attributed to the forced regressive systems tract (Hunt & Tucker 1992, 1995), or to the late highstand systems tract and/or early lowstand systems tract (Vail et al. 1977). Because of (commonly) very shallow depositional environments on the studied platform and lacking accommodation in elementary to medium-scale sequences, deposits representing a low relative sea level are very thin or not developed at all. Additionally, these deposits have the lowest preservation potential and tend to be eroded and reworked during the subsequent transgression. Thus, transgressive surfaces coincide in many cases with the sequence-bounding exposure surfaces.

This model accounts for depositional sequences on shallow marine platforms, but its principles are generally applicable. However, the assumption of a simple sinusoidal sea-level curve is highly unrealistic and does not correspond to the dynamics and the complexity of natural systems (Smith 1994). A composite curve taking into account the superposition of relative sea-level changes of different orders is a much more realistic approach. This implies a high-resolution stratigraphic expression of the relative sea-level changes and a characteristic stacking pattern of corresponding discontinuities and depositional sequences (Montañez & Osleger 1993, Pasquier & Strasser 1997, Hillgärtner 1998, Hillgärtner et al. 1998).

Superposition of relative-sea-level changes

The occurrence and distribution of specific types of discontinuity in cyclic successions, mainly reflecting relative sea-level changes (TS, MFS, SB), is a function of the amplitudes and frequencies of superimposed relative sea-level changes and their relative position on this composite sea-level curve (Fig. 5.6). Therefore, types of small-scale depositional sequence, defined by their bounding discontinuities and internal facies evolution, and their distribution in the succession will vary accordingly (Arnott 1995).

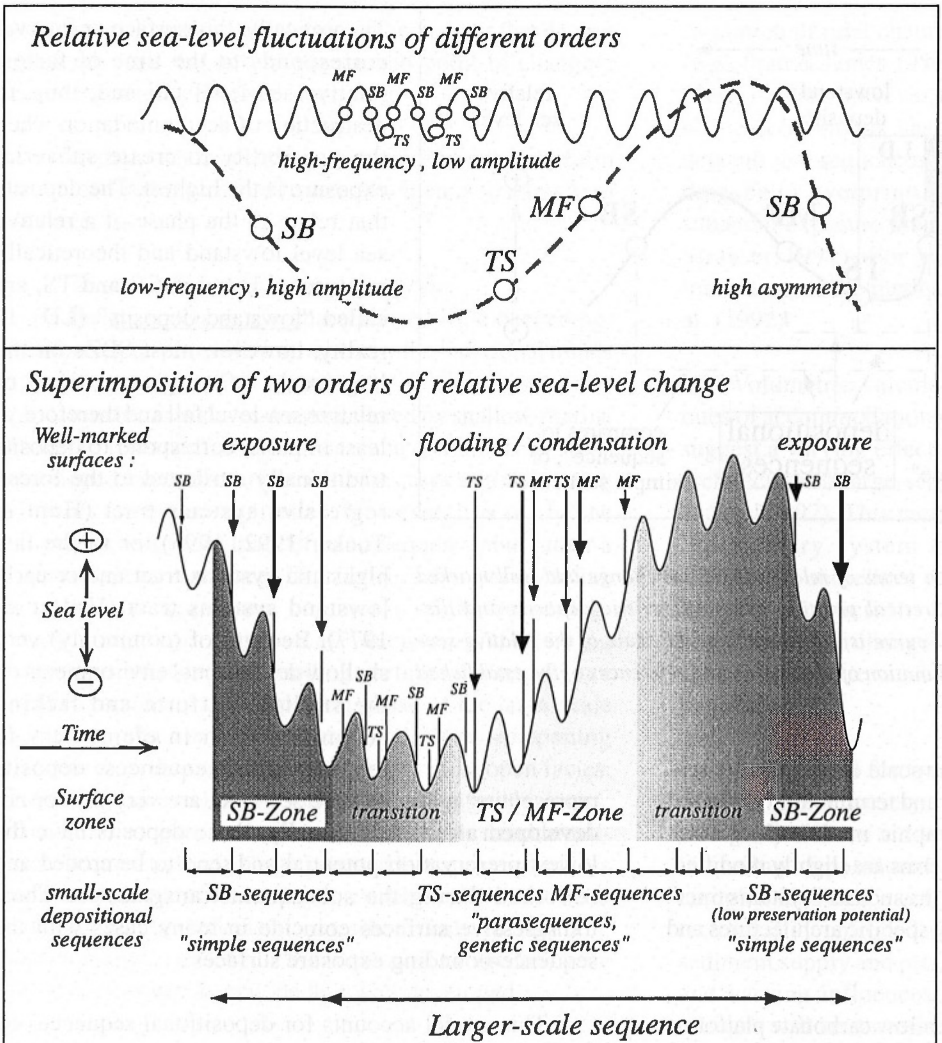


Fig. 5.6 a. Two different orders of sea-level change with theoretical positions of characteristic discontinuity surfaces. b. Superimposed relative sea-level changes of different orders lead to enhancement or attenuation of sea-level trends, forming zones of characteristic discontinuity surfaces rather than single unconformities on the large scale. The intensity of the inferred condensation or exposure depends on the rate of the larger-scale sea-level rise or drop, respectively (thicker arrows indicate higher intensity). Different types of discontinuity surfaces define characteristic smaller-scale depositional sequences, which compare to "parasequences" (Van Wagoner et al. 1988) and "simple sequences" (Vail et al. 1991). Small-scale depositional sequences in intervals of sea-level lowstand or highstand on the large scale theoretically can display all types of discontinuity surfaces (TS, MF, SB) and have a transitional character.

Superimposition of the same trends in sea-level variations leads to enhancement of characteristic signatures. When a small-scale relative sea-level fluctuation is superimposed on a larger-scale sea-level rise the flooding will be enhanced, whereas the small-scale sea-level fall will be attenuated. Consequently, MF surfaces will be marked better in the stratigraphic succession than sequence boundaries. Sequence boundaries, in contrast, will be enhanced when falling trends of different orders are combined, and TS surfaces will be underlined when different orders of initial flooding are synchronous.

Superimposition of different orders of relative sea-level change thus leads to multiplication of characteristic discontinuity surfaces, rather than the formation of one well-marked sequence-bounding unconformity of the larger-scale sequence (Montañez & Osleger 1993, Pasquier & Strasser 1997, refer also to chapter 6). These clusters of surfaces are referred to as "maximum-flooding zone" (**MFZ**), "sequence-boundary zone" (**SBZ**) and the rarely observed "transgressive-surface zone" (**TSZ**) (Fig. 5.6). High or low amplitudes, asymmetry, and varying phase relationships of the sea-level curves can enhance or attenuate the formation of such zones. If high-amplitude fluctuations are superimposed on a rapidly falling general sea-level trend, accommodation allowing sediment formation and deposition will be available for a short time during high-frequency sea-level rises, but the preservation potential of such sequences will be very low. The sedimentary record thus will be incomplete ("missing beats" of Goldhammer et al. 1990). During moderate- to low-amplitude fluctuations, however, the intensity of condensation and exposure in a surface zone should theoretically increase towards the strongest rate of sea-level rise and fall on the long-term trend, respectively (Figs. 5.6, 5.7).

The multiplication of characteristic discontinuities has the effect that depositional sequences punctuated by initial flooding surfaces, condensation, and exposure surfaces predominately occur during transgression, maximum-flooding, and maximum regression on the larger scale, respectively. Therefore, and on the basis of their

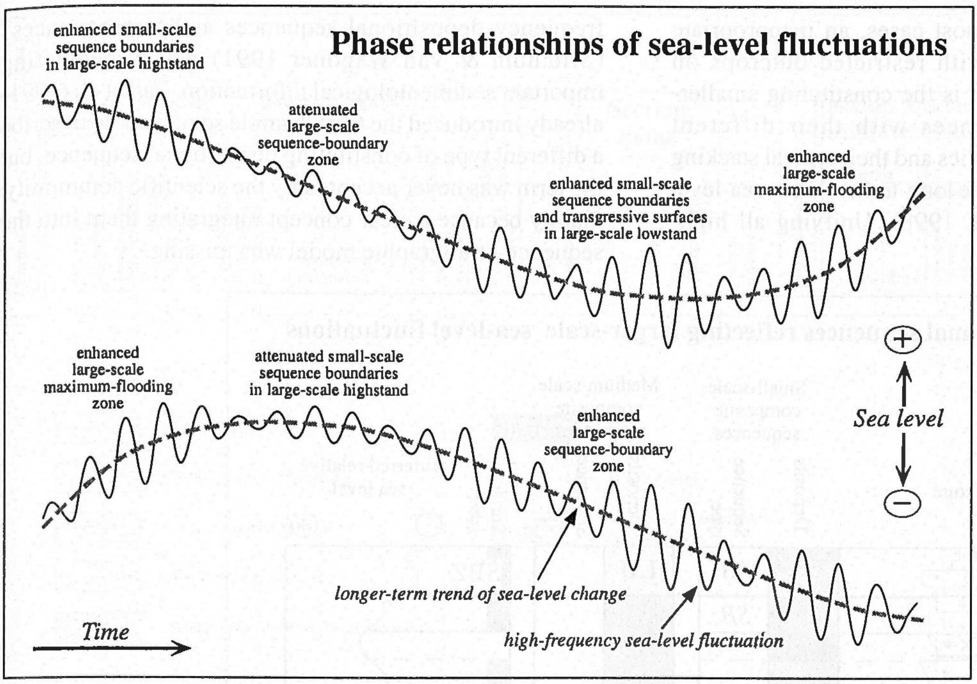


Fig. 5.7. Variations in the amplitude of small-scale sea-level fluctuations can lead to enhancement or attenuation of characteristic surface zones depending on the phase-relationship with larger-scale trends of sea-level change (after Strasser et al. 1998).

correlatability, the different sequence-types described above on account of bathymetric changes are classified in terms of relative sea-level changes.

TS-sequences correspond to the deepening-shallowing (transgressive-regressive, in terms of sea-level fluctuation) sequences defined by initial flooding surfaces. Here, the deepest/most open facies corresponding to the MF occurs in the center of the sequence. The subsequent shallowing not necessarily culminates in subaerial exposure. These depositional sequences can be compared to "parasequences" (Van Wagoner et al. 1990).

MF-sequences display a regressive-transgressive facies trend punctuated by condensation surfaces. The shallowest facies in the center of the sequences will, most probably, not be marked by subaerial exposure either, because of an attenuated sea-level fall. MF-sequences commonly correspond to "genetic sequences" (Homewood et al. 1992).

SB-sequences are bounded by exposure surfaces and show a transgressive-regressive evolution. They correspond to "simple sequences" (Vail et al. 1991) or the classical "sequences" (Vail et al. 1977), depending on their scale.

Fig. 5.8 shows the different sequence types and their typical stacking interpreted in terms of superimposed sea-level fluctuations.

Any classification of such type is, of course, a rather crude simplification of complex natural systems. The sequence types have to be seen as end-members of a continuum of sequences that may own a variety of characteristics of all sequence-types. For example, in transitional zones between superimposed relative sea-level rise and superimposed relative sea-level fall, corresponding to a sea-level highstand or lowstand on the long term, all types of discontinuities can be well expressed. It may therefore be that a MF-sequence overlaps with a subsequent SB-sequence. The attribution to either sequence type has to be done according to the most appropriate interpretation of the long-term sea-level trend, on the basis of lateral correlations and the best marked bounding surfaces (Fig. 5.8). In the same way, aggrading sequences do not show typical characteristics for any sequence type. However, seen in the context of relative sea-level changes and surrounding, unequivocally defined sequence types, all sequences can be interpreted as being on the more transgressional or more regressional long-term trend. Consequently, and following the model, most sequences can be attributed to TS-, MF-, or SB-sequences. In places where all discontinuities are equally well expressed, SBs are chosen as delimiting boundaries.

The main difference of this model to the sequence-stratigraphic model (Vail et al. 1977, Vail et al. 1991) is the self-similarity of depositional sequences on all scales. The nomenclature introduced for high-frequency sequence analysis (Mitchum & Van Wagoner 1991), already taking into account the superimposition of different orders of sea-level change, seems too clumsy to be useful. Systems tracts and sequence sets, for example, are defined by different progradation - aggradation ratios and reflection terminations (Mitchum & Van Wagoner 1991, Emery & Myers 1996) and are a measure of larger-scale sea-level trends. Their direct observation is extremely difficult in most field studies. The concept of systems tracts was born on the basis of seismic signatures. Sedimentologically speaking they are a virtual and, in most cases, an inappropriate feature for field studies with restricted outcrops on carbonate platforms. Here it is the constituting smaller-scale depositional sequences with their different sedimentological characteristics and their typical stacking pattern, which visibly translate long-term relative sea-level evolution (Hillgärtner et al. 1998). Unifying all high-frequency depositional sequences as "parasequences" (Mitchum & Van Wagoner 1991) means neglecting important sedimentological information. Vail et al. (1991) already introduced the term "simple sequence" to describe a different type of constituting depositional sequence, but the term was never accepted by the scientific community, mainly because a clear concept integrating them into the sequence-stratigraphic model was missing.

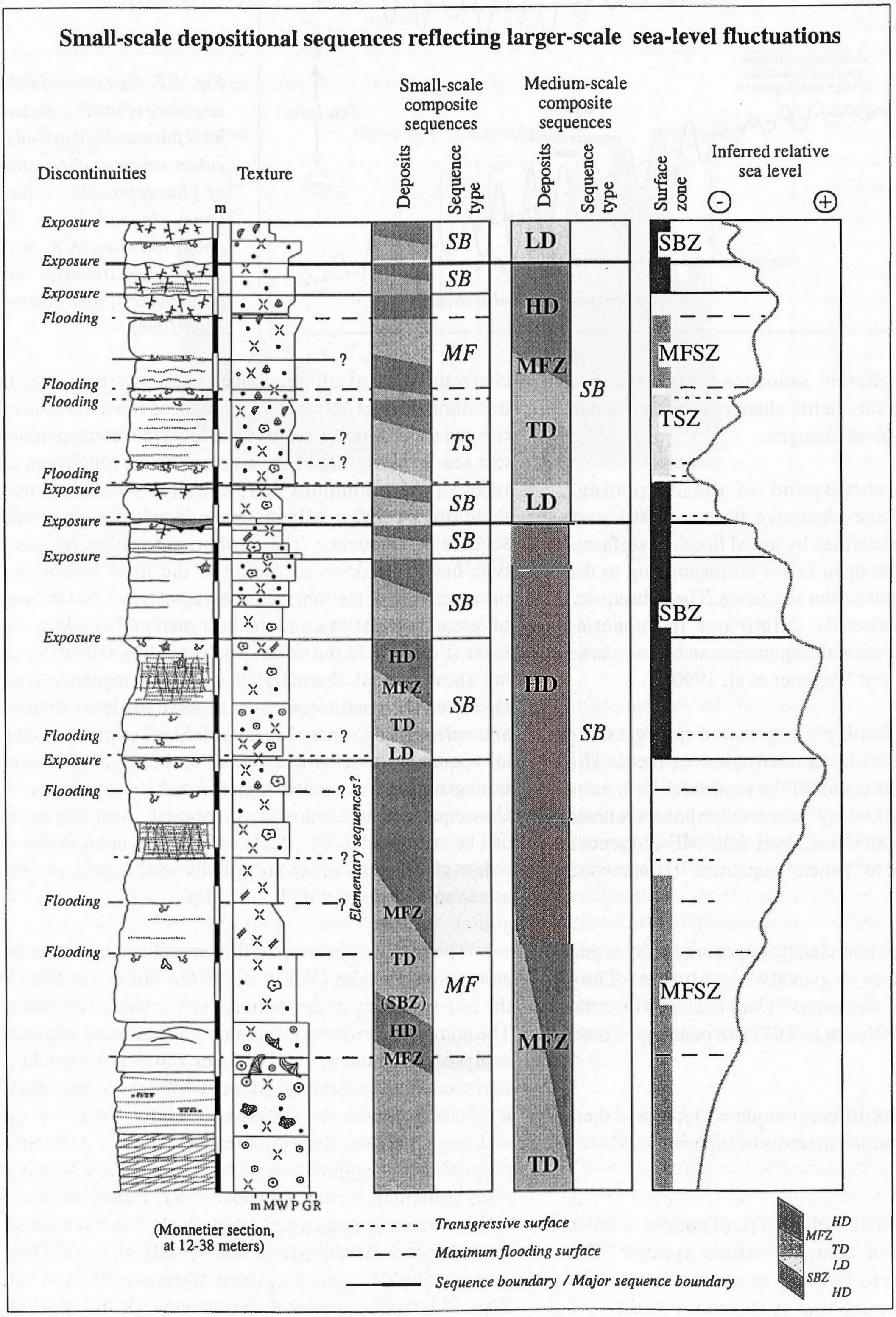


Fig. 5.8. Different orders and types of depositional sequences and their typical stacking pattern. The succession of smaller-scale sequences in an "ideal" SB-sequence on the medium scale is ...SB-TS-MF-SB..., and therefore, according to the sequence model, reflects larger-scale sea-level variations (upper part of the section). Note the different succession of deposits in small-scale MF-sequences (MFZ-HD-(inferred SB)-TD-MFZ) compared to small-scale SB-sequences (LD-TD-MFZ-HD) (outlined in lower part of the section). The representation of this interval in the field is given in Plate 8.5.

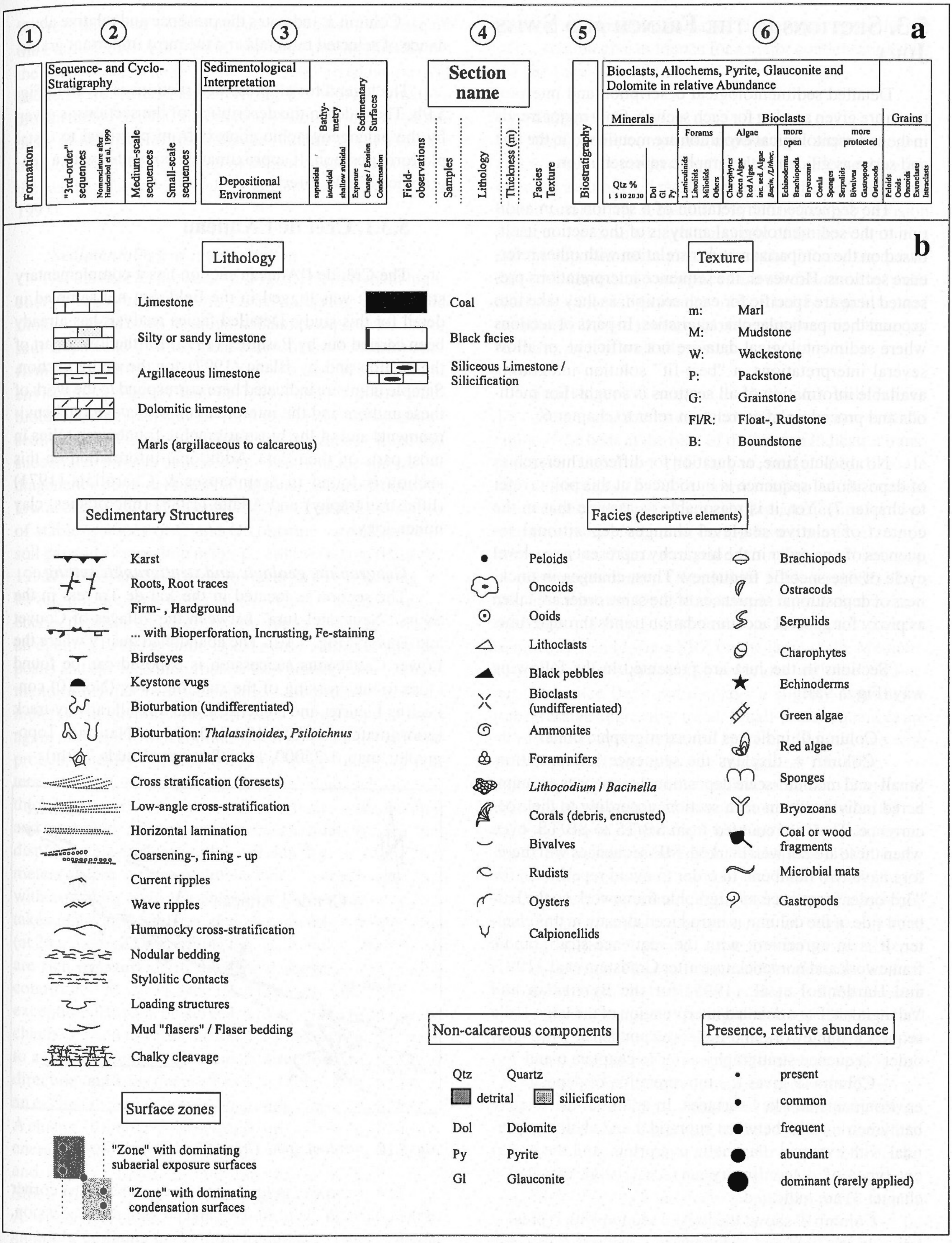


Fig. 5.9. Legend for all studied sections.

5.3. Sections in the French and Swiss Jura

Detailed sedimentological description and interpretation are given as a log for each section. Only major trends in the sedimentological evolution are mentioned in the text and serve as guide for the graphic representation.

The sequence interpretation of a section is, in addition to the sedimentological analysis of the section itself, based on the comparison and correlation with other reference sections. However, the sequence interpretations presented here are specific for each section, as they take into account their particular characteristics. In parts of sections where sedimentological data are not sufficient, or allow several interpretations, a "best-fit" solution integrating available information of all sections is sought. For methods and procedure of correlation refer to chapter 6.

No absolute time, or duration for different hierarchies of depositional sequence is introduced at this point (refer to chapter 7). Yet, it is reasonable to assume that in the context of relative sea-level changes depositional sequences of one order in the hierarchy represent a sea-level cycle of one specific frequency. Thus, changes in thickness of depositional sequences of the same order are taken as proxy for general accommodation trends through time.

Sections in the Jura are presented in the following way (Fig. 5.9a):

- Column ① indicates lithostratigraphic units.
- Column ≠ displays the sequence interpretation. Small- and medium-scale depositional sequences are numbered individually in each section, according to their occurrence. They are counted from SB/TS to SB/TS, even when these are not well marked. MF-sequences will therefore have two numbers. In order to avoid repetitions, the "3rd-order" sequence-stratigraphic framework on the left-hand side of the column is introduced already in this chapter. It is in agreement with the sequence-stratigraphic framework and nomenclature after Gradstein et al. (1995) and Hardenbol et al. (1999) for the Berriasian and Valanginian. For a detailed interpretation of the large-scale sequence framework and the correspondence with "3rd-order" sequence stratigraphy, refer to chapters 6 and 7.
- Column ③ gives the interpretation of depositional environments and key surfaces. In addition, the relative bathymetric trend between supratidal and shallow intertidal with eventual diagenetic overprints, and the different types of discontinuity surface (Hillgärtner 1988, chapter 3) are indicated.
- Column ④ shows the logged section with representation in the field and sedimentary structures (left side), and general facies and texture (right side).
- Column ∞ outlines important biostratigraphic markers.
- Column ± indicates the presence and relative abundance of selected minerals and the most important grains.

The legend for all presented sections is given in Fig. 5.9b. The order in the description of the sections is given by the paleogeographic context from proximal to distal platform position. It approximately coincides with a geographic NE-SW direction (Fig 1.1).

5.3.1. Crêt de l'Anneau

The Crêt de l'Anneau section has a complementary status, as it was logged in the field, but not sampled in detail for this study. Detailed facies analysis has already been carried out by Pasquier (1995) for the lower part of the section and by Blanc (1997) for the entire section. Sample numbers indicated here correspond to the work of these authors and the interpretation of depositional environments and of the biostratigraphic framework relies in most parts on their data. Additional information on this section is found in Steinhauser & Charollais (1971) (lithostratigraphy) and Adatte (1988) (microfacies, clay mineralogy).

Geographic, geologic and stratigraphic setting

The section is located in the Val de Travers in the Swiss "Neuchâtel-Jura" between the villages of Couvet and Travers (Fig. 5.10). The abandoned quarry where the Lower Cretaceous succession is exposed can be found close to the crossing of the state highway (No. 10) connecting Fleurier and Neuchâtel, and a small railway-track (coordinates 543.050/199.700, Swiss National Topographic map, 1/25000, 1163 Travers, altitude 732m).

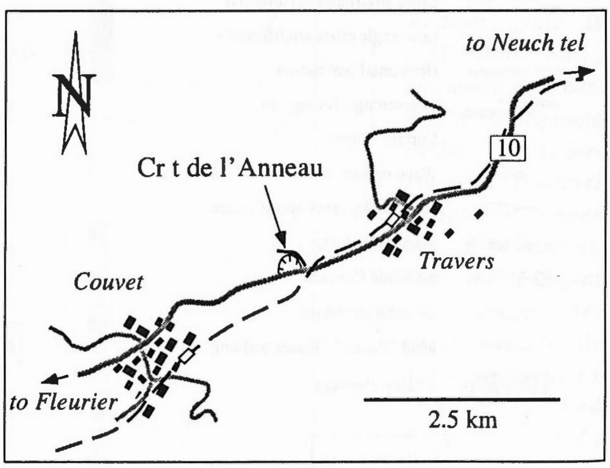


Fig. 5.10. Location of the Crêt de l'Anneau section.

The section is logged from the southeastern corner of the quarry to the northwest and crosses the succession, which forms the southern limb of an anticline plunging 40-50° to the South. Benthic foraminifers confirm a Lower Cretaceous age of the succession (Fig. 5.11, Blanc 1997). The dominantly calcareous rocks are attributed to the

Pierre-Châtel and Vions Formations, whereas the base of the Pierre-Châtel Formation, normally resting directly on the Purbeckian, is not exposed. Commonly, reddish, cross-bedded carbonate sands overlying the Vions Formation are attributed to the Calcaire Roux Formation (s.l.) on the basis of lithological resemblance (Adatte 1988). Ammonites found in the marly top of the succession (5 m above the last bioclastic bar) indicate a Hauterivian age (Blanc 1997).

Sedimentological interpretation

At the base of the section (Fig. 5.11), dominant high-energy lagoonal deposits clearly display a repeated shoaling trend with terminal subaerial exposures. Dinosaur tracks and a subsequent paleosol (at 7 meters) confirm the shallow-marine, near-coastal depositional environment. Energy then decreases and the marly, well-bioturbated deposits (at 10-13 meters) point to condensed sedimentation in open lagoonal environments. Repeated subaerial exposures indicated by a karst surface and vadose diagenetic overprints attest for lagoonal sedimentation with low accommodation potential (at 14-17 meters). Analysis of stable isotopes of C and O confirms the influence of soil gas and evaporation in this part of the section (Pasquier 1995). Exposures continue to occur in the following interval, but are less well marked. Thick-bedded, higher-energy deposits and open-marine fauna and flora indicate an opening of the system. Local marly and/or muddy deposits and elevated bioturbation then point to reduced sedimentation rates in lower-energy conditions (at 18-22 meters). An interval with well-marked exposure surfaces again indicates low accommodation potential in lagoonal, probably tidally influenced environments (at 23-24 meters). Another short opening of the system indicated by thick bioclastic carbonate sands terminates with shoaling and intensively bioturbated lagoonal to tidal deposits, that display intense alteration by soil development (at 25-26 meters). Here, a peak of siliciclastic influence is marked with a quartz content as high as 30 %. A sharp facies change takes place at a partly eroded and reworked hardground (at 26 meters). The overlying bioclastic-oolitic grainstones are rich in echinoderm debris and suggest open-marine conditions (at 27-31 meters). However, thin beds with exceptionally well preserved tabular cross-bedding and channelization towards the top, all with marly joints, point to a tidal influence. Sedimentary structures indicate flow directions towards the northeast and southeast suggesting an exclusive preservation of incoming tides (Blanc 1997). A change to more condensed deposition with slightly lower energy conditions is witnessed by elevated bioturbation and higher carbonate mud content (at 32-34 meters). Siliciclastic influence ceases completely in this interval. A well-marked, iron-stained erosion surface marks the change to a thick, echinoderm-rich bioclastic bar, again pointing to high-energy, shallow-lagoonal environments with siliciclastic influence. The thin-bedded, deeper-marine marls (ammonite fauna) of Hauterivian age at the top of the section give evidence for a major condensation during the Late Valanginian.

Sequence evolution

Small-scale composite depositional sequences generally are well marked in this section. Elementary sequences are not evident. For example, the first ooid-rich bed in small-scale sequence 1 is defined only by erosion surfaces. Here, an allocyclical origin of the discontinuities and the depositional sequence cannot be inferred unequivocally. Medium-scale sequences are all delimited by well-marked subaerial exposure and/or marly deposits, and commonly show a transgressive-regressive trend that may be more, or less well expressed.

Small-scale SB-sequences (1-4), with well-defined TSs, the general facies evolution, and progressive thickening of the beds at the base of the section indicate a transgressive trend on the larger scale. Small-scale MF-sequences (5-7) in combination with intense condensation and initial arrival of quartz testify for lowered carbonate production during fastest sea-level rise on the long term. This corresponds to a "give-up" situation in platform evolution (Kendall & Schlager 1981). Two thin SB-sequences (8-9) that indicate rapid loss of accommodation and filling up of available space point to a very reduced HD and form a SBZ on the large scale. Medium-scale sequence 4 practically coincides with the "3rd-order" sequence Be5, and displays a slightly asymmetric transgressive-regressive trend. Small-scale sequences are all of the SB-type, sequence 11 being the only one with well-expressed and possibly multiplied TSs (TSe = elementary TS). Small-scale sequences 12 and 13 are very thin and again point to low accommodation in a SBZ on the longer-term relative sea-level trend. The short opening of the system indicated in SB-sequence 14 is terminated with a condensation and multiple, amalgamated small-scale SB-sequences. Each bed is interpreted as one reduced small-scale sequence, since all show important discontinuities at their top, and testify for subaerial exposure and erosion. The occurrence of *Pfenderina neocomensis* only 4 meters above the last occurrence of *Pavlovecina allobrogensis* suggest a strongly reduced thickness and probably partial non-deposition and/or erosion of the Upper Berriasian in this interval. The erosion surface just below the bioclastic bars is interpreted as major TS on the large scale. The highly dynamical depositional systems of tidal sand waves and bars do not allow an identification of small-scale sequences at the top of the section. General facies evolution suggests a large-scale transgressive trend with a MF indicated by reduced sedimentation rates at 32-33 meters. *Montsalevia salevensis* in this interval suggests a correlation with sequence Va1 (chapter 6). By means of lateral correlation (chapter 6) the erosion/exposure surface below the last bioclastic bar is interpreted as amalgamation of two sequence boundaries

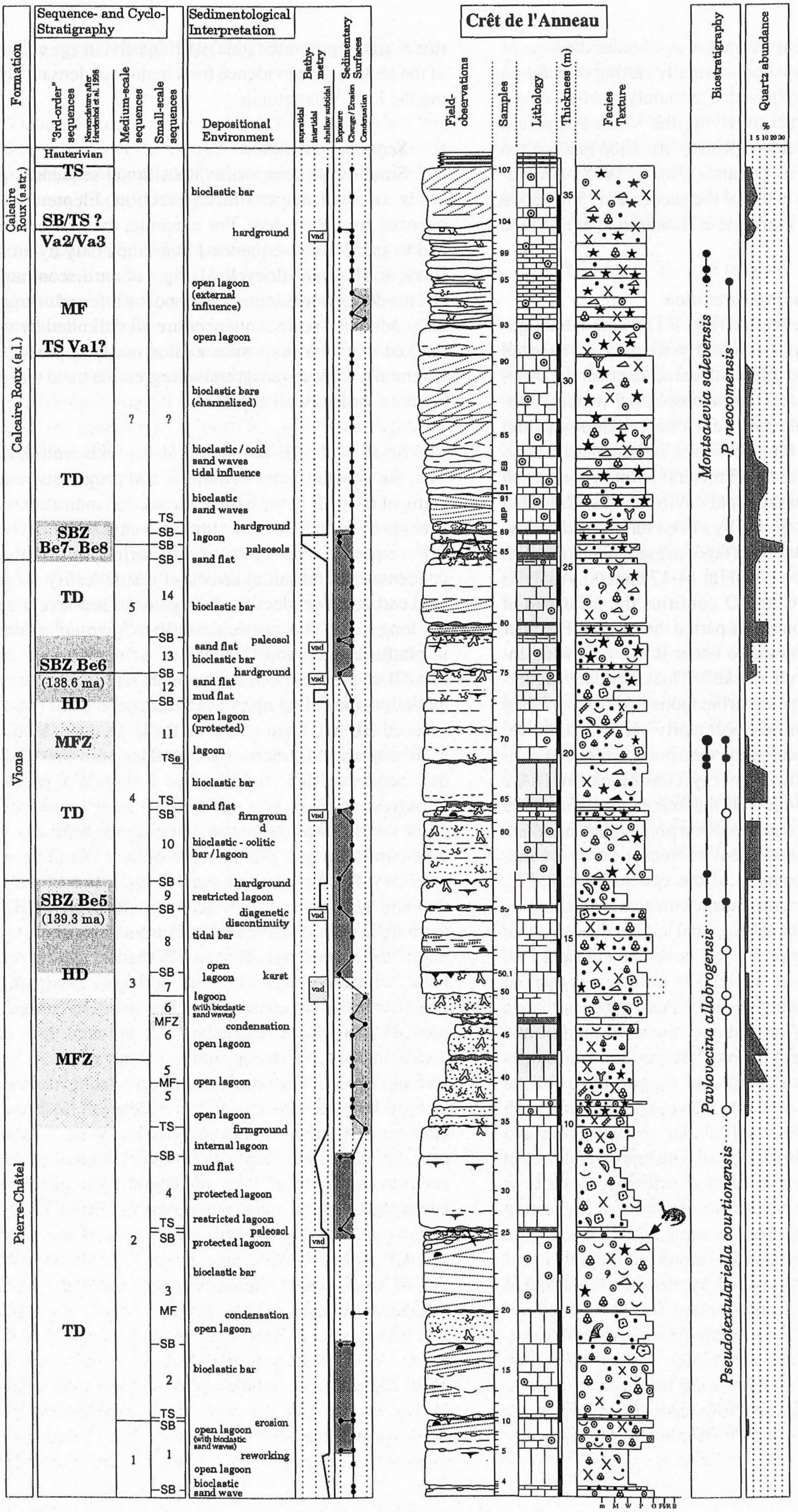


Fig. 5.11. Crêt de l'Anneau section (lower part of the section modified from Pasquier 1995).

on a longer-term relative sea-level variation. This, and the facies with an elevated quartz content suggest a correlation of the terminal bioclastic bar with the Calcaire Roux (s.str.). The extremely condensed Upper Valanginian and deeper-water Hauterivian deposits do not allow a sequence analysis but suggest an important transgressive trend on the large scale.

5.3.2. Chapeau de Gendarme

The Chapeau de Gendarme section is a reference section for the Middle Berriasian through Lower Valanginian strata in a proximal platform position. However, the Pierre-Châtel Formation was not sampled because detailed facies analysis has already been carried out by Waehry (1988).

Geographic, geologic and stratigraphic setting

The section crops out along the departmental road (D436) between St. Claude and the village of Septmoncel in the French Jura (Fig. 5.12). The base of the section is located on the left side of the road just behind the sharp U-turn that faces the "Chapeau de Gendarme" (coordinates 874.130/156.600, IGN map, 1: 25000, St. Claude). "Chapeau de Gendarme" is a landmark and describes an isoclinal hinge of a small fold in the limb of a large NE-SW oriented anticline. The section cuts through its eastern limb, where the continuously exposed succession plunges with 130/25. Charophyte-ostracod assemblages and benthic foraminifers indicate a Middle Berriasian to Early Valanginian age (Fig. 5.13a, 5.13b). The top of the marly Purbeckian is followed by carbonates of the Pierre-Châtel, Vions, and Lower Chambotte Formations. The top of the competent Chambotte limestones, which form a morphological ridge, abruptly give way to a small valley. Although the contact is not exposed, morphology suggests an overlying formation with incompetent lithologic properties, such as the Early Valanginian Guiers Member and Arzier Marls.

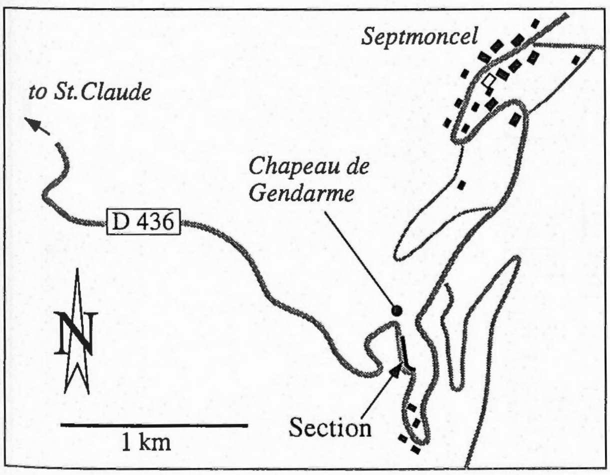


Fig. 5.12. Location of the Chapeau de Gendarme section.

Sedimentological interpretation

The base of the section is defined by greenish marls of the Purbeckian. Charophyte-ostracod assemblages (M2-M3) point to an Early to Middle Berriasian age (*Grandis / Subalpina* ammonite zones, Detraz & Mojon 1989). The following 8 meters of the Pierre-Châtel Formation are dominated by shallowest-marine to tidal environments with recurrent brackish influence and repeated subaerial exposure. This is indicated by presence of charophytes (M4 assemblage, *Privasensis* ammonite zone) and by desiccation features and/or vadose diagenetic overprints (Waehry 1988). The system then opens up to normal-marine conditions with a more or less protected lagoonal character (at 9-13 meters). Protected to slightly restricted lagoonal sediments with oncoid formation, omission surfaces, and desiccation features point to reduced sedimentation rates and short-lived exposures, respectively (at 14-17 meters). Definitely more open-marine, but still protected lagoonal conditions then prevail in the following, massive limestones. They terminate in microbial mud-flat (thrombolitic) facies and attest for a shallowing-up tendency (at 18-25 meters). A subsequent, renewed lagoonal sedimentation is abruptly capped by a microkarst testifying for a drastic loss of accommodation, but probably limited time of subaerial exposure (at 26 meters).

Siliciclastic influx of quartz and clays directly follows the karst and marks the base of the Vions Formation. It indicates an important environmental change of climatic and/or tectonic origin. Initially, very condensed lagoonal sedimentation then passes into shallow lagoonal environments with more open marine character (at 27-30 meters). Repeated shoaling, omission and/or exposure, and high contents in quartz, clay, and pyrite (associated with blackening by organic matter) are interpreted as signs of varying tidal influence (at 31-37 meters). A clearly near-coastal to continental lacustrine environment at 38 meters is evidenced by chalky carbonates rich in charophytes and ostracods. Shallow lagoons to tidal sand/mud flats dominate the succession at 38-52 meters. Tidal influence is evidenced by flaser bedding, channelization, high contents of intraclasts, mud pebbles, black pebbles (reworking), organic matter (blackening, pyrite), plant debris, the trace fossil *Psiloichnus*, and occurrence of large specimens of *Pholadomya*. An important exposure surface is marked by microkarst and paleosol formation at 41 meters. The increasing content of echinoderm debris towards the top of this interval points to a successive opening of the system. However, it probably also reflects the particular taphonomic behavior of echinoderm debris, which is characterized by a low density and is easily transported over large distances by tidal currents (Neumeier 1998). This probably leads to the characteristic mixture of continentally-derived and open-marine constituents. Variable mud content and omission surfaces also confirm varying energy conditions and highly discontinuous sedimentation, typical for tidal environments. Open lagoonal sedimenta-

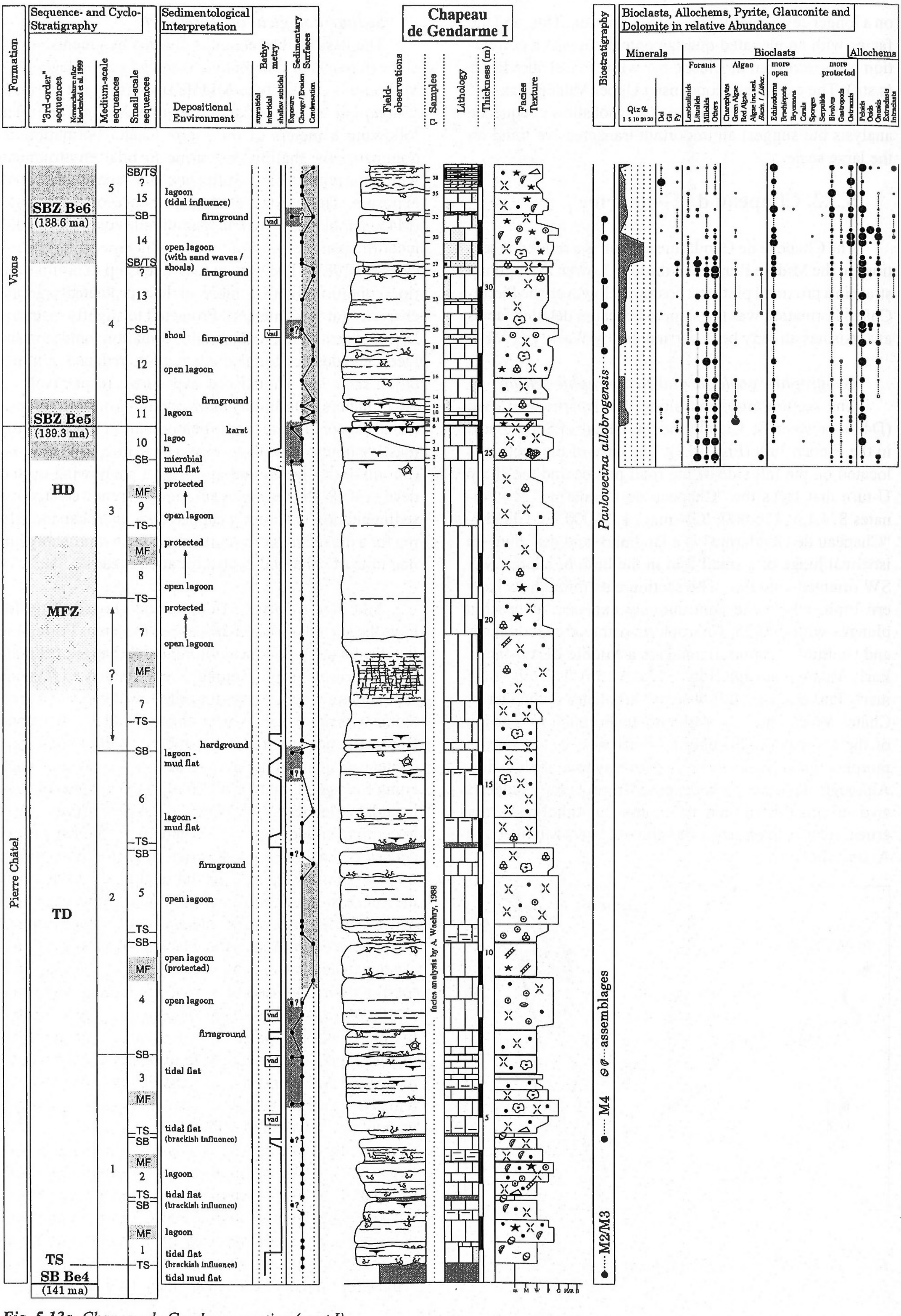


Fig. 5.13a. Chapeau de Gendarme section (part I).

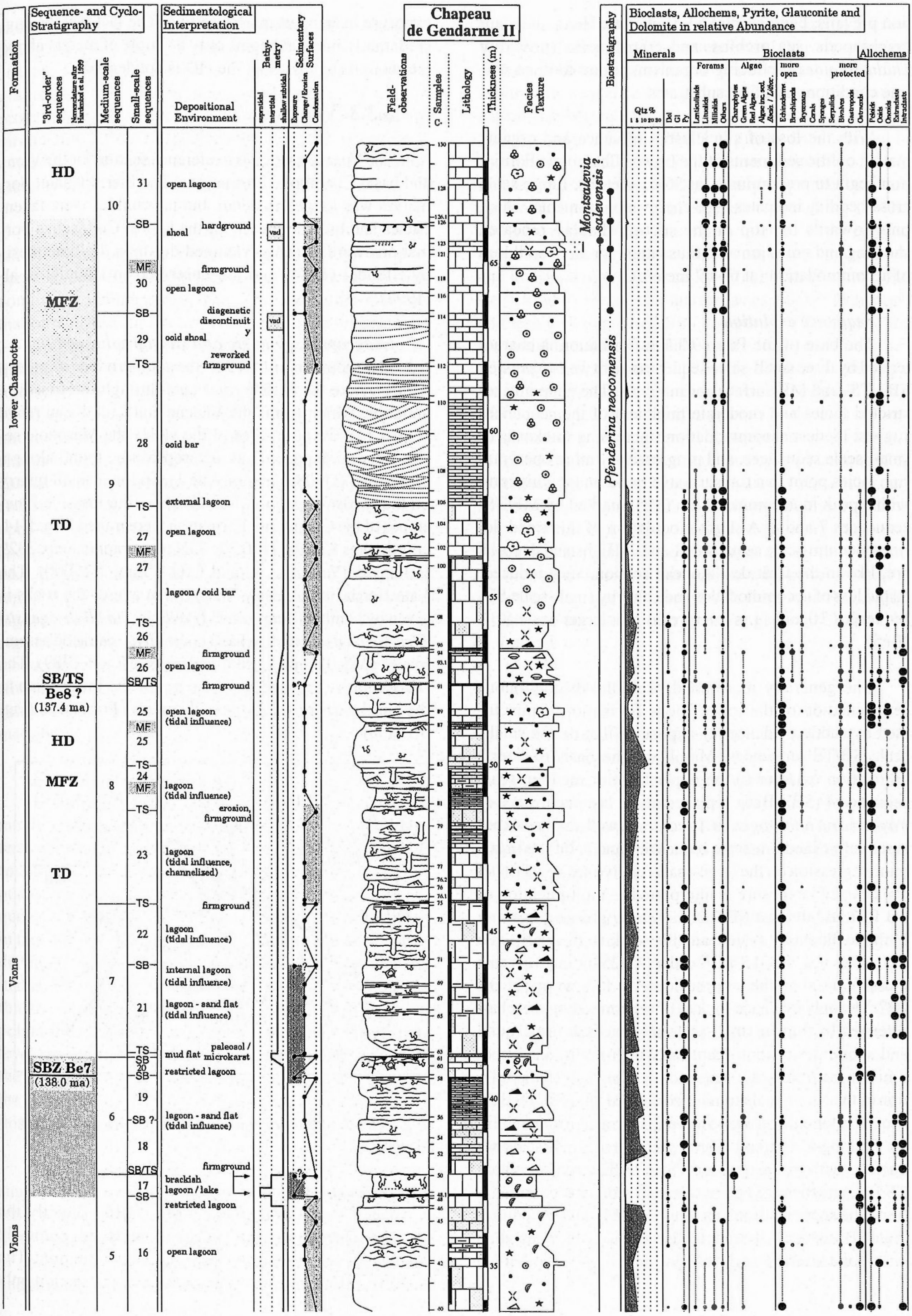


Fig. 5.13b. Chapeau de Gendarme section (part II).

tion prevails, beginning at 53-55 meters. Here, abundant brachiopods, sea urchins, and crab pincers (probably *Thalassinoides* producing organisms) point to open marine conditions and firm substrates.

With the loss of siliciclastic influence and organic matter, oolitic sediments of the Lower Chambotte Formation begin to predominate (at 56-63 meters). Large-scale cross-bedding indicates high-energy environments, which only towards the top of the section display repeated shoaling and omission and, thus, attest for an overall loss of accommodation (at 64-67 meters).

Sequence evolution

The base of the Pierre-Châtel Formation is characterized by three small-scale sequences with well-expressed SB-, TS- and MF-surfaces or intervals. The generally restricted facies and moderate thickness of the sequences suggest modest accommodation. Increasing thickness of small-scale sequences, and progressively more open-marine facies point to a transgressive trend on the long term with a peak in accommodation (MF) marked by MF/TS-sequences 7 and 8. A slight modulation of this trend defines medium-scale sequences 1 and 2. Highstand deposits are, like in the Crêt de l'Anneau section, very reduced. Rapid loss of accommodation indicated by small-scale SB-sequences 10 and 11 is interpreted as a larger-scale SBZ (Be5).

The generally more condensed and discontinuous sedimentation of the lower Vions Formation is punctuated by subtle evidence for exposures that define small-scale SB/TS-sequences. Mainly on the basis of lateral correlation (chapter 6) the upper limit of medium-scale sequence 4 (SBZ Be6) is placed at the last occurrence of *Pavlovecina allobrogensis*. In contrast, well-marked exposures define medium-scale sequence 6 and point to a maximum regression on the large-scale relative sea-level trend. This twofold exposure is interpreted as multiplication of SB Be7 and defines SBZ Be7. The progressive opening of the sedimentary system and the stacking of small-scale sequences in a "SB-TS-MF" order (21-25) point to a transgressive trend on the large scale in the following of SBZ Be7. The only evidence for loss of accommodation is characterized by thinner small-scale sequences, a firmground and subtle desiccation features, which define the upper limit of medium-scale sequence 8. This subtle modulation of the larger-scale trend is correlated with SB/TS Be8. Much better expressed, however, is a transgressive pulse on the long-term trend. Three successive small-scale sequences with well-expressed TS and MF, and a very thick TS/SB sequence (28) point to a high gain in accommodation. The superposition of a long-term MF and a medium-scale SB (sequence 9) leads to discontinuous sedimentation with condensation surfaces and occasional, short-lived exposure in an open-marine setting. The fact that the covered, marly interval begins only a couple of meters above the section suggests that the HD is condensed.

5.3.3. Vuache

The Vuache section is a reference section for the Vions and basal Chambotte Formations. The Pierre-Châtel Formation was logged in detail, but no samples were taken. Facies and biostratigraphy of the Lower Cretaceous Formations were already examined during a diploma thesis by Blondel (1984) and are described in Blondel et al. (1992).

Geographic, geologic and stratigraphic setting

The section is situated on the southern face of a steep valley where the Rhône river cuts through the first Jura anticline, which forms the Vuache and Crêt d'Eau range (Fig. 5.14). On both sides of the valley the Mesozoic series are well exposed. The outcrop can be found along a small road (D908a, section line 1), and in a small quarry on the roadside (section line 2) where the massive limestones of the Chambotte Formation were mined (Fig. 5.14, coordinates 874.98/2130.37, IGN topographic map, 3329 Bellegarde/Valserine, Grand Crêt d'Eau, 1/25000). The Late Berriasian to Early Valanginian age of the rocks is given by benthic foraminifers (*Pavlovecina allobrogensis, Pfenderina neocomensis*) and charophyte-ostracod assemblages (M5, Blondel 1984, Detraz & Mojon 1989). The succession is exposed from the top of the Purbeckian to the basal parts of the Lower Chambotte Formation (Fig. 5.15a, b).

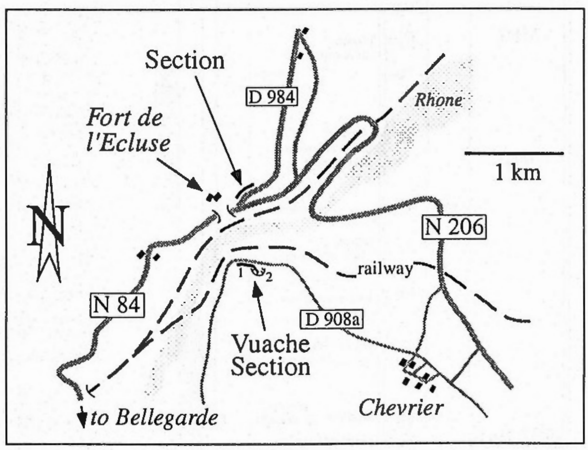


Fig. 5.14. Location of the Vuache and Fort de l'Ecluse sections.

Sedimentological interpretation

The Pierre-Châtel Formation overlies directly the greenish, charophyte-rich Purbeckian marls and commonly displays a homogeneous lagoonal sedimentation. The massive, white limestones at the base of the section only show few change/erosion and exposure discontinuities. Marly joints are the only indicator for episodic, more restricted sedimentation (at 1-12 meters). The pack- to grainstones then give way to a muddy sedimentation in protected and/or deeper (below wave base) lagoonal environments (at 13-25 meters). Shoaling and exposure is evidenced with higher-energy conditions and desiccation features at the top of the Formation (at 26-30 meters).

A drastic change in facies with abrupt influx of siliciclastics marks the onset of the Vions Formation. A very reduced range of *Pavlovecina allobrogensis* and discontinuous sedimentation with firm- and hardground formation suggests important condensation at the base (at 30-33 meters). A short phase of more continuous sedimentation terminates in an interval of repeated paleosol formation, and intercalation of charophyte-rich marls (at 33-43 meters). The alternating presence of charophytes and echinoderm-debris points to important salinity changes. Echinoderm debris probably is derived from nearcoastal hard bottom dwellers, such as various sea stars and sea urchins, and does not necessarily signify fully open-marine conditions. The rich content in siliciclastics and plant debris indicate a dominating terrestrial influence. Slightly more open-lagoonal conditions towards the top of the formation are indicated by green algae and a higher foraminifer diversity (at 43-49 meters). High-energy conditions and fully open-marine lagoonal environments then take over, along with a diminishing content in siliciclastics. This change, here expressed as a gradual transition, traditionally forms the base of the Chambotte Formation.

Sequence evolution

Following the top of the Purbeckian, which commonly is defined by an important SB (Be4) on the long-term sea-level trend (Strasser & Hillgärtner 1998), 6 small-scale SB/TS sequences delimited by erosion and exposure surfaces are identified. Two medium-scale sequences are defined by the most obvious discontinuities in this interval. In contrast to the more proximal Chapeau de Gendarme section no brackish influence is evidenced in this interval. The last two of the 6 small-scale sequences thicken upwards and display well-marked MF, which points already to highest accommodation and relative sea-level rise on the long term. This trend is culminating in the two following MF sequences (7 and 8). As in most of the other sections shoaling and exposure occur rapidly after this trend (small-scale SB-sequence 9, SBZ Be5).

Lateral correlation and the short range of *Pavlovecina* suggest condensation (non-deposition/erosion) of larger-scale sequence Be5 in a small SB-sequence (10). A primary inhomogeneity such as a well-marked discontinuity or paleokarst may have served as conduit for pore fluids and favored the Cenozoic karstification and siderolite infiltration at this place. In the condensed, quartz-rich interval two small-scale SB-sequences (11/12) are inferred, mainly on the basis of lateral correlation. One thicker small-scale sequence (13) points to increased accommodation, just before numerous exposures define three small SB-sequences (15-17). TS/SB sequence 14 is inferred in comparison with the Fort de l'Ecluse section (Fig 5.16a). Sequence 16 displays 4 elementary sequences that are made up of carbonate-marl couplets representing TD/HD and LD, respectively. The entire interval with repeated exposure is interpreted as a major loss of accommodation and lowstand of relative sea-level on the large scale (SBZ Be7). Slightly more open-marine facies, thicker sequences, and less well expressed SBs compared to TSs, stand for the initial transgressive phase on the long term. Preservation potential of marly LDs is also the highest during a longer-term transgressive trend. A well expressed, marly MF in sequence 19 additionally points to elevated accommodation. This marly interval is interpreted as MF instead of LD on the basis of absence of charophytes compared to other marls below, and lateral comparison of stacking pattern. Accommodation then stays moderate almost until the top of the section. Here, massive, high-energy lagoonal deposits again point to elevated rates of accommodation. A small modulation of this general transgressive trend is indicated only by one thin bed with firmground formation (small-scale sequence 22) Here, longer-term SB/TS Be8 is inferred.

5.3.4. Fort de l'Ecluse

The Fort de l'Ecluse section is chosen to complement the nearby Vuache section for the Lower Valanginian and to reveal the short-distance variability of facies, discontinuities, and stacking pattern. Facies and biostratigraphic studies have already been carried out by Blondel (1984) and others, and are described in Charollais & Wernli (1995).

Geographic, geologic and stratigraphic setting

The section is located directly along the French departmental road (D984) between Bellegarde and Gex on the northern side of the Rhône valley between Vuache and Crêt d'Eau ranges (Fig. 5.14, coordinates 875.25/131.00, IGN topographic map, 3329 Bellegarde/Valserine, Grand Crêt d'Eau, 1/25000). The succession makes up part of the eastern limb of the first Jura anticline and plunges 75 to 85 degrees towards the northeast. The section is situated approximately 600m north of the Vuache section. In contrast to the Vuache section, however, the Pierre-Châtel Formation is only partly exposed, but the complete Chambotte Formation and the lower part of the Guiers Member/Arzier marls are well accessible. Biostratigraphy has been established mainly on the basis of benthic foraminifers (Fig 5.16a, b, Charollais & Wernli 1995).

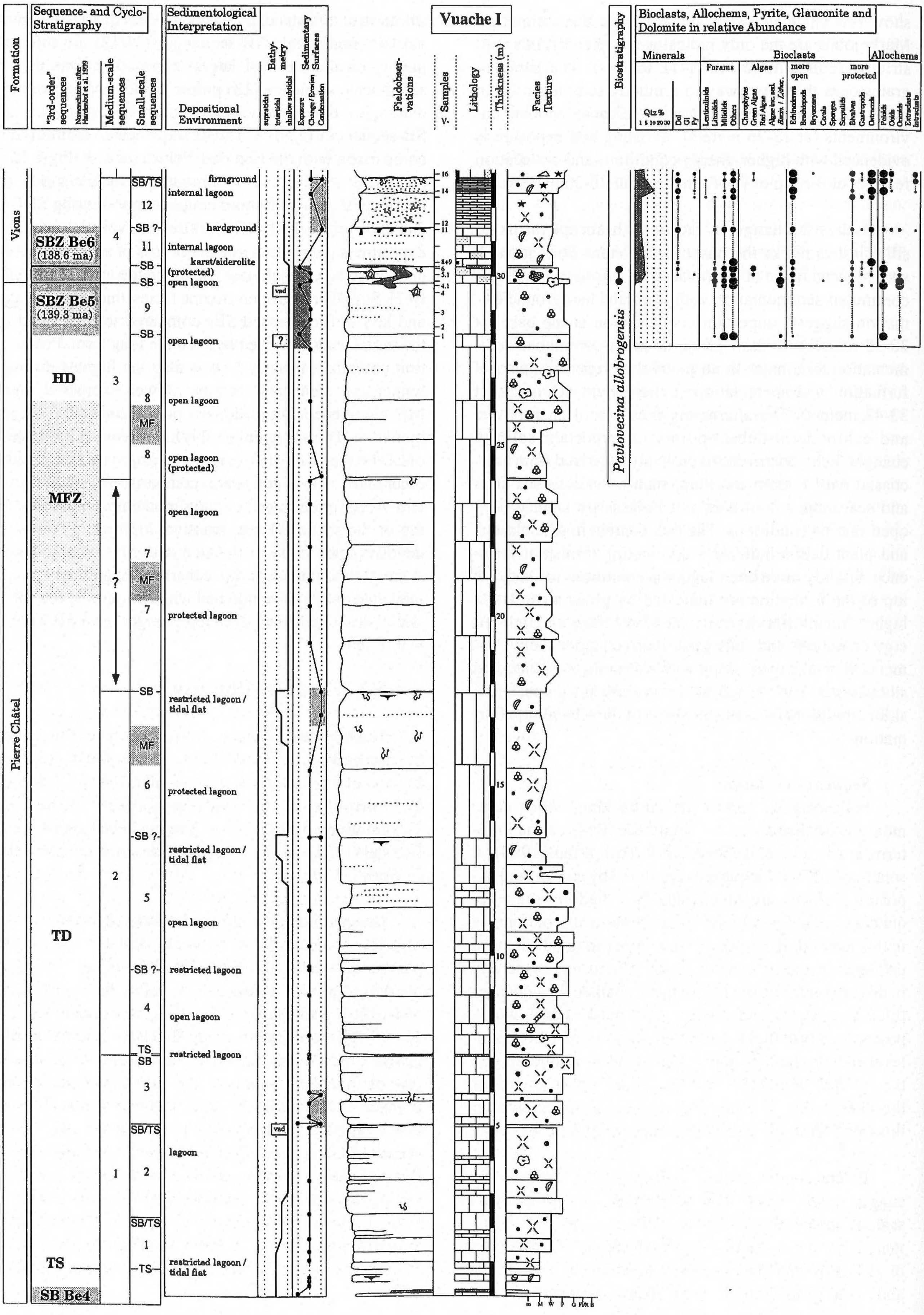


Fig. 5.15a. Vuache section (part I).

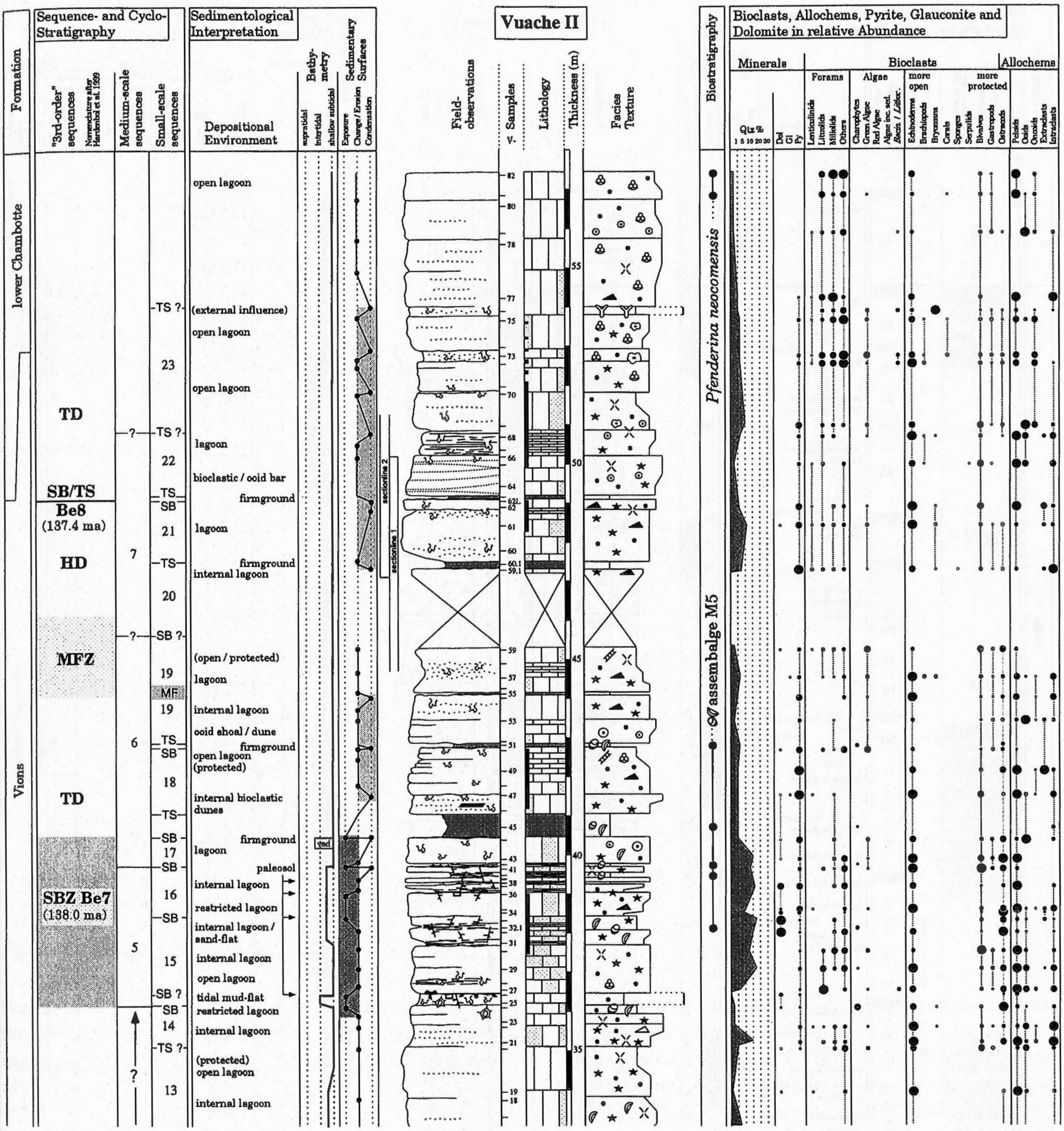


Fig. 5.15b. Vuache section (part II).

Sedimentological interpretation

General facies evolution of the entire section and especially the Pierre-Châtel Formation is well comparable with that of the Vuache section. The boundary between Pierre-Châtel and Vions Formations is as sharp. Initial condensation is indicated by firm- and hardground formation and, as in the Vuache section, by a reduced range of marker foraminifers (Fig. 5.16a). A short interval of open lagoonal environments is followed by marly carbonates punctuated by abundant exposure and condensation discontinuities.

Expression of discontinuities and depositional environments vary over a distance of 1 km. An oolitic (tidal) sand-wave (at 15 meters), for example, is unambiguously correlatable with restricted lagoonal sediments containing charophytes in the Vuache section (at 36 meters). The hardground and paleosol developed there are not observed in the Ecluse section. Such lateral variability over a short distance is rather the rule than the exception in modern, comparatively small platform environments. Here, however, it points to morphological structuration and (exceptionally?) diverse depositional environments on the large Jura platform in the Upper Berriasian. It has to be noted,

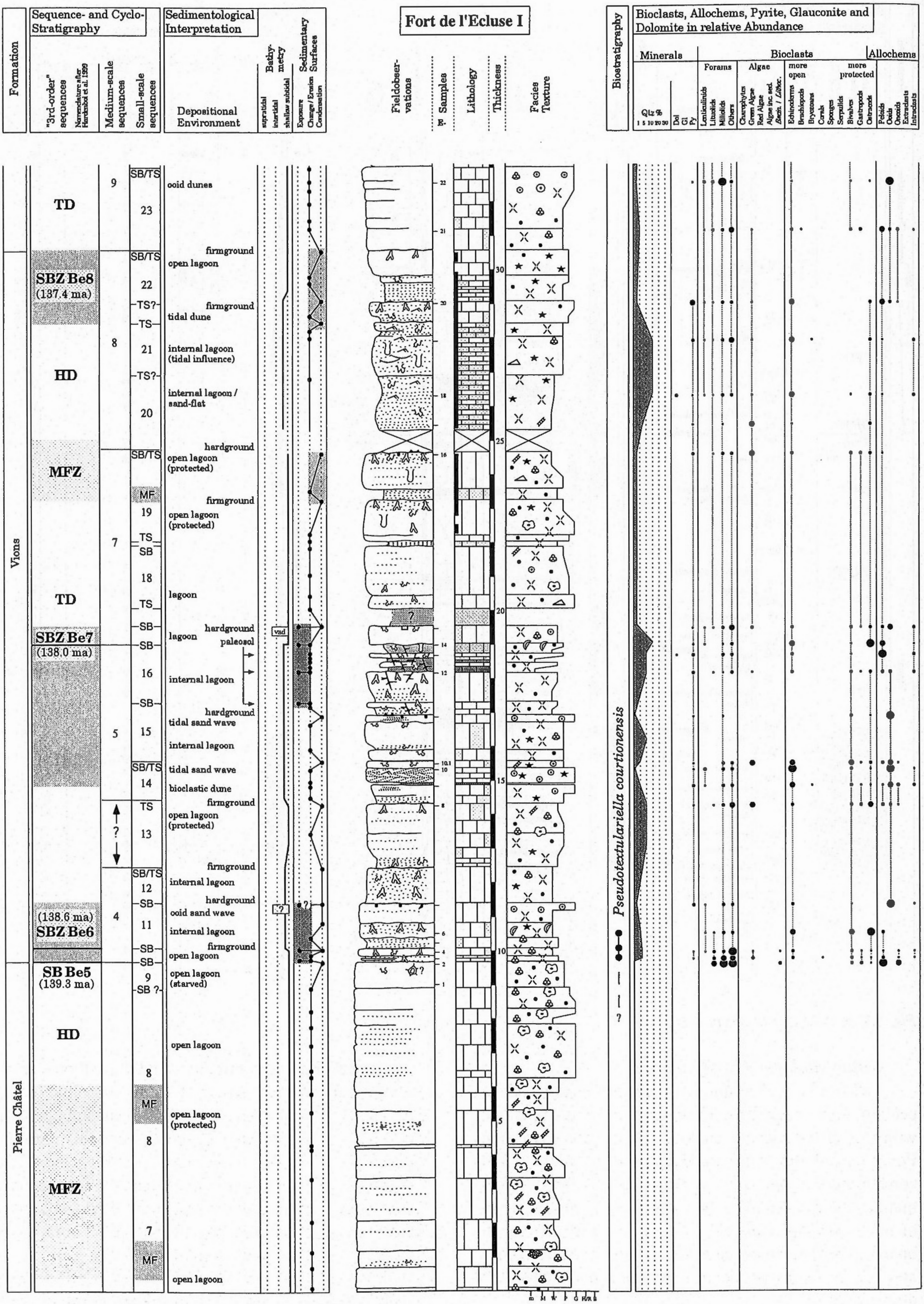


Fig. 5.16a. Ecluse section (part I).

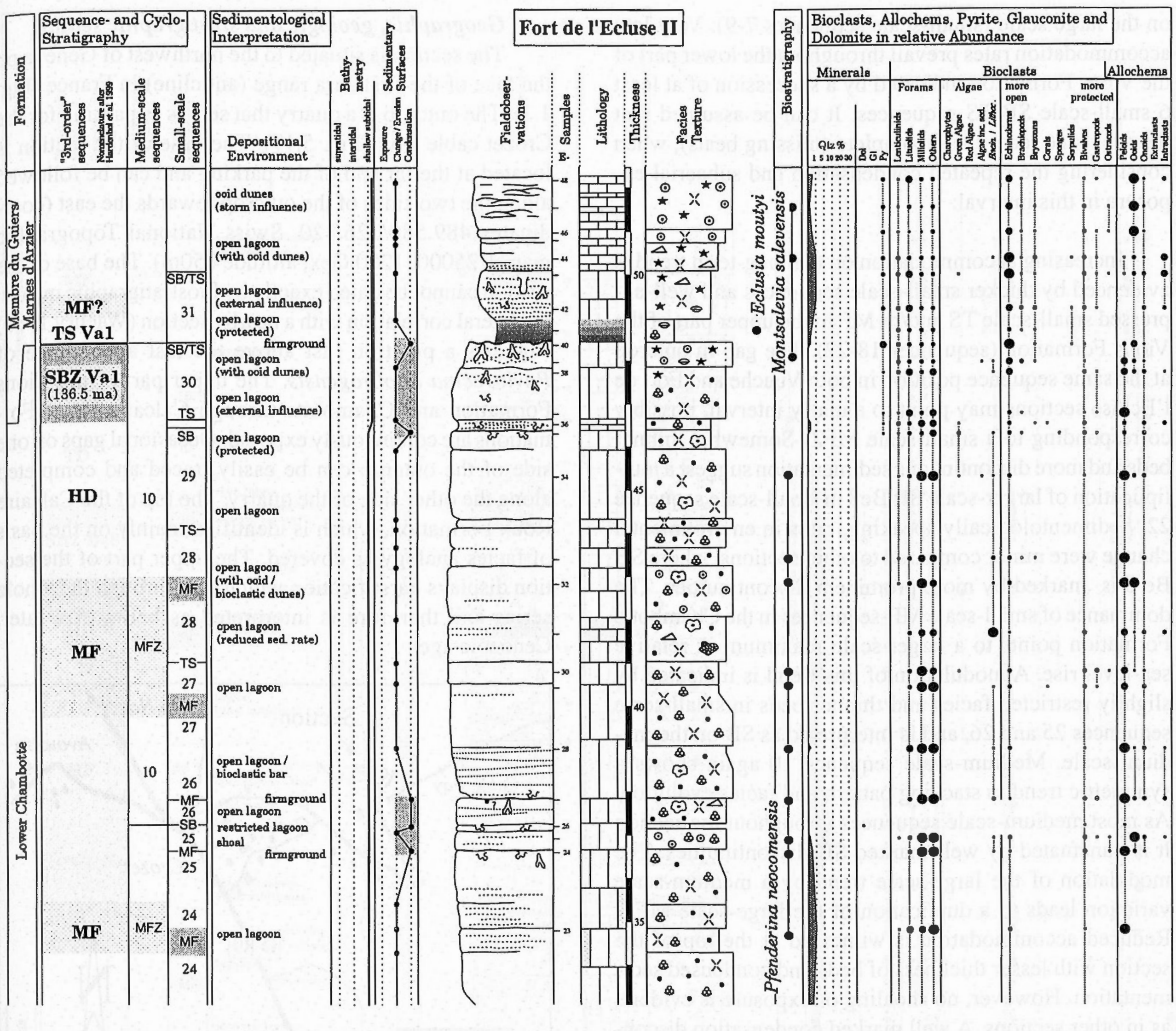


Fig. 5.16b. Ecluse section (part II).

though, that such small-scale variability does not change the correlatability of general facies evolution and stacking pattern and, thus, does not question the important role of allocyclical forcing.

After this interval with strong terrestrial influence more open lagoonal environments take over (at 20-25 meters). A shallowing and/or elevated influence of tidal currents is evidenced by renewed siliciclastic input, flaser bedding, and reworking. This facies compares well to tidalites observed in the Chapeau de Gendarme and Crozet sections. Open lagoonal sediments and massive beds of the Chambotte Formation (base at 30 meters) then indicate an increase in accommodation. Facies is very homogeneous and displays only slight variations in mud content. One exception is a restricted lagoon documented by a thin bed in the middle of the Formation (at 37 meters). The top of the Formation is announced by increasingly discontinuous (condensed) sedimentation and is punctuated by an extensively bioturbated firmground.

Facies then indicates a general opening of the lagoonal environment with increasing content of echinoderms, brachiopods and bryozoans. The initially protected, marly, open-lagoonal environment gradually passes into high-energy environments. Renewed influx of siliciclastics and first occurrence of storm influence (HCS) suggest a climatic change to more humid and seasonal conditions. Marker foraminifers in this interval (Fig. 5.16b) give a Lower Valanginian age and allow correlation with the Guiers Member/Arzier Marls.

Sequence evolution

In order to allow the comparison with the Vuache section the sequence framework is numbered identically, even though the base of the Pierre-Châtel Formation is missing (Figs. 5.15 and 5.16).

In the same way as in the Vuache section, the upper part of the Pierre-Châtel Formation is characterized by maximum accommodation followed by rapid regression on the large scale (small-scale sequences 7-9). Very low accommodation rates prevail throughout the lower part of the Vions Formation indicated by a succession of at least 6 small-scale SB/TS sequences. It can be assumed that the sequence record is incomplete (missing beats), when considering the repeated condensation and subaerial exposure in this interval.

Increasing accommodation on the long-term trend is evidenced by thicker small-scale sequences and well-expressed small-scale TS and/or MF in the upper part of the Vions Formation (sequences 18-22). The gap in outcrop at the same sequence position in both Vuache and Fort de l'Ecluse sections may point to a marly interval, possibly corresponding to a small-scale MFZ. Somewhat thinner beds and more discontinuous sedimentation suggest a multiplication of larger-scale SB Be8 in small-scale sequence 22. Sedimentologically speaking, rates in environmental change were minor compared to other sections, where SB Be 8 is marked by more prominent discontinuities. The dominance of small-scale MF-sequences in the Chambotte Formation points to a larger-scale maximum in relative sea-level rise. A modulation of this trend is indicated by slightly restricted facies and thinner beds in small-scale sequences 25 and 26, and is interpreted as SB on the medium scale. Medium-scale sequence 10 again shows a symmetric trend in stacking pattern and facies evolution. As most medium-scale sequences throughout the section it is punctuated by well-marked SB-discontinuities. The modulation of the large-scale trend by a medium-scale variation leads to a duplication of the large-scale MFZ. Reduced accommodation is witnessed at the top of the section with lesser thickness of beds and condensed sedimentation. However, no shoaling or exposure is evident, as in other sections. A well marked condensation discontinuity, defining the onset of external (offshore) conditions, is interpreted as large-scale TS (Va1). By means of lateral correlation it appears that the MF of the following large-scale sequence is almost directly superimposed and correlates with the Arzier marls.

It can be stated that this section allows exceptionally well to establish a small-scale sequence framework in the Chambotte Formation. This may be the result of more protected lagoonal environments, in contrast to high-energy lagoons and ooid bars in other sections that are less sensitive for short-term allocyclical signals.

5.3.5. Crozet

Compared with other sections in the Jura, the Crozet section has not been extensively studied. Only once, the newly exposed outcrop was examined during a diploma thesis mainly concerned with facies and biostratigraphy (Böker 1994).

Geographic, geologic and stratigraphic setting

The section is situated to the northwest of Geneva on the face of the first Jura range (anticline) in France (Fig. 1.1). The outcrop is a quarry that serves as parking for the Crozet cable way (Fig. 5.17). The base of the section is located at the far end of the parking and can be followed along the two sides of the outcrop towards the east (coordinates 489.580/126.820, Swiss National Topographic map, 1/25000, 1280 Gex, altitude 650m). The base of the section cannot be dated exactly by biostratigraphic means, but lateral correlation with a nearby section (Waehry 1988) suggests a position just above the last appearance of *Pavlovecina allobrogensis*. The upper part of the Vions Formation and Chambotte through Calcaire Roux Formations are continuously exposed. Occasional gaps on one side of the outcrop can be easily traced and completed along the other side of the quarry. The top of the Calcaire Roux Formation, which is identified mainly on the basis of facies analogy, is covered. The upper part of the section displays karstification which cuts through the whole series and therefore is interpreted as being of a later, Cenozoic age.

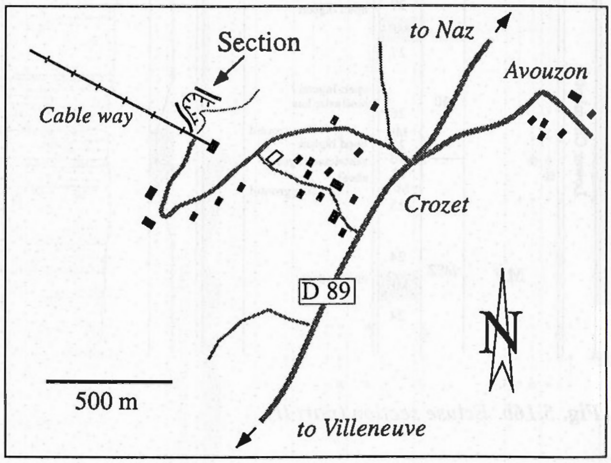


Fig. 5.17. Location of the Crozet section.

Sedimentological interpretation

The basal part of the section is characterized by an extremely discontinuous sedimentation in near-coastal environments. High content in siliciclastics (clay and quartz) and charophytes point to restricted lagoonal (tidal) environment with occasional brackish influence. The repeated, abrupt change from muddy-lagoonal to skeletal-calcareous sediments with loading structures and intensive bioturbation is interpreted as wash-over sedimentation (at 4 meters). More continuous open-marine conditions are indicated by initially condensed and, then moderate- to high-energy lagoonal environments with elevated contents in echinoderms and ooids (at 7-11 meters). A short, repeated phase of restricted-lagoonal conditions (sterile marls at 11/12 meters), is followed by open-lagoonal environments of similar bathymetry but with influence of tidal currents. Resembling the Chapeau de Gendarme sec-

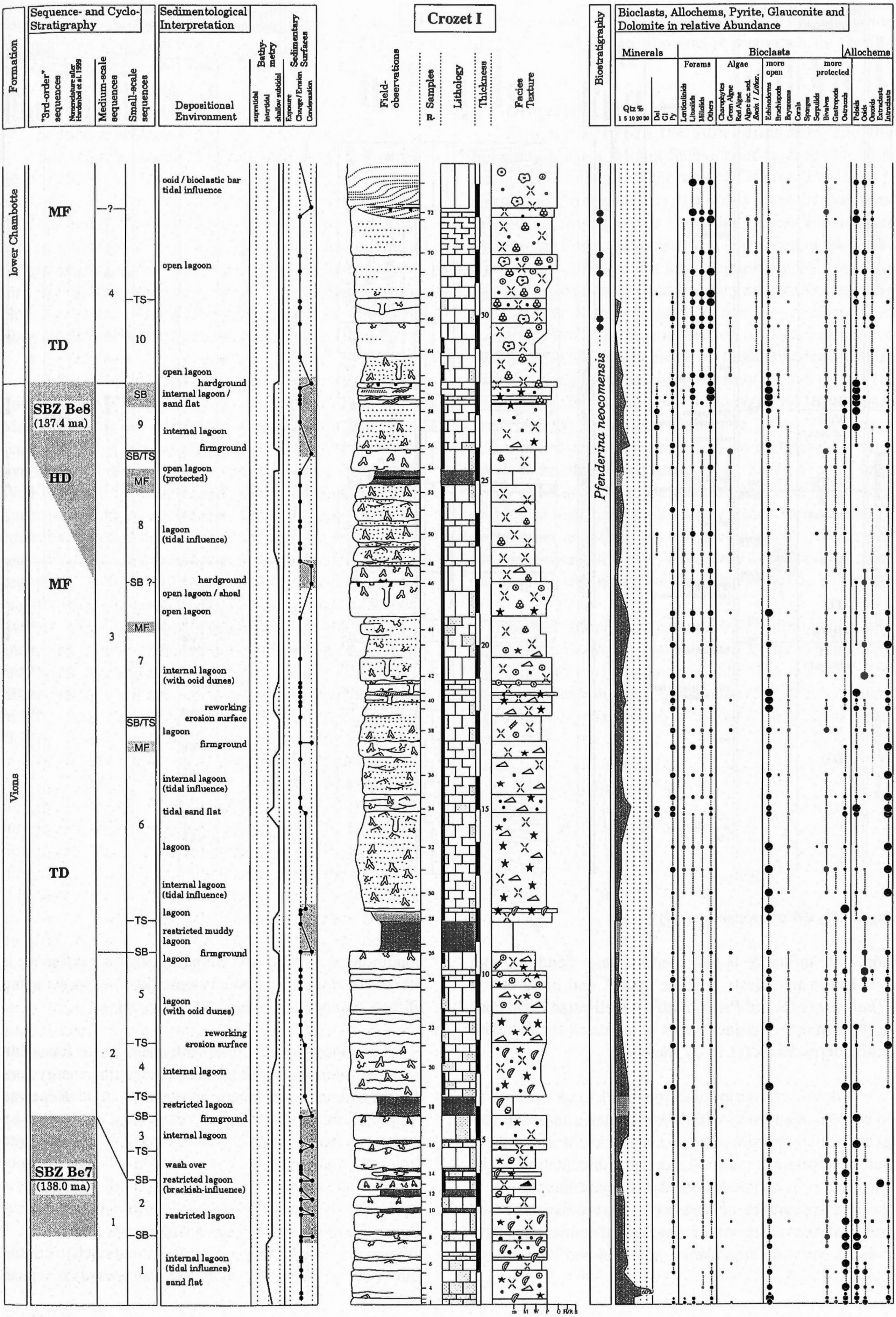


Fig. 5.18a. Crozet section (part I).

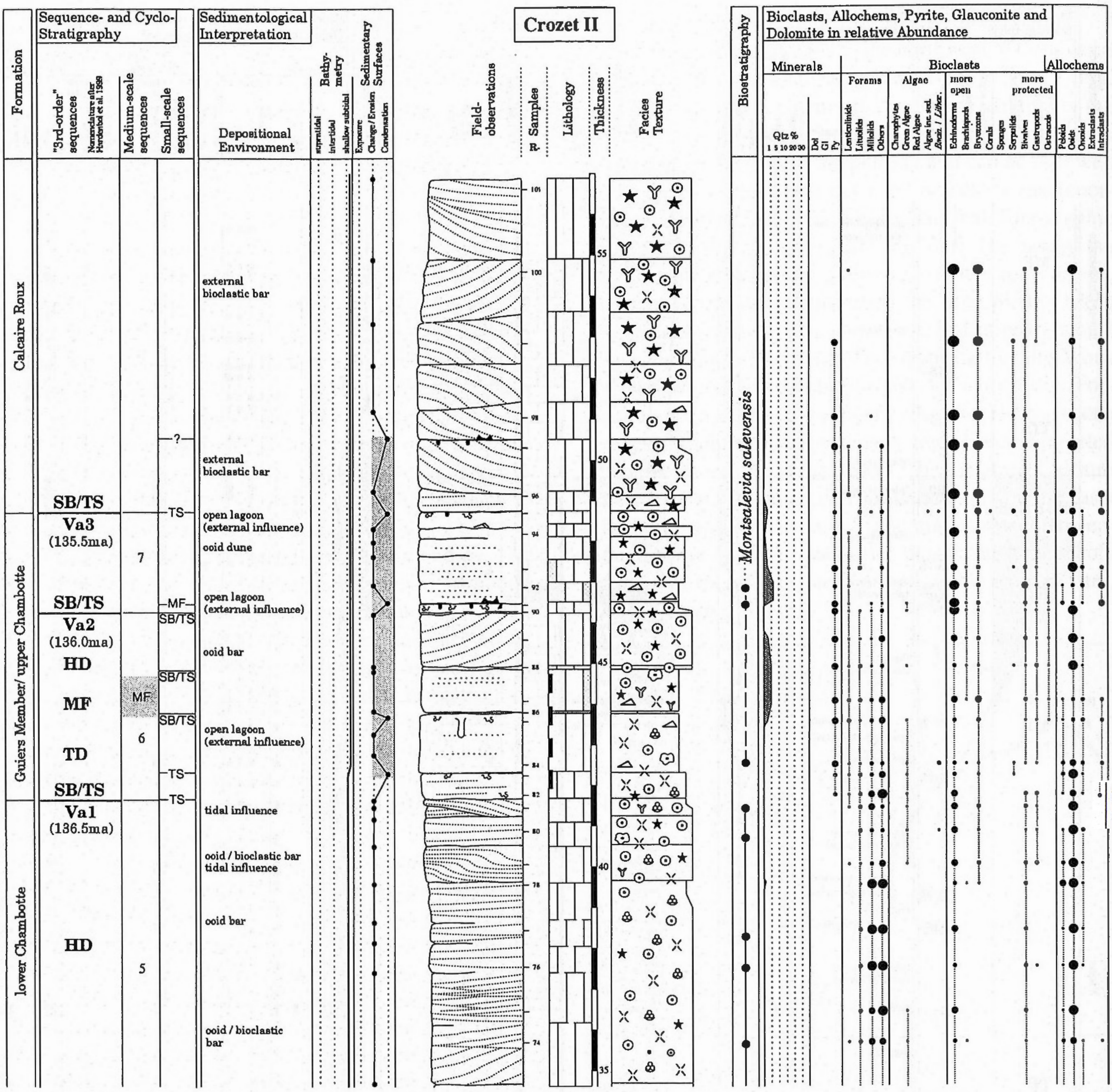


Fig. 5.18b. Crozet section (part II).

tion, tidal influence is indicated by flaser bedding, high contents in intraclasts, organic matter, and bioturbation (*Thalassinoides* and *Psiloichnus*). A well-established connection to open-marine waters is indicated by abundant echinoderm debris (at 12-27 meters).

A drastic change in color from black (dark red-brown) to beige is associated with lower contents in pyrite (organic matter) and siliciclastics and marks the initiation of purely calcareous, marine-lagoonal sedimentation of the Chambotte Formation (at 28-33 meters). It coincides with the first appearance of *Pfenderina neocomensis*. Thick beds that pass into massive cross-bedded oolitic limestones indicate high accommodation potential and high-energy conditions in an ooid belt. Shoaling and tidal influence is marked by thinning beds and sigmoidal foresets at the top of the Chambotte Formation (at 42 meters).

Sedimentation continues with open-marine fauna, but is more discontinuous and points to changing energy conditions with varying contents in carbonate mud. Renewed influx of siliciclastics and reworking indicated by intraclasts and abraded, iron-stained ooids (black/red facies) also suggests an environmental change of probably climatic origin (refer also to chapter 8). This facies is typical for the Lower Valanginian (*Montsalevia salevensis*) Guiers Member and/or Upper Chambotte Formation. A different facies with massive bioclastic bars rich in crinoid and bryozoan debris (from 49 meters onward) is typical for the external, open-marine sedimentation of the Late Valanginian Calcaire Roux Formation. However, no biostratigraphic data are available to confirm this attribution.

Sequence evolution

The sequence analysis of this section alone is somewhat problematic, since subaerial exposures are not well marked and small-scale bathymetric trends are rather subtle. This, and an elevated thickness of the Upper Berriasian succession, suggest generally higher subsidence rates at this particular locality, and/or erosion of exposure indicators. Therefore, the sequence framework is established in close comparison with stacking patterns of the nearby sections (Vuache, Fort de l'Ecluse, Chapeau de Gendarme).

In the restricted environments at the base of the section three small-scale SB-sequences are inferred (1-3). Marly intervals are interpreted as elementary and small-scale LDs. The general facies evolution points to a relative lowstand of sea level on the large scale (SBZ Be7). Well-marked TSs, increasing thickness of small-scale sequences, and more open facies testify for increasing accommodation and deepening in the following part of the section (small-scale sequences 4-5). Thick TS- and SB-sequences in the tidally-influenced interval point to a massive increase in accommodation and portray a fast relative sea-level rise on the long-term trend (6-8). Thin beds and shoaling point to modulation of the transgressive long-term trend and slight loss in accommodation (SBZ Be8, SB-sequence 9). Renewed transgression is indicated by more open facies, thicker beds and high-energy conditions. However, small-scale sequences are difficult to identify in these mobile high-energy depositional systems (ooid/bioclastic bars). Channelization with hardground formation in the middle of the Lower Chambotte Formation is interpreted as indicator for a slight loss in accommodation and as upper limit of medium-scale sequence 4. Thinning beds, shoaling, and tidal influence at the top of the following interval again point to loss in accommodation and are correlated with SB/TS Va1. A symmetric transgressive-regressive evolution is manifested in medium-scale sequence 6. Four small-scale SB/TS sequences can be distinguished and point to a transgressive trend on the large scale. Mainly on the basis of biostratigraphy, facies, and sequence evolution the "3rd-order" SB/TS Va2 and Va3 are placed at the best expressed discontinuity surfaces in this interval. The reduced thickness of these sequences compared with other, more distal sections, suggest important condensation in the Guiers Member and Upper Chambotte Formation.

5.3.6. Monnetier

The Monnetier section was already studied as early as 1913 by Joukowsky & Favre and is described in their monograph about Mount Salève. Samples from their study were looked at in terms of biostratigraphy by Zaninetti et al. (1988) and the section was re-examined by Deville (1991).

Geographic, geologic and stratigraphic setting

Situated between the Jura Mountains and the Subalpine Chains, Mount Salève rises from the Molasse Basin just southeast of the city of Geneva (Fig. 1.1). It constitutes the frontal part of a thrust sheet that has been moved northwestwards over the Lower Freshwater Molasse of Early Oligocene age. The steep, west-northwest facing cliffs exhibit an exceptionally well exposed and almost continuous sedimentary succession reaching from Upper Kimmeridgian to the Lower Barremian (Lombard 1967). Monnetier section is located just above the village of Monnetier on the northeastern edge of Grand Salève, facing its "small brother" Petit Salève (Fig. 5.19, coordinates 504.450/112.800 Swiss National Topographic map, 1/25000, 1301 Genève, altitude 760m). The section is composed of 4 parts (section-lines) in different mining levels of an abandoned quarry (Fig. 5.19). Section-lines either overlap and can be unambiguously correlated, or individual beds can be physically traced from one place to the other and, thus, provide a continuous overview of the succession. However, the access to the vertical cliffs can be difficult and a rope is needed at certain levels.

Biostratigraphy by foraminifers, calpionellids, and charophyte-ostracod assemblages confirms the Late Berriasian to Early Valanginian age (Fig. 5.20a-c). The succession exhibits the Upper Pierre-Châtel to Lower Chambotte Formations in an undeformed, "layer-cake" fashion.

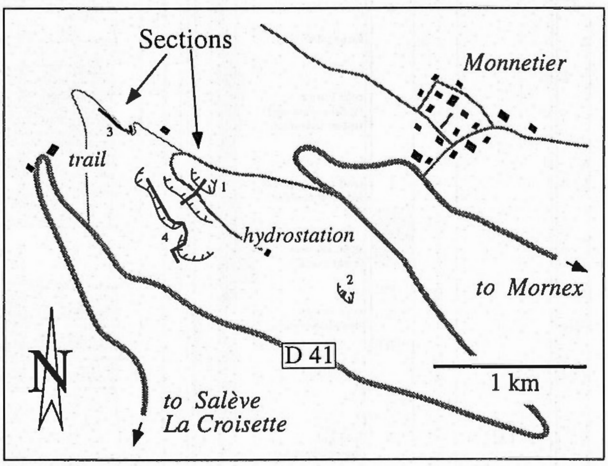


Fig. 5.19. Location of the Monnetier section.

Sedimentological interpretation

In comparison with the Salève section (La Corraterie) it can be assumed that the high-energy oolitic/bioclastic depositional environments at the base of the section belong to the middle part of the Pierre-Châtel Formation.

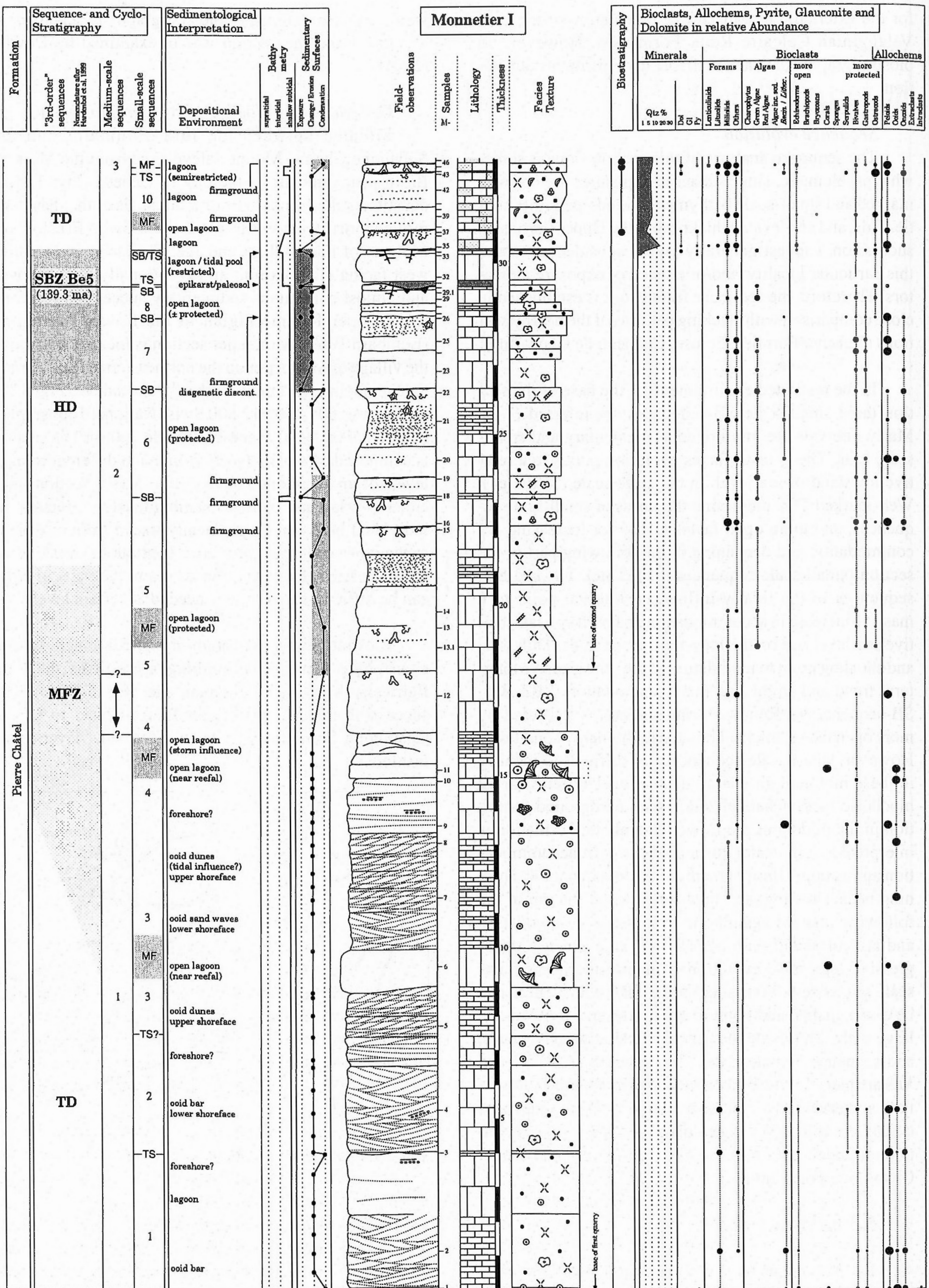


Fig. 5.20a. Monnetier section (part I).

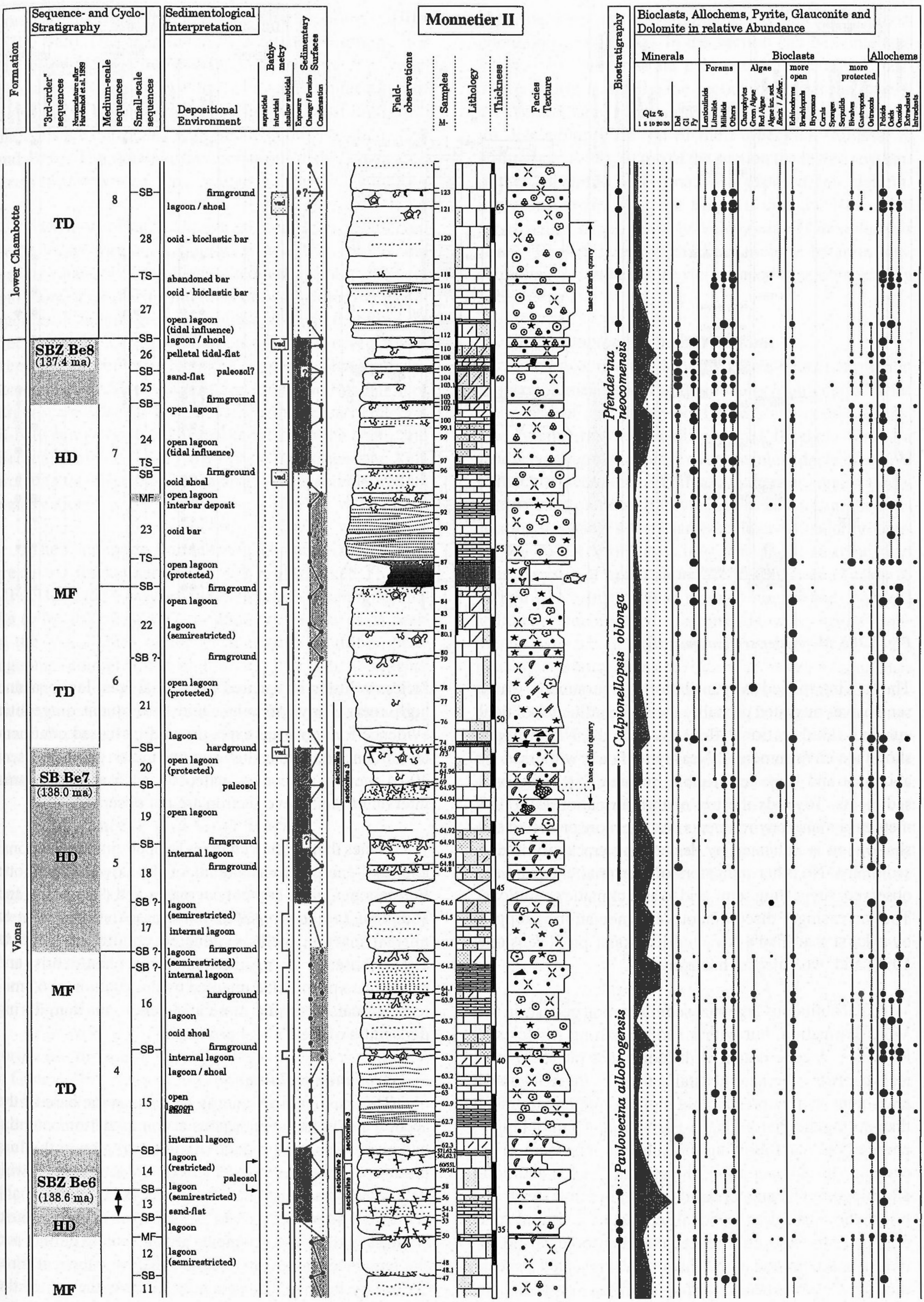


Fig. 5.20b. Monnetier section (part II).

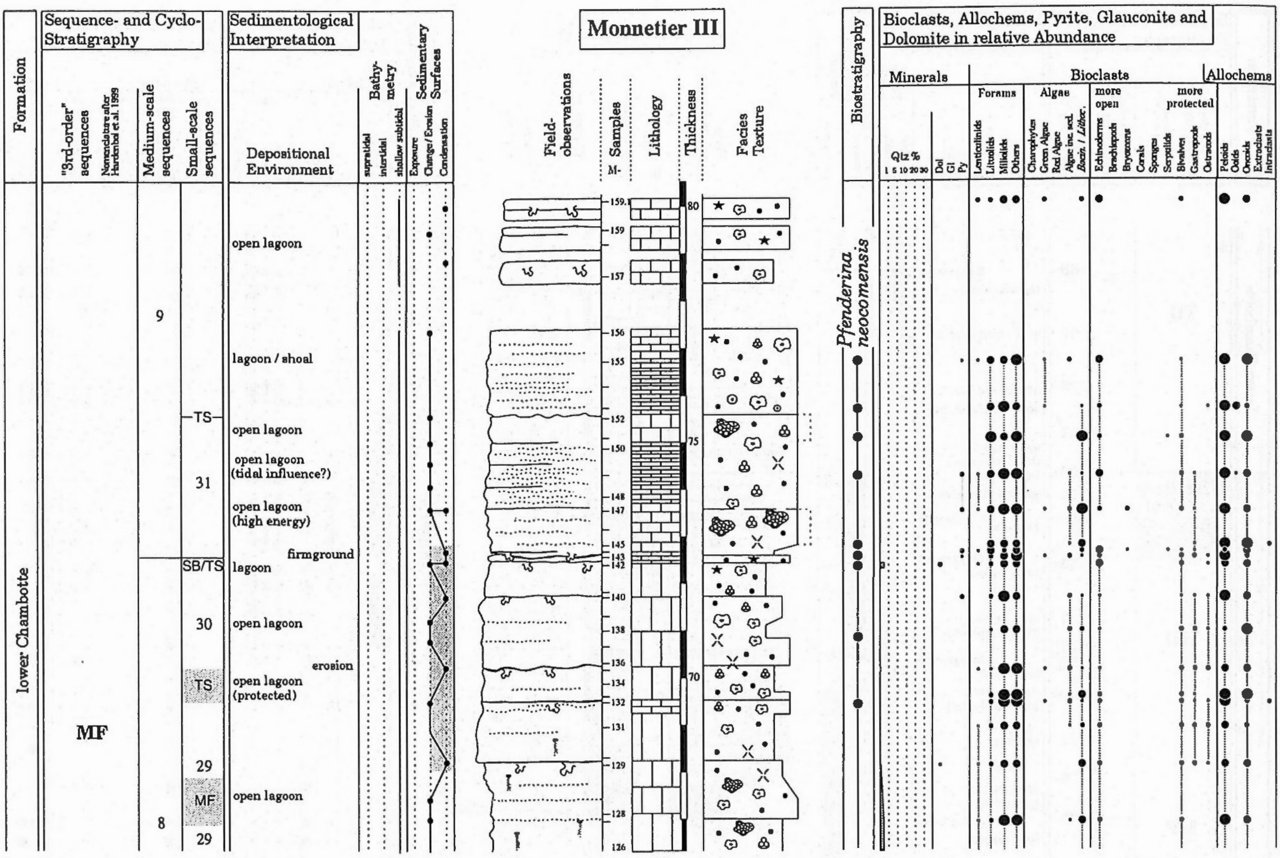


Fig. 5.20c. Monnetier section (part III).

They are interpreted as open platform environments where sand waves migrated probably under the influence of tidal currents (unidirectional flow towards the NE in upper shoreface environments). Nearby reefs are witnessed by abundant and large reef rubble in intercalated lagoonal sediments. Towards the top of the Formation (at 17-28 meters) lagoonal environments become more protected and shallowing is indicated by desiccation cracks in muddy sediments. No other indicators of subaerial exposure are observed, suggesting semi-arid to arid climatic conditions. The increasingly discontinuous sedimentation is topped by a karst that displays cylindrical karst pits (Vanstone 1998) and two paleosol horizons.

Discontinuous lagoonal sedimentation persists in the Vions Formation, but with a sudden, strong siliciclastic influence. A zone dotted with root traces (at 35-37 meters) confirms the near-coastal setting of these lagoonal environments in a presumably more humid environment than during the Pierre-Châtel Formation. An initial high-energy ooid bar following the subaerial exposure gives evidence for an opening of the system. This is additionally suggested by an increased content in echinoderm debris in the subsequent lagoonal sediments. A second zone dominated by exposure surfaces and paleosols again points to loss in accommodation. This interval revealed a specimen of *Calpionellopsis oblonga* and suggests a Late Berriasian age (*Picteti* subzone, Deville 1991). A similar facies evolution is repeated with initial open-lagoonal and high-energy environments, and subsequent diagenetic evidence for subaerial exposure and increased continental influence (siliciclastics, organic matter, plant debris, at 49-61 meters). However, paleosols such as in the same interval of the Salève section are not observed.

Facies then visibly changes to open-marine, lagoonal and high-energy environments of the Lower Chambotte Formation. Quartz content decreases and disappears, and abundant *Lithocodium/Bacinella* associations point to normal marine, well-oxygenated conditions (Dupraz 1988). Absence of organic matter and plant debris, and subaerial exposure documented by desiccation (at 65 meters) indicate probably more arid conditions than during deposition of the Vions Formation.

Sequence evolution

The massive high-energy deposits at the base of the section testify for an enormous accommodation potential and high carbonate productivity in this region of the Jura platform. Relative sea-level rise on the long term is well illustrated by small-scale TS-sequences (1-2) and small-scale MF-sequences (3 to 5). However, bathymetric changes in these environments are difficult to define, and the sequence framework is somewhat hypothetical. Progressively thinner SB-sequences towards the top of the Formation characteristically indicate loss of accommodation, particularly with well-documented exposure surfaces (SBZ Be5). One medium-scale sequence is inferred for the following symmetric, transgressive-regressive trend. The sequence coincides with the biostratigraphic range of *Pavlovecina allobrogensis,* and the two most regressive intervals are correlated with SBZs on the 3rd-order (Be5 and Be6). The detailed interpretation of this interval in terms of relative sea-level change is given in Figure 5.8.

Two more of such symmetrical trends can be observed in the Vions Formation, but facies as well as number and stacking pattern of small-scale sequences point to longer and more open-marine intervals of two medium-scale sequences. The transgressive-regressive trend therefore reflects larger-scale modulation of relative sea level. Repeated exposures in the most regressive phase punctuate small-scale SB-sequences and imply multiplication of larger scale SBs (SB Be7, Be8). The thick, marly interval in-between (sequence 23) marks a LD on the small and medium scales, but falls in a MF on the large scale. This probably explains the open-marine facies and favored its preservation.

The Chambotte Formation is characterized by progressively thicker sequences with well marked TSs and MFs. Here, the transgressive trend already announced in the following of SB Be7 continues and achieves its peak in the lower third of the Formation with thick open-lagoonal sediments. The high-energy bioclastic sediments at the top of the section are interpreted as prograding HD on the large scale. In the Salève section nearby, these deposits display shoaling and give clear evidence for lowered accommodation and slowed rise or beginning fall in relative sea-level.

5.3.7. Salève (La Corraterie)

The stratigraphy of Mount Salève in the Corraterie region is well known due to its good exposure and has been subject of numerous studies (e.g., Joukowsky & Favre 1913, Salvini 1982, Salvini-Bonnard et al. 1984, Zaninetti et al. 1988, Deville 1991).

Geographic, geologic and stratigraphic setting

Salève (La Corraterie) section is situated on the steep northwestern face of Mount Salève, high above the city of Geneva (Fig 5.21). The section is located about 4 km to the south of Monnetier section and displays an excellent lateral control of facies variability and discontinuities due to almost horizontal strata and well-exposed, accessible cliffs. The base of the section lies in the middle of a steep slope about 100 meters below Corraterie trail and 200 meters to the north of the very steep and rocky Etournelles trail (coordinates 501.770/110.115, Swiss National Topographic map, 1/25000, altitude 1115 m). The section crosses the Corraterie trail and ends close to the Trou de la Tigne, where the cliffs give gradually way to the meadow covering the top of the mountain. The outcrop is almost continuous, and breaks in the section can be completed laterally by tracing strata below and above (Figs. 5.21, 5.22). Stratigraphically, the section begins at the base of the Pierre-Châtel Formation that is well defined by a major discontinuity on top of the Purbeckian (Strasser & Hillgärtner 1998). At top of the section the lowermost part of the Calcaire Roux Formation is exposed, thus delimiting a chronostratigraphic range from Late Berriasian (*privasensis* zone) to Early Valanginian (*Campylotoxus* zone). Biostratigraphic markers include foraminifers, calpionellids, and charophyte-ostracod assemblages (Fig 5.22).

Sedimentological interpretation

The lower two thirds of the Pierre-Châtel Formation show open-marine and predominantly high-energy facies composed of alternating ooid shoals and open lagoons. The oosparites with well-marked foresets imply a sandwave migrated rapidly over the intertidal to emersive Purbeckian sediments. Echinoderms herein suggest a fully marine environment. Bathymetric trends are implied according to changes from large-scale cross bedding, typical for lower shoreface, into low angle stratification characteristic of upper shoreface to foreshore (Tucker & Wright 1992). Intercalated, strongly bioturbated, lagoonal sediments indicate condensation and low-energy environments. After about one meter of covered section (at 29 meters), an abrupt change to tidal-flat / restricted-lagoonal facies takes place. The following interval of about 5 meters is covered. Some beds protruding from the steep grassy hillside also show tidal-flat facies with birdseyes. Correlation with the Monnetier section suggests that this facies continues all the way up to the karst whose equivalent is, without doubt, located somewhere in the covered interval in the Salève section. This regionally important surface is marked by a karst over wide areas of the northern Jura Mountains (Pasquier 1995). The top of the Formation is

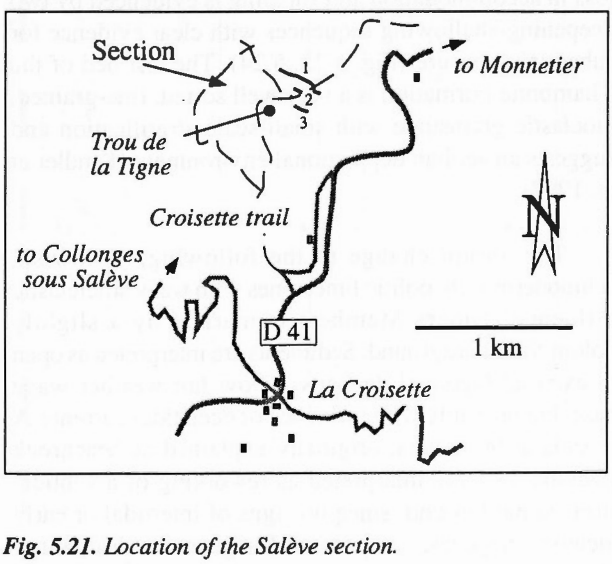


Fig. 5.21. Location of the Salève section.

characterized by the onset of lagoonal sedimentation and shows intense bioturbation (*Thalassinoides*) which indicates low accumulation rates.

Characteristic for the Vions Formation is its quartz content, in some beds up to 30%, and the reddish color caused by oxidation of pyrite-rich lithologies. Many of the well-stratified, relatively thin beds show intense bioturbation (*Thalassinoides*), indications of firm- and hardgrounds, as well as numerous paleosol horizons. Facies varies from open lagoons with local shoals to tidal flats. The lowest part of the Vions Formation shows lagoonal sediments with four horizons of paleosol formation pointing to rapid loss of accommodation. The facies evolution then indicates a short opening with high-energy shoals and open lagoons. Important emersive facies occurs again at 50 to 57 meters. Isotope analysis (see below) also confirms that this interval was strongly influenced by continental processes. Renewed lagoonal sedimentation with elevated content in echinoderms and less siliciclastics then again attests for open marine conditions punctuated only by minor subaerial exposure. Paleosols, coal beds, and increased siliciclastic influence are the last well-documented signs of subaerial exposure and near-coastal environments at the top of the Vions Formation.

The overlying, massive white limestones just below the top of Mount Salève belong to the Lower Chambotte Formation. The abrupt change to massive beds with high-energy facies indicates opening of the sedimentary system towards the Tethys ocean and implies high accommodation. The base of the Formation is dominated by deposition of ooid shoals that locally show vadose cementation. Subsequently, thick-bedded, lagoonal sediments indicate highest accommodation. The uppermost, thinner-bedded sediments of the Chambotte Formation again represent high-energy shoal to coarse-beach deposits and display abundant keystone vugs. Here again, loss in accommodation and shoaling is evidenced by two deepening-shallowing sequences with clear evidence for subaerial exposure (Fig. 5.23, 5.24). The last bed of the Chambotte Formation is a very well sorted, fine-grained, bioclastic grainstone with small-scale stratification and suggests an aeolian depositional environment (Kindler et al. 1997).

The abrupt change to the following, yellowish, echinoderm-rich, oolitic limestones with weak siliciclastic influence (Guiers Member) is marked by a slightly dolomitized hardground. Sediments are interpreted as open to external lagoonal deposits, below fair-weather wave base, but probably with influence of deep tidal currents. A breccia at 99 meters, originally explained as beachrock (Deville 1991) is interpreted as reworking of a subtidal firm- to hardground, since no signs of intertidal or early meteoric diagenesis are present. The generally discontinuous sedimentation in the Guiers Member points to low accumulation rates, probably due to repeated reworking by episodic high-energy events (storms) and a lower rate of carbonate production (negatively influenced by elevated siliciclastic input).

The oolitic, echinoderm-rich limestones in the last ten meters of the section are interpreted as subtidal ooid bars on an open to external platform. A major facies change is observed three meters below the top of the section, where a bryozoan- and crinoid-rich limestone indicates another rapid opening. The typical red-colored facies with quartz nodules is regionally attributed to the Calcaire Roux Formation. It implies that the oolitic bars below represent an atypical Upper Chambotte Formation. However, in the Salève section no biostratigraphic markers supporting this hypothesis were found.

Isotope analysis

Analysis of stable isotopes of carbon and oxygen was carried out in two parts of Salève section in order to confirm sedimentological interpretation of subaerial exposure and compare the influence of soilgas and evaporation in the Vions and Chambotte Formations (Fig. 5.24).

In the Vions Formation drastic negative shifts of $\delta^{13}C$ of up to 4‰ (PDB) with simultaneous positive shifts of $\delta^{18}O$ (sample Sa 7 and Sa 7.4) clearly evidence subaerial exposure with strong influence of soil gas (Allen & Matthews 1982, Joachimski 1991). Similar evolution of isotope ratios is observed in samples Sa 8.1 and Sa 9.7. Here root traces sedimentologically confirm the environmental significance of the isotope values. Samples Sa 5.5 and Sa 8.5, however, show less-evident isotope signatures. Considering the extreme shifts which occur in the entire interval, variations of 1‰ seem negligible but can point to exposure of shorter duration and/or less intense colonization by land plants. The generally irregular curve and strong shifts below an exposure indicate low pedogenetic homogenization of isotopic values by soil gas and point to generally short-lived subaerial exposure (Joachimski 1994).

The Lower Chambotte Formation is characterized by important positive shifts in $d^{18}O$ but comparably low variations in $d^{13}C$. Here, evaporation possibly was more important and colonization by land plants was absent or played only a minor role. Positive shifts of 1 to 2 ‰ in $d^{18}O$ are observed in samples Sa 115.1, Sa 127u, Sa 130, Sa 142, and Sa 146. Negative shifts in $d^{13}C$ of less than 1‰ in Sa 127u and values of 0 to 1‰ PDB are not sufficient evidence for soilgas influence and may simply reflect facies variations (Joachimski 1994). Considering the low over-all variability of $d^{13}C$, the total shift between sample Sa 142 and 145 of more than 1 ‰ may be significant and indicate subaerial exposure with minor plant colonization at the top of the Chambotte Formation. Strong parallel shifts of $d^{13}C$ and $d^{18}O$ (sample Sa 148 to Sa 153)

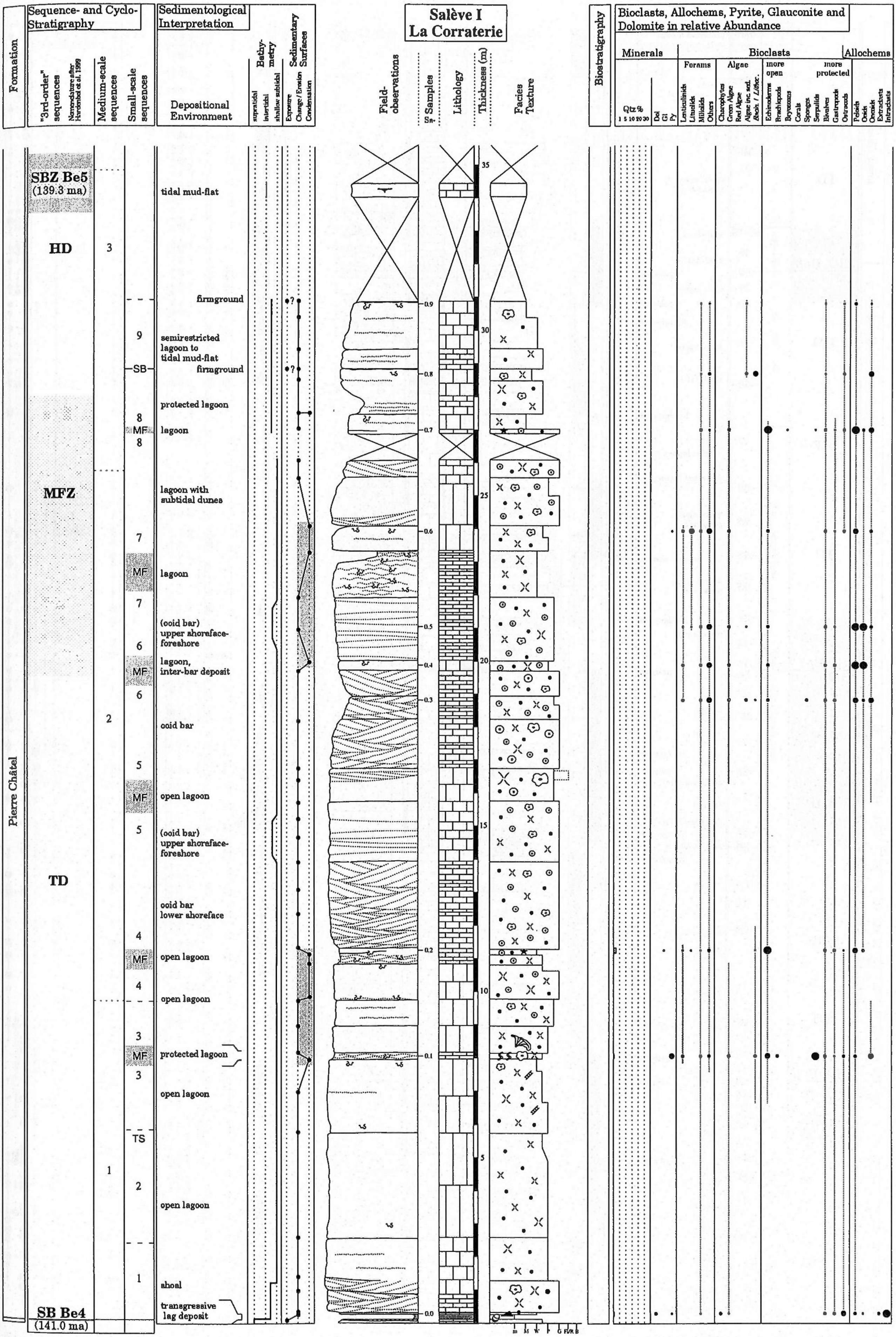


Fig. 5.22a. Salève section (part I).

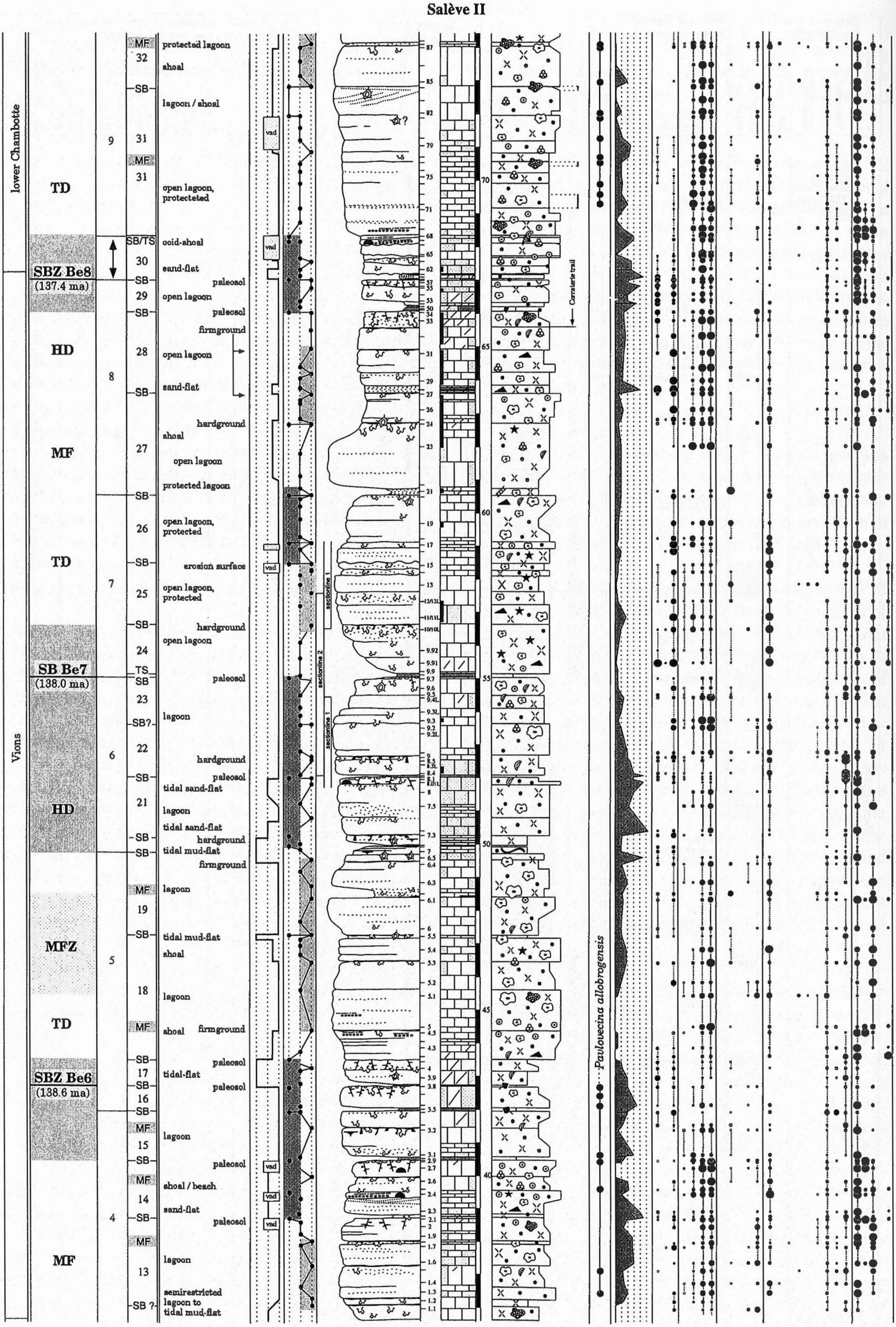


Fig. 5.22b. Salève section (part II).

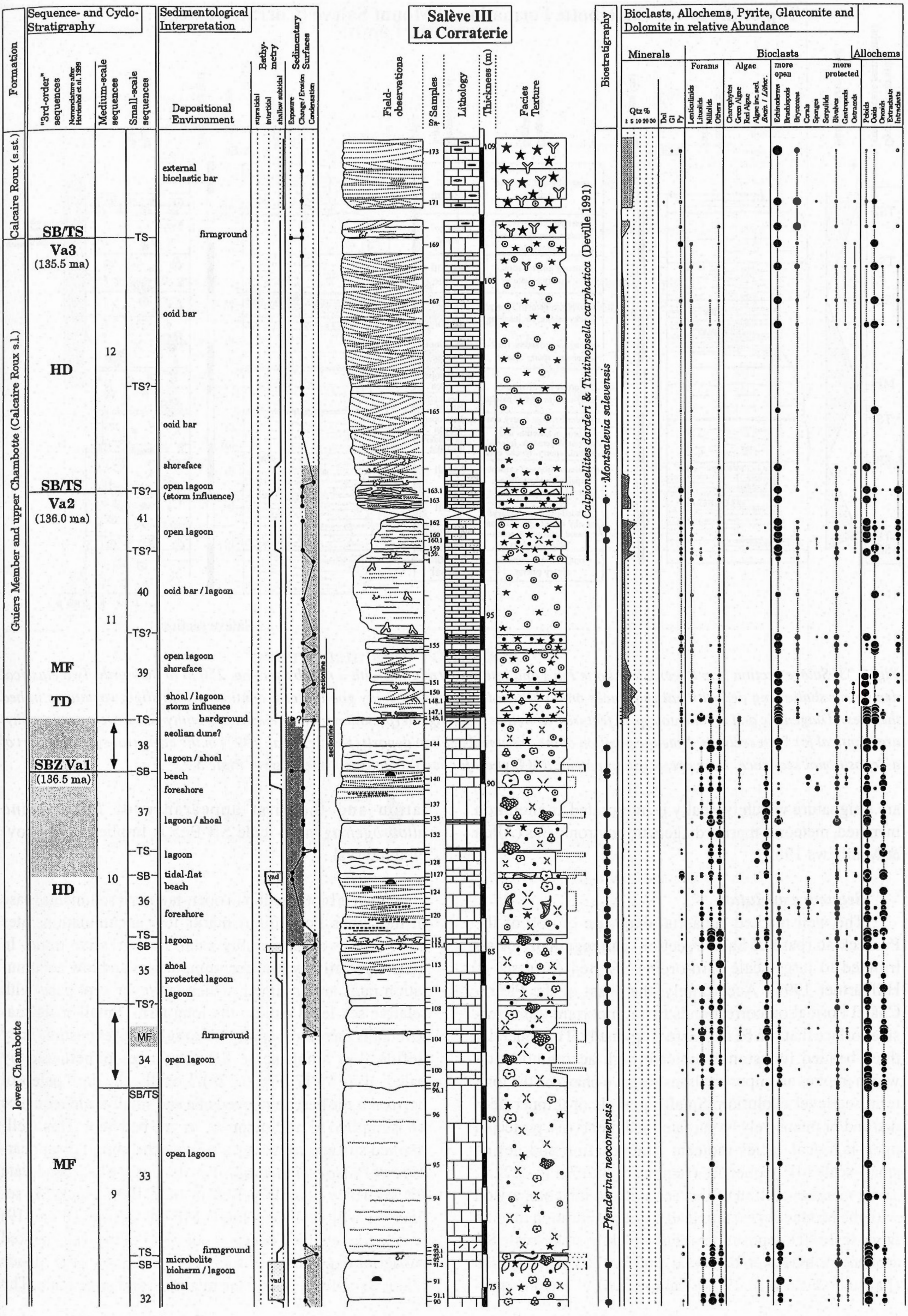


Fig. 5.22c. Salève section (part III).

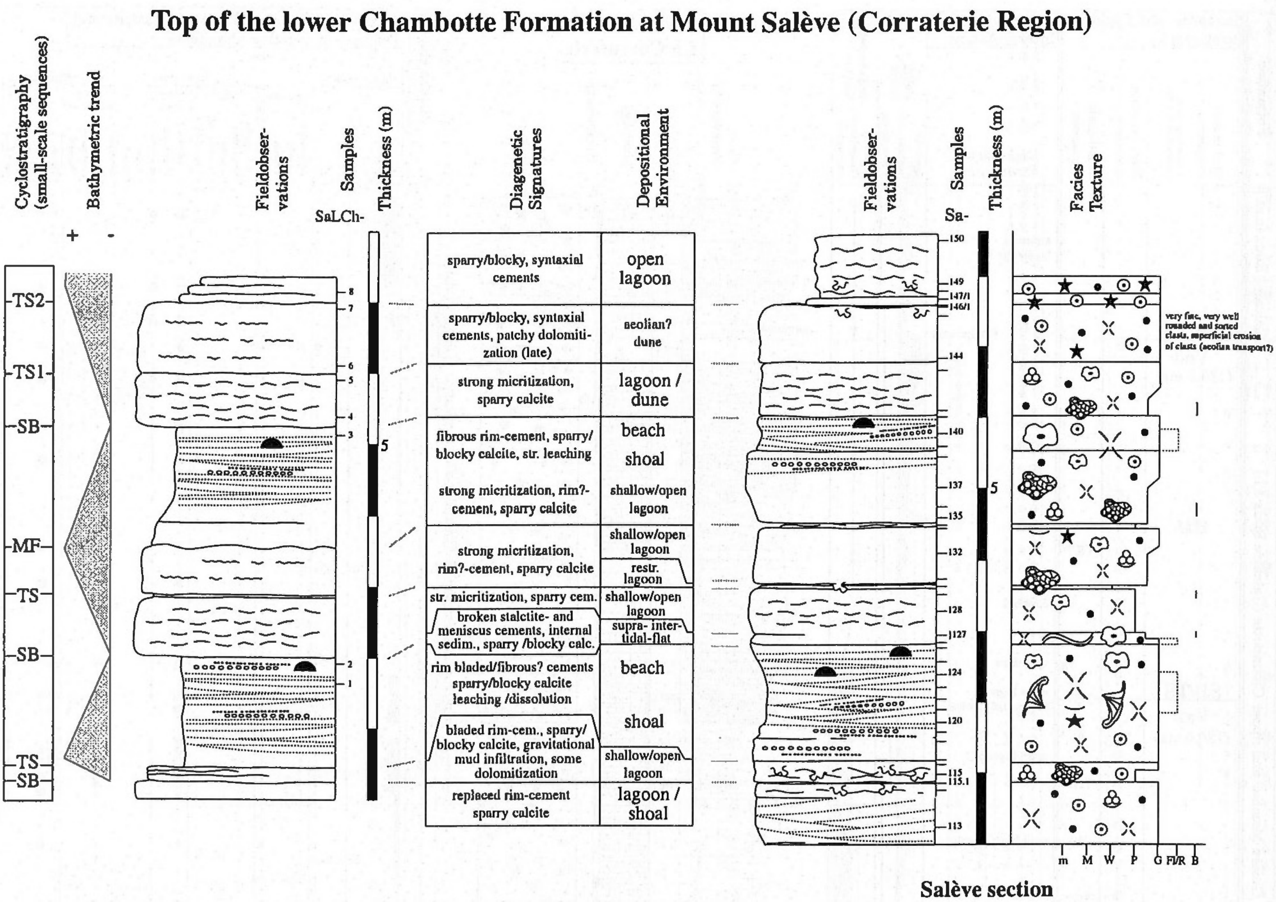


Fig. 5.23. Salève section (top lower Chambotte Fm.) and lateral correlation with a nearby section, 250 m to the north. Two classical deepening-shallowing depositional sequences on a carbonate shoreline display good lateral continuity with slight variations in bed thickness. Diagenetic and sedimentological features, as well as resulting interpretation (dep. environment, bathymetry, cyclostratigraphy) are identical for both sections. Note the position of the presumed aeolian deposit (Kindler et al. 1997) in the early transgressive part of a depositional sequence. Here preservation potential is highest for such sediments. Refer also to Plate 5.7.

are a signature which typically is interpreted as alteration in mixed, meteoric-marine diagenetic environments (Allen & Matthews 1982).

Sequence evolution

The well-marked subaerial exposure on top of the Purbeckian (base of Salève section) can regionally be attributed to large-scale (3rd-order) SB Be4 (Strasser & Hillgärtner 1998). Accordingly, the onset of the Pierre-Châtel Formation corresponds to an important transgressive surface that can be correlated regionally. The generally thick-bedded limestones indicate high accommodation which implies an important transgressive phase on the long term sea level evolution. Small intervals of thinner-bedded and/or intensively bioturbated sediments of a protected open lagoonal facies indicate condensation and define small-scale MF-sequences (sequences 3 to 7, Fig. 5.22a). The general evolution of this section and the comparison with the Monnetier section lead to the assumption that the large-scale HD-deposits are relatively thin (small-scale SB sequence 9 and covered interval) and may locally be eroded (Pasquier & Strasser 1997). On the basis of lateral correlation and the first appearance of *Pavlovecina allobrogensis* large-scale SB Be5 is implied in the covered interval.

Intense bioturbation in open-lagoonal sediments (base of the Vions Formation) marks low accumulation rates and portrays a comparably small transgressive trend. In general, relatively low accommodation and low accumulation rates are assumed, which however kept pace with relative sea-level rise on the long term. Small-scale sea-level falls therefore rapidly led to subaerial exposure. They define four small-scale SB-sequences punctuated by paleosols (14 - 17, Fig. 5.22b). Locally, the third paleosol horizon is the best developed one and small-scale sequence 16 is implied to be condensed to one bed here. This well-marked surface coincides also with the abrupt disappearance of *Pavlovecina allobrogensis*. It is the best candidate for sequence boundary Be6 in SBZ Be6. Thick SB-sequences with well-developed MF (sequences 18 and 19) testify for regained accommodation before well-marked subaerial exposure surfaces (Fig. 5.22b, 5.24) evidence relative sea-level fall on the medium- and large scale. The

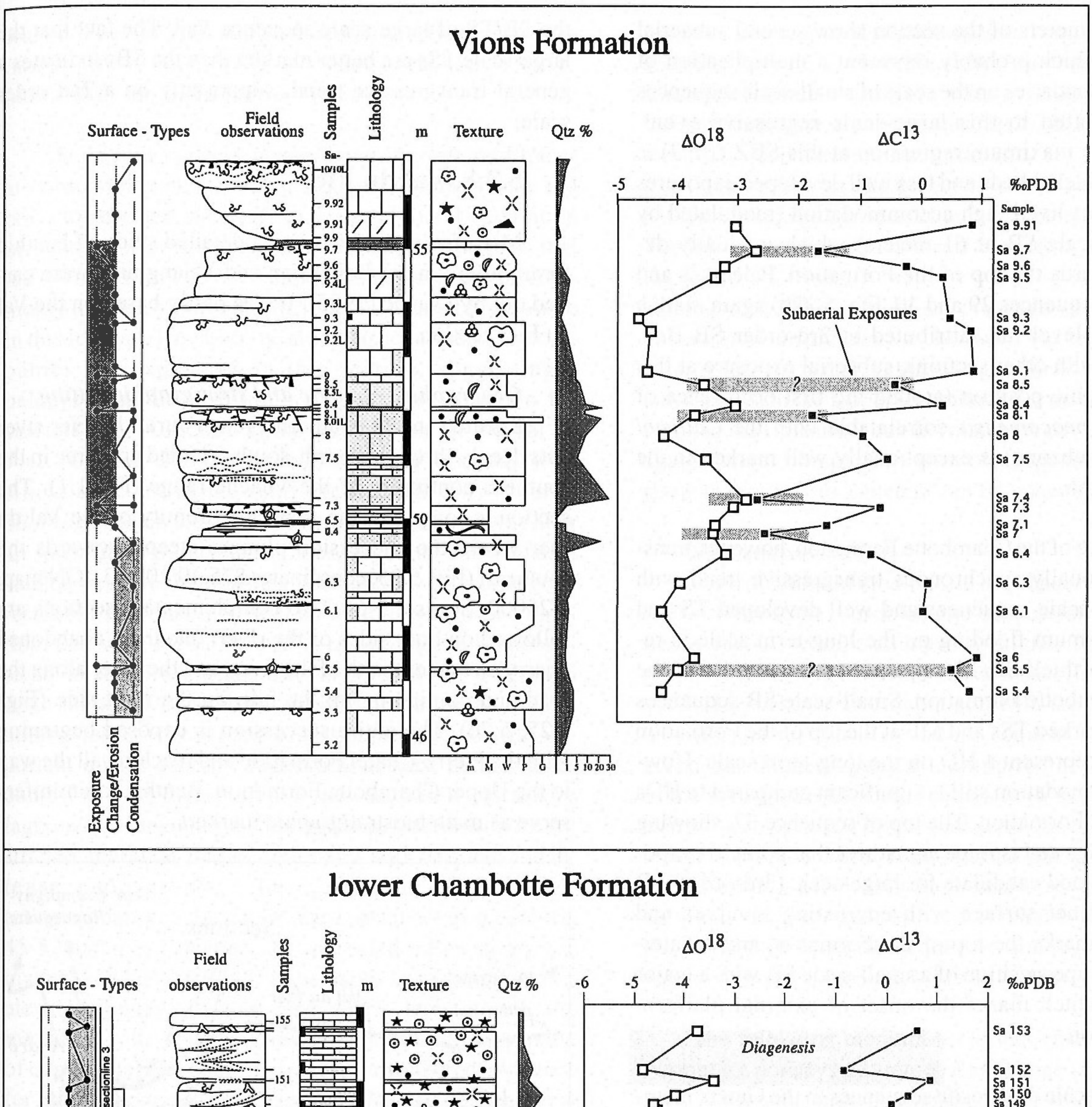


Fig. 5.24. Stable isotope values in selected intervals of the Salève section.

following 7 meters of the section show several subaerial exposures which probably represent a multiplication of sequence boundaries on the scale of small-scale sequences (20-24), related to this large-scale regressive event. Regionally a maximum regression at this SBZ (Be 7) is observed. Thicker beds and less well developed exposures indicate short-lived high accommodation (modulated by a medium-scale SB at 61 meters) which gradually decreases towards the top of the Formation. Paleosols and coal beds (sequences 29 and 30, Fig. 5.22b) again mark a relative sea-level fall, attributed to 3rd-order SB Be8. Compared with other sections, subaerial exposure at this biostratigraphic position (around the first occurrence of *Pfenderina neocomensis* correlatable with the *Callisto/ Alpillensis* subzone) is exceptionally well marked in the Salève section.

The base of the Chambotte Formation, however, translates a regionally synchronous transgressive trend with thick small-scale sequences and well developed TS and MF. A maximum flooding on the long-term scale is recorded by a thick SB/TS sequence in the middle of the Lower Chambotte Formation. Small-scale SB-sequences with well marked TSs and MF at the top of the Formation (Fig. 5.24) represent a HD on the long-term scale. However, accommodation still is significant compared to HDs in the Vions Formation. The top of sequence 37, showing keystone vugs and isotope signatures that point to evaporation, is a good candidate for large-scale (3rd-order) SB Va1. A distinct surface with encrusting bivalves and bioerosion marks the top of the Formation and is interpreted as a superposition of a small-scale SB with a major TS (Va1) which marks the onset of external platform sedimentation.

Small-scale composite sequences in the Guiers member show characteristics different from the ones observed in the Formations below. Typically, the base is made up of echinoderm-rich sediments which pass into oolitic sediments with HCS (sequence 41). They are interpreted as subtidal cycles (Osleger 1991). Their base indicates deposition below storm-wave base during rapid relative sea-level rise. The top, in contrast, is influenced by storms and indicates sedimentation during slowing relative sea-level rise and/or relative sea-level fall above storm-wave base. Small-scale sequences are less well marked than in the Formations below, probably due to dynamics intrinsic to these high-energy sedimentary environments. This is especially true for the thick ooid bars at the top of the section, which are only delimited by a level of intense reworking at the base and by a firmground at the top. The reworking is interpreted as relative lowering of wave base during larger-scale relative sea-level fall. According to biostratigraphic markers (Fig 5.22c), this level is correlated with SB/TS Va2. The iron-stained surface at the top, where the change to external platform sedimentation (Calcaire Roux Formation) takes place, is interpreted as the SB/TS of large-scale sequence Va3. The fact that the large-scale TSs are better marked than the SBs indicates a general transgressive trend, supposedly on a 2nd-order scale.

5.3.8. Val du Fier

Microfacies analysis and a detailed study of benthic foraminifers in the Berriasian and Valanginian were carried out by Darsac (1983). It was partly based on the Val du Fier section.

Geographic, geologic and stratigraphic setting

Located in the southern French Jura, the Fier river cuts deeply through a north-south oriented anticline in the southern continuity of the Vuache range (Fig 1.1). The section is logged at the eastern extremity of the Val du Fier where the succession plunges steeply towards the southeast (Fig. 5.25, coordinates 875.40/109.63, IGN map, 1/25000, Seyssel 5-6). Two complementary sections are followed on both sides of the river: one in an abandoned quarry on the left bank of the river, and the other along the road and continuing up the hill, on the right side (Figs 5.25, 5.26). The entire succession is exposed beginning with the Pierre-Châtel Formation and reaching all the way to the Upper Chambotte Formation. Benthic foraminifers serve as main biostratigraphic markers.

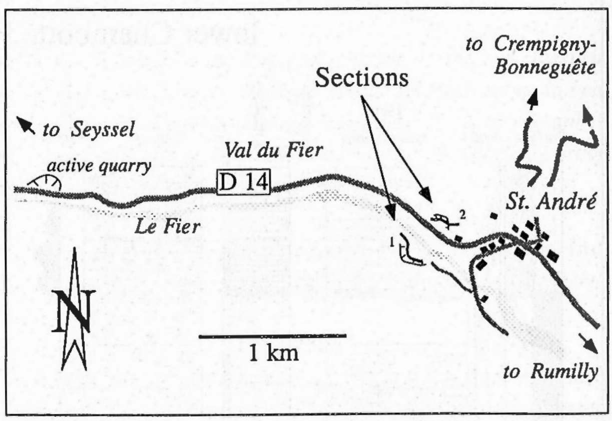


Fig. 5.25. Location of the Val du Fier section.

Sedimentological interpretation

The base of the section is covered, but lateral comparison with the cliffs along the river show that it correlates approximately with the top of the Purbeckian. In contrast to other distal sections (Monnetier and Salève), the Pierre-Châtel Formation in the Val du Fier displays no high-energy ooid bars. Here, the depositional environment is lagoonal, with open-lagoonal deposits at the base and dominantly protected open-lagoonal sediments towards the top of the Formation. Abrupt changes from mud- and wackestones to grainstones essentially reflect variations in water movement, probably related to changes in wave-base and/or modified barrier systems.

At 28 to 32 meters several horizons with root traces are observed in shallow to restricted lagoonal facies and point to repeated subaerial exposure. Lateral correlation of bundles of exposure surfaces (chapter 6) suggest that the change from Pierre-Châtel to Vions Formations, commonly marked by siliciclastic input, is located somewhere in this interval. However, no siliciclastic influence accompanies the exposures in this section, suggesting a paleogeographic position with a bypass situation for siliciclastics. A section at the western end of the Val du Fier (Darsac 1983) displays a larger interval with siliciclastic influence (and presence of *Pavlovecina allobrogensis*) in the same stratigraphic position. Only after an interval of purely calcareous, lagoonal limestones a sudden peak of siliciclastics, associated with a change to restricted lagoonal/intertidal conditions, paleosol formation, and with appearance of *P. allobrogensis* is encountered in the present section (at 39 meters). Although open-lagoonal conditions already prevailed before, *P. allobrogensis* is restricted to this short interval with important quartz content, suggesting at least partial facies control over its distribution. Renewed, purely calcareous lagoonal sedimentation then again passes into more restricted, marginal-marine sediments with frequent and intense pedogenesis. These deposits witness low accommodation potential in near-coastal environments (at 45-52 meters). However, the increased influx in organic matter (black facies) and siliciclastics is accompanied by elevated content of echinoderm debris, lenticulinids, and intraclasts, all suggesting reworking and intense mixing of continental and open-marine particles. Direct evidence for tidal currents (as in Crozet and Chapeau de Gendarme sections) is not observed in this section, which points to less accommodation and lower accumulation rates, or to a different paleogeographic position.

The change to open-lagoonal carbonate sedimentation of the Lower Chambotte Formation is transitional and marked by gradual decrease in siliciclastics and organic matter. Open-marine influence and condensation are indicated by sediments rich in brachiopods and increasing abundance of benthic foraminifers (at 62 meters). The massive limestone beds suggest elevated accommodation rates, which are interrupted only once by discontinuous sedimentation and weak siliciclastic input in the middle of the Formation (at 72-74 meters). Shoaling is evidenced by coarse beach deposits and vadose diagenesis (e.g., pendant cements) at the top of the Formation.

A drastic change in facies to sponge- and bryozoan-rich, marly sediments with intense bioturbation (*Thalassinoides*) marks the onset of the Guiers Member. Rapid deepening and significant condensation with contemporaneous input of siliciclastics and organic matter are recorded here. This probably reflects a change to a more seasonal climate, witnessed also by storm deposits in slightly shallower environments (HCS, at 85 and 98 meters). The depositional environments are situated on a proximal ramp around storm-wave base. Here, a bathymetric range between 20 and 40 meters is interpreted. The top of the section is characterized by open to external lagoonal sediments with a varying influence of siliciclastics. The continuous presence of *Montsalevia salevensis* indicates a Lower Valanginian age (*Campylotoxus* zone). Lithostratigraphically this interval corresponds to the Upper Chambotte Formation. The transition to the Calcaire Roux Formation is covered, but once again, lateral comparison with outcrops along the river show that only 1 to 2 meters can be missing.

Sequence evolution

Identification of small-scale sequences in the Pierre-Châtel Formation is problematic because of the homogeneous facies. Abrupt changes to higher-energy facies, probably reflecting lowering of wave base, are interpreted as SB/TS surfaces. In comparison with the stacking pattern of other sections, five SB/TS sequences, which progressively thicken upwards are inferred in the basal part of the section. Small-scale sequences (6 to 8) are characterized by intervals of discontinuous sedimentation in protected lagoonal environments, that are interpreted as MFs. They define a MFZ on a large scale. As in most other sections, the large-scale HD is comparably small and depicted by four small SB-sequences and rapid loss of accommodation (sequences 9-12, SBZ Be5).

The following medium-scale sequence in the Vions Formation portrays a symmetrical change in relative sea-level with four first thickening, then thinning SB sequences. It is punctuated by well-marked subaerial exposure. The best developed paleosol is the best candidate for large-scale SB Be6 in the SBZ, which is defined by three, very condensed SB-sequences. The next transgressive-regressive trend on the large scale is composed of 2 medium-scale sequences (5,6), whereby only the lower one reflects a clearly symmetrical relative sea-level trend. Medium-scale sequence 6 is composed of 4 small-scale SB sequences, all punctuated by subaerial exposures that become progressively more intense towards the top. This evolution of multiplied SBs culminates with a thick bed (two amalgamated sequences) topped by a paleosol with massive root trunks. It is the best candidate for SB Be7 in the large SBZ. By comparison with the stacking pattern in the other sections it becomes clear that the two relatively massive intervals in the following large-scale sequence (Be7) correspond to two medium-scale sequences. Here, however, they are composed of only three small-scale sequences each. Other small-scale sequences possibly are eroded at the base (in SBZ Be7) and/or amalgamated and condensed in the upper part and at the top of

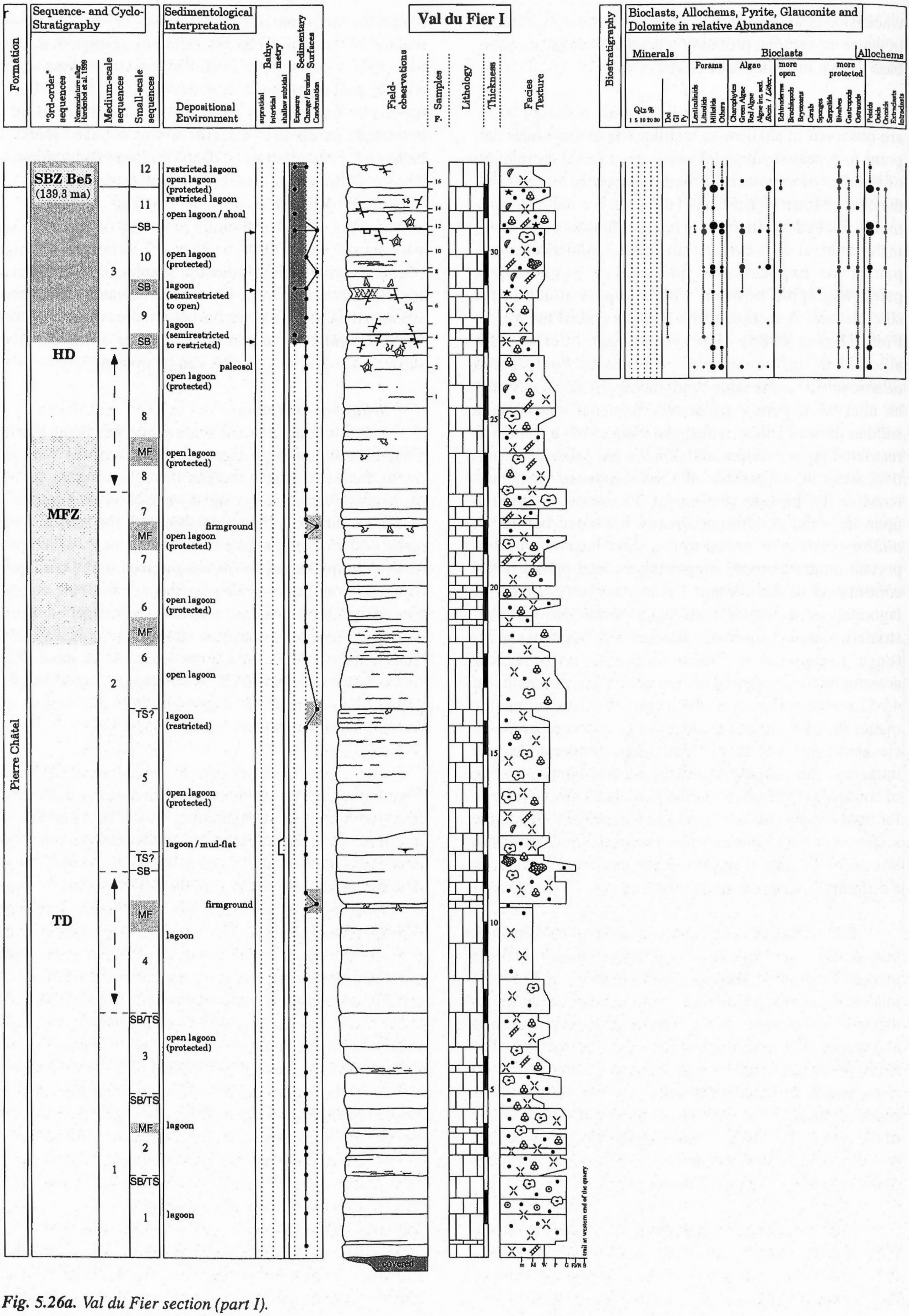


Fig. 5.26a. Val du Fier section (part I).

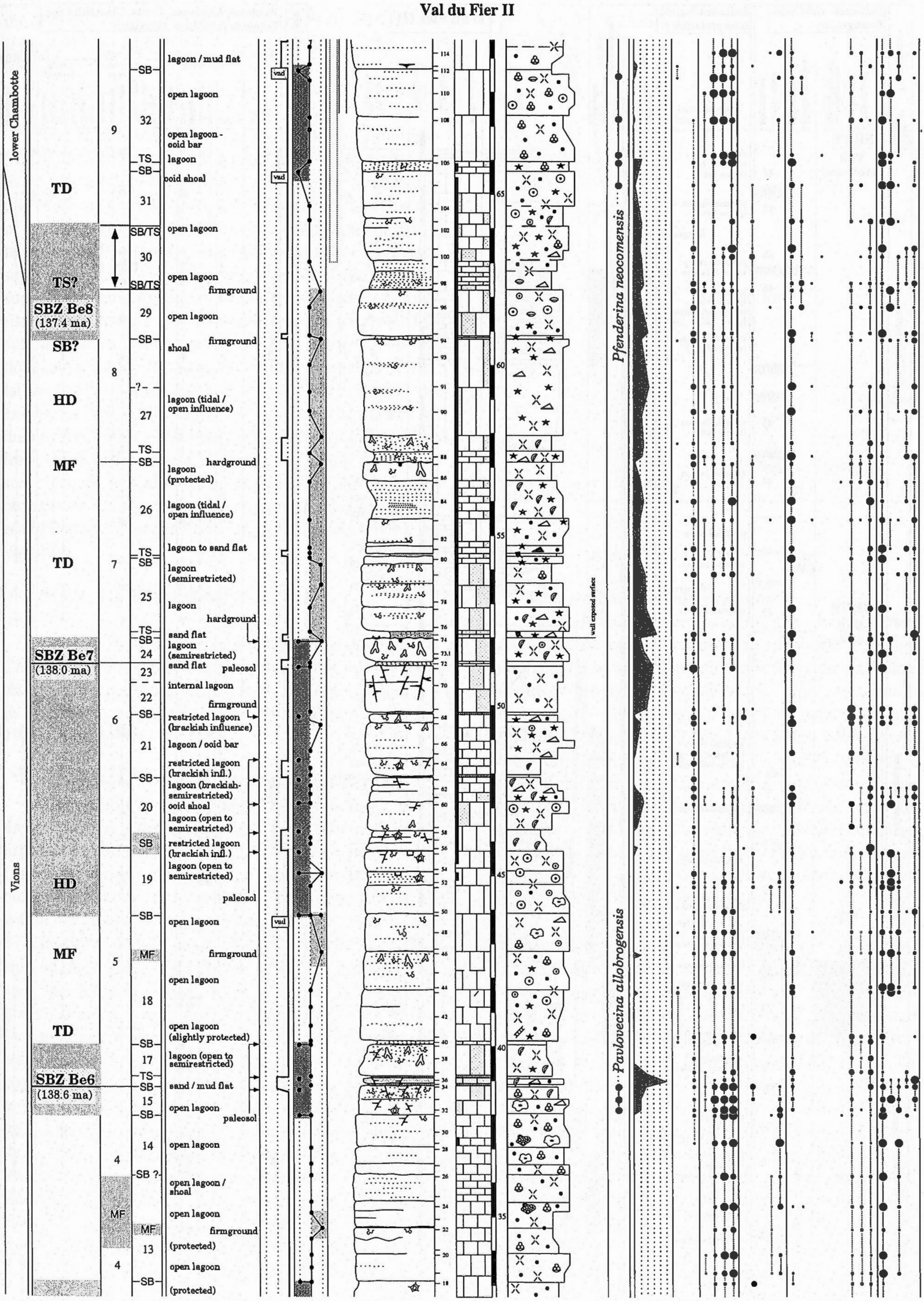


Fig. 5.26b. Val du Fier section (part II).

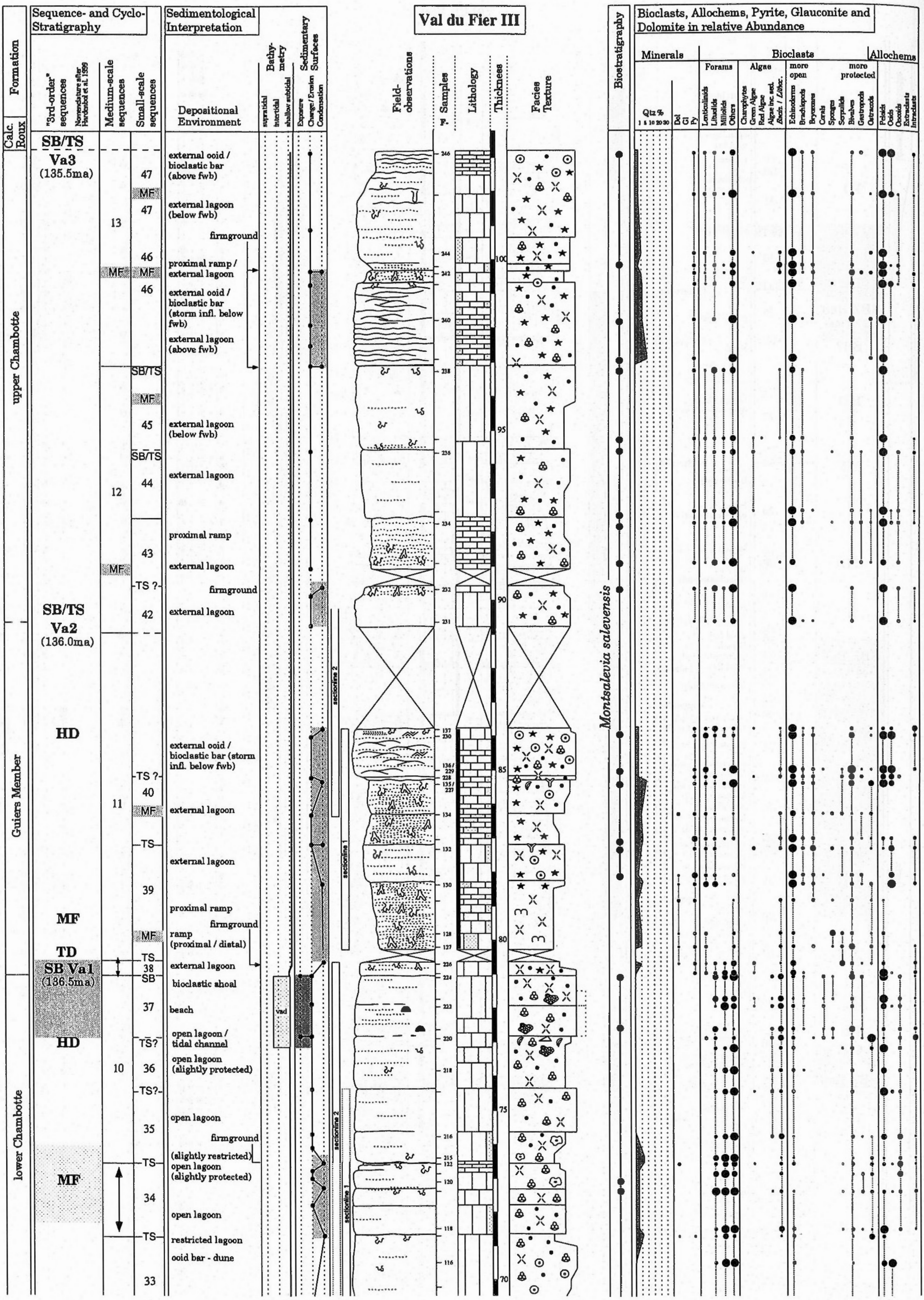


Fig. 5.26c. Val du Fier section (part III).

the interval that is associated with large-scale SB/TS Be8. It has to be noted that SB Be8, in contrast to Salève and Monnetier sections, is not marked by subaerial exposure at all, but by condensation in open-marine environments.

The large transgressive trend, already beginning at SBZ Be7, is continued with thicker small-scale sequences in the Lower Chambotte Formation. They continue to display well-expressed TSs and MFs. The shoaling at the top of the Formation testifies for loss of accommodation in a large-scale HD. However the well-expressed TS of large-scale sequence Va1 and the immediately following MF suggest condensation during an important relative sea-level rise. The following two TS-sequences are interpreted as subtidal cycles, comparable to the ones in the Salève section. Large-scale SB Va2 is only inferred by lateral correlation and biostratigraphic data. No field observations except for the abrupt change to pure limestone-facies indicate an environmental change. Bathymetric changes are less evident in the Upper Chambotte Formation. Yet, small-scale sequences inferred on the basis of subtle facies changes and condensation discontinuities suggest a generally transgressive trend on a large-scale relative sea-level variation.

5.3.9. La Chambotte

The La Chambotte section is the type section for the Chambotte Formation. It has been subject of various studies, mainly concerning facies, litho-, and biostratigraphy (e.g., Steinhauser 1969, Darsac 1983, Charollais et al. 1992, Blanc 1996).

Geographic, geologic and stratigraphic setting

The section is situated about 20 km to the south of the Val du Fier on the western flank of the same Jura range, high above Bourget lake (Fig. 1.1). Large parts of the Lower Cretaceous, including the "Urgonian facies", crop out along a small road (N 491B), that climbs steeply towards the village of La Chambotte (Fig 5.27, coordinates 874.350/93.450, IGN map, 1/25000, Rumilly 5-6). The studied interval is logged in two parts: first along the road and, in contrast to other studies, following up a steep, grassy slope to avoid tectonic disturbance (section-line 1). The upper part of the section again follows the roadside (section-line 2, Fig. 5.27). Marly intervals tend to be covered in the section, but all such intervals were dug out to obtain unaltered sample material and continuous control of stacking pattern and facies evolution. The base of the section is constituted of dolomitized carbonates attributed to the lower Pierre-Châtel Formation. The top is defined by the change to the Calcaire Roux Formation. Biostratigraphy is established on the basis of benthic foraminifers and a rare occurrence of calpionellids (Fig. 5.28).

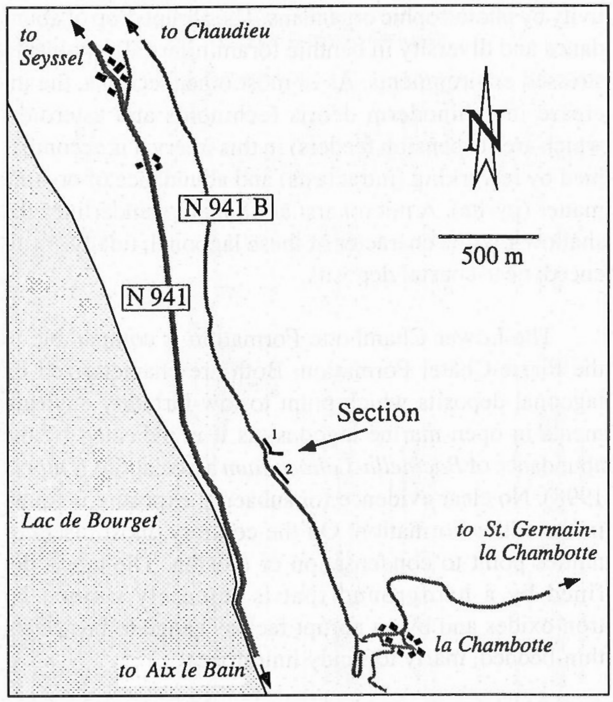


Fig. 5.27. Location of the La Chambotte section.

Sedimentological interpretation

The Pierre-Châtel Formation in the Chambotte section is characterized by a dominance of moderate- to high-energy lagoonal sedimentation. In several intervals, large oncoids and abundant *Bacinella-Lithocodium* associations point to open-marine conditions with reduced sedimentation rates. This is also indicated by an abundance of condensation discontinuities (at 10-11, and 16 meters). A morphological change from a massive, protruding limestone cliff to thin-bedded, morphologically retreating sediments points to an environmental change. These high-energy deposits at the top of the Formation indicate shoaling by foreshore/beach sediments (horizontal lamination, rapidly changing grain sizes).

The change to the Vions Formation is marked by a generally more discontinuous sedimentation, a color change from light-beige to beige-brown, and the beginning siliciclastic influence. The lithological change is transitional and, therefore the base of the Formation is defined at the first appearance of *Pavlovecina allobrogensis*. The subsequent marly interval with restricted lagoonal to intertidal facies is marked by paleosol formation and peak-abundance of siliciclastics and particulate organic matter. It gives evidence for a first, major emersive phase in the *Paramimounum* ammonite zone. Renewed lagoonal sedimentation, which still is characterized by stratigraphic condensation, then again evolves into restricted-lagoonal to tidal deposits with dominant microbial mats. This microbial activity probably is related to renewed influx of siliciclastics and nutrients that stress carbonate productivity by phototrophic organisms. The abrupt drop of abundance and diversity in benthic foraminifers also points to stressed environments. As in most other sections, the increase in echinoderm debris (echinoids and asteroids, which are suspension feeders) in this interval is accompanied by reworking (intraclasts) and abundance of organic matter (pyrite). A microkarst at 48 meters underlines the shallow-marine character of these lagoonal, tidally-influenced, near-coastal deposits.

The Lower Chambotte Formation is comparable to the Pierre-Châtel Formation. Both are characterized by lagoonal deposits which point to low-turbidity environments in open-marine lagoons, as it is indicated by the abundance of *Bacinella-Lithocodium* associations (Dupraz 1998). No clear evidence for subaerial exposure is found in the entire Formation. On the contrary, most discontinuities point to condensation or erosion. The top is defined by a hardground that is intensely stained by iron-oxides and by an abrupt facies change to brownish, thin-bedded, marly to sandy limestones.

Although in many respects similar to the upper Vions Formation (color, content in siliciclastics, echinoderms, intraclasts, and pyrite), the continuing abundance and diversity of foraminifers and storm influence (HCS) in the Guiers Member suggest a position in a deeper, more open-marine environment. The very discontinuous sedimentation gives way to entirely calcareous, lagoonal, sediments attributed to the Upper Chambotte Formation. The disappearance of siliciclastics and lower contents in echinoderm debris despite of the open-marine character of the deposits can be interpreted in different ways. They suggest either environmental changes of climatic origin (lower seasonality), or a bypass situation for siliciclastics by which nutrient-poorer oligotrophic conditions prevail that are less favorable for abundant growth of echinoderms. Storm deposits and more external sediments in the topmost part of the section already announce the drastic environmental change to deep, external platform conditions (offshore) of the Calcaire Roux Formation. The abrupt change is denoted by an intensely stained hardground witnessing major condensation.

Sequence evolution

For the Pierre-Châtel Formation sequence analysis can only be carried out for the upper two thirds, where a succession of small-scale MF-sequences testifies for an important rise in accommodation on the large scale (sequences 2-5). The following shoaling is associated with a short HD and large-scale SBZ (Be5, small-scale SB-sequences 6-8). In contrast to most other sections, the interval appears to be condensed (considering thickness) but does not evidence significant subaerial exposure.

Slightly thicker SB-sequences culminate in a major loss of accommodation attributed to SBZ Be6, just above the last appearance of *Pavlovecina allobrogensis*. The following interval displays a symmetrical transgressive-regressive evolution on the medium scale (sequence 5). On the basis of lateral comparison of stacking pattern and facies evolution it is assumed that the entire zone of large-scale SBZ Be7, which is composed of restricted lagoonal to tidal-flat environments (medium-scale sequence 6 and adjacent small-scale sequences), is very condensed. Similarly, medium-scale sequences 7 and 8, delimited by the most prominent discontinuities in the interval, display a relatively reduced thickness. Although facies suggests an opening of the sedimentary system, accommodation apparently remained low to moderate at this particular platform position. SB/TS Be8 is inferred in the marly- sandy interval and marked by an important condensation surface.

Accommodation begins to increase with the onset of the Chambotte Formation and with thicker small-scale sequences characterized by well-developed TSs and MFs. The large-scale transgressive trend is only slightly modulated on the medium scale (slightly marly, discontinuous sedimentation). Large-scale SB/TS Va1 is only indicated by a deeply cut erosion surface. The major change in facies at the top of the Formation is interpreted as multiplied TS, which points to a rapid rise in relative sea-level on the large scale.

The following small-scale sequences are interpreted as subtidal shallowing-up cycles (Osleger 1991), delimited by SB/TS surfaces. In contrast to the Vions Formation, the Guiers Member and the Upper Chambotte Formation are thicker than in other sections. They display more small-scale sequences than elsewhere, which points to less important condensation and/or erosion, and higher accommodation. In the transition between these two lithostratigraphic units the discrimination of small-scale sequences is difficult and carried out in comparison with adjacent sections. Bounding surfaces are interpreted to mark amalgamated SBs and TSs. In the same way, large-scale SB/TS Va2 can only be inferred and is placed at the top of the Guiers Member, where a major facies change takes place. A moderate, relatively constant accommodation rate on a larger-scale transgressive relative sea-level trend is indicated by the constant thickness of SB/TS sequences in the Upper Chambotte Formation. The indicator for shallowest facies (channelization during base level drop?) is attributed to SB/TS Va3. The following small-scale sequence (47) does not really correspond to a LD and rather results from a duplication of TS Va3.

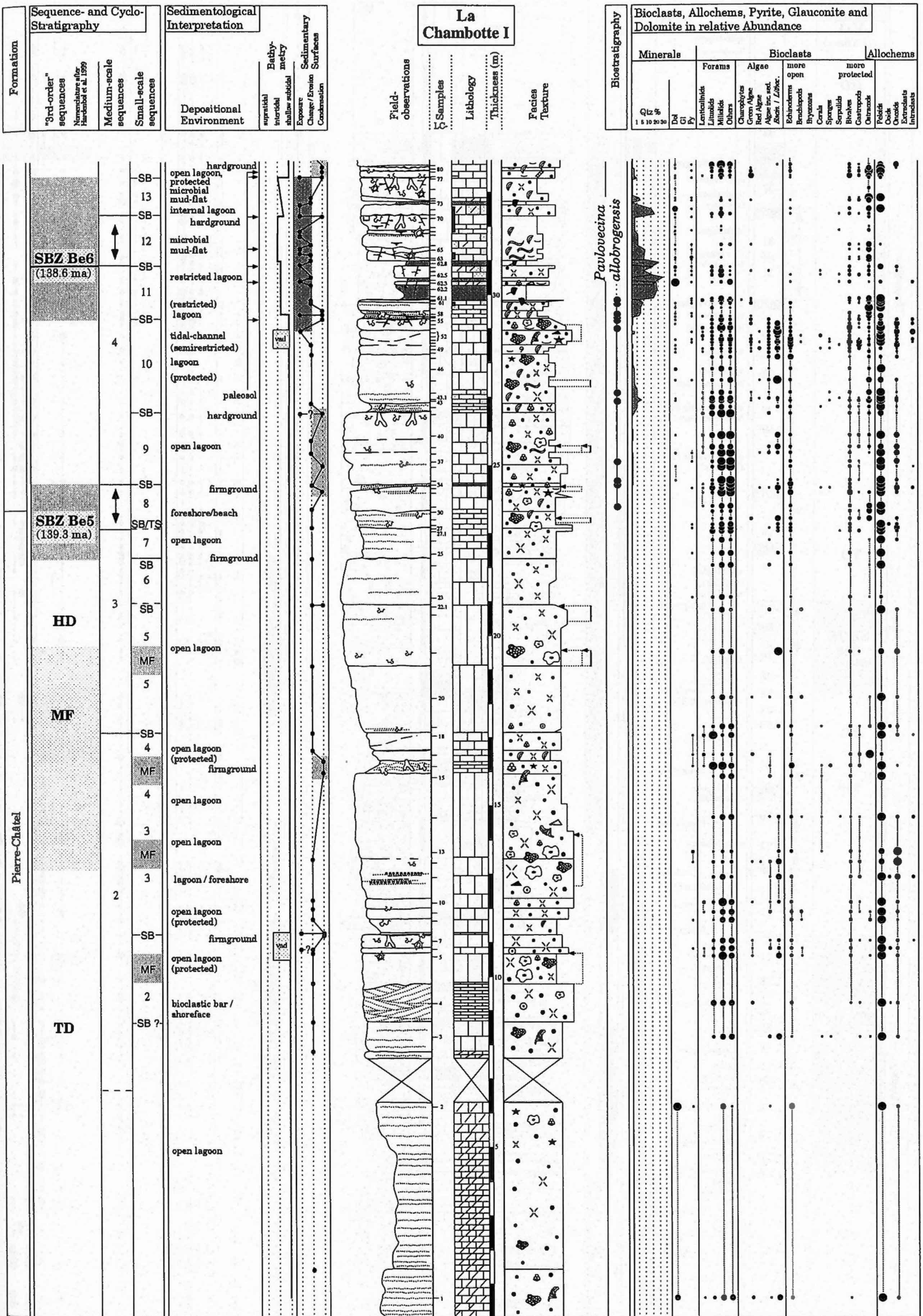


Fig. 5.28a. La Chambotte section (part I).

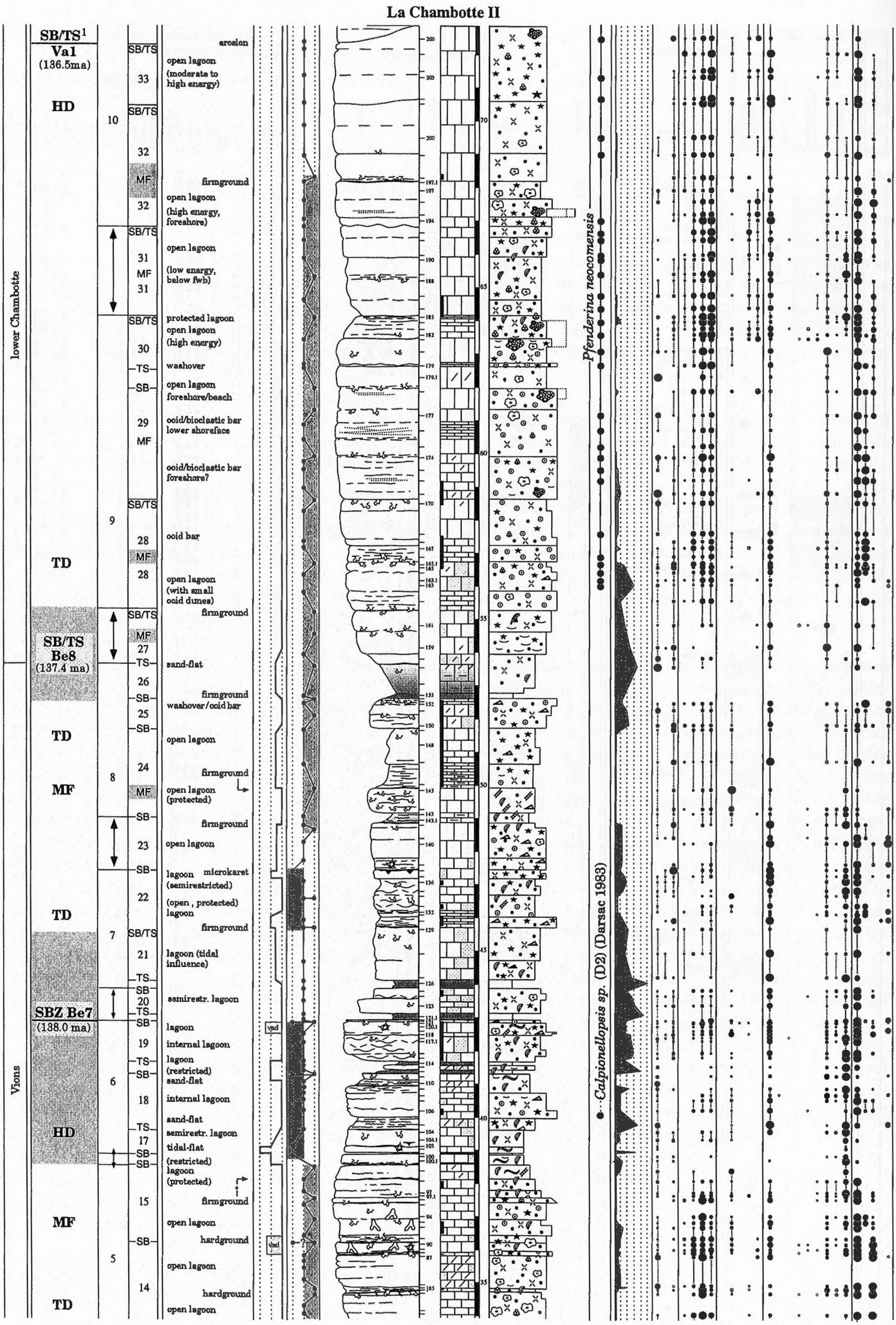


Fig. 5.28b. La Chambotte section (part II).

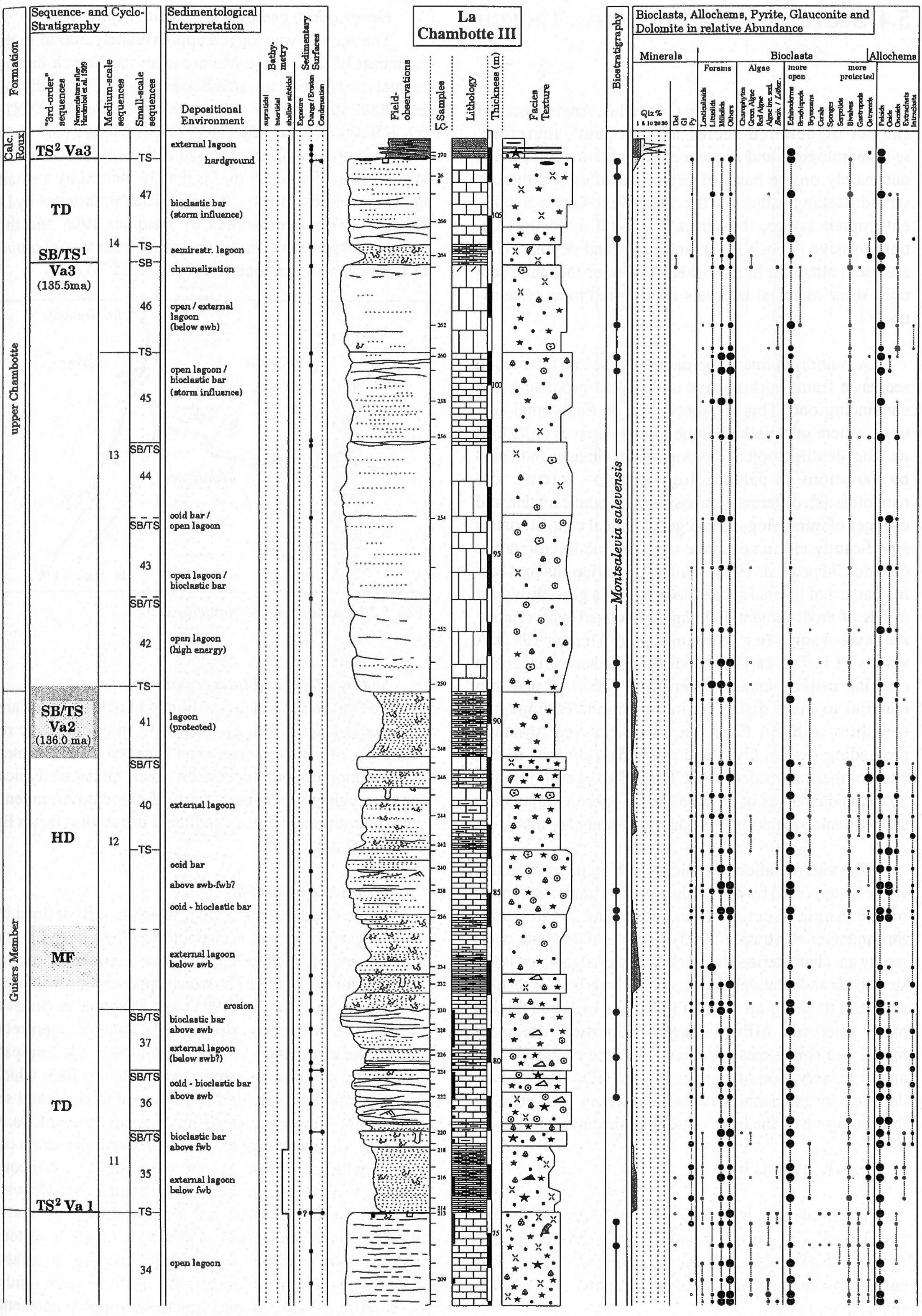


Fig. 5.28c. La Chambotte section (part III).

5.4. Sections in the Vocontian Trough

w The two basinal sections (Montclus, Angles) are not subject of detailed facies analysis and, therefore, sedimentological and sequence interpretation are carried out mainly on the basis of existing studies and the observed stacking pattern. In the objective to keep a consistent nomenclature, the terms lowstand deposit (LD), transgressive deposit (TD), and highstand deposit (HD) are used, although in published studies of the same sections some identical intervals are referred to as systems-tracts.

A multidisciplinary approach in the search for a sound sequence framework always is the most promising and convincing one. This is especially true for basinal sections, where one method alone quickly arrives at its limits. The identification of environmental changes indicated by variations in paleoecology (macro-, micro-, and nannofossils), different composition of organic matter, and changes of mineralogical and geochemical characteristics significantly aid in sequence-stratigraphic interpretation (Jan du Chêne et al. 1993). However, variations in stacking pattern of basinal successions can be a good proxy for trends of environmental change associated with relative sea-level changes (e.g., Strohmenger & Strasser 1993). A variety of factors can be responsible for such changes in stacking pattern, namely intensified export of platform material to more distal, basinal positions (slumps and turbidites in basin floor fan, slope fan and lowstand prograding wedge, highstand shedding), changes in calcite compensation depth (CCD) and oxygen-minimum zones, and changes of climate and/or oceanic circulation that influence planktonic productivity and clay input.

The interpretation of typical stacking patterns in this study is supported by a multidisciplinary analysis of parts of the Angles section (Jan du Chêne et al. 1993, Strohmenger & Strasser 1993). Lowstand deposits commonly are characterized by thick, often amalgamated limestone beds and relatively low content in marls. TDs display a general thinning-up trend of limestone beds and thicker marly interbeds. MFs display condensation, dominant marls, and sometimes a darker color due to enrichment and/or preservation of organic matter. HDs typically are depicted by limestone beds with constant thickness, or thickening-up on the large scale and thick marly intervals.

5.4.1. Montclus

The ammonite biostratigraphy of Montclus section is well studied by Le Hégarat (1971), and the basal part of the section (*Subalpina* to *Picteti* ammonite zones) was subject to sequence- and cyclostratigraphic interpretation by Pasquier (1995).

Geographic, geologic and stratigraphic setting

The section was logged approximately 500 m to the northeast of the village Montclus on the French departmental road (D994) towards Serres (Fig. 5.29, coordinates 863.400/240.850, IGN topographic map, Serres XXXII-39, 1/50.000, cf. Le Hégarat 1971). The lower part of the section crops out along the road and then follows up the steep flanks of a valley that is deeply incised by a small, ephemeral creek. Biostratigraphic data (ammonites by Le Hégarat 1971, calpionellids by Pasquier 1995 and this study, see Appendix 2) indicate a range from the *Subalpina* to *Campylotoxus* ammonite zones (Fig. 5.30 a, b).

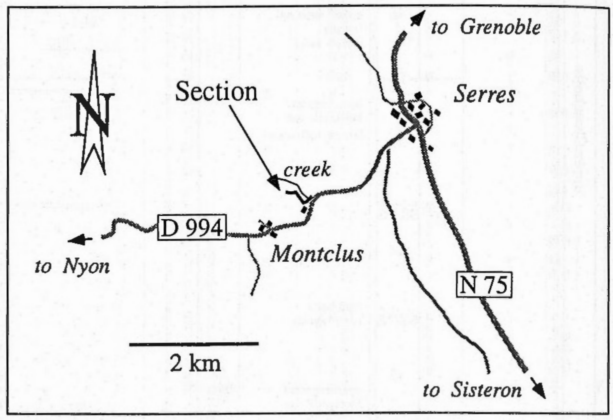


Fig. 5.29. Location of the Montclus section.

Sedimentological interpretation

The entire section is constituted of pelagic marls and limestones (Fig. 5.30a, b). Only few, thin, bioclastic resediments of platform origin and slumped intervals interrupt the monotonous succession. Such facies are typical for hemipelagic to pelagic base-of-slope environments with intermittent phases of sediment instability (facies Bp 1-4; Fig. 2.5c).

Sequence evolution

Large-scale sequence Be4, which is well defined by biostratigraphic means, is characterized by a large LD and thin TD and HD. Three medium-scale sequences are interpreted for the LD and TD, which opposes four sequences interpreted by Pasquier (1995) and Pasquier & Strasser (1997). The difference arises from a changed interpretation of the elementary sequences and their stacking-pattern within small-scale sequences. Sequence Be5, which is eroded and/or condensed in a slumped interval, and sequence Be6 show characteristics of amalgamated LDs. A first well-visible change to more marly sedimentation coincides with a TS in large-scale sequence Be7. A second change, well visible in the field, introduces a yellowish color to the succession, and a change to thick marly interbeds with homogeneous thickness of the limestone beds. This indicates generally higher accumulation rates and possibly a transgressive trend on a large scale (under the condition that each marl-limestone couplet represents

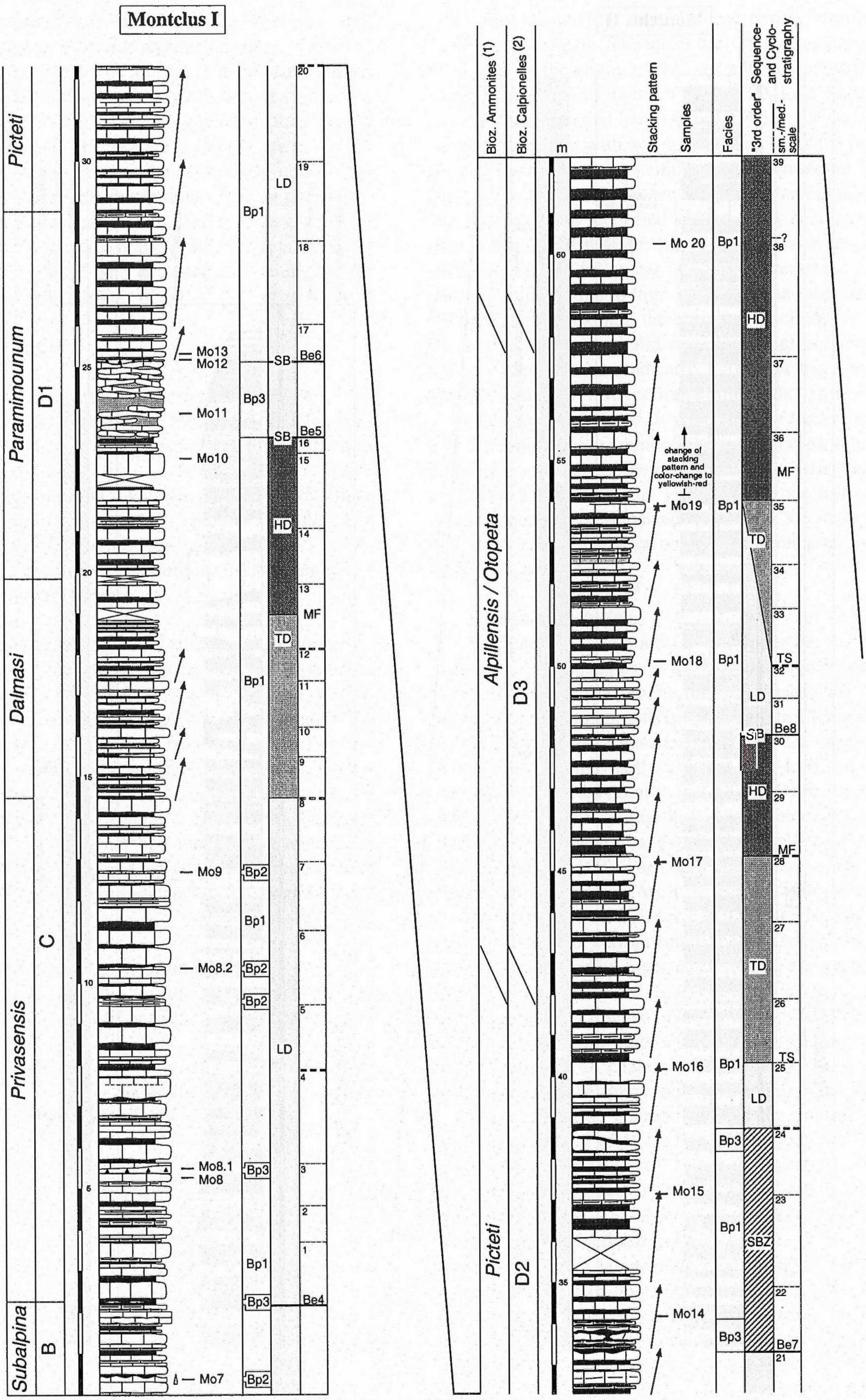


Fig. 5.30a. Montclus section (part I).

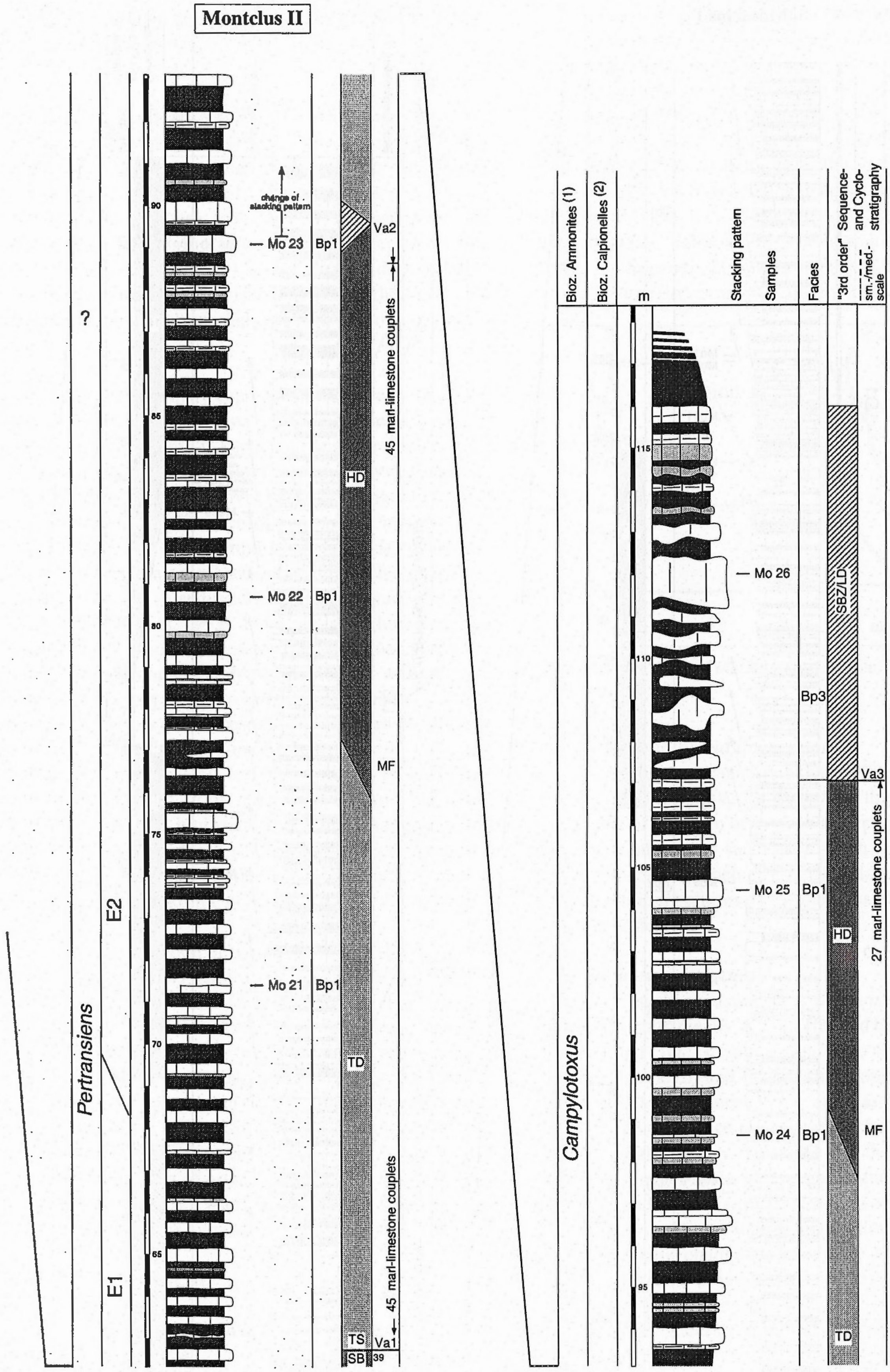


Fig. 5.30b. Montclus section (part II).

the same amount of time; see Chapter 7). Whereas a well-marked change in stacking style at the base of a thick, calcareous interval can be attributed to SB Be8, SB/TS and MF of sequence Va1 can only be inferred at positions with discreetly abnormal stacking pattern (thick marls, slump). SB/TS and MF Va2 are marked more obviously with thick limestone beds and a thick marly interval, respectively. Although biostratigraphic data are poor in the upper part of the section, and the *Verrucosum* zone is not identified, the base of the massive slump can presumably be correlated with SB Va3 and probably belongs to the upper *Campylotoxus* zone.

5.4.2. Angles

The Angles section is the hypostratotype of the Valanginian stage (Busnardo & Thieuloy 1979). It has been extensively studied in many respects of bio-, sequence- and cyclostratigraphy (e.g., Le Hégarat & Ferry 1990, Atrops & Reboulet 1993, Quesne & Ferry 1995, Blanc 1996, Reboulet & Atrops 1997, and part of Bull. Cent. Rech.-Expl. Elf-Aquitaine, Pau-SNPA, vol. 17/1, 1993, with several publications on multidisciplinary sequence-stratigraphic interpretation).

Geographic, geologic and stratigraphic setting

The section can be found along the small departmental road (D 33) to the village of Angles (Fig. 5.31). It begins at the intersection with French National Road N 202 and continues for about 600 m towards the east. The base is formed by massive limestones of Early Berriasian age (*Jacobi/ Grandis* zone) age and continues all the way up to the Hauterivian (Le Hégarat & Ferry 1990, Busnardo et al. 1979). The original bed numbers as applied by Busnardo et al. (1979), and which are continuously used in other publications, are indicated in the logged section (Fig. 5.32a, b).

Sedimentological interpretation

Facies is comparable to the Montclus section. The basal part of Angles section displays a concentration of

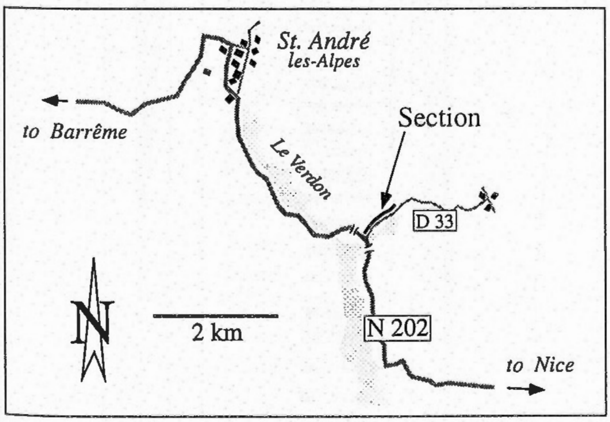


Fig. 5.31. Location of the Angles section.

calcareous mass-flow deposits that reveal important biostratigraphic gaps. The upper part of calpionellid zone B, as well as the entire zone C are missing below a big slump at the base of the section (bed 90, Le Hégarat & Ferry 1990). A second hiatus is observed in the slumped interval beginning with bed 120, and includes the upper D1 and lower D2 calpionellid zones. Calpionellid zone D3s (upper D3), partly constituting the *Otopeta* ammonite zone, was not identified and probably is condensed around bed 187 (Bulot, cf. Blanc 1996). The absence of ammonites in the following interval is interpreted as an effect of dilution due to high sedimentation rates (Blanc 1996). This may explain the abnormal thickness (10 m, Blanc 1996) of calpionellid zone D3t. High sedimentation rates are certainly indicated by thicker marly intervals and slightly thicker limestone beds. This trend continues and is reflected in progressively thicker intervals corresponding to ammonite zones of comparable duration (*Pertransiens, Campylotoxus* zones; Gradstein et al. 1995). The base of the *Verrucosum* zone follows a slumped interval and is located around bed 305 (Atrops & Reboulet 1993). It marks the onset of the dominantly marly Upper Valanginian.

Sequence evolution

Sequence interpretation for the Upper Berriasian is partly based on the results of the multidisciplinary analysis of characteristic geochemical, palynofacies, and nannofossil signatures, published in Jan du Chêne et al. (1993). Large-scale sequences Be4 to Be7 are characterized by stacked and partly amalgamated LDs. Slightly more marly intervals point to small TD/HDs in sequence Be4 and Be 6. Thickening-up trends of limestone beds reflect environmental changes on the small scale (beds 100-118 and 133-160). The continuously increasing content in marls (beginning with SB Be7) is interpreted as transgressive trend on a larger scale. SB Be8 is indicated only by a lenticular, current-rippled bed. Sea-level fall was probably attenuated by superimposition on a larger transgressive trend and did not lead to instabilities on the platform slope and deposition of mass flow deposits. However, the thick limestone beds and slightly thinning marl interbeds display all characteristics of a small lowstand wedge and/or higher planktonic productivity. The TD of sequence Be8 shows a modulation of the transgressive trend on a medium scale leading to a duplication of the TS. All following SBs (Va1 - Va3) are marked by small LDs with slumps directly followed by marl-rich intervals, interpreted as TDs. The absence of well-developed LDs and the generally increasing marl content suggest a transgressive trend on a larger scale (2nd-order) that is even intensified in the marl-dominated Hauterivian. The slumps do not contain platform detritus and indicate instabilities on the lower slope of the shelf or, more probably, instabilities in the morphologically structured basin itself (e.g., Pasquier 1995). Hence, they can also indicate episodic tectonic activity that triggered mass flows.

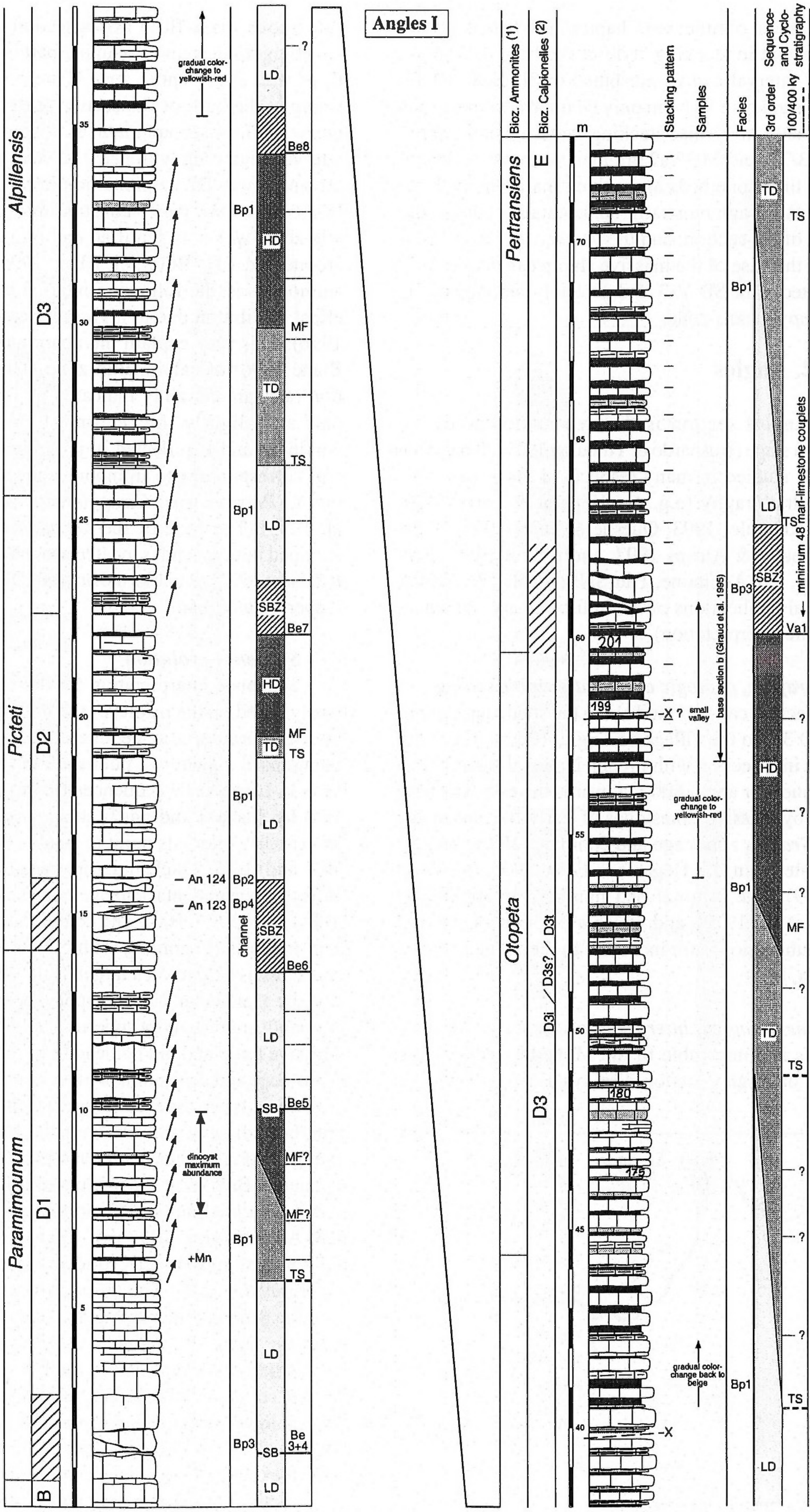


Fig. 5.32a. Angles section (part I).

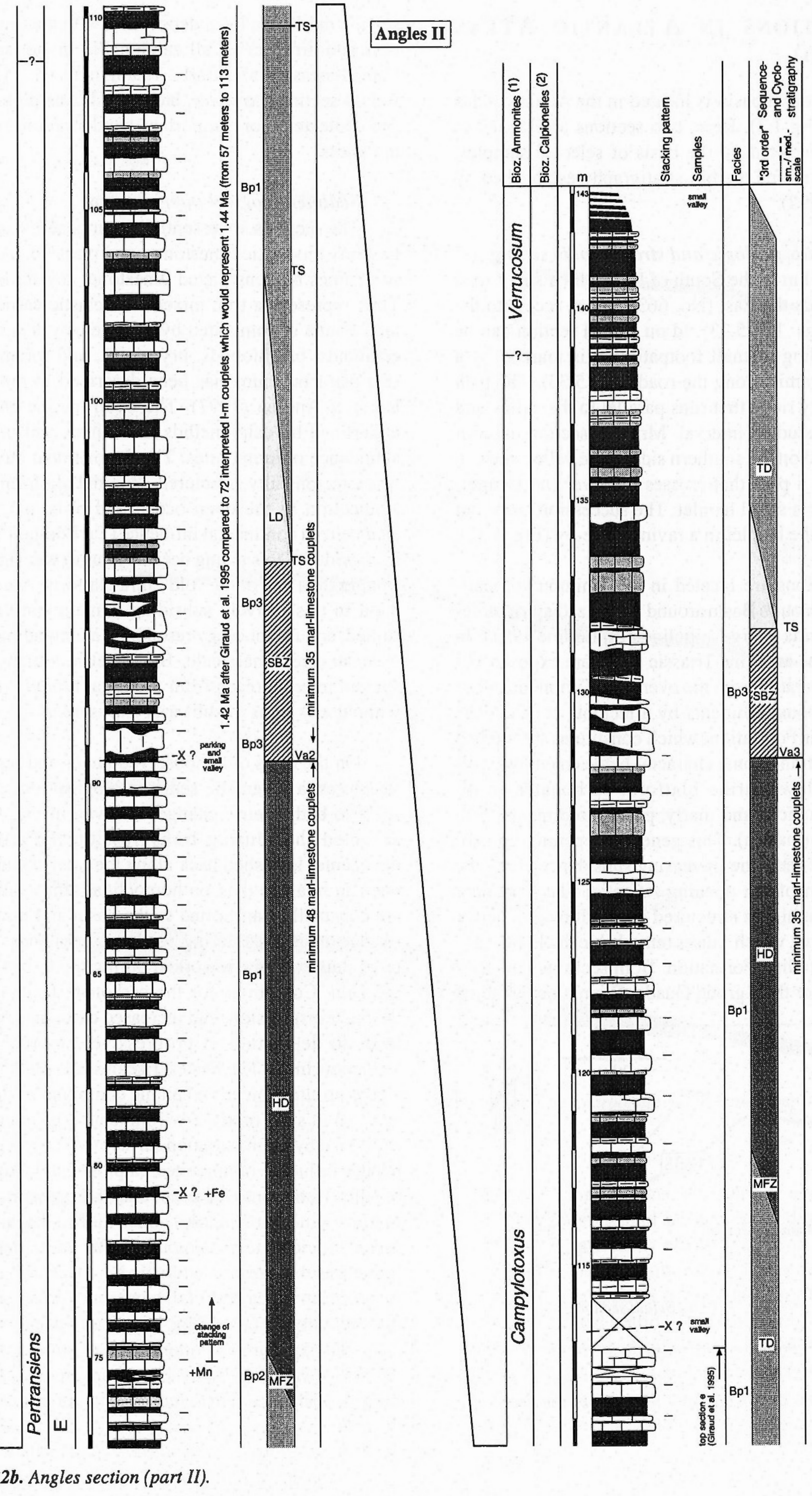


Fig. 5.32b. Angles section (part II).

5.5. Sections in Atlantic Atlas (Morocco)

The Essaouira basin is located in the Atlantic Atlas of Morocco (Fig 1.3). Here, two sections are logged in detail and interpreted on the basis of selected samples, stacking pattern, and the bio-stratigraphy established by Taj Eddine (1992).

Geographic, geologic and stratigraphic setting

About 20 km to the South of Essaouira a small road heading towards the East (No. 6638) gives access to the study area (Figs. 1.3, 5.33). Id ou Belaid section can be found descending a small footpath that is marked by a small rock pyramid along the road (Fig. 5.33). The path leads to a rocky ridge that runs parallel to the valley and makes up the studied interval. Mradma section, in contrast, is situated on the southern side of the valley and can be reached by a path that crosses the river (no bridge!) and leads up to a small hamlet. The succession crops out to the west of the hamlet in a ravine close-by (Fig. 5.33).

Both sections are located in the Smimou compartment of the Essaouira Basin around the Tidzi diapiric structure, which forms a large anticline (Taj Eddine 1992). In the center of the anticline Triassic series are exposed and commonly separated from the overlying Tithonian and/or Lower Cretaceous sediments by a tectonic contact. The Agroud Ouadar Formation, which constitutes the studied interval, has a transitional character between entirely calcareous, shallow-marine platform-carbonates (Cap-Tafelney Formation) and marly, pelagic sediments (Sidi Lhousseine Formation). This general deepening-up evolution partly reflects the increasing subsidence along the passive margin of the opening Atlantic. The Essaouira Basin itself is a highly structured and tectonically-active shallow platform, which causes remarkable thickness variations in the studied Formation. Both sections are reference sections for the Agroud Ouadar Formation. Mradma section marks the local depocenter with maximum thickness and presence of all calpionellid biohorizons in the Late Berriasian and Early Valanginian (B - D3). Id ou Belaid section, however, lacks the D2 calpionellid zone and indicates more important condensation by reduced thickness.

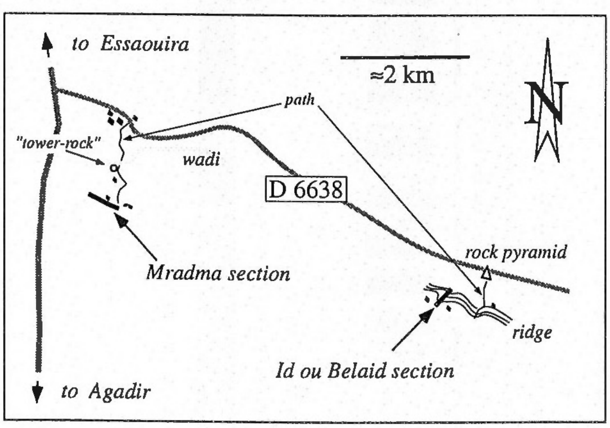


Fig. 5.33. Sketch illustrating the location of the Moroccan sections.

Sedimentological interpretation

The sections are essentially composed of marls, silty to sandy bioclastic limestones, and massive, bioclastic to sometimes bioconstructed limestones (oyster bioherms). They represent a true mixed siliciclastic-carbonate system. Fauna is dominated by bivalves, oysters, serpulids, echinoids, brachiopods, bryozoans, and foraminifers. A rich ostracod fauna has been described in the marls by Rossi & Andreu (1997). The open marine character is underlined by calpionellids, cadosines, and intermittent abundance of ammonites. The depositional environment was a tectonically structured, external platform with free connection to the open ocean and proximal, probably channelized continental influence (Taj Eddine 1992). Platform width of this young oceanic margin was significantly smaller than that of the "old" Jura platform. A steeper gradient in this actively subsiding domain can thus be assumed, and morphology resembled a homoclinal ramp, at least on a regional scale. Bathymetric variations in the studied interval reach from proximal ramp to deep ramp without any signs of subaerial exposure.

On the basis of the observed facies and sedimentary structures, a schematic sedimentological model with respect to bathymetric changes is given in Fig. 5.34. It is assumed that during transgression the siliciclastic depocenter is pushed back along the ramp. Conceptually, one can imagine than on homoclinal ramps siliciclastics are constantly transported to the basin and no important storage areas exist during sea-level lowstands. The sediment thus could be reworked during the following transgression. Conditions for the carbonate factory improve during transgression, and increased carbonate production leads to deposition of progressively more calcareous sediments in the TD. Fastest rise in sea level then leads to a give-up situation in the benthic communities that remain stressed in such mixed environments. This commonly is expressed by a condensation surface (MFS) and maximum pelagic influence (ammonites, calpionellids). Progradation of siliciclastics onto the deep ramp then generates initially marly (sometimes sterile), then more silty and finally mixed siliciclastic-calcareous (photic zone) depositional sequences that are characteristic for a HD. The maximum rate of relative sea-level fall commonly is not depicted by discontinuity surfaces. This is certainly due to the lack of subaerial exposure and attenuation of smaller-scale relative sea-level falls by a larger transgressive trend and higher subsidence rates. Such a trend, or at least a maxi-

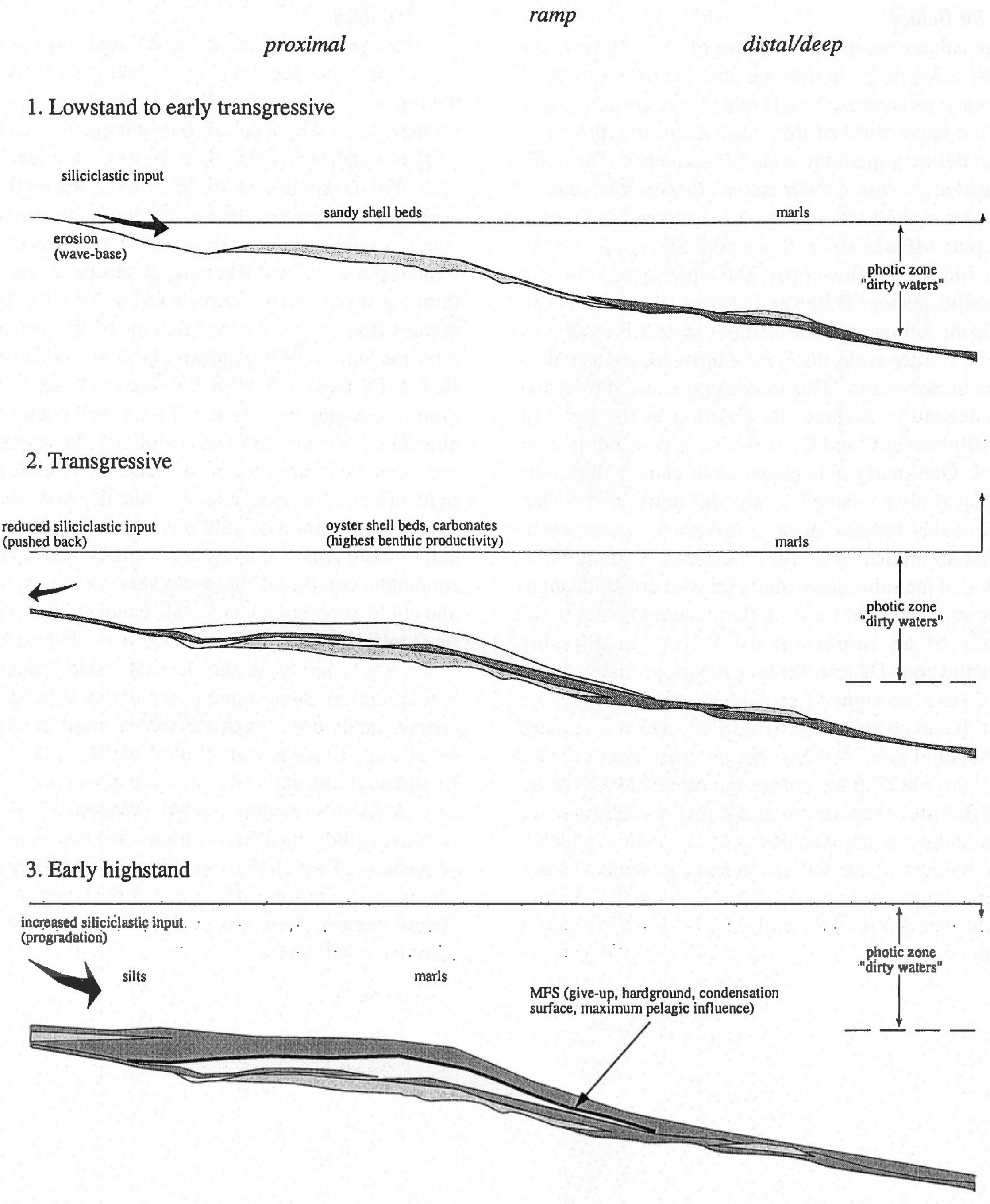


Fig. 5.34. Conceptual sedimentological model for the facies distribution on mixed carbonate-siliciclastic ramp environments of the Essaouira platform with respect to relative sea-level variations.

mum pelagic influence (ammonite, calpionellid abundance) can be observed in both sections in calpionellid zones C to basal D3.

Sequence evolution

Sequence analysis in the Moroccan sections is carried out only to the medium-scale level of depositional sequences. Limestone-marl alternations on the bed scale and on the scale of bed clusters are present in both sections, but sedimentary dynamics of the system are not known well enough to argue for an allocyclical control on these bedding rhythms. It cannot be excluded either, but more sections would be needed to distinguish between regionally and locally effective controls. However, on the medium scale, depositional sequences are punctuated by major condensation surfaces and correlate well between the two studied sections (see chapter 6). Commonly, they can be depicted in other parts of the basin as well (Taj Eddine 1992). The condensation surfaces are interpreted as expression of maximum flooding, and medium-scale sequences thus represent typical MF-sequences.

Id ou Belaid

The calcareous unit at the base of Id ou Belaid section (Fig. 5.35) nicely shows the possible overlapping of depositional sequences. Two levels of erosion at the base and in the upper third of this interval are interpreted as SBs that define a medium-scale SB-sequence. The MFS positioned at the top of the massive cross-bedded unit (at 5 meters) and the hardground (at 8.5 meters) define the overlapping MF-sequence. Since both SBs are located in calpionellid zone B (lower part and close to top) they are good candidates for SB Be3 and SB Be4 (Hardenbol et al. 1998). In the following, three medium-scale MF-sequences constitute a larger-scale, thickening-up trend, and a gradual increase in carbonates. This trend is punctuated by a major condensation surface. Its position at the limit of calpionellid zones C and D1 make it a good candidate for MF Be4. One marly MF-sequence in calpionellid zone D1 displays only a subdued, sandy SBZ (attributed to SBZ Be5), probably because of open (deeper-) marine conditions that are indicated by the abundance of ammonites. At the top of the subsequent marly interval drastic changes in color and disturbed bedding (laminations) point to the presence of an important discontinuity. Missing calpionellid zone D2 testifies for a hiatus, probably due to erosion, since no signs of prolonged non-deposition are evident. It can be attributed to SB Be8, which is situated in calpionellid zone D3. Significant erosion and drastic facies change at 27.5 meters are interpreted as MF of sequence Be8 and a superimposed SB on the medium scale. The major discontinuity at the top of the section, which is marked by intense mineralization and a pavement of ammonites (*Pertransiens* zone), witnesses condensation of superimposed SB, TS, and MF Va1 in the Early Valanginian.

Mradma

The general sedimentological and sequence evolution of Mradma section (Fig. 5.36a/b) is very similar to the one observed in Id ou Belaid. The base of the section is characterized by a calcareous unit and topped by well-marked condensation. It directly follows a level of erosion that is attributed to SB Be4. The well-marked condensation can thus be the effect of superposition of a smaller-scale MFS on a larger-scale TS. This large-scale trend is potrayed by thickening-up of calcareous beds and thinning-up of marly intervals and defines the TD of sequence Be4. Apparent modulations of this trend lead to three medium-scale sequences. The intervals between MFS Be4, MFS Be5, and MFS Be6 are made up of one medium-scale sequence each. MFs are well documented by abundance of ammonites and calpionellids, whereas SBZs are attenuated. Presence of calpionellid zone D2 (missing in Id ou Belaid section) allows to identify MFS Be6. However, manifestation of only one depositional sequence in calpionellid zone D2 suggests condensation and/or environmental conditions that were below a critical threshold and could not record small-scale changes. The reason can be significant bathymetry, which is evidenced by thick, green marls implying the deepest environments of the whole section. Subsequent progradation is punctuated by coarse, sandy deposits that evidence a fall in relative sea level and, thus, are attributed to SB Be8. Intensely bioturbated limestones located just above are interpreted as condensation around the MF. Sequence Be8, situated in calpionellid zone D3, is composed of two medium-scale sequences, of which the upper one is bounded by the same major condensation surface that is observed in the Id ou Belaid section. Here, it is interpreted as enhanced MF attributed to MF Va1.

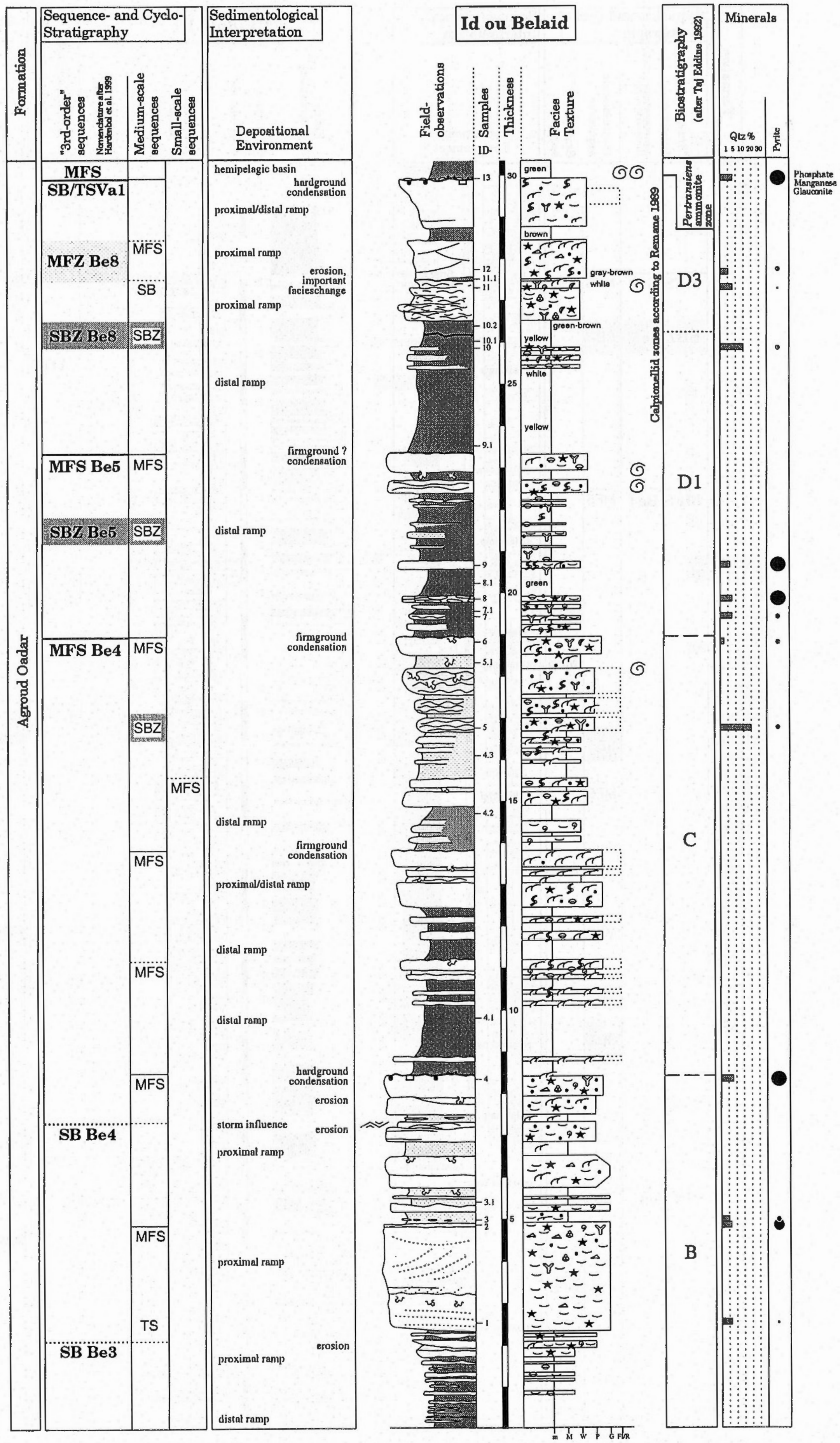


Fig. 5.35. Id ou Belaid section.

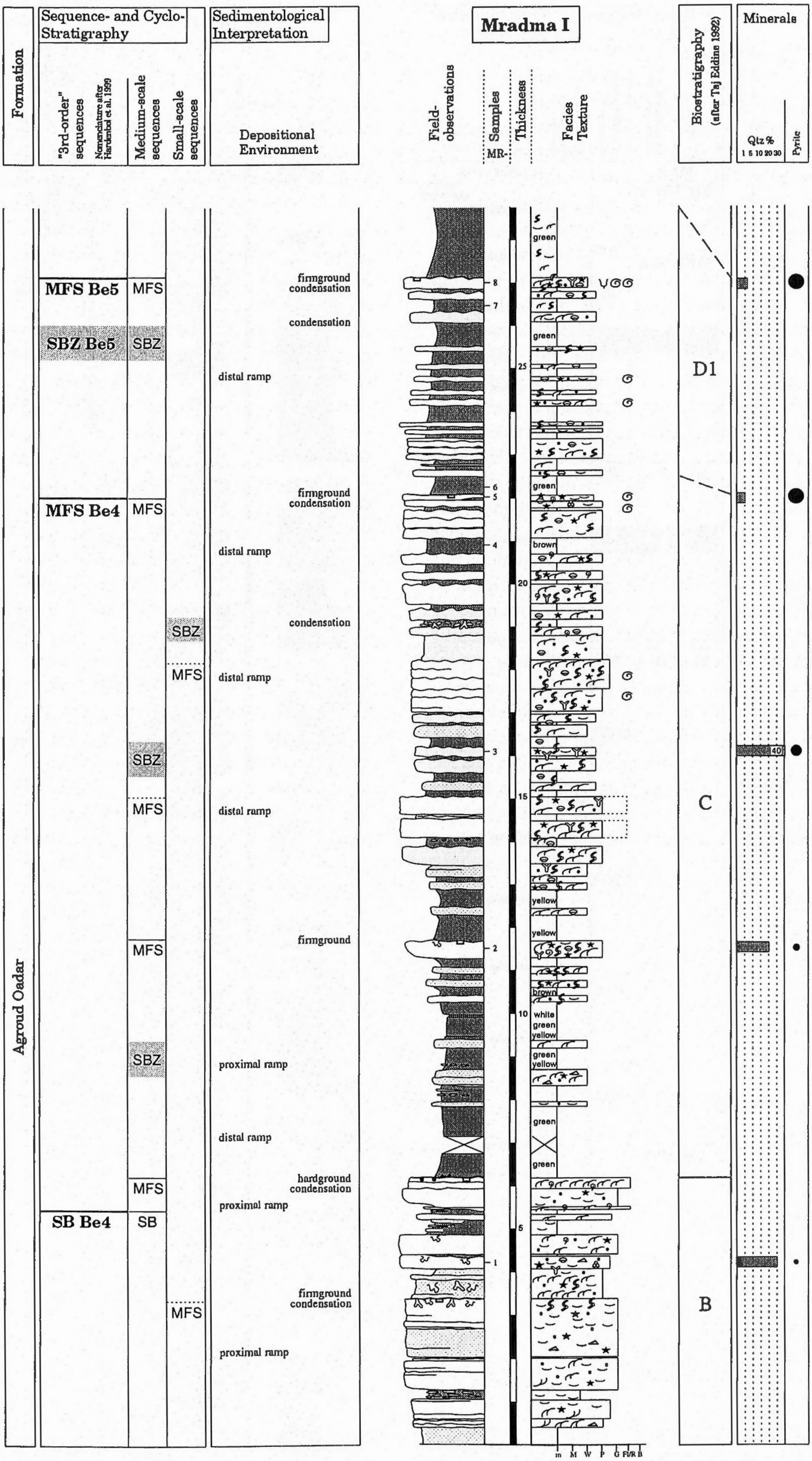


Fig. 5.36a. Mradma section (part I).

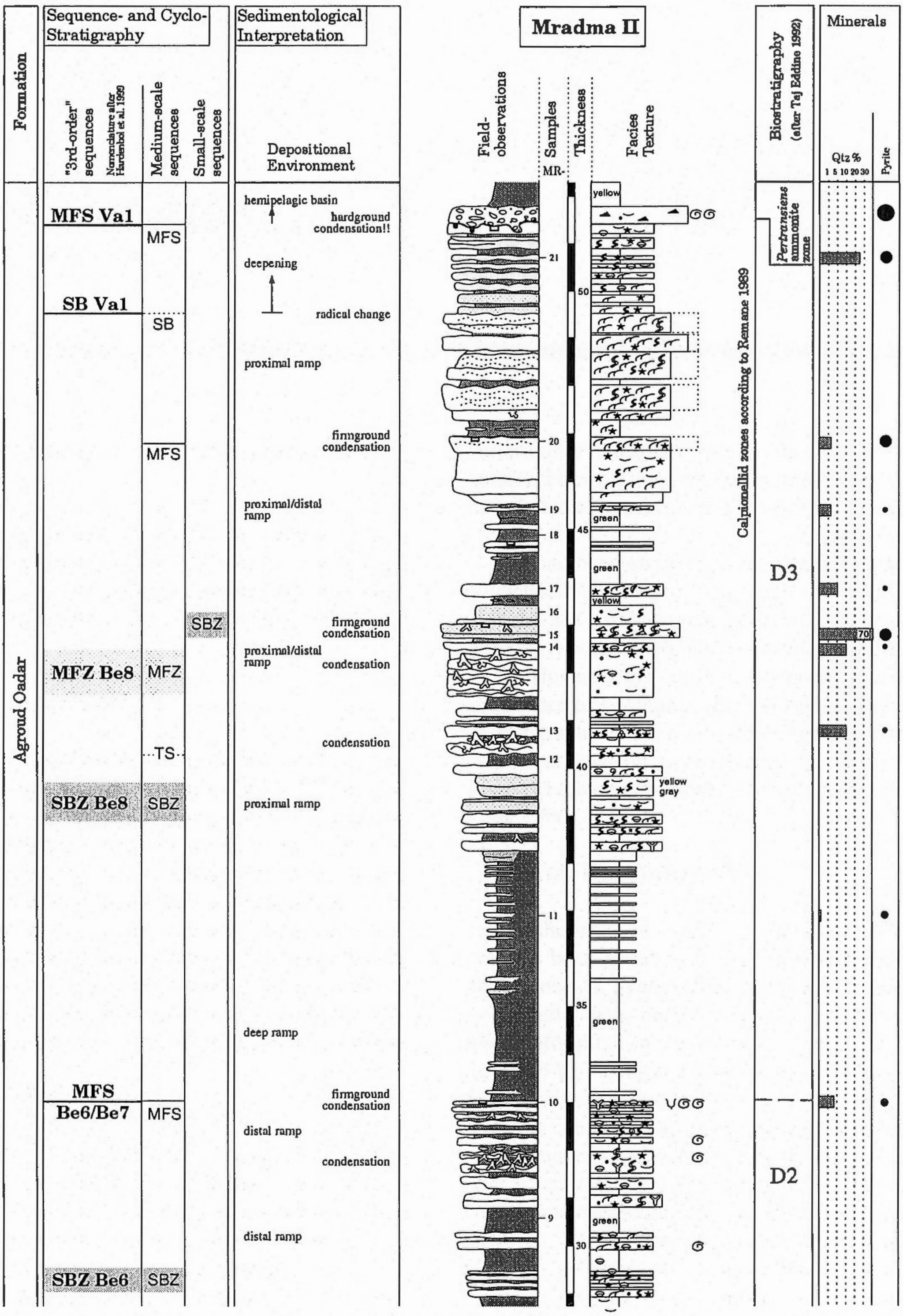


Fig. 5.36b. Mradma section (part II).

6 - CORRELATIONS

6.1. METHODS OF STRATIGRAPHIC CORRELATION

The basis of the chronostratigraphic correlations presented here is a relatively well established biostratigraphy. Other methods such as magnetostratigraphy have been proven to be ineffective in the studied formations (Blanc 1996).

6.1.1. Biostratigraphy

Biomarkers used in this study are described in chapter 4. On a regional basis (Jura platform) benthic foraminifers are the main tool for correlation purposes. Their first and last appearances are believed to be synchronous on a regional scale (across the platform), but taphonomic and ecological factors, as well as bias by irregular sampling, introduce a certain degree of error (Johnson 1992). This means that in combination with other criteria (see below) the general interval of presence of specific species is correlated, but first and last occurrences are not necessarily found in the same beds in all sections.

6.1.2. Lithofacies

Well-marked changes in lithofacies have always been used as criteria to define regionally correlative lithostratigraphic units. Such changes can reflect major environmental overturns and, on a local to regional scale, can have a more or less isochronous character. However, the small-scale diachronism introduced by lateral migration of facies belts is difficult to evaluate without the help of other chronostratigraphic tools.

Major lithostratigraphic changes occur regularly on the Jura platform. The most obvious ones are the transitions to and from the siliciclastically influenced Vions Formation and Guiers Member. It is reasonable to assume that large-scale climatic, tectonic and/or eustatic changes reflected in such sudden siliciclastic influence have an effect on the entire platform within a relatively short span of time. Paleogeographic disposition may lead to local variations in first appearance of siliciclastics (bypass situations), but on a low-resolution scale such changes are interpreted as synchronous on the Jura platform.

It has been suggested that peaks in allochthonous siliciclastic abundance (quartz, clays) in the Vions Formation are correlatable, considering *Pavlovecina allobrogensis* as a chronostratigraphic biohorizon (Jaquet 1974). However, clay-mineral composition, and especially abundance of kaolinite, which also have been suggested as stratigraphic markers (Persoz & Remane 1976), are very much dependent on hydrodynamic conditions on the platform and are dominantly facies-controlled (Adatte 1988, Pittet 1996).

6.1.3. Discontinuities

In the wake of sequence stratigraphy a great attention has been given to the chronostratigraphic value of discontinuities. Sequence boundaries, commonly marked by subaerial exposure on the platform and their "correlative conformity" in the basin (Vail et al. 1977) are believed to be isochronous markers. The same is inferred for maximum flooding surfaces which commonly are depicted by condensation. As already discussed before (chapter 5) these concepts were developed on the basis of seismic lines where reflectors, interpreted as unconformities, can be followed over large distances. It has been demonstrated, however, that the limited resolution of the seismic tool produces reflection terminations that do not correspond to unconformities in outcrop (e.g., Schlager 1995, 1997). On shallow-marine platforms, slightly irregular platform morphology, varying amplitudes of relative sea-level change, variations in sediment supply, accumulation, and redistribution easily lead to contemporaneous deposition beyond a theoretical lap-out position. This has the consequence that discrete lap-out surfaces do not develop. More likely, closely related families of surfaces are formed that individually are of limited extent (Cartwright et al. 1993).

On the Jura platform, for example, single discontinuity surfaces with the same sedimentological characteristics, and/or with precise biostratigraphic dating, could not be correlated unequivocally between all sections. This stands in contrast with other studies, where single surfaces indicating subaerial exposure were correlated across the platform from the Swiss Jura into the Vercors area in Southern France (Darsac 1983, Boisseau 1987, Adatte 1988, Blanc 1997, refer also to chapter 6.2, Fig. 6.3). For this reason a discontinuity analysis for each platform section was carried out and the distribution of the different surface types (exposure, condensation, and erosion/facies change) in all sections was established (Hillgärtner 1998). Figure 6.1 shows an example form the Monnetier section, where interpretation of depositional environments and discontinuity surfaces is illustrated in detail (refer to chapter 5 for other sections). It can be observed that exposure surfaces and surfaces indicating stratigraphic condensation tend to occur repetitively in certain intervals. They form zones where either exposure or condensation in subtidal environments is predominant. Surfaces indicating facies change or subtidal erosion occur in-between and commonly do not form such distinct clusters. Within the established biostratigraphic framework, a correlation of the surface zones (but not of single surfaces) across the platform becomes possible (see chapter 6.2).

6.1.4. Depositional sequences

Integration of the sequence evolution in each section, of biostratigraphic data, and major lines of correlation given by surface zones, allows the establishment of a high-resolution sequence-stratigraphic framework. Beginning with large-scale trends, the correlation of medium-scale and small-scale sequences becomes progressively more difficult, since local paleogeographic differences have more effect on the short-term record of environmental change and can lead to missing beats. Furthermore, biostratigraphic resolution is lower than the inferred time span represented by small-scale sequences and most medium-scale sequences (the biostratigraphic range of marker fossils reaches over more than one sequence). Therefore, correlation on the small scale can only represent a reasonable best-fit solution. It is a solution with the least contradictions on the basis of the various analyses and interpretations carried out.

Correlation of depositional sequences in and between the different domains studied are presented in the following way:

- small- and medium-scale sequences interpreted to be correlative are numbered homogeneously for all sections in each correlation chart, regardless of their numbers in preceding figures.

- well-marked discontinuities (or zones of discontinuities) that can be attributed to "3rd-order" sequences in the sequence framework of Hardenbol et al. (1999) are named homogeneously throughout the entire manuscript. The relation of the "3rd-order" correlation template to the hierarchy of depositional sequences is discussed later.

6.2. Correlation in the Jura domain

Correlation of discontinuity-surface zones

The stratigraphic distribution and correlation of surface zones between studied sections are shown in Figure 6.2. The sections are arranged along a transect generally proceeding from more proximal to more distal platform positions, but refer to Figure 1.1 for their exact location. The base of most sections is well defined by a regional sequence boundary (Be4) dated as *Privasensis* subzone by ammonites and charophyte-ostracod assemblages (Detraz & Mojon 1989, Strasser 1994, Pasquier 1995). Exposure zones 4 and 5 are well constrained by *Pavlovecina allobrogensis*. Calpionellids and charophyte-ostracod assemblages in the La Chambotte, Salève, and Monnetier sections allow to attribute exposure zone 6 to the *Picteti/Alpillensis* subzones (Fig. 6.1, Deville 1991, Blanc 1996). Charophyte-ostracod assemblages indicate a latest Berriasian age for exposure zones 7 and 8 (Deville 1991, Detraz & Mojon 1989), whereas exposure zone 9 and Early-Valanginian surface zones are constrained by benthic foraminifers *(Pfenderina neocomensis, Montsalevia salevensis)*. Late first occurrences of *Pfenderina neocomensis* in some sections (Fig. 6.1) probably are due to unfavorable environmental conditions (restricted facies). Occurrences of the three most important marker foraminifers are noted in the correlation chart (Fig. 6.2). Within this biostratigraphic framework, a correlation of the surface zones across the platform becomes possible, and the following important trends can be observed:

(1) The upper Middle Berriasian is dominated by widespread condensation, which can be correlated in many cases all across the platform. Exposure occurred only locally and mainly in proximal parts of the platform (exposure zones 2 and 3, Fig. 6.2).

(2) The lower half of the Upper Berriasian shows alternating zones of exposure and condensation. Exposure, however, prevails and can be correlated throughout the platform from proximal to distal positions (exposure zones 4, 5 and 6). Yet, exposure zones 5 and 6 correlate with zones indicating condensation and small-scale erosion in proximal platform positions and in the Crozet section, respectively. Although facies generally suggest a shallowing trend, no signs of subaerial exposure are detected in these localities. Accommodation increases slightly towards the outer platform during the lower half of the Upper Berriasian.

(3) The top of the Berriasian and the Lower Valanginian are marked by a dominating trend of stratigraphic condensation in subtidal environments. Exposure zones occur repetitively but are restricted to parts of the outer platform (exposure zones 7 to 10). Especially during the Late Berriasian these exposures correlate laterally with condensation surfaces in the most proximal and most distal parts of the platform (exposure zones 7 and 8). In the Crêt de l'Anneau section, the most proximal locality, the complete Upper Berriasian is condensed and/or eroded down to a few tens of centimeters of remaining sediment.

Interpretation of surface zones

Comparing the distribution of the surface zones with the "global" sequence-stratigraphic framework established by Hardenbol et al. (1999) (Fig. 6.3), it becomes clear that most zones with dominating exposure correspond to their "3rd-order" sequence boundaries. Zones with dominating condensation fall between sequence boundaries. Accepting the global significance of this sequence framework would imply that the repeated occurrence of specific zones of discontinuities reflect larger-scale (second- to third-order) eustatic sea-level changes. Exposure-dominated zones between the "3rd-order" sequence boundaries (exposure zones 7 and 9) may reflect higher-order sea-level fluctuations on the order of several 100 ky, when taking into account the relative ages of sequence boundaries (Fig. 6.3, Hardenbol et al. 1998). A multiplication of diagnostic surfaces forming zones of repetitive short-lived exposure or condensation features, rather than distinct single surfaces marked by a long time gap, has already been described by Montañez and Osleger (1993), Elrick (1996), and Pasquier & Strasser (1997). This phenomenon is explained by a superposition of high-frequency sea-level variations on a larger-scale trend of relative sea-level change. This creates the maximum-flooding or maximum-regression zones on a third-order scale (refer to chapter 5 and Fig. 5.6). They are suggestive of a greenhouse climate mode, when small-scale, short-term relative sea-level changes show low amplitudes and tend not to exceed larger-scale, long-term accommodation changes (Read 1995). This does not imply, however, that all discontinuities represent hiatuses

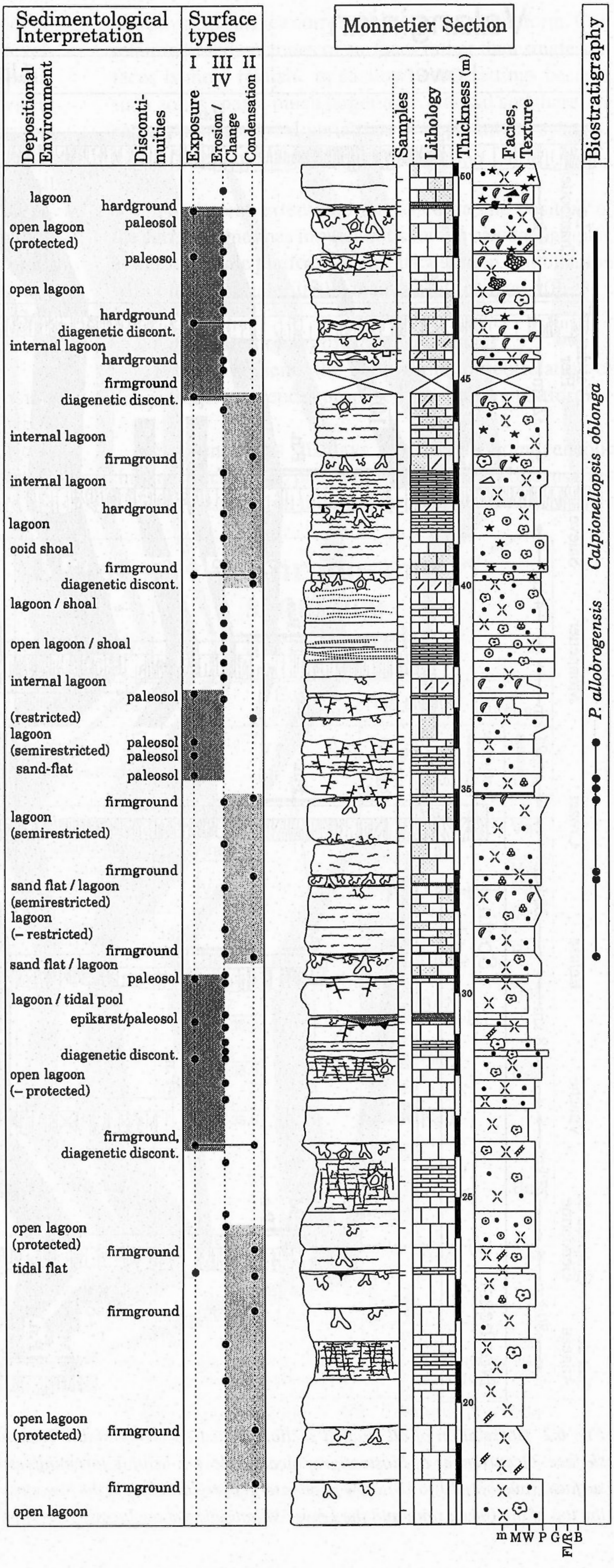


Fig. 6.1. Part of Monnetier section with detailed facies evolution and interpretation of depositional environments. Exposure- and condensation-related discontinuity surfaces form zones of repetitive occurrence. Refer also to Plate 8.5.

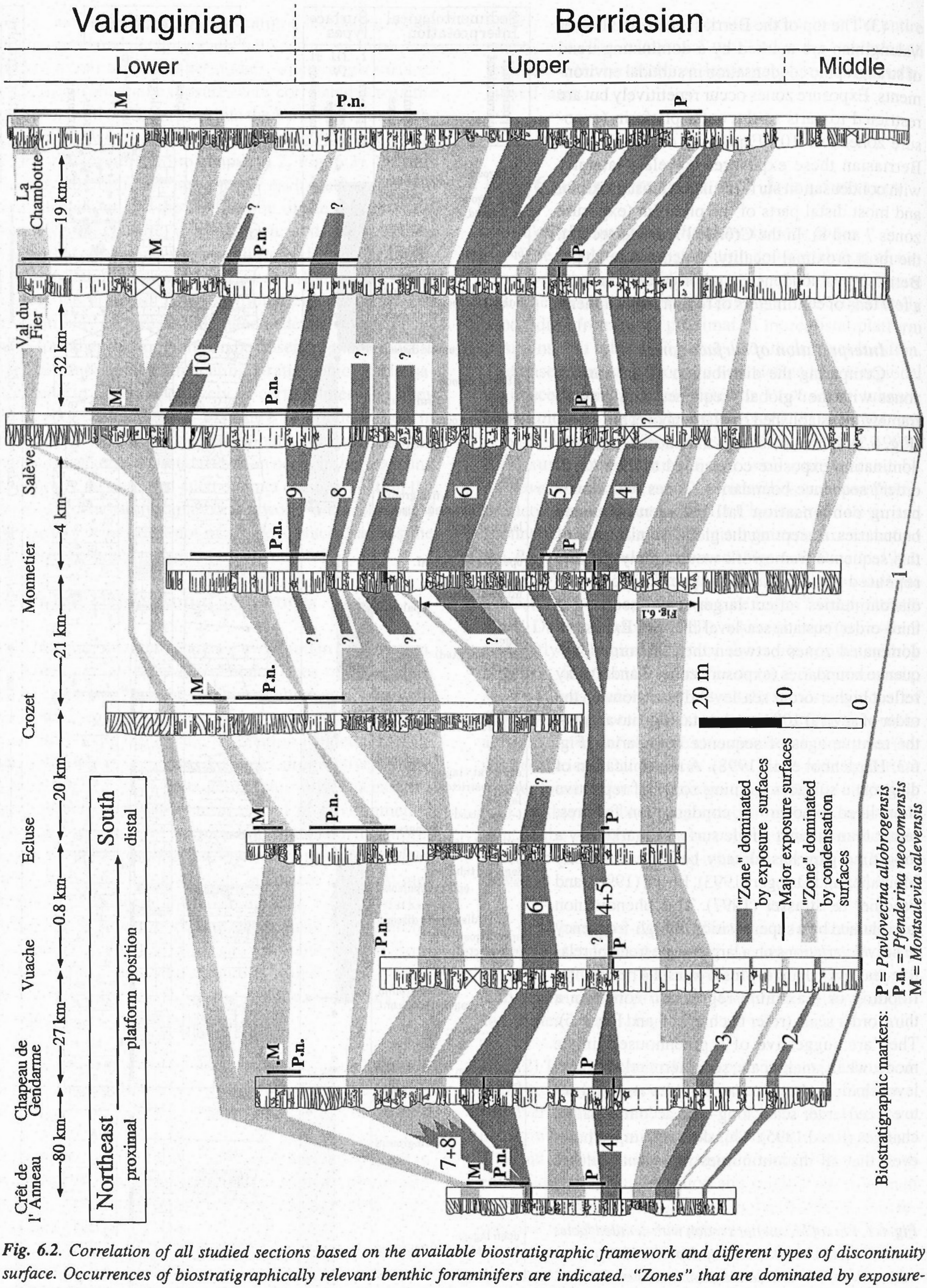


Fig. 6.2. Correlation of all studied sections based on the available biostratigraphic framework and different types of discontinuity surface. Occurrences of biostratigraphically relevant benthic foraminifers are indicated. "Zones" that are dominated by exposure-surfaces (numbered 1 to 10) show good lateral correlatability in the lower part of the Upper Berriasian, whereas the Middle Berriasian, the top of the Berriasian, and the Lower Valanginian show locally restricted exposures and widespread condensation.

of equal and exclusively short-lived duration. Well developed exposure surfaces (karst, exceptionally well-developed paleosols) with assumed longer durations of subaerial exposure occur locally and in different positions within exposure zones (Fig. 6.2). This suggests that local factors such as pre-existing morphology and differential subsidence modified allogenic signals. The assumption that all the observed discontinuities reflect high-frequency variations in relative sea-level cannot be proven. However, the sedimentary expression of any high-frequency cyclic process, be it autogenically or allogenically controlled, was certainly influenced or even forced by larger-scale (3rd-order) accommodation changes.

Most of the discrete, large-scale exposure discontinuities described by others (Fig. 6.3) are correlative with exposure-dominated zones observed here. However, in my opinion, they represent the best expressed discontinuities within a zone in one section, but are not one physical surface correlatable across the platform. Correlation based on zones of surfaces rather than single surfaces is more realistic in shallow-water settings because such zones span a much larger time interval and therefore can account for local variations of depositional systems, morphology, and lateral facies changes.

The lateral extent and the paleographic position of the surface zones has implications for the paleogeographic evolution of the platform and is discussed in combination with other data later in the manuscript (chapter 10).

Correlation of depositional sequences

Fig. 6.4 presents the high-resolution correlation of depositional sequences proposed for the Jura platform.

Sequence Be4 displays a relatively homogeneous thickness across the platform with minimal condensation towards the platform interior. Pasquier (1995) described a

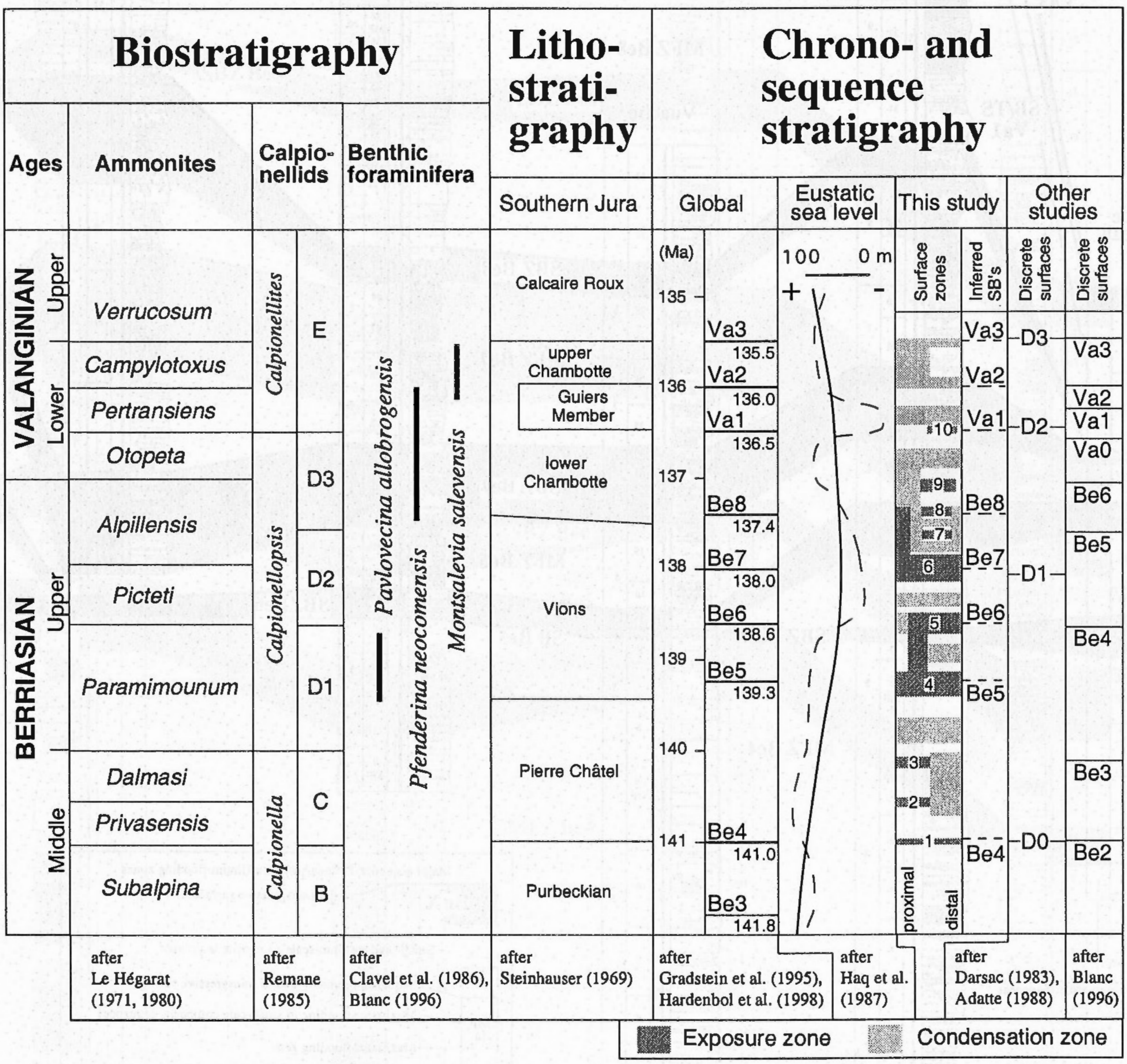


Fig. 6.3. Biostratigraphic, lithostratigraphic, chronostratigraphic, and sequence-stratigraphic framework of the studied interval. The positions of the observed zones of discontinuity surfaces and the presumptively correlative sequence boundaries are indicated. Exposure zones are numbered and correspond to those indicated in Figure 6.2. For a detailed interpretation refer to text.

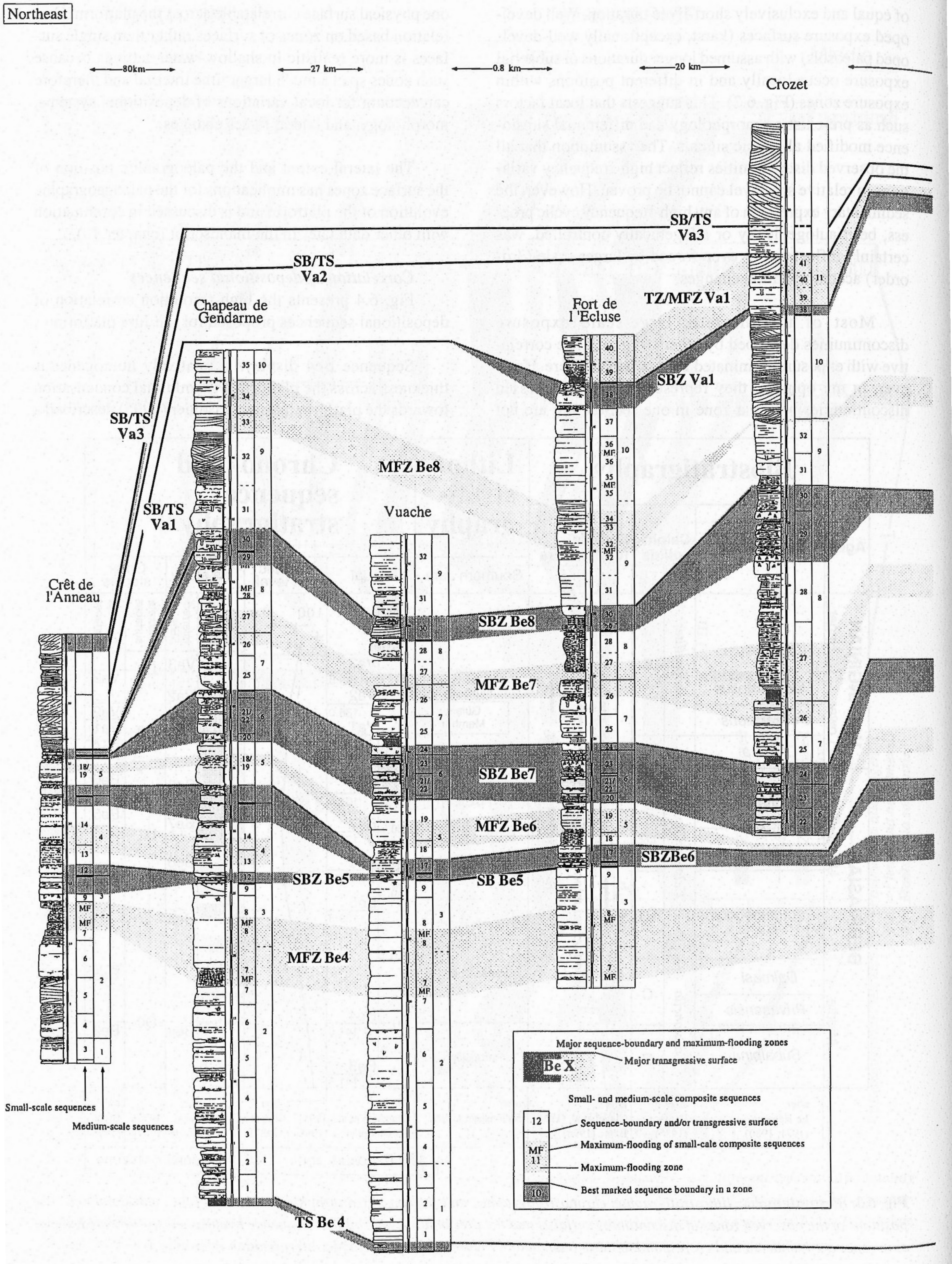
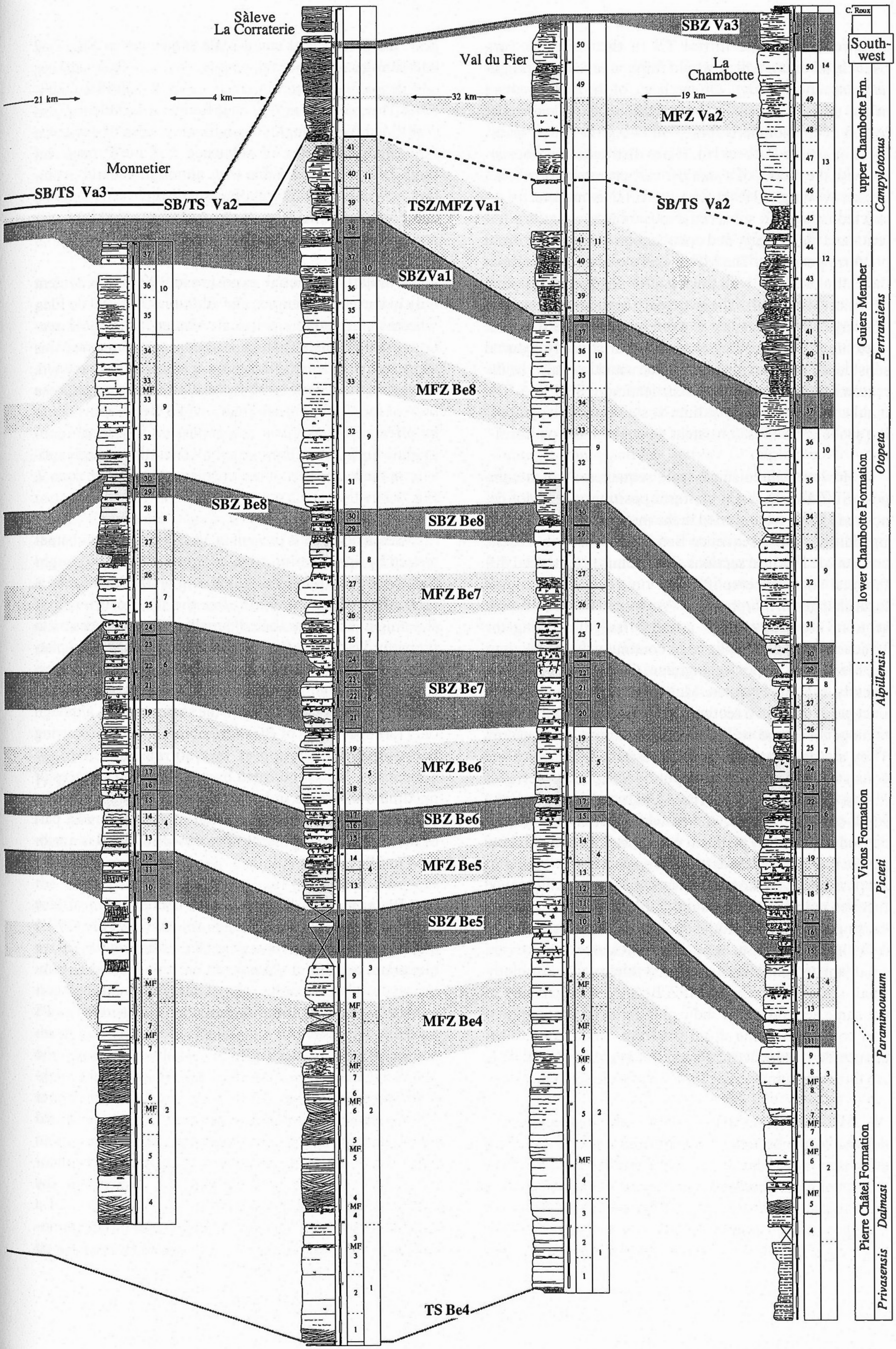


Fig. 6.4. Correlation of depositional sequences on the Jura platform. Correlative small- and medium-scale sequences have the same numbers in all sections. Nomenclature of "3rd-order" sequence-stratigraphic interpretation according to Hardenbol et al. (1999).


Sàleve
La Corraterie
Monnetier
Val du Fier
La Chambotte
21 km
4 km
32 km
19 km
SB/TS Va3
SB/TS Va2
TSZ/MFZ Va1
SBZ Va3
MFZ Va2
SB/TS Va2
SBZVa1
MFZ Be8
SBZ Be8
SBZ Be8
MFZ Be7
SBZ Be7
MFZ Be6
SBZ Be6
MFZ Be5
SBZ Be5
MFZ Be4
TS Be4
C. Roux
South-
west
upper Chambotte Fm.
Guiers Member
lower Chambotte Formation
Vions Formation
Pierre Châtel Formation
Campylotoxus
Pertransiens
Otopeta
Alpillensis
Picteti
Paramimounum
Dalmasi
Privasensis

diachronism of the initial TS in the northern Jura (Neuchâtel area), which would imply a preceding (angular) ramp morphology. On the basis of the data obtained in this study it appears to be synchronous, at least in the region of the southern Jura. Following the initial transgression on the Purbeckian, facies distribution is consistent with the model of a shallow carbonate platform (e.g., Tucker & Wright 1990). The thick TD is marked by restricted facies and subaerial exposure in the platform interior and high-energy and open-lagoonal deposits in distal platform positions. The MF is well expressed throughout the entire platform with thick and/or condensed lagoonal deposits (MFZ Be4). The extremely asymmetric shape of this large-scale sequence is expressed with the thin HD. Tectonic activity leading to an uplift or strongly reduced subsidence could explain this pattern and may also be the reason for the widespread karstification in SBZ Be5. The number of small- and medium-scale sequences composing sequence Be4 is consistent throughout the platform.

However, three small-scale sequences, which compose SBZ Be5 in distal platform positions, were not deposited and/or were eroded in the most proximal sections. In addition, entire sequence Be5 is condensed in Vuache and Fort de l'Ecluse sections. This points to enhanced differential subsidence/uplift of tectonic origin that caused local changes in morphology. Vuache and Fort de l'Ecluse sections are located near a major fault line along the Vuache range, which has been continuously active since the Mesozoic (Upper Valanginian, Charollais et al. 1983, as early as Kimmeridgian, M. Meyer, pers. comm. 1997). Except for these two sections, sequences Be5 and Be6 are homogeneous and well correlatable across the platform. They are delimited by well-expressed SBZs, which are consistent with zones of repetitive exposure surfaces (zones 4 and 5, Fig 6.2). These SBZs are composed of 3 small-scale sequences, one of which commonly is condensed in proximal platform positions. Yet, subaerial exposure is not as evident as in distal positions. Two scenarios are possible: 1. erosion and reworking of sedimentological features pointing to exposure, 2. higher accommodation causing constantly subtidal conditions. The first scenario is favored in this case, because small-scale sequences are amalgamated and display reduced thicknesses. Similarly small-scale sequences in MFZ Be6 are amalgamated in the same sections, also indicating reduced accommodation compared to the distal platform. In contrast to sequence Be4, sequences Be5 and Be6 (to base of SBZ Be7) are thin and composed of 1 medium-scale sequence each.

SBZ Be7 spans up to 5 small-scale sequences and is marked by the best-expressed emergence features across the platform. Exceptions are the complete condensation/erosion of the entire Be7 sequence in the Crêt de l'Anneau section (superimposition of SB/TS Be8 on SBZ Be7), and no exposure indicators in SBZ Be7 of the Crozet section. In the case of the Crozet section, the stacking pattern suggests the presence of small-scale sequences in SBZ Be7 with similar or higher thicknesses than in other sections, and deposition in continuously shallow-subtidal conditions. This points to important differential subsidence, also confirmed by the highly variable thickness of sequence Be7 (e.g., medium-scale sequence 8 in the Crozet section). It is also remarkable that, although general evolution indicates a MF, correlative small-scale sequences 26 and 27 are delimited by exposure surfaces in the Monnetier and Salève sections (exposure zone 7, Fig. 6.2).

Similarly, subaerial exposure in SBZ Be8 is evident only in these two sections, and additionally in Val du Fier, whereas correlative small-scale sequences in other sections point to condensation in subtidal environments. This suggests that the platform was morphologically structured. However, the drastic, platform-wide facies change at the base of the Chambotte Formation (TS Be8) seems not to be affected by platform morphology. Only local topographic highs are still indicated by laterally limited exposure in the lower part of the Formation (exposure zone 9, Fig. 6.2).

Sequence Be8 is present in all sections and is characterized by open-marine, high-energy carbonates throughout the entire platform. This suggests absence of a protecting barrier, at least after the last occurrence of shoaling in exposure zone 9 (small-scale sequences 31 to probably 32). Thereafter, ooid bars occur in proximal platform positions (Chapeau de Gendarme, Fort de l'Ecluse, Crozet). The high-energy dunes in the Crêt de l'Anneau section probably correspond to the same interval, although only rare specimens of *Pfenderina neocomensis* occurring below and above support this correlation biostratigraphically. MF is indicated by condensation surfaces or maximum thickness in lagoonal sediments all across the platform. Shoaling and/or erosion in the uppermost part of the Formation is present in most sections and is attributed to SB Va1.

Better marked, however, is the condensation surface defining the upper limit of the Formation. Deville (1991) described subaerial exposure at this surface on the Salève mountain, but this could not be confirmed and is not observed in other sections. The condensation surface certainly evidenced by this discontinuity is interpreted as TS marking the onset of a tectonically enhanced TD in sequence Va1. Accelerated differential subsidence across the platform can be implied on the basis of significant thickness variations of the Guiers Member and the Upper Chambotte Formation. The enhanced TD is distinguished by the deepest facies at its base, probably because of an abrupt tilt and flooding of parts of the platform. The theoretical MF corresponds to the bioturbated interval in the center of the Guiers Member (outcropping in La Chambotte section). After rapid infill of created accommodation the HD-sediments were winnowed (wave base on a ramp profile) and condensed and/or eroded during a possible fall of base level that can be attributed to SB/TS Va2. It is expressed by condensation and reworking in more proximal platform positions (Salève, Crozet, Crêt de l'Anneau). In more distal areas (Chambotte, Val du Fier), however, SB Va2 is not marked by an important discontinuity, probably due to sufficient accommodation, compensating and attenuating any possible relative sea-level fall. The rather subtle character of this discontinuity certainly points to a long-term transgressive trend.

Sequence Va2 constitutes the top of most sections. It displays a MFZ that is only observable in the most distal sections. The subtle shoaling and erosion in these sections, attributed to the HD of sequence Va2 and SBZ Va3, laterally pass into a major erosion surface towards the northern Jura platform. In the Neuchâtel area this surface can cut as far down as to the Pierre-Châtel Formation (Steinhauser & Charollais 1971, Adatte 1988, Blanc 1996). This again testifies for important differential subsidence across the platform, which leads to the formation of a tectonically enhanced TS. This surface is well exposed in the La Chambotte section where it is overlain by marls that indicate deposition in deep-ramp environments. In more proximal areas, the laterally equivalent Calcaire Roux Formation is composed of high-energy bioclastic sands, which display progressively reduced thickness towards the north. This lateral facies relationship points to a certainly diachronous onlap of deeper-water sediments onto a ramp-type morphology.

6.3. Correlation in the Vocontian Domain

Correlations of pelagic and hemipelagic limestone-marl alternations on various scales in the Vocontian Trough have evidenced their wide lateral continuity throughout the entire basin, especially in marl dominated intervals (e.g., Cotillon 1991). The sequence framework established for the two sections studied is presented in Fig. 6.5. The pattern of correlation lines quickly reveals that general accumulation rates and depocenter in the basin changed through time. Whereas in the lower part (*Privasensis* to *Paramimounum* ammonite zones) the Montclus section displays a thicker succession, the Angles section testifies for higher accumulation rates in the *Alpillensis/Otopeta*, upper *Pertransiens* and *Campylotoxus* zones. As a general trend, accumulation rates increase towards the top. This can be deduced from comparison of relative thickness with relative durations of ammonite zones, according to the chronostratigraphic scale (Gradstein et al. 1995, Fig. 6.3).

Differential accumulation mainly concerns sequence Be4, which displays a reduced LD in Angles section. Sequence Be5 consists of a very condensed deposit with characteristics of a SBZ/LD in both sections and is directly amalgamated with the SBZ/LD of sequence Be6. Individual small-scale sequences observed in Montclus cannot be identified in the Angles section where slumping is more important, and the TD is very condensed. SBZ/LD Be7 follows shortly after the MF of sequence Be6. In both sections this points to a continuing general sea-level fall and/or lowstand on the large scale. Sequence Be7 displays thinner limestone beds and slightly thicker marl interbeds in the Montclus section, whereas marl content first significantly increases after SBZ/LD Be8 in the Angles section (Fig 6.5). This can, in a speculative way, be interpreted as different behavior of source areas of clays (Jura platform vs. Provence platform), variations of their distribution patterns, variations in planctonic productivity rates within the basin, or more probably, a combination of all. Sequence Be8 appears to be reduced in Montclus, considering the range of the *Alpillensis/Otopeta* ammonite zones and the number of implied small-scale sequences. A condensation at the MF is inferred, because SBZ Va1 does not display any signs of erosion in the Montclus section. This is in contrast to the small slump in Angles. In a very similar way SBZ Va2 is indicated by few, thick limestone beds in Montclus, and a slump in Angles. Sequences Va1 and Va2 are dominated by marl deposits, and no unequivocal LD can be distinguished. A weak thickening-up tendency of limestone beds gives the impression of multiplied TSs, all pointing to the importance of a long-term transgressive trend. In this interval correlation of individual limestone-marl bundles between the sections seems possible (Fig. 6.5 blow up). It suggests homogeneous environmental conditions throughout the basin. The series of 5 thickening-up limestone beds probably reflect correlatable environmental changes in response to allocyclical forcing on a Milankovitch-type frequency, which is known to occur in the Vocontian Trough (Giraud et al. 1995). Slumps in both sections drastically interrupt this regular stacking pattern and apparently rework the entire HD in Montclus. Such an event on a generally transgressive trend, which is continued in the marly Upper Valanginian (Cotillon et al. 1979, Giraud et al. 1995), rather points to sudden slope instabilities of tectonic origin than to a major LD related to a drop in relative sea level.

6.4. Correlation of Platform and Basin

Correlation of key surfaces and intervals from platform to basin, which define the larger-scale trends, is carried out with high confidence on the basis of comparable biostratigraphic resolution. On the smaller scale, however, only the number of depositional sequences is compared and a tentative correlation is proposed (Fig. 6.6).

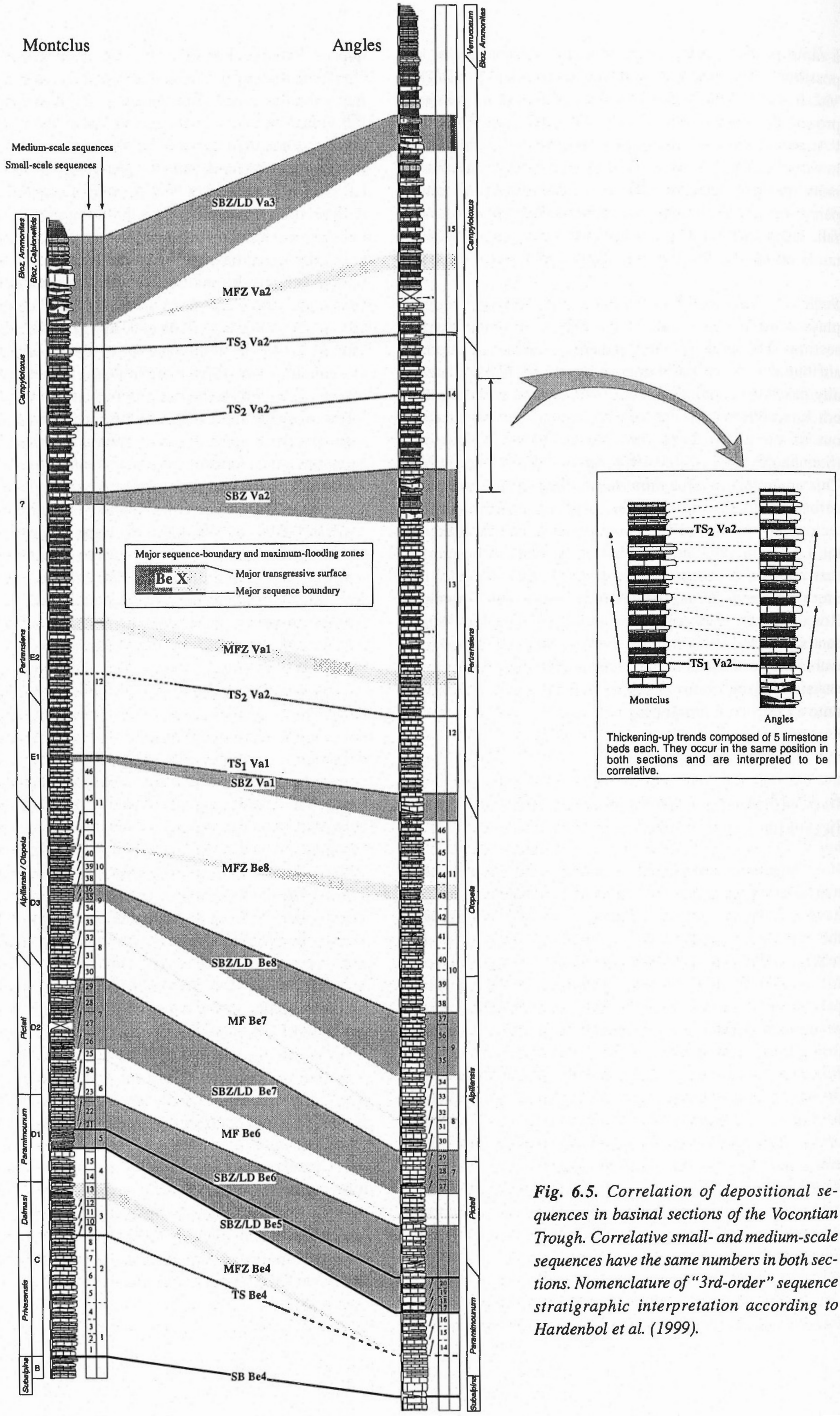


Fig. 6.5. Correlation of depositional sequences in basinal sections of the Vocontian Trough. Correlative small- and medium-scale sequences have the same numbers in both sections. Nomenclature of "3rd-order" sequence stratigraphic interpretation according to Hardenbol et al. (1999).

All larger-scale sequences distinguished on the platform can be identified in the basin. However, sequence Be4 is heavily condensed in Angles section, sequence Be5 in both Montclus and Angles. The base of sequence Be4 is condensed on the platform and the initial TS on the platform (TS_1) correlates with deposits of lowstand character in the basin. This corroborates the model by Pasquier & Strasser (1997) who interpret an initial flooding of the platform in the long term with contemporaneous deposition of characteristic LDs in the basin. The basinal TS does not have an equally well-expressed counterpart on the platform, since there accommodation was already high enough to allow deposition and preservation of shallow-marine facies. However, in some sections (Salève, Monnetier) a change to high-energy deposits points to more open-marine conditions, which, in context with the basinal sections, can be interpreted as a second (duplicated) flooding surface (TS_2).

Four small-scale sequences between best candidates for SB Be5 and SB Be6 on the platform correspond to four small, basinal, thickening-up sequences, constituting an amalgamated SBZ/LD (sequences 17-20, Angles section). This suggests that the same allocyclical signals were recorded in the basin, and that condensation was caused rather by low sedimentation rates than by erosion. A generally regressive trend between Be5 and Be7, which is recorded by restricted lagoonal facies, local subaerial exposure, and elevated abundance of siliciclastics in the platform sections, is well mirrored by calcareous and thick-bedded basinal LDs.

The first increase in relative abundance of clays, following SBZ Be7 in the basin, is coeval with a beginning transgressive trend on the platform. Gradual disappearance of siliciclastics on the platform in sequence Be8 is reflected by further increase in clay abundance in the Vocontian Trough. SBZ Be8, which is only locally marked by subaerial exposure and which delineates the significant facies change between Vions and Chambotte Formations, is portrayed inconspicuously in the Montclus and Angles sections. This supports the hypothesis that SBZ Be8 reflects at most an attenuated smaller-scale sea-level fall on a larger-scale transgressive trend, if any sea-level fall at all. In the same way, SBs Va1, Va2, and Va3 only slightly modulate the long-term trend of increasing clay content in the basin. However, discontinuities attributed to sequences Va1 and Va3 commonly are well expressed on the Jura platform and suggest episodic phases of tectonic activity and differential subsidence.

Two possibilities to correlate the basinal TD Va1 with the major discontinuity on top of the Chambotte Formation can be imagined (Fig. 6.6): 1. correlation of the discontinuity with the basinal MF, which implies condensation of the entire TD and a significant hiatus, 2. correlation with a TS, implying more or less continuous sedimentation, depending on which TS (multiplied in the basin) is chosen.

The discontinuity on the platform certainly indicates stratigraphic condensation, but it is impossible to deduce any hiatus duration. No time-equivalent, well-expressed LD is formed in the basin, which favors the interpretation as flooding surface, and not as sequence boundary. The number of implied medium-scale sequences in sequence Va1 on the distal platform (La Chambotte section only) points to more or less continuous sedimentation when compared with the presumably complete basinal record. Good correlatability of the three small-scale sequences above SBZ Va1 in the distal parts of the platform (Fig. 6.4) suggests drastic increase in differential subsidence and deposition of deepest facies on top of the discontinuity. Subsequently, the created space was filled with sediment compensating for subsidence, and building up into the zone of the storm wave-base. Hence, smaller-scale relative sea-level fluctuations were recorded in more distal positions. Erosion and reworking at (storm) wave-base level predominated in more proximal settings during slowing relative sea-level rise and/or beginning relative sea-level fall that can be attributed to SB Va2. On the basis of these reflections a correlation of one of the basinal TS_1 Va1 or TS_2 Va1 with the discontinuity above SBZ Va1 on the platform is favored.

The sudden influx of siliciclastics in sequence Va1 on the platform, which does not obviously coincide with a relative sea-level lowstand (ramp facies) as in sequences Be5 to Be7, additionally points to tectonic activity (uplift in the hinterland), and probably to climatic changes on the long term. However, siliciclastics on the platform diminish in sequence Va2, whereas in basinal settings marls begin to dominate the facies entirely. This is explained with a continuing transgressive trend and slowing differential subsidence, allowing the sediment to be winnowed of clays in high energy platform environments in combination with decreasing influx of coarse siliciclastics, and/or a bypass situation. A drastic change is evidenced with SB and TS Va3, again pointing to accelerated differential subsidence on the platform. It is expressed in the same fashion as at SB/TS Va1 with proximal erosion and distal condensation. This evolution is accompanied by major slumps in basinal areas, which also indicate tectonic instability (absence of platform detritus, superposition on a larger trend of sea-level rise).

6.4.1. The conceptual model linking platform and basin

An interesting point to discuss is the distribution through time and relative abundance of siliciclastics on the platform and of clays in the basin. As already sug-

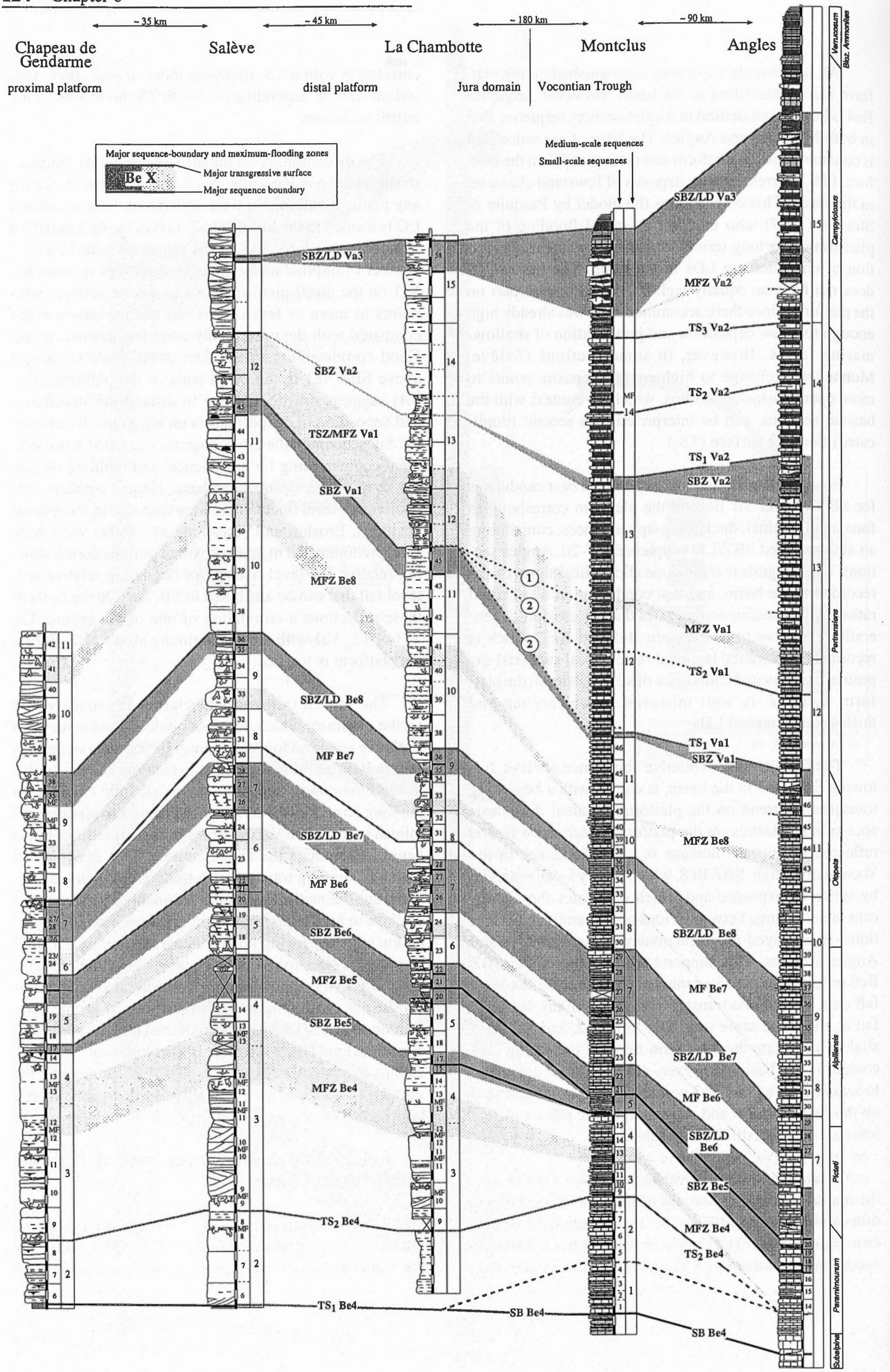

~ 35 km
~ 45 km
~ 180 km
~ 90 km
Chapeau de Gendarme
proximal platform
Salève
La Chambotte
distal platform
Jura domain
Montclus
Vocontian Trough
Angles
Major sequence-boundary and maximum-flooding zones
Be X
Major transgressive surface
Major sequence boundary
Medium-scale sequences
Small-scale sequences
SBZ/LD Va3
MFZ Va2
TS3 Va2
SBZ Va2
TS2 Va2
TSZ/MFZ Va1
TS1 Va2
SBZ Va1
SBZ Va2
MFZ Be8
MFZ Va1
TS2 Va1
SBZ/LD Be8
TS1 Va1
SBZ Va1
MF Be7
MFZ Be8
SBZ/LD Be7
MF Be6
SBZ Be6
SBZ/LD Be8
MFZ Be5
MF Be7
SBZ Be5
SBZ/LD Be7
MFZ Be4
MF Be6
SBZ/LD Be6
SBZ Be5
TS2 Be4
MFZ Be4
TS2 Be4
TS1 Be4
SB Be4
SB Be4
Verrucosum
Campylotoxus
Pertransiens
Otopeta
Alpillensis
Picteti
Paramimounum
Subalpina
Bioz. Ammonites

Fig. 6.6. Correlation of depositional sequences from platform to basin. Correlative small- and medium-scale sequences have the same numbers in all sections (but do not correspond to the numbers in Fig. 6.4). Nomenclature of "3rd-order" sequence stratigraphic interpretation according to Hardenbol et al. (1999).

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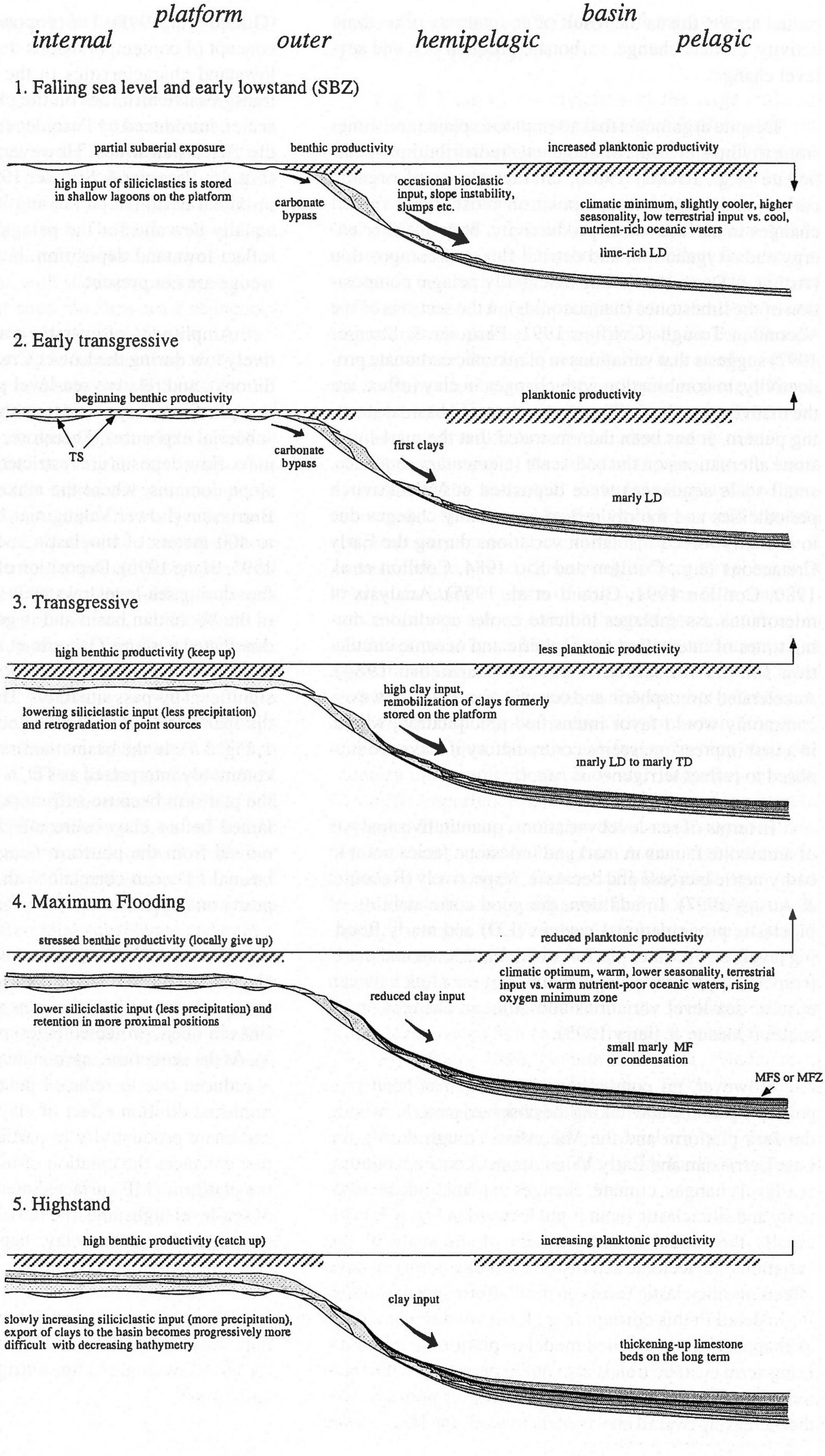


Fig. 6.7. Conceptual model linking trends in relative abundance of limestones and siliciclastics on the platform with those in the basin. The main controlling factors are relative sea-level variations and climatic changes, both influencing carbonate productivity and fluxes of siliciclastics. The concept reflects a very simplified configuration of Jura platform and Vocontian Trough during the Early Cretaceous (Late Berriasian-Early Valanginian), which is one of a wide platform or low-angle ramp during low amplitude sea-level variations. The concept for contemporaneous deposition of basinal LD and TD on the platform (stage 2 and 3) is based on Pasquier & Strasser (1997).

gested above, this is the result of an interplay of tectonic activity, climate change, carbonate productivity, and sea-level change.

Despite arguments that attempt to explain marl-limestone rhythms in terms of diagenetic redistribution of carbonate (e.g., Ricken 1986), characteristics of organic carbon and inorganic composition most likely reflect changes in surface-water productivity, bottom-water energy and oxygenation, and detrital flux and composition (Arthur & Dean 1991). The essentially pelagic composition of the limestones (nannoconids) in the sections of the Vocontian Trough (Cotillon 1991, Pasquier & Strasser 1997) suggests that variations in planktonic carbonate productivity, in combination with changes in clay influx, are the main factors that influence the observed basinal stacking pattern. It has been demonstrated that the marl-limestone alternations on the bed-scale (elementary sequence, small-scale sequence) were deposited at Milankovitch periodicities and mainly reflect seasonality changes due to orbitally forced insolation variations during the Early Cretaceous (e.g., Cotillon and Rio 1984, Cotillon et al. 1980, Cotillon 1991, Giraud et al. 1995). Analysis of microfauna assemblages indicate cooler conditions during times of intensified atmospheric and oceanic circulation for the limestone intervals (Darmedru 1984). Accelerated atmospheric and oceanic circulation however, commonly would favor intensified precipitation, which, in a first impression, seems contradictory if clays are supposed to reflect terrigeneous runoff.

In terms of sea-level variations, quantitative analysis of ammonite faunas in marl and limestone facies point to bathymetric increase and decrease, respectively (Reboulet & Atrops 1997). In addition, the good correlatability of bioclastic progradational wedges (LD) and marly flooding levels on the outer platform with limestones and marls (respectively) in the pelagic realm suggest a link between eustatic sea-level variations and climatic changes on all scales (Quesne & Ferry 1995).

However, no comprehensive model has been proposed, explaining and linking the observed patterns in both, the Jura platform and the Vocontian Trough during the Late Berriasian and Early Valanginian. A concept relating sea-level changes, climate, changes in planktonic productivity and siliciclastic input is put forward in Fig. 6.7. Principally the model is independent of the scale of the variations, but tectonic activity on the longer term can have effects on siliciclastic fluxes on the platform, not accounted for in detail in this concept (e.g., Early Valanginian, refer to chapter 10 for a detailed model of platform evolution). Long-term eustatic trends also put in perspective the relative abundance of limestone and marls, as portrayed by the increasing overall clay content towards the Hauterivian (Giraud et al. 1995). The proposed model incorporates the concept of contemporaneous deposition of deposits with lowstand characteristics in the basin and deposits with transgressive affinities on the platform (on the long-term scale), introduced by Pasquier & Strasser (1997) for Middle Berriasian times. However, their conceptual model (Fig. 15, Pasquier & Strasser 1997) mainly illustrated the proximal basin (slope fan and lowstand wedge) but may equally be valid for the pelagic domain where deposits reflect lowstand deposition, but slope fan and lowstand wedge are not present.

Amplitudes of eustatic sea-level change were relatively low during the Lower Cretaceous (greenhouse conditions), and relative sea-level probably never fell below the platform edge (no synchronous, platform-wide subaerial exposure). Therefore, bioclastic deposition and mass-flow deposits are restricted to the outer platform and slope domains, where the maximum thickness of Upper Berriasian (Lower Valanginian) series is attained with up to 400 meters of bioclastic and oolitic facies (Pasquier 1995, Blanc 1996). Deposition of characteristic basin-floor fans during sea-level lowstands (stage 1) was not observed in the Vocontian basin and is generally rare in carbonate dominated systems (Jacquin et al. 1991). Fluvial systems had not the potential to cut down into the shelf and create significant by-pass situations. This favored storage of their dissipated load in shallow lagoons on the platform (stage 1, Fig. 6.7). In the basin, the first increase of clay content, commonly interpreted as TD, is diachronous to the TD on the platform because sufficient bathymetry had to be attained before clays were effectively winnowed and removed from the platform (stage 3, Fig 6.7). Therefore, basinal LDs can correlate with clearly transgressive deposits on the platform.

During MF, siliciclastic influx in platform areas and clay input into the basin commonly are reduced, due to retrogradation of point sources and possible deposition of fines in open, protected lagoons below wave base (stage 4). At the same time, carbonate accumulation in the basin is reduced due to reduced pelagic productivity. The diminished dilution effect of clays due to reduced benthic carbonate productivity in partial give-up situations, further enhances the creation of marly, condensed levels on the platform (MF surfaces/intervals, chapter 5). In times of sea-level highstand, the initially high potential for winnowing and export of clays begins to be hindered by decreasing bathymetry (stage 5).

The concept is based entirely on this case study and may only be valid for wide, carbonate-dominated platforms or low-angle ramps during low amplitude sea-level variations.

6.5. Correlation in the Atlantic domain

The large-scale sequence correlation of the two Moroccan sections clearly displays important condensation in the Id ou Belaid section (Fig. 6.8). Whereas sequences Be3 and Be5 are congruent in their thickness, sequence Be4 displays thicker marly intervals, which probably is related to its more distal position. The first, well-marked condensation surface in sequence Be4 marks the onset of a large-scale transgression with superimposition of a medium-scale MFS. Four of such medium-scale sequences compose sequence Be4 in both sections. The main difference between the two studied locations arises from condensation of the entire interval between MFS Be6 and SB Be8 in the Id ou Belaid section. No evidence for prolonged times of non-deposition are recognized, which points to deep-reaching erosion. The interval that is condensed in Id ou Belaid is represented by marly, deep-ramp deposits in the Mradma section. This configuration rather points to strong differential subsidence and possible uplift in more proximal positions than simply a fall in relative sea level that exposed the inner shelf. The latter would have caused deposition of a LD in Mradma section, which would be indicated by coarse siliciclastics in this domain (chapter 5). Tectonically accentuated morphology, which leads to important lateral thickness variations, is well known in this paleogeographic region (Taj Eddine 1992). Tectonic activity took place in the time interval between MF Be6 and SB Be8. It probably also led to the condensation of sequence Be7 in the Mradma section due to accelerated relative sea-level rise. Sequence Be8 is deposited in both sections but displays a thinner succession with shallower and higher-energy deposits and significant erosion in the proximal domain. It points to a relative transgressive trend and probably slowing differential subsidence and culminates in a major condensation surface. This discontinuity can be recognized in the entire basin and is the result of a tectonically enhanced, accelerated transgressive trend on the long term that eventually led to drowning of the Morocco platform in Middle Valanginian times (Schlager 1991, Taj Eddine 1992).

6.6. Correlation of Atlantic and Tethyan domains

Fig. 6.9 shows a correlation of the large-scale sequence evolution between sections from the Jura platform, the Vocontian Trough, and the Atlantic margin. Their correlation is possible on the basis of biostratigraphic data calibrated to ammonite subzones (Le Hégarat 1971, 1980, Bulot 1995) and calpionellid biozones (Remane 1985), coherent in all of these regions. It is evident that the correlation over such large distances, and in such different paleotectonic and paleobathymetric conditions is not unproblematic and has to be looked at as a best-fit solution. However, the more remarkable it is to observe that the sequence framework is very similar in these different environments. Despite the fact that key surfaces or intervals are expressed differently (MF dominating in Morocco vs. SB dominating in the Tethyan region), the thickness evolution of the individual sequences displays a comparable trend. This cannot be explained by overall constant sedimentation rates. For example, calpionellid zone C corresponds to a thick lithostratigraphic interval but is of shorter duration in terms of absolute time than zone D1, which corresponds to a thin lithostratigraphic interval in all domains (Fig. 6.10). Furthermore, the number of depositional sequences composing sequence Be4, Be5, and Be8 in Morocco is the same as the number of medium-scale sequences observed in the Tethyan domain (Fig. 6.9). This implies that the external factor forcing the sedimentary environments had a similar frequency and was of similar importance in both the Tethyan and the Atlantic domains. However, one apparent contradiction is manifest between the studied regions. A maximum of pelagic influence between SB Be4 and SB Be8 and more proximal sedimentation in sequence Be8 is expressed on the Atlantic margin. On the Jura platform, in contrast, the interval between Be5 and Be7 displays the most regressive facies. Yet, these opposite trends only modulate the long term transgressive evolution from the Middle Berriasian to the Hauterivian, which is well evident in both domains (Fig. 6.10, Blanc 1996, Jacquin et al. 1991, Wiedmann et al. 1982, Schlager 1989, Taj Eddine 1992). In addition, phases of increased differential subsidence due to tectonic activity occur in similar time intervals in both domains, during the Late Berriasian and in the Early Valanginian (Fig. 6.10).

Figures 6.8 and 6.9 on the follwing pages ➔

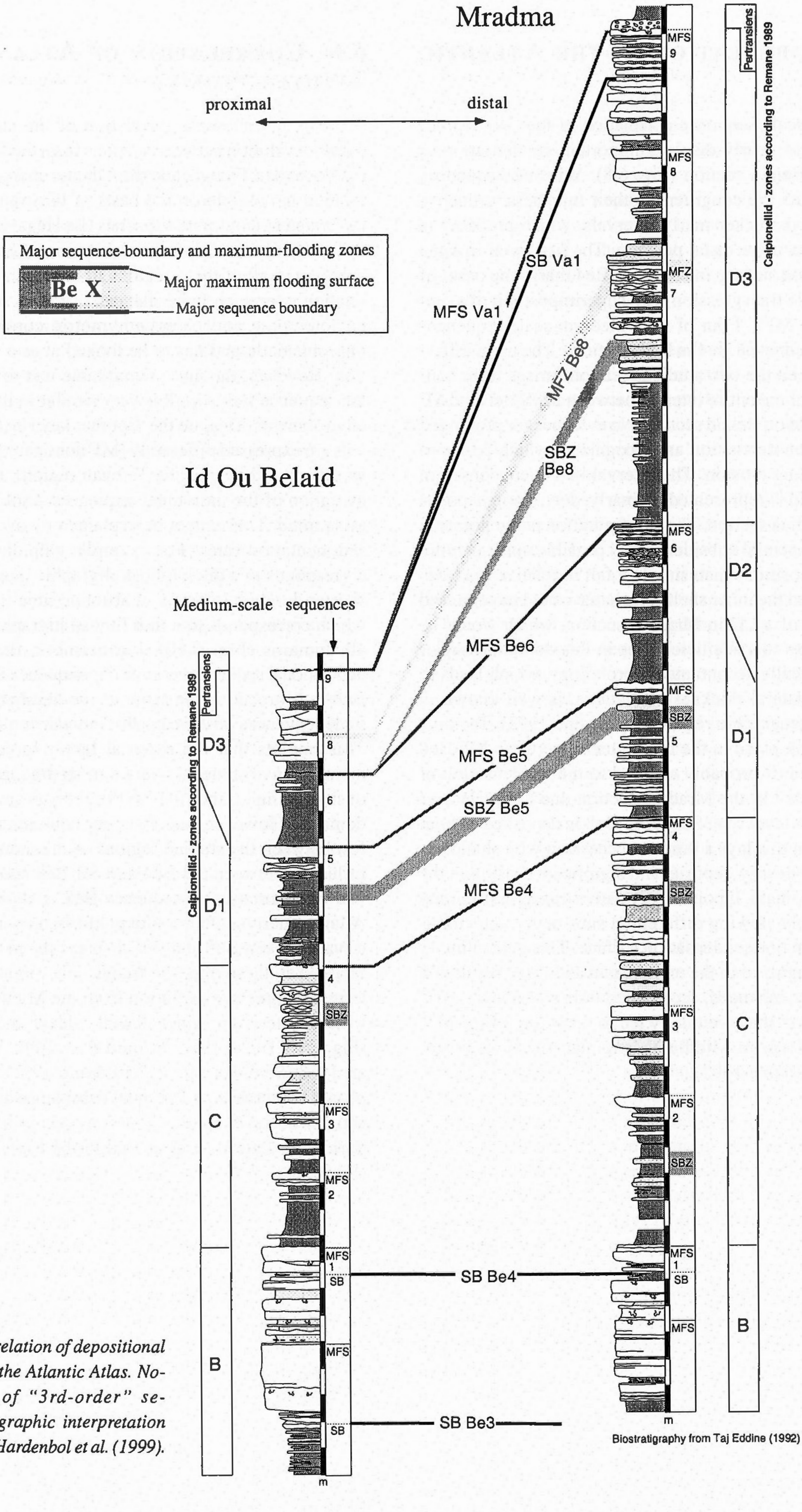


Fig. 6.8. Correlation of depositional sequences in the Atlantic Atlas. Nomenclature of "3rd-order" sequence-stratigraphic interpretation according to Hardenbol et al. (1999).

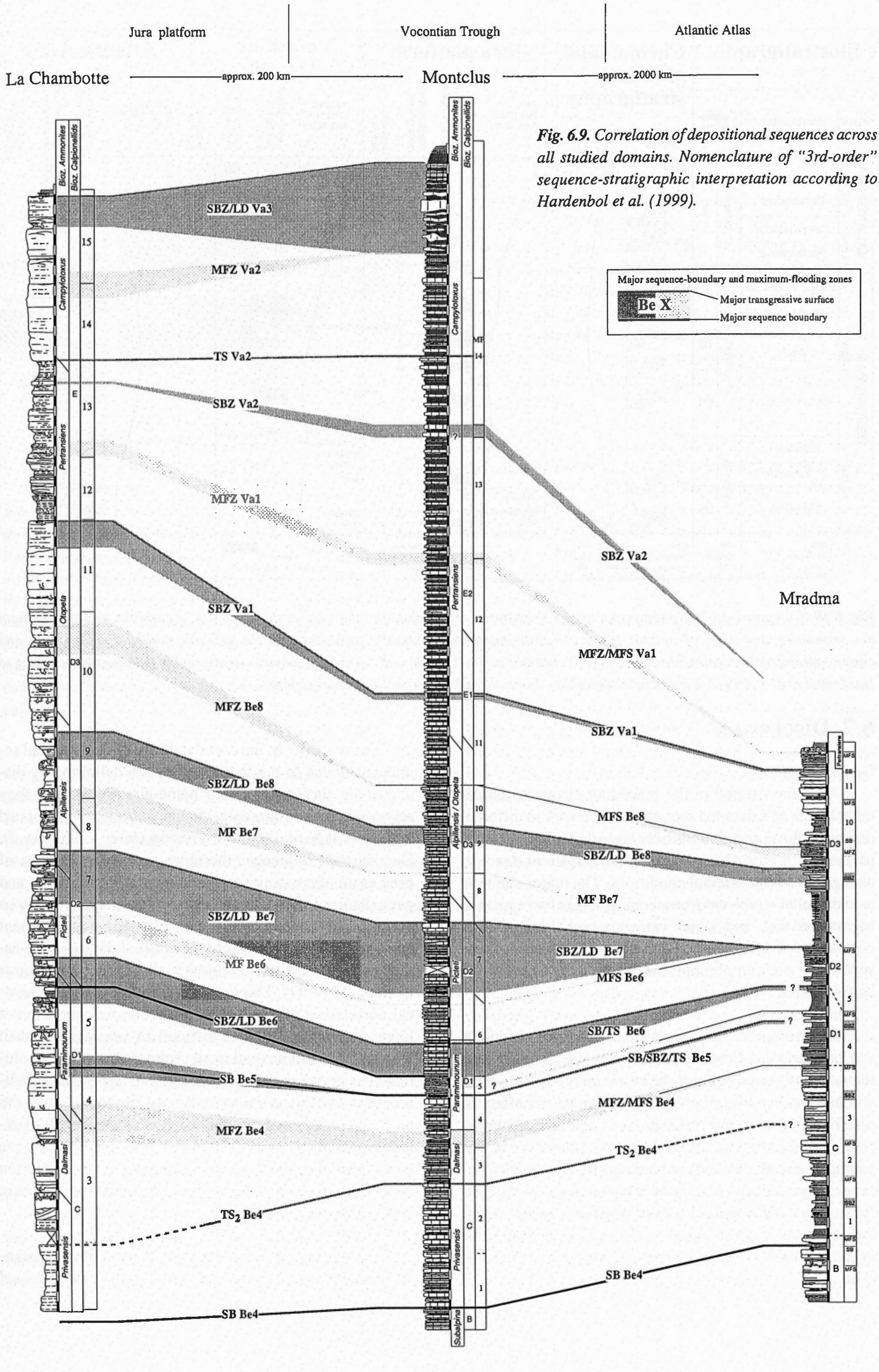


Fig. 6.9. Correlation of depositional sequences across all studied domains. Nomenclature of "3rd-order" sequence-stratigraphic interpretation according to Hardenbol et al. (1999).

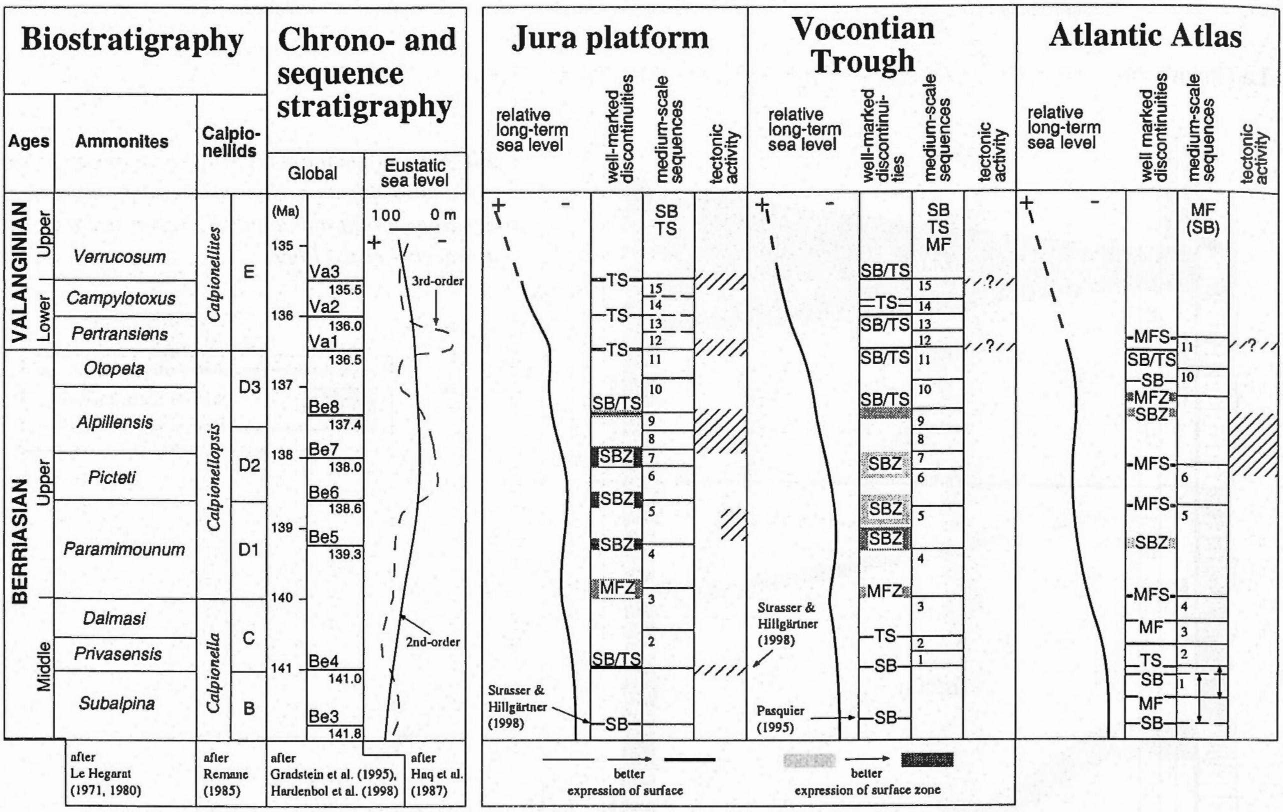


Fig. 6.10. Comparison of long-term relative sea-level trends integrated over the studied time interval, occurrence and type of main discontinuities, and times of tectonic activity (mainly increased differential subsidence) in the different studied domains. Bio- and chrono- stratigraphic scales, Haq et al. (1987) eustatic sea-level curve, and "global" sequence-stratigraphic framework according to Hardenbol et al. (1999) are given for comparison. For a detailed interpretation refer to text.

6.7. Discussion

General topics

As demonstrated in the preceding chapters, the establishment of a detailed sequence framework in different domains allows correlations over large distances and comparison of how different sedimentary systems react to changes in environmental conditions. The major aim now is to decipher which environmental changes had a local to regional extent, and which reflect a global signal. The correlation of smaller-scale depositional sequences is essential for such an attempt, but a few things have to be kept in mind:

- environmental changes, such as sea-level variations, can be marked differently in different sedimentary systems. In one region MF may be better marked, whereas in another locality SBs develop. This may be the effect of different platform morphology, sediment accumulation rates, subsidence patterns, and biological thresholds. For example, basinal sequences bounded by TS- or MF-surfaces are genetically related to SB-sequences on the platform. As a consequence, the number of sequences may be the same, but their bounding surfaces have not the same environmental significance and did not form at the same time.

- uncertainty in the correlatability of depositional sequences is due to the fact that they are delimited by discontinuity surfaces. As their name already implies, they represent an environmental change commonly associated with non-deposition and/or erosion (refer to chapter 3). Depositional sequences therefore record the passage of time as an alternating set of sedimentary increments and gaps (Sadler 1981). The time span of the gap can rarely be specified. The time increment represented by the sediment can only be specified within the entire duration of the sequence, when sufficient chronostratigraphic resolution is attained (Fig 6.11). This can have the effect that unequivocal correlation of smaller-scale sequences composing a larger-scale one becomes difficult or impossible, when their number is not equal in all sections. However, a solution that is coherent with the logic of the general sedimentary evolution commonly can be found, and the probable position of condensation can be defined. Missing depositional sequences, in fact, are important pieces in the puzzle of platform evolution and in the definition of differential accommodation (differential subsidence, platform morphology).

- sensitivity of sedimentary systems to environmental changes is critical for the formation of depositional

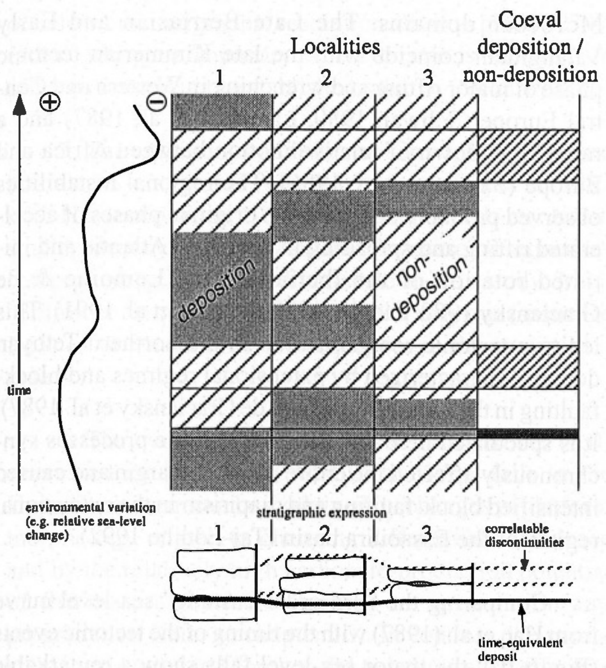


Fig. 6.11. Times of deposition and non-deposition/erosion in three different, hypothetical localities in response to the same environmental change (on any scale). The effective time-equivalent deposit in all three sections makes up only a very small part of the depositional sequence and cannot be correlated unless its extent is larger than the chronostratigraphic resolution. Discontinuities are easier to correlate, although different time-spans are represented. However, one environmental variation can create different numbers of discontinuities in different localities.

sequences. In this study, small-scale (composite) depositional sequences are the smallest correlatable unit. Elementary sequences on the bed scale are too much influenced by autocyclic processes in the moderate- to high-energy lagoonal environments, which prevailed in the study area. Other depositional settings such as the shallow restricted lagoons of the Purbeckian were sensitive enough to record climate and sea-level variations on the precessional frequency of the Earth, and elementary sequences can be correlated over large parts of the Jura platform (Strasser 1988a, 1988b, 1994).

Problematic correlations

Identification and, consequently, the correlation of sequences on the small scale are problematic in the Guiers Member and the Upper Chambotte Formation in the Jura, and in the sections located in Morocco.

Significant condensation, more important bathymetry, and dynamic depositional environments of the Lower Valanginian Formations in the Jura make it difficult to prove the lateral continuity of the small-scale depositional sequences. Therefore, the distinction between autocyclical and allocyclical control becomes difficult. However, the average thickness of small-scale sequences in the rest of the succession is matched by the thickness of sedimentary packages defined by surfaces of facies change and erosion in these Formations (sequences 39 to 47 in Fig. 5.26; sequences 35 to 47 in Fig. 5.28). Although differential subsidence suggests momentarily higher subsidence rates in distal positions, these sequences did not deepen continuously as indicated by generally shallowing facies that follows the initial deepening. By means of indirect reasoning it is assumed that these sedimentary packages (aggrading sequences) reflect relative sea-level variations on the small scale. This is evidenced by the basinal sections, which display elementary- and medium-scale sequences throughout, and small scale sequences in certain intervals. There they are interpreted to reflect climatic/sea-level variations on the Milankovitch frequency band (e.g., Giraud et al. 1995)

In Morocco, only medium-scale sequences are inferred, although smaller-scale variations in the stacking pattern are present. The regional sedimentary dynamics are not sufficiently studied to imply allocyclical controls, but the occurrence of such patterns indicate that environmental variations on a similar order to that observed in the Jura influenced these depositional environments.

Different hierarchies of sequences and their significance

Small-scale sequences can be well correlated on a regional scale on the Jura platform, and in many cases they can be correlated with their counterparts in the basin. Therefore, the impact of the climatic/relative sea-level variations to which the depositional environments reacted was effective at least in the Vocontian Trough and the surrounding platform areas.

Medium-scale sequences show comparable patterns and relative thicknesses regardless of different environmental conditions. They are correlatable between all studied regions and suggest a pertinent allocyclical forcing with significance at least for the Tethyan and Atlantic domains.

Large-scale sequences ("3rd-order" sequences according to Hardenbol et al., 1998) are a lot more complex. They are composed of 1 to 4 medium-scale sequences, and when compared to the extent and duration of biozones they span between 0.5 and 1.5 Mys (Fig. 6.10). They are delimited by well-expressed discontinuities or discontinuity zones that, in fact, coincide with medium-scale, sequence-bounding surfaces (Fig. 6.10). This means that they define larger-scale relative sea-level variations, which have an important effect on the expression of medium-scale sequences. Comparing these trends in the different domains, it becomes clear that it varies considerably. Whereas a maximum regression is witnessed in the Jura domain in the interval between Be5 and Be7, the same interval corresponds to a time of maximum pelagic influence in Morocco. The Vocontian Trough does not display large lowstand deposits during this time and evidences of condensation occur only in the interval from Be5 to Be6, with a continuous transgressive trend from there on. This points to a regional importance of the "3rd-order" trend.

An even larger-scale general sea-level evolution is clearly transgressive in all of the studied domains. It compares well to sea-level curves observed in other parts of the world and compiled "global" sea level evolution (Fig. 6.12). The only exception is the "global" sequence-stratigraphic curve by Haq et al. (1987), which displays major sea-level drops of more than 100 meters amplitude during the Late Berriasian and Early Valanginian. In the studied domains, no evidence supports such an interpretation of global sea-level evolution. Even though maximum regression on the Jura platform coincides with the first sea-level drop indicated on the curve, the platform was neither exposed entirely, nor was such an amplitude of sea-level fall compensated for by elevated subsidence rates (Jacquin et al. 1991). Schlager (1991) already discussed the "Haq-curve" in this time interval and argued for a misinterpretation of drowning unconformities at the Moroccan Atlantic margin (source data in Vail et al. 1977) as sequence boundaries. This study supports his hypothesis. Particularly the Moroccan sections display well-developed MF discontinuities and no evidence for major subaerial exposure has been found in the entire Essaouira basin (Taj Eddine 1992).

Time intervals of increased differential subsidence, as indicated by sedimentological and sequence analysis, are remarkably correlatable between the Jurassic and Moroccan domains. The Late Berriasian and Early Valanginian coincide with the late Kimmerian tectonic phase of major rifting and wrenching in Western and Central Europe (Schwan 1980, Lambeck et al. 1987) and a rapid slow-down of relative motion between Africa and Europe (Savostin et al. 1986). The regional instabilities observed probably had the same origin in phases of accelerated rifting and spreading in the North Atlantic and initiated rotation of the Iberian block (Lemoine & de Graciansky 1988, Ziegler 1988, Jacquin et al. 1991). This led to a tectonic reorganization in the northern Tethyan domain, characterized by extensional regimes and block-faulting in the Latest Berriasian (de Graciansky et al. 1987). It is speculated here that these large-scale processes synchronously affected the entire Atlantic margin and caused intensified block-faulting and diapirism in the extensional regime of the Essaouira basin (Taj-Eddine 1992).

Comparing the "3rd-order eustatic" sea-level curve from Haq et al. (1987) with the timing of the tectonic events (Fig. 6.10), the major sea-level falls show a remarkable correlatability with times of increased differential subsidence. It may be speculated that the "Haq-curve" in the studied interval rather reflects relative sea-level changes due to tectonic activity than truly eustatic sea-level variations.

"3rd-order sequences"

This leaves to be explained what "3rd-order" sequences, as identified by Hardenbol et al. (1999), reflect in the studied interval. They can neither be explained alone by global eustatic sea-level variations, nor by the tectonic events described above. It is suggested that they represent a superimposition of short-term, eustatic sea-level fluctuations (responsible for the formation of medium-scale depositional sequences), of long-term (2nd-order) eustatic sea-level rise from the Berriasian to Valanginian (Fig. 6.12), of different growth potential of the platforms, and possibly of differential flexural behavior of the plate margins (intra-plate stress sensu Cloetingh 1986, 1988). This interpretation is based on the following arguments:

1. "3rd-order" sequences can be correlated between the studied domains, but effectively it is a correlation of medium-scale sequence-bounding surfaces or entire sequences (e.g., SBZ Be7) in the "same" biostratigraphic position.

2. The fact that medium-scale sequences have an extent close to

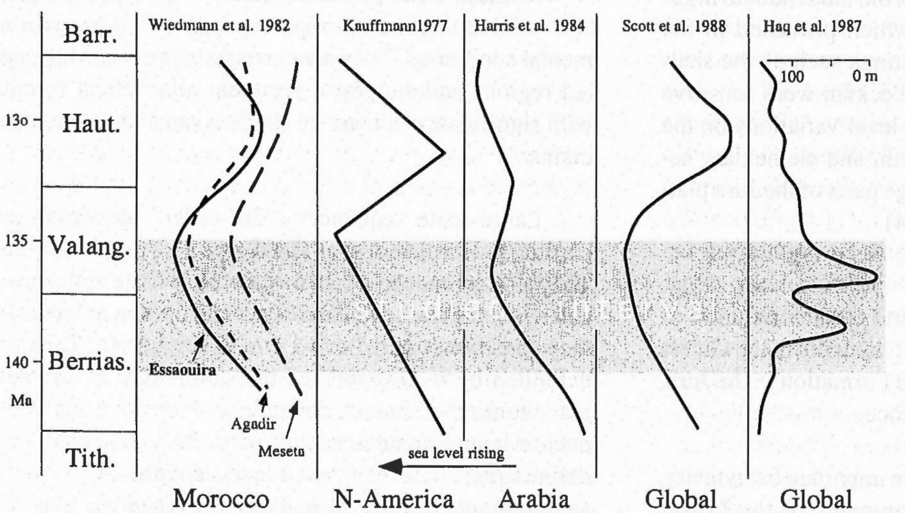


Fig. 6.12. Early Cretaceous long-term sea-level curves from various parts of the world show parallel trends, all indicating long-term eustatic control and sea level rise during the studied interval. The sequence-stratigraphic curve of Haq et al. (1987) is, however, in contradiction with the conventional record. (modified after Schlager 1991).

or below biostratigraphic resolution allow to place a "3rd-order" sequence boundary wherever one is "needed" in order to satisfy a pre-established scheme (Miall 1997).

3. The continuous addition of "3rd-order" sequence boundaries on "global" charts (compare Haq et al. 1987 with Hardenbol et al. 1998) suggests that high-resolution studies identify constantly smaller depositional sequences corresponding to the medium-scale in this study. Most "3rd-order" sequences are composed of only one (sequence Be5) or two medium-scale sequences in the studied domains.

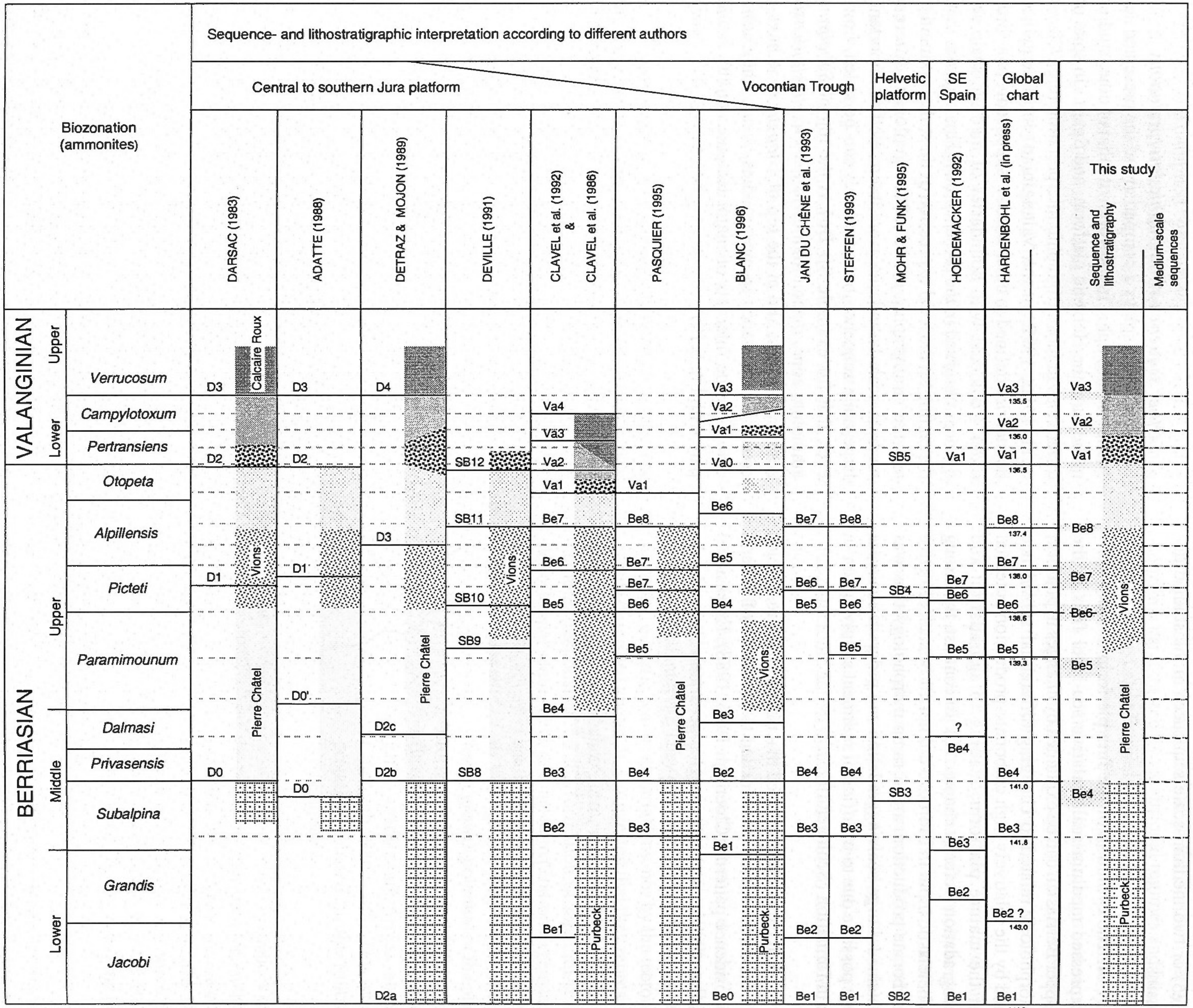
4. The evolution of the Jura platform and the well-expressed medium-scale relative sea-level falls in the Upper Berriasian can be explained by changes in platform morphology related to changing differential subsidence and by the relatively high carbonate-production potential of the mature platform, leading to aggradation and progradation (refer to chapter 10). In contrast, the strong siliciclastic input in Morocco did not allow such effective carbonate production, and a ramp morphology thus was retained. Large-scale aggradation and progradation were not possible due to insufficient sediment supply and/or in situ formation (Schlager 1993).

5. It cannot be excluded that differential stress fields in the three studied domains led to variations of general subsidence patterns (Cloetingh 1986, 1988). Extensional stress regimes at basin margins generate enhanced subsidence and an increase in the rate of onlap (Cloetingh 1988). Interpreting the observed tectonic phases as a result of changed in-plane stress patterns in an extensional tectonic regime (see above), the transgressive trends commonly succeeding the tectonic events (above Be4, Be7, Va1, Va3) may well owe a certain component to this mechanism. Comparable thicknesses of the studied successions in the different domains, however, suggest that such trends were either similar in both, the Tethyan and Atlantic domains or of limited importance.

Other sequence-stratigraphic interpretations

Fig. 6.13 displays a comparison of the presented sequence-stratigraphic framework with that of other studies in the Jura, the Helvetic platform, and Spain. In spite of differences in biostratigraphic interpretation (e.g., Clavel et al. 1986, 1992, in the Valanginian) most sequence boundaries defined in other studies (supposedly of a "3rd-order") coincide with boundaries of medium-scale sequences observed in the present study. This suggests that, depending on the local expression of these discontinuities and the stratigraphic model of the geologist, different medium-scale boundaries were interpreted as important, large-scale sequence boundaries. It can be noted, however, that for example SB Be4 at the top of *Subalpina* ammonite zone, discontinuities in the *Picteti/Alpillensis* ammonite zones, and at the top of *Otopeta* zone have a common importance in most studies. Again, they coincide with the intervals of major tectonic activity (compare with Fig 6.10).

Figure 6.13 on next page ➔



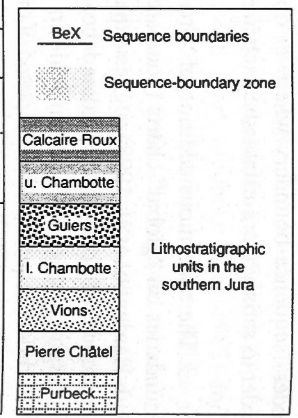


Fig. 6.13. Comparison of sequence-stratigraphic interpretations from different regions of the Tethyan realm. The presented framework of medium-scale sequences (stippled lines) displays good correlation with most other sequence boundaries indicated. For further explanation refer to text.

7 - Cyclostratigraphy

7.1. Introduction

The main quest in cyclostratigraphic analysis is to determine periodicities of identified cyclothems in relating them to a known forcing factor. The goal is to obtain a much more precise chronostratigraphic framework than biostratigraphy, in most cases, can provide. It allows for the estimation of rates of ecological, sedimentary, and diagenetic processes and gives a high-resolution time scale for the reconstruction of basin evolution (Strasser et al. 1998).

Concepts

A simple way to obtain any kind of regular rhythm is to relate it to a cyclical motion. It has already been suggested by Gilbert (1895) that sedimentary patterns correspond to parameters of planetary movement. Nowadays it is established knowledge that climatic variations on Earth, related to the distribution of insolation patterns by latitude and by season, are controlled by quasi-periodic variations of orbital parameters (Milankovitch 1941, Berger 1978, 1988). The main parameters are the Earth's cycle in axial obliquity, the cycle of precession of the equinoxes, and the changing eccentricity of the Earth's orbit around the sun (e.g., Fischer et al. 1990). The periodicities have been calculated with high confidence for the past 200 million years and indicate an actual mean duration of 20 ky for the precession, with peaks at 19 ky and 23 ky, 41 ky for the obliquity, and 100 ky and 400 ky for the two principle cycles of eccentricity (Berger & Loutre 1994). Since precession is related to spin rate it has slowed down during geological time. This results in periodicities between 18 ky and 22 ky for the precession and 37.5 ky for the obliquity in Early Cretaceous times (Berger & Loutre 1994). No evidence has been revealed that astronomical influences have significantly fluctuated in intensity during geological time or were absent at all. However, the sedimentary record of the astronomical signal may be blurred or reinforced by a multitude of other factors (De Boer & Smith 1994).

Excellent chronostratigraphic control allowed to establish a direct link between orbitally controlled insolation changes and glaciation/deglaciation cycles for the Quaternary (De Boer & Smith 1994). Polar ice-caps play an important linking role between orbital forcing and stratigraphically significant effects such as climate and sea-level change. During the Cretaceous, however, global greenhouse conditions prevailed and no evidence for polar ice caps has been revealed (Frakes & Francis 1990). Yet, the climate was not uniformly warm but consisted of intervals of warming and cooling. The early Cretaceous "cool greenhouse mode" was characterized by ice-rafting (Francis & Frakes 1993), but ice-volumes in high latitudes were not sufficient to induce purely glacio-eustatic fluctuations (Frakes et al. 1992). Other mechanisms must have contributed to eustatic variations with orbital frequencies that are widely observed in Cretaceous times (e.g., Fischer et al. 1990). They include evaporation of isolated ocean basins (see Strasser 1988a), thermal expansion of superficial ocean waters (Gornitz et al. 1982), thermally induced volume changes in deep-water circulation (Schulz & Schäfer-Neth 1998), volume changes of alpine glaciers (Fairbridge 1976), and water retention and release in lakes and aquifers (Jacobs & Sahagian 1993). High-frequency, low-amplitude (± 2 to 10 meters) changes in eustatic sea level that are thought to be responsible for the stratigraphic patterns observed in this study can be explained by such processes.

Methods

Periodicities in ancient sedimentary successions (predating the Quaternary and the Tertiary) commonly are sought by different methods, including the division of the interval between chronostratigraphic anchor points by the number of cyclothems, mathematical methods (e.g., time series analysis), and analysis of accommodation changes by means of Fischer plots. However, most of these methods comply only with relatively simple sedimentary systems, with low variability in facies, regular shallowing-up to a known bathymetric level, and without major hiatuses.

On shallow-marine carbonate platforms such methods are inappropriate or can lead to serious errors for different reasons (Strasser et al. 1998):

- bed thickness and accommodation do not have a linear relationship because of facies-dependent differential compaction. Observed bed thickness consequently does not correspond to the original accommodation;

- one cycle of environmental change can lead to the formation of multiple beds and, therefore, no constant time span can be attributed per bed (Pittet & Strasser 1998);

- difficulty in estimating accommodation potential if the sequence was entirely subtidal;

- loss of record of high frequency sea-level and/or climatic fluctuations when sedimentological and/or biological thresholds were not passed and no facies contrast was created;

- presence of autocyclic processes;

- uncertainty about the completeness of the stratigraphic record and non-deposition or non-preservation of depositional sequences;

- interference of long-term and short-term periodicities leading to enhancement and attenuation of high-frequency signals;

- imprecise chronostratigraphic data that introduce enormous error ranges in mathematical calculations of periodicities.

Sadler (1994) demonstrated that with best empirical estimates for rates of accumulation, subsidence and sea-level change, the stratigraphic record of shallow-marine carbonate environments is dominated by accommodation cycles with periods of 10 ky to 100 ky. This is without assuming regular periodic processes. Consequently, if precise chronostratigraphic data are not available (such as in the Paleozoic and most of the Mesozoic) hierarchical bundling of beds and the resulting stacking patterns are the most promising argument for Milankovitch forcing.

7.2. Stacking pattern

In the present study a consistent 4:1 or 3:1 relationship between small-scale depositional sequences and medium-scale depositional sequences can be observed (Fig. 7.1). Small-scale sequences commonly are composed of 3 to 6 beds. Although it has not been proven that all of these beds in platform environments are elementary depositional sequences due to allocyclical forcing, these hierarchical relationships are astonishingly regular. Small-scale sequences in the basin show a comparable stacking pattern, and the presence of a precessional signal was evidenced for individual marl-limestone couplets (e.g., Giraud et al. 1995). A more regular 5:1 relationship between small-scale- and elementary sequences was illustrated for the Lower- and Middle Berriasian Purbeck facies, and a precessional orbital signal was inferred (Strasser 1988a, 1994). There is no reason to assume that this orbital signal ceased in the studied interval, but environments were certainly less sensitive, and critical thresholds allowing the formation of elementary sequences were not attained.

As a working hypothesis, elementary sequences (where evidenced, and mainly basinal) are assumed to correspond to the 20-ky precession cycle of the equinoxes, the small-scale sequences to the 100-ky eccentricity cycle and the medium-scale sequences to the 400-ky eccentricity cycle. Stacking related to obliquity cycles (ca. 38 ky) could not clearly be identified, although bundling of two beds (supposedly elementary sequences) appears on the platform.

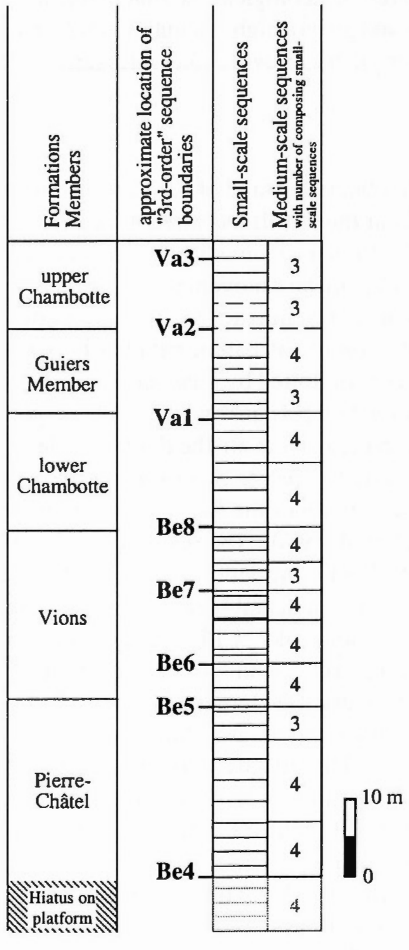


Fig. 7.1. Illustration of the dominant 4:1 relationship of small- and medium-scale sequences. The figure displays observations in distal platform positions with relative thickness of individual sequences (taken from Val du Fier and La Chambotte sections).

7.3. Timing

Absolute/relative dating of ammonite zones and "3rd-order" sequence boundaries by Hardenbol (1998), based on Gradstein (1994), allow a comparison with cyclostratigraphically deduced duration of sequences. This gives an additional reference point, although the error range in radiometric dates commonly is large. Fig. 7.2 displays the cyclostratigraphically deduced time spans. It includes the cyclostratigraphic framework for the Early and Middle Berriasian (Strasser & Hillgärtner 1998) in order to indicate the two radiometrically dated sequence boundaries Be1 and Va1. All other dates in the scale by Hardenbol et al. (1999) are relative ages based on these anchor points.

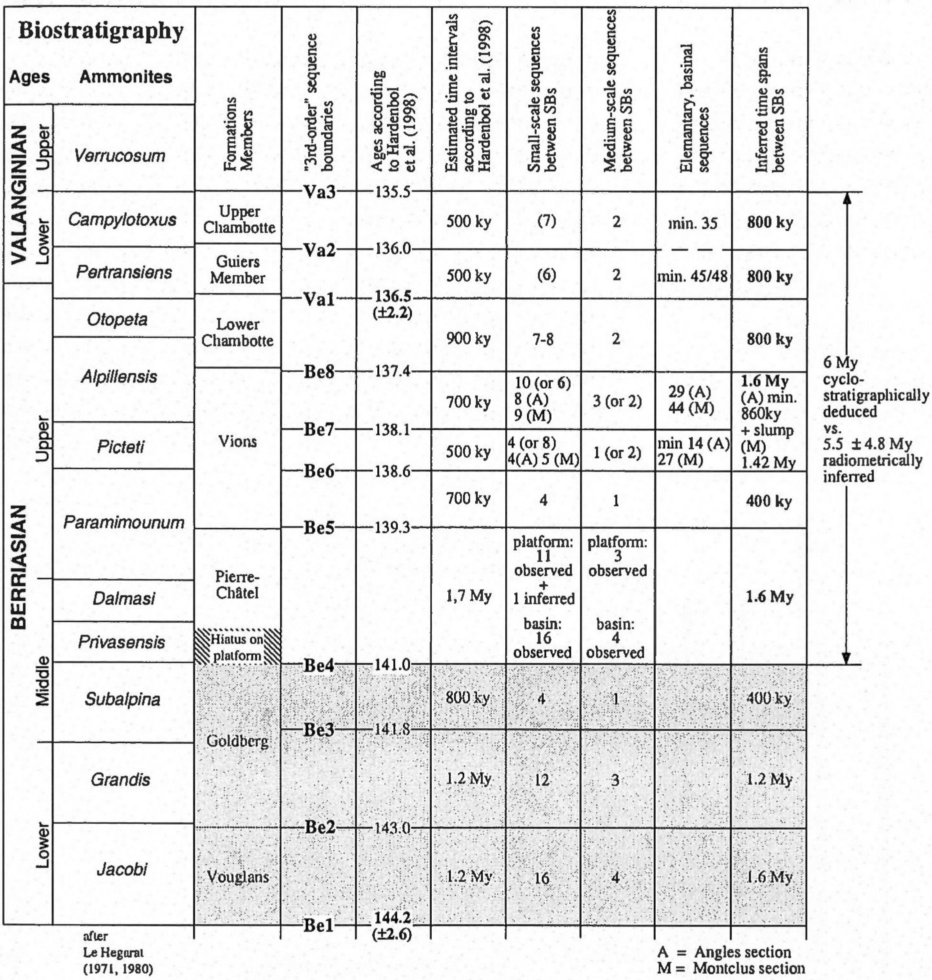


Fig. 7.2. Comparison of time represented in the studied interval. For discussion refer to text. Lower (gray) part from Strasser & Hillgärtner (1998).

Cyclostratigraphy

The duration of "3rd-order" sequences, as indicated in Fig. 7.2, can only be approximate, since it depends on the placement of the "major" boundary within a sequence boundary zone. However, most SBZs are of relatively short duration (200 ky), except for SBZ Be7.

Sequence Be4 displays only 3 medium-scale sequences on the platform compared to 4 observed in the basin (Fig. 6.6). This indicates a hiatus of at least 400 ky at this widely recognized discontinuity (Strasser & Hillgärtner 1998). The duration of sequences Be6 and Be7 cannot be defined precisely. Depending on the position of SB Be7 in the large SBZ one obtains durations between 400 ky and 800 ky, and 600 ky and 1 Ma respectively. The time interval between SB Va1 and SB Va2 is tentatively interpreted in platform environments. However, basinal sections well display marl-limestone couplets related to Milankovitch frequencies.

Giraud et al. (1995) calculated periodicities for the Valanginian of the Angles section on the basis of corrected curves of $CaCO_3$ and gray-level variations. For this calculation different durations of the Valanginian according to different authors were considered (Table 1 in Giraud et al. 1995). The result was a dominating precessional signal up to the top of the *Pertransiens* ammonite zone and a dominating obliquity signal from there on. However, recalculating their data with a duration of 5 Ma for the Valanginian, using the more recent time scale of Gradstein et al. (1995), one obtains a dominating precessional signal up to the top of the *Campylotoxus* zone. The change in orbital signature at this position coincides with a well-visible facies change to dominantly marly sedimentation. The drastic facies change, confirmed by the appearance of boreal faunas probably reflects a change in marine circulation patterns, now favoring the imprint of the obliquity signal in much lower latitudes than commonly encountered (Giraud et al. 1995).

The number of observed marl-limestone couplets consequently can give an estimate of the duration of the sequences inferred. A minimum of 45 and 48 couplets, and 35 couplets was counted for sequence Va1 and Va2, respectively (Figs. 7.2, 5.30, 5.32). Consequently, a duration of approximately 900 ky and 700 ky result, which correspond reasonably well to the duration of 2 eccentricity cycles each (800 ky). It has also to be kept in mind that the identification of SB Va2 in the basinal sections (Montclus and Angles, Figs. 5.30 and 5.32) is not evident. Superposition of a 100-ky cyclicity on the 400-ky cycle may easily have shifted the best-visible expression of the medium-scale SB.

The entire duration of the studied interval deduced on the basis of cyclostratigraphic analysis sums up to 6 My which compares reasonably well with the duration of 5.5 My estimated on the basis of radiometric data (Fig. 7.2). The total error of ±4.8 My for the radiometric data does, however, relativize this result, but at the same time reveals the potential for cyclostratigraphy to refine chronostratigraphic frameworks. Despite the imprecision inherent in stratigraphic data and errors introduced by the personal perception of the geologist, cyclostratigraphic analysis certainly attains lower error ranges.

However, the author is well aware of the danger of circular reasoning. This is easily introduced when using chronostratigraphic data to deduce periodicities, and in return justifying the chronostratigraphic scale by counting marl-limestone couplets. In this study, absolute time spans are therefore used rather to test and confirm periodicities implied by stacking patterns. In any case, however, cyclostratigraphic methods have to be used with care.

8 - Paleoclimate and environmental change

8.1. Introduction

Sediment composition and diagenetic effects are good proxies for changes in the biosphere and hydrosphere, which often are directly related to climatic changes. Such changes may include reactions of the carbonate factory to changes in oceanic circulation, nutrient input, energy conditions, salinity, and other factors (Hallock & Schlager 1986, James 1997). Changes in precipitation patterns, seasonality, and temperature most likely influence siliciclastic input and early diagenetic effects in the sediment. In the present chapter, sedimentological, geochemical, and mineralogical evidence for environmental changes, as related to variations in climate, is presented and discussed (Fig. 8.1).

8.2. Subaerial exposure and diagenesis

Climate is an important factor in controlling the type and the distribution of early-diagenetic signatures in carbonate sediments (Tucker 1993). Differences are best expressed during subaerial exposure, when other factors such as sediment accumulation rate are of minor importance. In the studied succession of the Jura domain, early-diagenetic alteration during subaerial exposure varies significantly (Fig 8.1).

The Middle Berriasian Purbeck facies displays calcrete formation, desiccation, and evaporite pseudomorphs typical for sabkha environments (Strasser 1988a). This points to a warm, equable climate and semi-

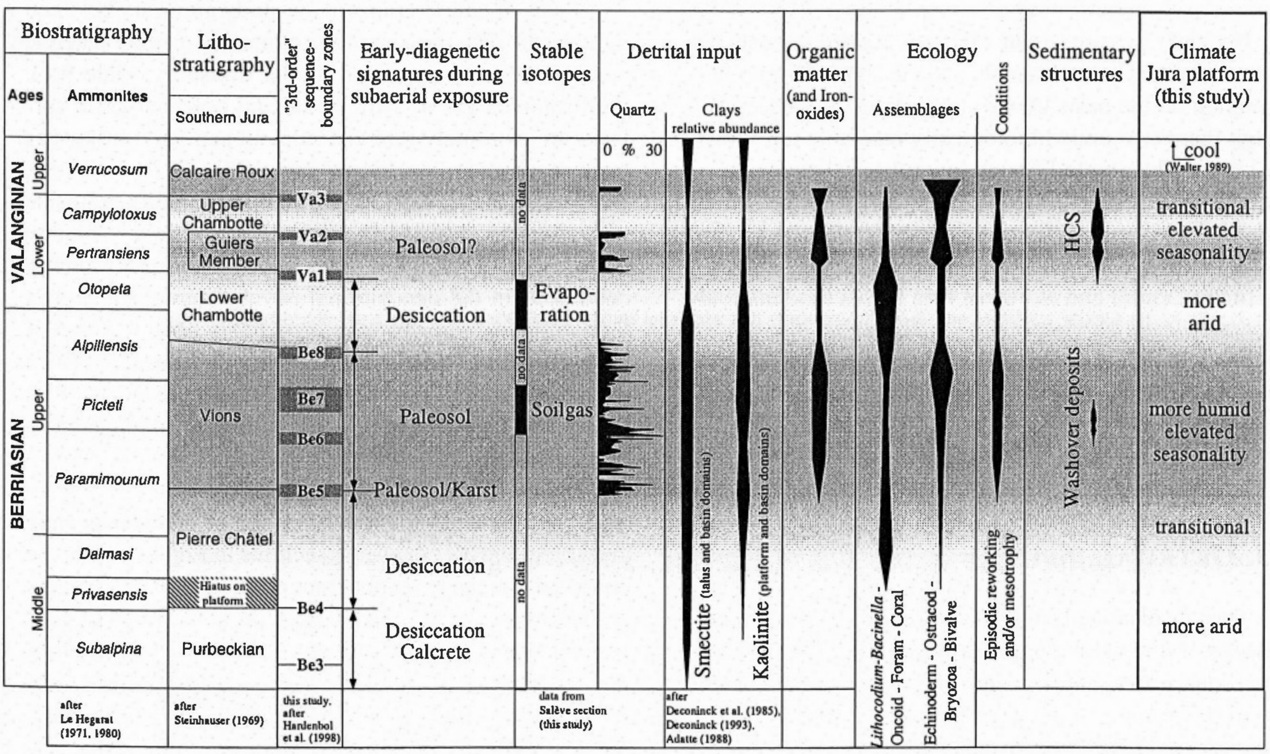


Fig. 8.1. Sedimentological, geochemical, and mineralogical evidence relevant for the interpretation of climatic conditions in the Jura domain. For discussion refer to text.

arid conditions. Emersions are less evident in the Pierre-Châtel Formation, but commonly are indicated by desiccation cracks in muddy sediments. They occupy a somewhat transitional interval culminating in subaerial exposures, which are well documented by diagenetic alteration associated with paleosol formation and karstification in the Vions Formation. Although lower sedimentation rates and presumably lower accommodation rates in the Vions Formation may have favored colonization by land plants and accelerated marine cementation (abundant firm- and hardgrounds), the formation of coal seams (e.g., Salève section) very unlikely corresponds to arid to semi-arid climates. The coexistance of charophytes indicating freshwater lakes and early diagenetic dolomite point to a slightly more humid climate or at least seasonally elevated precipitation. Early diagenetic dolomitization probably occurred in the mixing zone between a well-established meteoric lens and marine waters, and/or within the zone of circulating water ahead of the mixing zone (Tucker 1993). A change occurs towards the Lower Chambotte Formation where exposures are marked again by desiccation and/or vadose diagenesis. No clear evidence of paleosol formation has been encountered, though, even in the most proximal platform positions. Although only limited data are available, stable isotope values furthermore point to dominating evaporation, in contrast to soil-gas influence observed in the Vions Formation (Chapter 5). This suggests less favorable conditions for vegetation in presumably more arid conditions, similar to those in the Pierre-Châtel Formation.

No clear picture about climatic conditions can be obtained for the Guiers Member and the Upper Chambotte Formation on the basis of early-diagenetic evidence. Although these two units lithologically resemble the Vions and Lower Chambotte Formations, subaerial exposure is rare and only subtle evidence for paleosol formation is found in proximal platform positions at the top of the Guiers Member (Blanc 1995). However, the deep truncation in the Central and Northern Jura in this time interval is not associated with important karstification. It suggests that precipitation was reduced and/or temperatures were lower, both slowing down or inhibiting extensive karst formation.

8.3. Detrital input

The rate of siliciclastic input is dependent on climate, lithology in the catchment area, relief and distance between source area and depositional basin, and base-level elevation (Leeder et al. 1998). Higher abundance of siliciclastics on a carbonate platform suggests more intense precipitation and higher run-off, and a more efficient transport of the detrital material into the sea. The distribution of siliciclastics on the platform is dependent on the depositional environments and current regimes. This is especially true for the fine fraction and expressed in facies-dependent clay-mineral assemblages (e.g., Adatte 1988). However, large-scale trends in kaolinite abundance on the platform and smectite abundance in basinal regions display a homogeneous distribution and, therefore, can be used as proxies for climatic conditions in the studied interval (Deconinck et al. 1985, Persoz & Remane 1976, Adatte 1988).

8.3.1. Detrital quartz and associated minerals

Detrital quartz and scarce associated minerals such as zircon, feldspar, pyroxene, and tormaline occur in the Vions Formation, the Guiers Member, and at the top of the Upper Chambotte Formation. The appearance of detrital quartz in the Vions Formation is explained by fluvial input from the northwest, probably connecting the Wealden facies in the Paris basin and the Jura platform (Steinhauser & Charollais 1971, Blanc 1996). The source of the siliciclastics has to be sought for in the emergent areas of the Rheno-Bohemian (Vosges) massif, and/or of the Central Armorican massif. The well-sorted and mature nature of the siliciclastics can be either the result of long transport paths and effective sorting in coastal currents, and/or arise from erosion and reworking of older siliciclastic deposits, such as the Triassic Buntsandstein, which formerly covered the Vosges massif.

Sizes of quartz grains in the study area vary between 50 µm and 200 µm. Crystals commonly display faceted, angular shapes typical for fluvial transport (Plate 9.1). Well-rounded grains with pitted and polished grain surfaces are extremely rare and rule out aeolian transport as main mechanism for detrital input (Plate 9.2). However, factor analysis reveals no correlation of quartz grainsize with abundance of quartz, nor with abundance of muddy matrix, which is looked at as approximation of energy conditions in the depositional environment. This allows to infer that quartz grainsize is probably not a function of simple proximal-distal relationships with respect to a point source, nor controlled purely by current/wave energy. This favors the hypothesis of a recycling of older, mature (presorted) siliciclastics.

The distribution and abundance of siliciclastics are presented for five correlated sections in Fig. 8.2. It can be observed that the first appearance of quartz in the sections is slightly time transgressive from proximal to distal platform positions, suggesting slow transport paths and/or inhomogeneous distribution on the platform. The highly variable quartz abundance on the small scale throughout the sections has the consequence that simple statistical analyses of distribution patterns are of low interpretative value (regression lines, Fig. 8.2). More promising is the distribution of positive peaks in quartz abundance (envel-

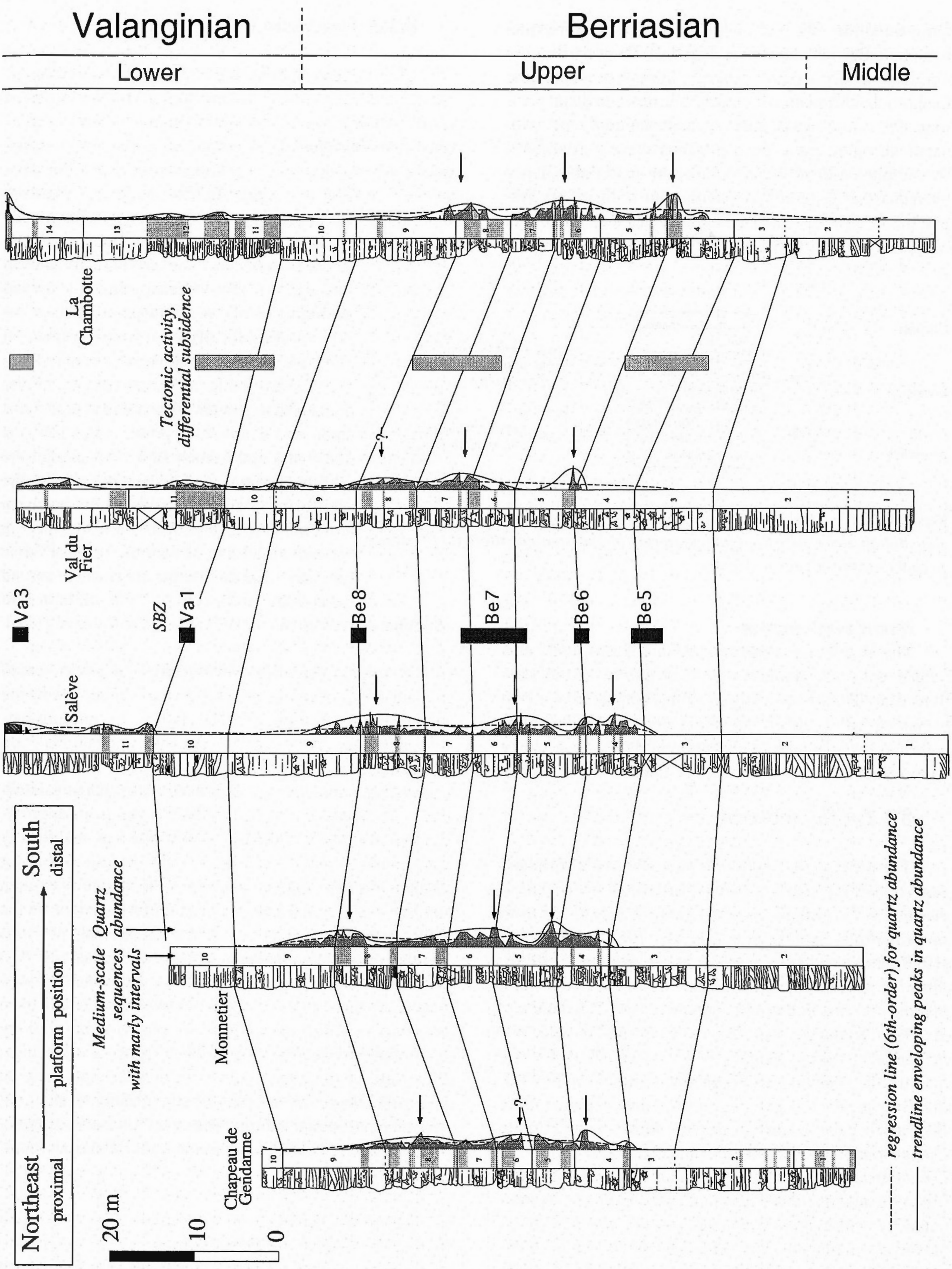


Fig. 8.2. *Abundance of siliciclastics in selected sections of the Jura platform in a correlation framework of medium-scale sequences (related to the 2nd eccentricity cycle, 400 ky). Intervals of low relative sea level and tectonic activity are indicated. For discussion of the trends in quartz abundance refer to text.*

oping trendline, Fig. 8.2), which displays three maxima in the Late Berriasian (arrows, Fig. 8.2). They are roughly correlatable between the sections, although most maxima display a diachronism with later occurrence in distal positions. On a small scale, quartz peaks commonly are associated with subaerial exposure or intertidal environments. This implies a local facies control on quartz abundance when quartz is generally available in the sedimentary system, but a relation of external input cycles with maximum availability of quartz (Fig. 8.3).

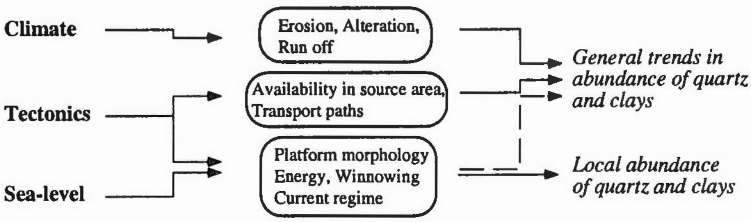


Fig. 8.3. Simple flow diagram illustrating the main factors that influence the distribution patterns of detrital material in the studied interval. Feedback mechanisms are not accounted for.

Quartz neoformation

The abundance of neoformed quartz positively correlates with abundance of detrital quartz in the Late Berriasian. The source of silica for this authigenesis is best explained with transformation and alteration of clay minerals and feldspar, and/or alteration of detrital quartz in a humid tropical climate (see also chapter 2.4.3).

8.3.2. Clay minerals

The general distribution of kaolinite in platform and basin domains and smectite in talus/basin domains during the studied interval is illustrated in Fig. 8.1. Data are from investigations by Deconinck et al. (1985), Deconinck (1993), Adatte (1988), and Persoz & Remane (1976). Smectite specifically forms under warm climates with a marked contrast in humidity (Paquet 1970, Deconinck et al. 1993). Kaolinite is preferentially formed in paleosols in humid climates (Singer 1984, Curtis 1990). Consequently, the appearance of both clay minerals in the Middle Berriasian and maximum abundance in the Late Berriasian can be interpreted as indicator for climate change to more humid and seasonal conditions than in the Early Berriasian. The diminishing clay content in the Early Valanginian coincides with more arid conditions, as indicated by sedimentological and isotopic characteristics. However, it coincides also with the renewed input of detrital quartz and organic matter in the Guiers Member. Here, seasonally humid conditions can be inferred, but it can only be speculated if lower temperatures inhibited or retarded the formation of smectite and kaolinite.

8.3.3. Discussion

Clay mineral distribution suggests more humid climatic conditions starting already during Middle Berriasian times. Detrital quartz and organic matter, however, occur only in the *Paramimounum* zone. The coincidence of first quartz appearance with a phase of elevated differential subsidence (Fig. 8.2) suggests that changes in platform morphology and probably of morphology in the hinterland are the main factors for the abrupt input of coarser detrital material. The gradual, long-term change to a more humid climate provided enough run-off and transport capacity for siliciclastics, locally even coarse grained, in the Late Berriasian. However, detrital material was available only after tectonic movements led to greater exposure of source rocks and/or to a shift of a point source towards the Swiss and French Jura. This caused the abrupt appearance of detrital material on the platform. Grainsize trends in the northern Jura indicate a delta nearby and transport in graben structures of tectonic origin (Vallée de Joux - Val de Travers area) which served as conduits in direction of the southern Jura (Steinhauser & Charollais 1971, Persoz & Remane 1976).

A simple cause-effect relationship is not sufficient to explain the 3 peaks in general quartz abundance in the Upper Berriasian (Fig. 8.2). The slightly diachronous pattern with later occurrence of the peaks in distal platform positions may suggest a relation with input cycles due to repeated tectonic activity and/or climatic changes. However, the frequency of the cycles ranges from approximately 500 ky to 800 ky. This does not match any astronomical frequency, which could be responsible for climate forcing. Tectonic activity alone may only explain the lower and upper peaks and, therefore, cannot be the only controlling factor either. However, general lowstands of relative sea-level (sequence-boundary zones, Fig. 8.2) coincide reasonably well with the peaks in quartz abundance. As these relative sea-level lowstands result from an interplay of subsidence patterns, platform morphology (carbonate productivity), and eustatic sea-level change, the origin of the cyclic pattern in detrital input may be manifold. However, trapping of siliciclastics in shallow-marine environments during sea-level lowstand certainly was an effective mechanism in the Late Berriasian (compare with Fig. 6.7).

Gradual disappearance of quartz in the latest Berriasian (Chambotte Formation) is probably related to more arid conditions. The relative transgressive sea-level trend additionally caused winnowing of detrital material, dilution of siliciclastics by high carbonate productivity, and retrogradation of point sources to more proximal platform positions. This climatic trend is corroborated by clay-mineral assemblages (Fig. 8.1) and strongly diminishing content of iron oxides, both pointing to lower rates of weathering and soil formation.

The very abrupt appearance of siliciclastics in the Early Valanginian correlates well with a phase of enhanced differential subsidence (Fig. 8.2). This suggests a dominating tectonic control triggering the availability of detrital material, which then was rapidly distributed across the platform. Erosion of a coastal cliff, where rocks rich in siliciclastics and associated terrestrial material (soils) were exposed may be a mechanism to explain the elevated content of organic material, limonite, and reworked particles of paleosols in the Lower Valanginian sediments (Blanc 1996, Kuhn 1996). An erosion of Upper Berriasian rocks in the northern Swiss Jura (area of low or negative subsidence rates) presumably provided the main source of siliciclastics. In contrast, open-marine, high-energy environments in more distal platform positions (high subsidence rates) allowed for a rapid dispersal of detrital material by long-shore currents (Blanc 1996) and tempestites. Facies-related distribution of quartz in the Lower Valanginian is underlined by the relatively lower quartz content in high-energy environments pointing to effective winnowing. The overall lower abundance of quartz (Fig. 8.2) and low abundances of smectite and kaolinite (Fig. 8.1) indicate that climate (humidity/temperature) was probably less important as controlling factor during this time interval.

8.4. Other evidence

8.4.1. Sedimentary structures

Sedimentary structures pointing to specific climatic conditions include a few washover deposits in SBZ Be7 (Crozet section) and abundant hummocky cross-stratification (HCS) in the Lower Valanginian (Fig. 8.1). Both point to repeated, intense storm activity typical for seasonal climates (Duke 1985).

8.4.2. Organic matter and nutrients

The presence of particulate organic matter is intimately related to the appearance of siliciclastics. This suggests a primary terrestrial source for the organic matter, commonly underlined by the presence of charred plant debris. Associated with input of siliciclastics and organic matter is the increased input of nutrients, stimulating intense organic activity on the shallow platform, partial eutrophication, and local anoxic conditions in the water and sediment. Eventually this led to the observed strong pyritization and blackening of limestones (chapter 2.4.3). Together they indicate a more humid climate that is favorable for vegetation in the hinterland and for efficient run-off transporting siliciclastics, organic debris, and nutrients into the sea.

8.4.3. Ecology

Fauna and flora primarily respond to changes in bathymetry and water quality (salinity, turbidity, nutrient content, temperature, oxygenation) that can be affected by variations in climate. However, on the Jura platform the main influence on organisms appears to be in form of detrital input and nutrient levels as indirect consequence of changes in climate, bathymetry and tectonic activity.

The base of the Pierre-Châtel Formation and the Vions Formation (Upper Berriasian) locally reveal a charophyte fauna, clearly indicating a brackish influence or at least near-by fresh water. *Bacinella-Lithocodium* associations show a clear anti-correlation with input of detrital material and are indicative for well-oxygenated waters with low turbidity (Dupraz & Strasser 1998). Commonly, they correlate positively with oncoids, high diversity and abundance of foraminifers (except lenticulinids), and corals, all pointing to oligotrophic conditions. Intermittent mesotrophic conditions and low oxygenation levels in the sediment are related to changing accumulation rates, high nutrient levels, and repetitive restriction, typical for shallow-marine, tidally-influenced depositional environments. Such environmental conditions stimulate the growth of fleshy algae, ahermatypic suspension feeders such as bivalves and bryozoans, grazers and/or predators such as sea urchins and gastropods, and ostracods (Hallock & Schlager 1986, James 1997). Lower in-situ carbonate productivity is indicated by discontinuous sedimentation with intensive reworking, abundance of siliciclastics and organic matter, and abundance of semi-autochthonous echinoderm debris. Caplan et al. (1996) interpreted facies changes from carbonates with a phototrophic benthic assemblage to muddy sediments with a heterotrophic benthic assemblage and high abundance of organic matter and firmground surfaces as progressive eutrophication in the euphotic zone. This stands in contrast to the typical attribution of such changes to relative sea-level rises (cf. Caplan et al. 1996), but matches well the observed ecological and environmental evidence in this study.

Analysis of the bryozoan fauna in the southern Jura revealed that the same faunal assemblage extends from the top of the Middle Berriasian to the base of the Upper Valanginian (Walter 1997). A crisis in the bryozoan fauna at this upper limit is interpreted as major cooling event in the Jura domain (Walter 1989). This in turn may indicate that temperature variations were moderate during the Late Berriasian and Early Valanginian.

Small-scale climatic cyclicity has been evidenced by microfauna assemblages and variations in ammonite fauna in basinal marl-limestone alternations of the Vocontian basin (Reboulet & Atrops 1997, Darmedru 1984, see also chapter 6). During sedimentation of calcareous beds environmental conditions are homogeneous throughout the basin and oxygenation levels are higher, both resulting in higher diversity of the microfauna within the Vocontian Trough (Darmedru 1984). It can be interpreted as an effect of accelerated oceanic circulation coupled with intensified atmospheric circulation during cooler climates (Moore et al. 1980). This observation is corroborated by ammonite fauna suggesting higher sea-level for the marl intervals and sea-level lowstand with high photic layer fertility for limestone beds in the Late Valanginian (Reboulet & Atrops 1997). Sea-level high- and lowstands testify for warmer and cooler climates, respectively, when the mechanism of volumetric changes in oceanic watermass in response to variations in temperature and insolation is assumed (e.g., Warrick 1993).

8.5. Conclusion

The climate system is difficult to understand as it depends on the interaction of many factors with short- and long-term variability and regional and global importance (Fig 8.4). The complex feedback mechanisms involved are susceptible for amplification and/or attenuation of external forcing by non-linear responses and passing of thresholds. Reactions of sedimentary systems to such forcing are more pronounced in certain latitudes and in certain sedimentary environments than at/in others (Fischer et al. 1990). Environmental forcing related to orbital cycles, for example, may globally be marked by sea-level changes (chapter 7), but associated short-term climatic variations vary regionally or even locally according to latitudinal climate belts, paleogeographic position with respect to continental terrain, and local paleogeography (e.g., Perlmutter & Matthews 1992, Pittet & Strasser 1998a, Pittet 1996). Thus, sedimentary deposits are a rather imperfect record of climate change in a localized region. Short-term variations are highly influenced by local depositional and micro-climatic conditions, but changes on the long term (medium- and large-scale sequences) may reflect oscillations between different states of the global climatic system (Hay et al. 1997).

8.5.1. Long-term climate change

The general climatic evolution from hot and arid conditions in the Middle Berriasian to a more humid and seasonal, overall more temperate climate in the Late Berriasian and Early Valanginian is consistent with the global climatic evolution proposed for this time interval (Fig 8.5, e.g., Frakes et al. 1992, Kuhn 1996).

The Helvetic platform in Eastern Switzerland underwent an evolution very similar to the one in the Jura with respect to occurrence of siliciclastics and modes of carbonate production (Figs. 8.1 and 8.5). Carbon isotope values ($\delta^{13}C$) and strontium isotope ratios from the Helvetics and the Southern Alps indicate slowly increasing weathering and run-off in the Tethyan realm, with a marked acceleration in the Late Valanginian (Fig. 8.5, Föllmi 1994, Weissert & Mohr 1996). The progressively increasing influx of detrital material and nutrients consequently led to eutrophication of the environments. Long-term sea-level rise additionally passed a threshold in opening up continental seaways between the Boreal and Tethyan domains. This led to the already mentioned cooling event, a somewhat paradoxical situation during intensifying greenhouse conditions, catchily called "greenhouse on the rocks" by Kuhn (1996). The environmental stress introduced by cold boreal waters, strongly rising greenhouse-gas levels (CO_2, CH_4, water vapor) associated with intensified detrital input, and eustatic sea-level rise led to such stressful conditions for the carbonate producers that

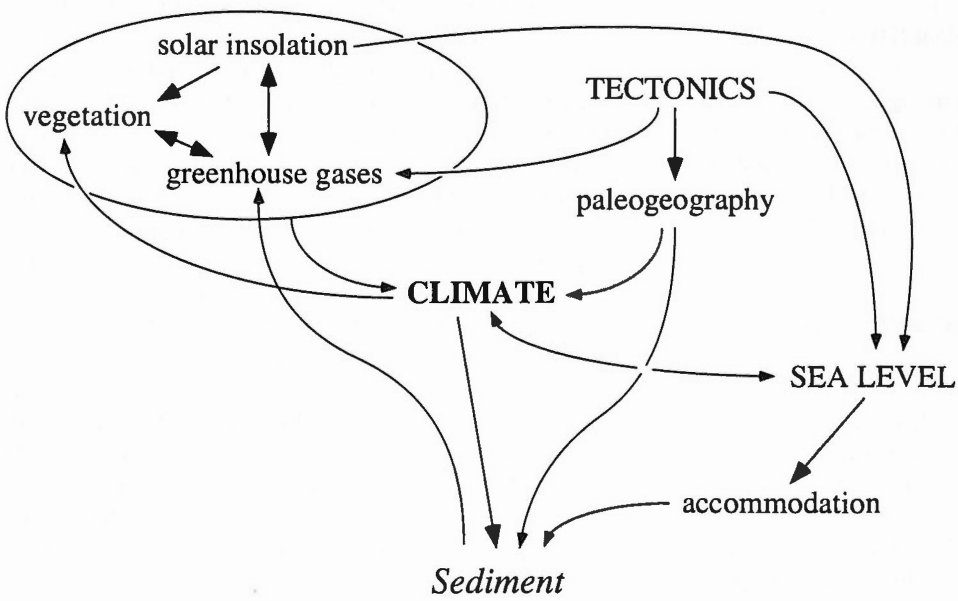


Fig. 8.4. Schematic illustration of relationships between globally and/or regionally effective parameters relating to climate. Climate and climate change are eventually recorded in the sediment by reactions of the biosphere and hydrosphere, both influencing sediment composition, diagenetic effects, and sedimentary structures (e.g. related to storm frequency).

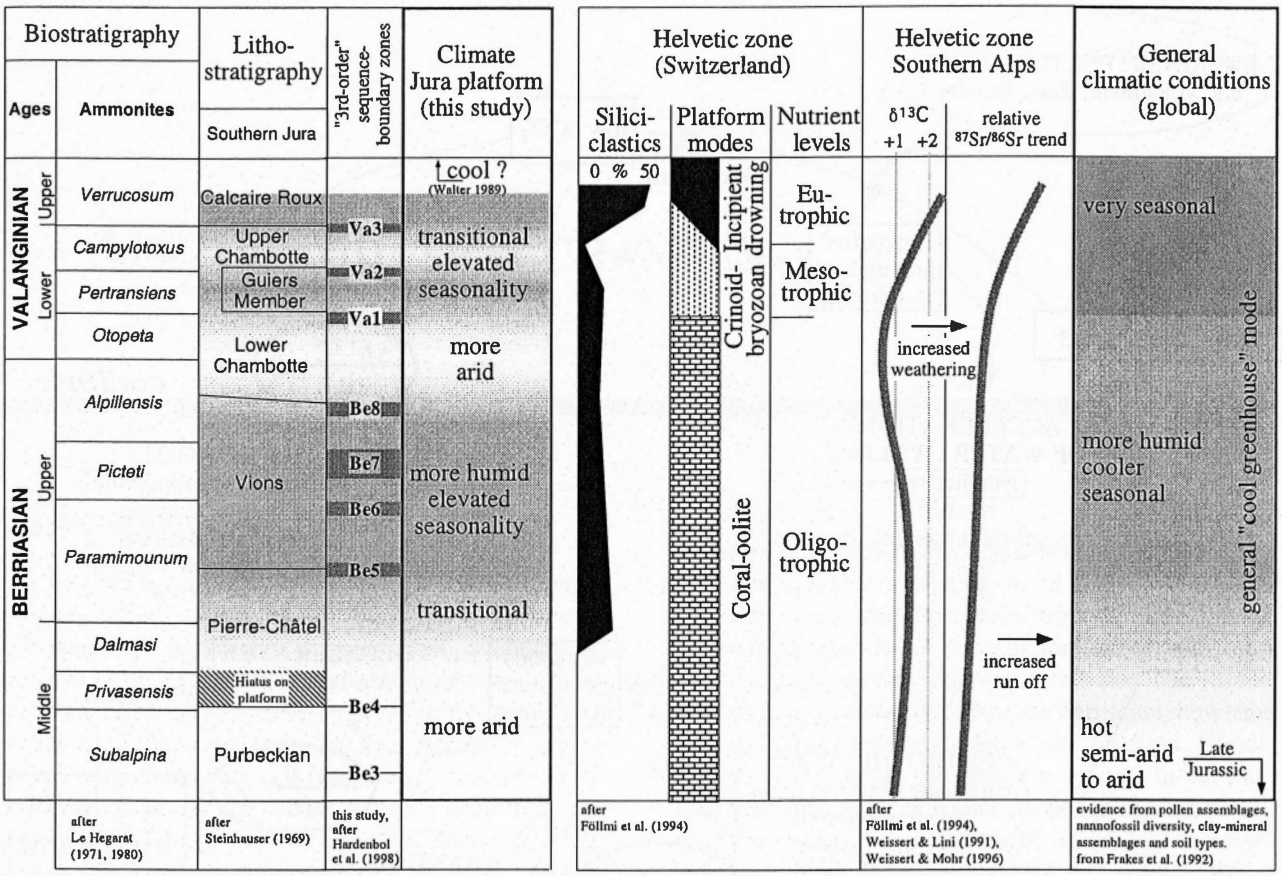


Fig. 8.5. Long-term climatic evolution in the Jura domain compared with evidence for climatic conditions from platforms in the Helvetic and the Southern-Alpine realm, and with the trend of global climatic evolution. For discussion refer to text.

large platform areas fringing the continents were suffocated and drowned in the Late Valanginian (Föllmi et al. 1994, Kuhn 1996).

Significant $\delta^{13}C$ excursions commonly correlate with platform drowning events in the Early Cretaceous and are interpreted to reflect partitioning of carbon between organic and carbonate carbon sinks (Weissert et al. 1998). This is triggered by extended episodes of intensified greenhouse climate with elevated CO_2 levels in the atmosphere. The chain of associated feedback mechanisms is illustrated in Figure 8.6. The smaller-scale variation from mesotrophic back to oligotrophic conditions indicating variable carbon burial rates observed in the Early Valanginian was not noticed in the $\delta^{13}C$ curves in the Helvetics, although oceanic $\delta^{13}C$ values display a weak shift to more positive values (Fig. 7 in Weissert & Mohr 1996).

Mesotrophic conditions in the Late Berriasian probably reflect an initial climatic cycle with elevated humidity, but with a lower intensity than in the Late Valanginian. An episode of volcanic activity, which begins at the top of calpionellid zone C (cf. Weissert et al. 1998), may have triggered an increase in atmospheric CO_2 and initiated the climate change to more humid conditions. However, the magnitudes of CO_2 input and associated climate change probably were not sufficient, so that the negative feedback leading back to oligotrophy (refer to Fig 8.6) took over before eutrophication and platform demise could occur. This demise eventually happened in the Late Valanginian (*Verrucosum* ammonite zone). Only then the carbonate carbon sinks were affected enough to cause the major shift in carbon partitioning, which is observed in wide areas of the Tethyan and the Atlantic realms (Weissert et al. 1998).

Greenhouse conditions lead to elevated evaporation rates which generally accelerates the global water cycle. Modern climate models show that in a first phase of a greenhouse climate high latitudes receive more precipitation, and that polar areas serve as water sink rather than water source (Warrick 1993). This may be seen as additional negative feedback on sea-level rise on the long term. Consequently, the formation of high-latitude ice on continental areas (suggested by ice-rafted debris; Francis & Frakes 1993) serving as water sink may point to an eustatic origin of the slowing in 2nd-order relative sea-level rise observed in the Upper Berriasian of the Jura.

The Upper Berriasian (Vions Formation) can, in conclusion, be looked at as a first, short episode of intensified

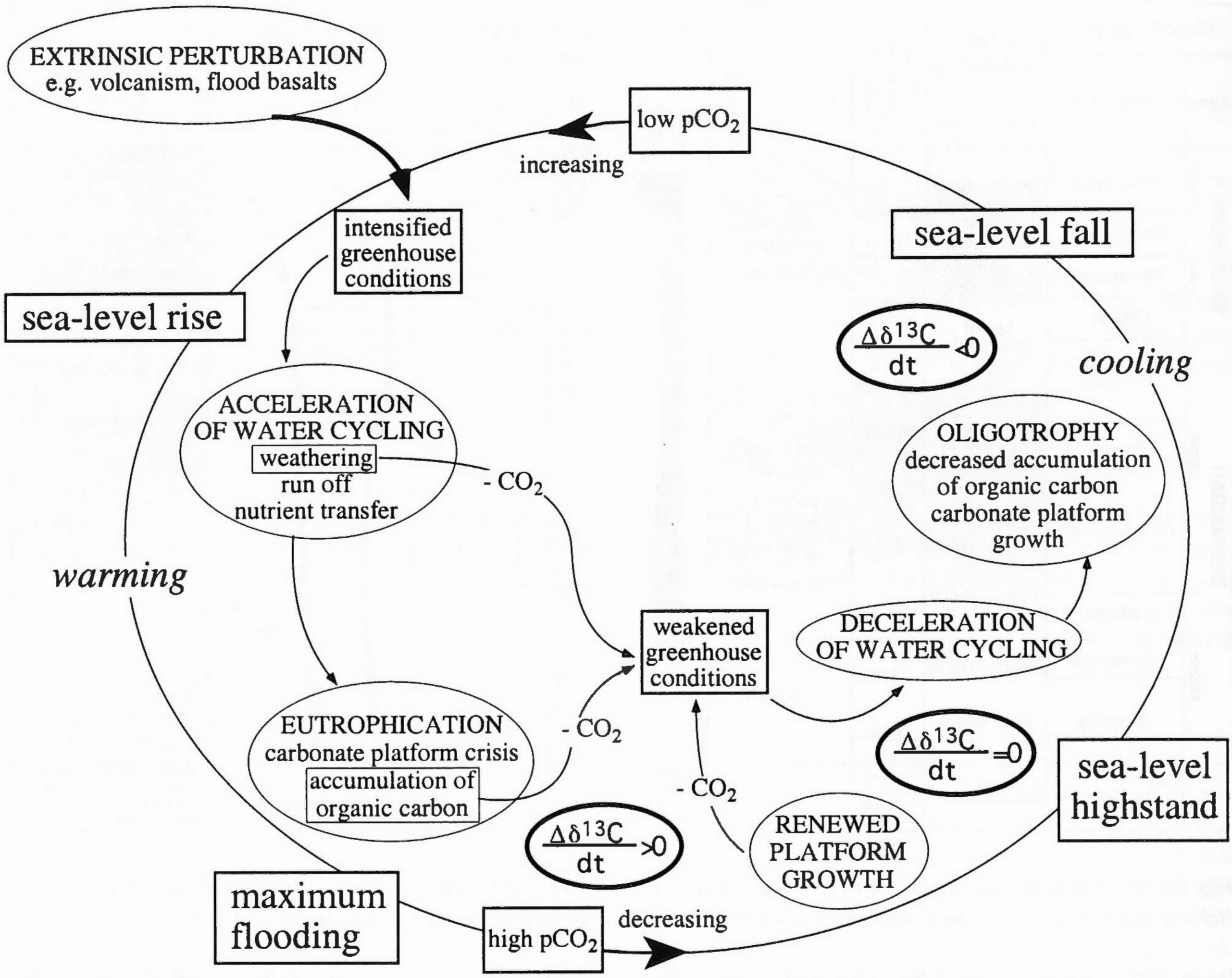


Fig. 8.6. Diagrammatic representation of links and feedback mechanisms between global carbon and global water cycle affecting climate and platform growth (after Föllmi et al.1994 and Weissert & Mohr 1996).

greenhouse conditions with elevated levels in humidity and seasonality, probably recorded only on a regional basis. However, it marks the first pulse on a large-scale trend in accelerated water cycling towards the greenhouse climax in the late Early Cretaceous (Aptian).

8.5.1. Short-term climate change

Short-term climatic changes in the studied interval are mainly recorded by the indirect effect of low-amplitude sea-level changes. Feedback relationships illustrated in Figure 8.6 possibly are independent of a time scale, but magnitude of change on a short term under general greenhouse conditions is small and unlikely to be directly recorded. Pittet (1996), however, suggested that relative abundances of bioclasts and peloids versus relative abundances of ooids, oncoids and corals can reflect lower and higher temperatures on 100 ky to 400 ky scales, respectively. In his study of the Middle Oxfordian, humidity was mirrored to a large part by the influx of siliciclastics. In the case of the Berriasian and Lower Valanginian, small-scale variations in abundance of siliciclastics and mesotrophic conditions on the shallow platform are interpreted as a function of the general climatic evolution and local dynamics of the depositional system. They are rather related to trapping and restriction during low relative sea-level, than the effect of climatically controlled input cycles. It appears that intrinsic dynamics of the sedimentary system partly obliterated shorter-term climate variations. Only very sensitive environments and/or ecosystems in platform environments, e.g., coral/microbialite bioherms, are susceptible to keep a detailed record of high-frequency climatic changes down to a precessional signal (Dupraz & Strasser 1998). In basinal environments, however, the effect of low amplitude sea-level changes and changes in water temperature in tune with the precessional frequency have large effects on circulation patterns and the nutrient cycle and, thus, are recorded basinwide.

9 - ACCOMMODATION CHANGES

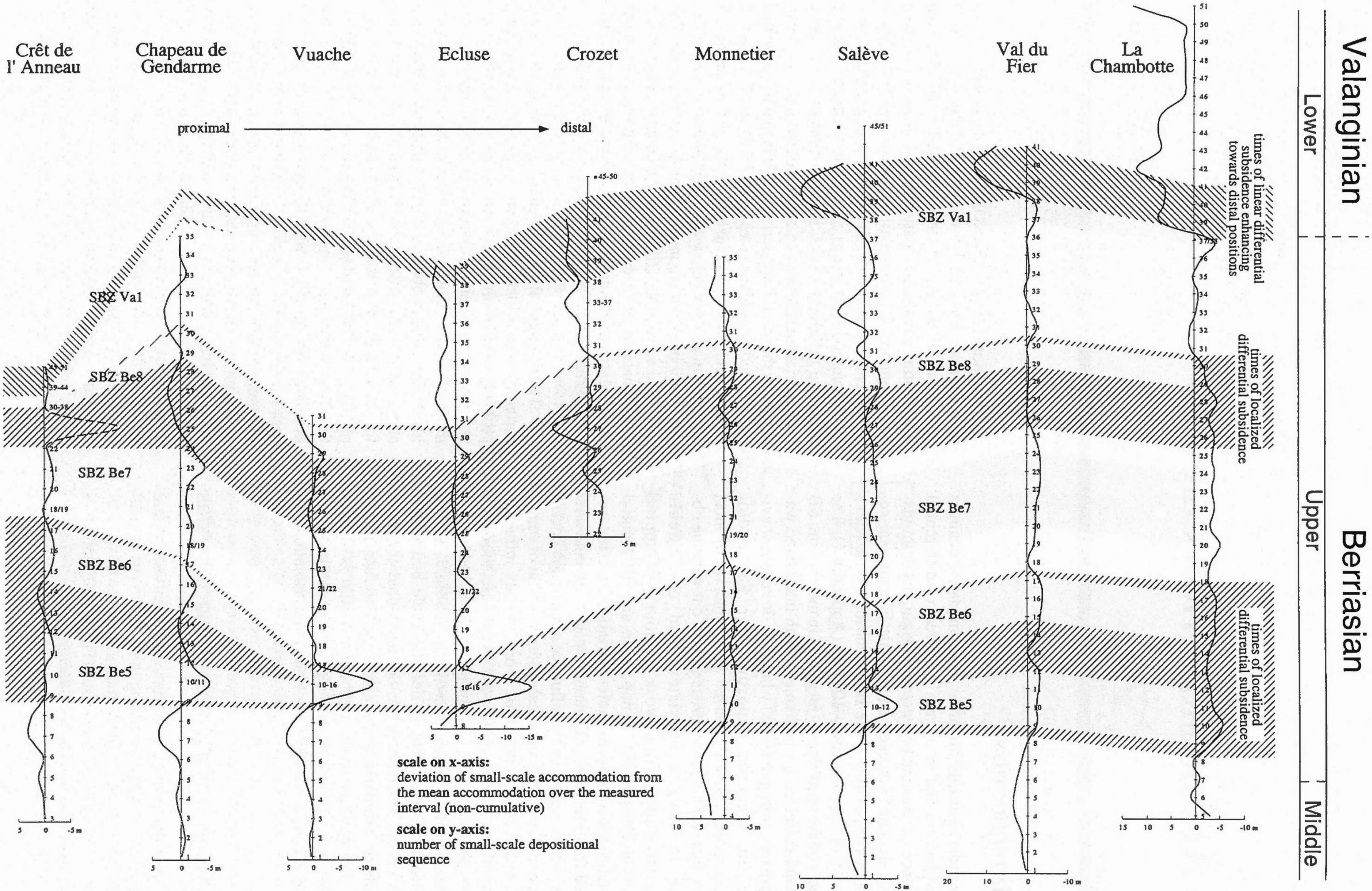
9.1 INTRODUCTION

In order to quantify relative accommodation changes on the platform and to discern the effect of differential subsidence of presumably tectonic origin, accommodation is calculated for each small-scale depositional sequence (corresponding to the 1st eccentricity cycle with a frequency of approximately 100 ky) (Appendix 1). The thickness of the sequences is measured in meters from SB to SB and corrected for bathymetry with interpreted water depth (according to facies) when SBs do not show evidence for subaerial exposure. For TS- and/or MF-sequences, SBs are inferred at the interval of most shallow facies. This procedure is carried out with the objective to have a consistent and correlatable cyclostratigraphic framework without overlapping depositional sequences. Compaction and differential compaction according to facies variations are not accounted for in this study. Content of carbonate mud and/or clay, which significantly influence compaction coefficients, commonly vary on a scale inferior to that of small-scale sequences and, therefore, would mainly influence the amplitude of the general trends. In any case, the calculated values can only be an approximation biased by limited precision of measurement in the field, errors introduced by placement of definite cycle boundaries (zones on the small-scale are not accounted for), and bathymetric estimates that are thought to reflect trends rather than absolute water depth. However, it is thought that this type of analysis provides a reasonably correct representation of trends in accommodation variation.

9.2. SHORT-TERM AND LONG-TERM TRENDS

Accommodation trends in the different sections are displayed as curves representing the deviation of accommodation from a mean accommodation value of the section for each small-scale sequence (Fig. 9.1). The mean is calculated by dividing the total accommodation of all sequences observed in the section by the maximum number of depositional sequences occurring in the corresponding time interval. The correlation of the deviation curves on the basis of cycle numbers clearly displays times of marked differential subsidence patterns and times with more homogeneous accommodation variations. The Middle Berriasian and the middle of the Late Berriasian show similar accommodation trends across the platform. Locally enhanced differential subsidence is evident in the early Late Berriasian and middle Late Berriasian and, less prominently, in the latest Late Berriasian. It is mainly expressed by low accommodation rates in SBZ Be5 and SBZ Be6 in Vuache and Fort de l'Ecluse sections, and enhanced accommodation rates below SBZ Be8 with contemporaneous low or negative accommodation (erosion?) in the Crêt de l'Anneau section. Differential accommodation with a more linear trend of enhanced accommodation towards distal platform positions is characteristic for the Early Valanginian (Fig. 9.1).

The differential accommodation on the small scale is certainly due to local variation in platform topography, probably as a result of tectonic activity and differential subsidence, and local variation in current regimes that modify platform morphology. In order to eliminate these local effects and to obtain a trend of accommodation variation, which better reflects the regional evolution, the mean for each sequence across the platform is calculated (Appendix 1). In the same way as described before, a deviation from mean accommodation is then calculated and presented in a graph. Two different ways to present the data are displayed in Figure 9.2. Small-scale variations in accommodation are best visible in the non-cumulative deviation curve compared with a cumulative deviation curve, which is comparable to a Fischer plot (e.g., Sadler 1993). Small-scale variations display changes in their trend with a frequency of 300 ky to 500 ky. Marked changes not coinciding with medium-scale SBs display a displacement of a maximum of one small-scale sequence, which is within the precision of the method (arrows in Fig. 9.2A). Regionally effective tectonic mechanisms, which could induce such short-term accommodation variations, are not


Crêt de l' Anneau
Chapeau de Gendarme
Vuache
Ecluse
Crozet
Monnetier
Salève
Val du Fier
La Chambotte
proximal
distal
SBZ Va1
SBZ Be8
SBZ Be7
SBZ Be6
SBZ Be5
times of linear differential subsidence enhancing towards distal positions
times of localized differential subsidence
times of localized differential subsidence
Valanginian
Lower
Berriasian
Upper
Middle
scale on x-axis:
deviation of small-scale accommodation from the mean accommodation over the measured interval (non-cumulative)
scale on y-axis:
number of small-scale depositional sequence

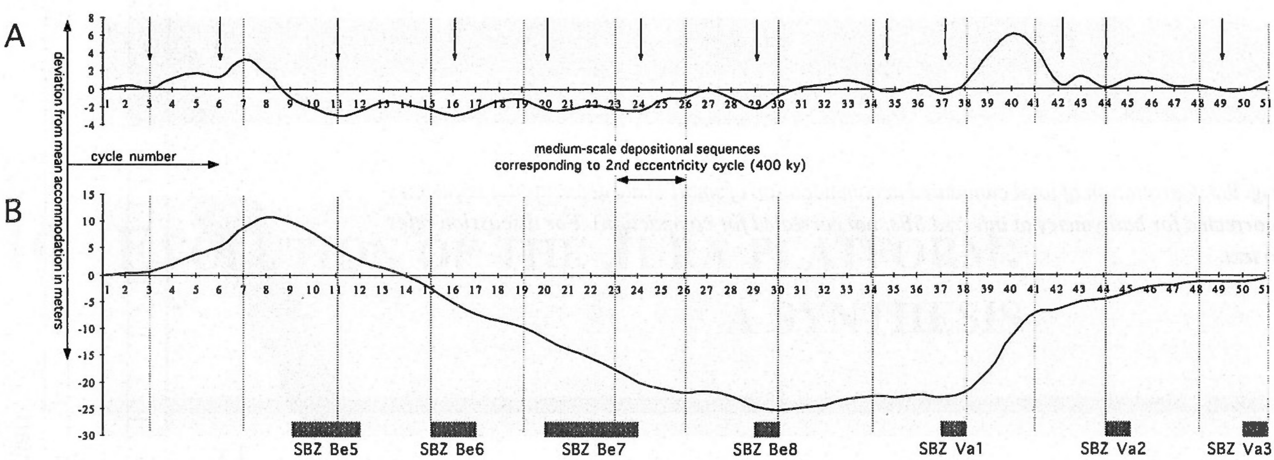


Fig. 9.2. Accommodation variation of small-scale sequences corrected for differential subsidence. (A) Non-cumulative deviation of mean accommodation across the platform (for each small-scale sequence) from mean vertical accommodation in the studied interval. (B) Cumulative deviation of mean accommodation across the platform (for each small-scale sequence) from mean vertical accommodation in the studied interval. This graph is comparable to a "Fischer plot" (Sadler et al. 1993). For discussion refer to text.

known (cf. Strasser 1991). The fact that these accommodation patterns are visible in spite of differential subsidence on the platform and show an affinity to the medium-scale sequence framework suggest that they reflect eustatic sea-level variations, probably corresponding to the 2nd eccentricity cycle. Variations in accommodation rate on the long term (Fig. 9.2B) are a composite of changes in general subsidence pattern and long-term tectono-eustasy.

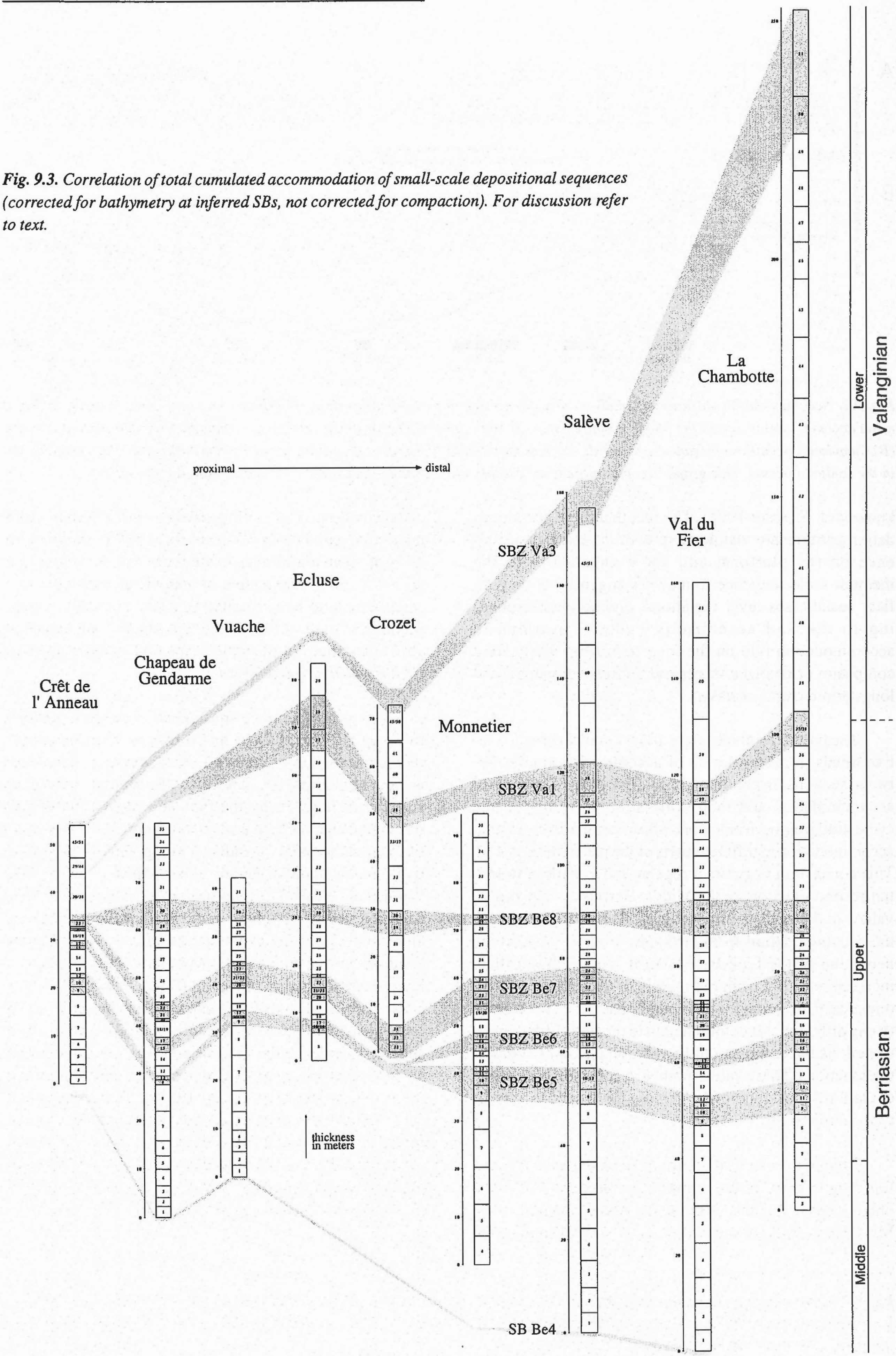
The presented graphs only allow a comparison of relative trends in the evolution of accommodation rates between sections. In contrast, cumulative plots of calculated accommodation and their correlation according to the cyclostratigraphic framework allow comparison of total accommodation in different parts of the platform (Fig. 9.3). This correlation suggests a regular and slightly differential accommodation for the Middle Berriasian with higher values in distal platform positions. The overall differential accommodation from proximal to distal areas at the beginning of the Late Berriasian is slightly lower, then increases steadily from the middle of the Late Berriasian onwards. Correction for bathymetry has the effect that the strong differential accommodation in the Early Valanginian is well visible compared to correlation plots of the measured sections. These values point to a considerable change of platform configuration from more flat-topped to a ramp-type morphology.

Roughly correcting the total measured accommodation (Fig 9.3) for 60% compaction (Strasser 1994) one obtains approximately 100 meters accommodation in 6 My for proximal platform positions, and 350 meters for distal positions. Subtracting general eustatic trends with a maximum estimate in sea-level rise of 50 meters (on the basis of mean amplitudes in the Early Cretaceous, Haq et al. 1987), 50 to 300 meters of tectonic and isostatic subsidence have to be accounted for. This results in a mean accommodation of 8 to 50 meters/My that are considerably lower than for example in the Oxfordian of the Jura (75 meters/My, Pittet 1996).

Analysis of tectonic subsidence on the Jura platform by Loup (1992b) showed that it may be seen as an indirect consequence of (crustal) extension that reactivated ancient structures. He states that "the ultimate control on subsidence is given by the type and orientation of past discontinuities", which corroborates the observed pattern of local differential subsidence along ancient fault lines (e.g., Vuache fault/strike-slip zone, Charollais et al. 1983, Wildi et al. 1989). The long-term rates of tectonic subsidence and crustal stretching in the southern Jura are generally low and display characteristics for a very landward position (on the stable European plate) with respect to a hinge zone located to the south (Loup 1992b). However, particularly the Lower Cretaceous is characterized by very low subsidence rates on the Jura platform compared to subsidence rates in the Triassic and Jurassic. Higher subsidence rates during Early Cretaceous times occur only in the southern Helvetics (Loup 1992a). This inversion of subsidence tendencies in the Early Cretaceous is interpreted by Wildi et al. (1989) and Loup (1992a) as an indication for early compressive movements related to the first eoalpine orogenic events in the Eastern Alps in the context of generally extensional tectonics.

← *Fig. 9.1. Accommodation variation calculated on the basis of thickness variation of small-scale depositional sequences (corrected for bathymetry at inferred SB's, not corrected for compaction, see Appendix 2). Times with marked differences in accommodation are indicated with cross-hatched pattern.*

Fig. 9.3. Correlation of total cumulated accommodation of small-scale depositional sequences (corrected for bathymetry at inferred SBs, not corrected for compaction). For discussion refer to text.



10 - Evolution of the Jura platform: A synthesis

"The number of variables conceptually available as controls on stratigraphic processes is almost infinite;...yet our perception in any given real geological situation is that a very small number of variables need be considered." (Smith 1994)

10.1. Introduction

The preceding chapters have already shown that factors controlling platform evolution are related in complex ways. This chapter aims to give an overview on the interplay and relative importance of tectonic, eustatic and climatic factors for platform evolution through time. Their relative role cannot be understood investigating them independently, but for the sake of clarity, variability of these factors on different scales and their potential role are summarized separately. Subsequently, their interplay and particular influence on sedimentary systems of the Early Cretaceous Jura platform are presented in chronological order. This synthesis is based on the studied sections and on data from the literature. It is, of course, to a large part hypothetical in that it presents simplified but sedimentologically most logical scenarios. Nevertheless, it allows to visualize the dynamic evolution of the study areas through time.

10.2. The role of tectonics, eustasy, and climate

10.2.1. Tectonics

Tectonic activity is the most important factor influencing the shape of the Earth's crust. On a global scale, plate-tectonic activity with changes in rates of spreading and/or rifting directly influence long-term sea-level changes (1st, 2nd, Vail et al. 1991, Mörner 1980). Similarly, such large-scale tectonic activity influences global climate changes (e.g., Föllmi 1995, Hallam et al. 1991) and local to regional climatic conditions through different paleoposition of continental and oceanic areas with respect to paleolatitudes and through mountain building processes (Matthews & Perlmutter 1994).

A general pulse in global tectonic activity (late Kimmerian phase) with accelerated spreading and rifting, and local volcanism in the North Atlantic, the Canada Basin, and the Indian Ocean may have contributed to a global second-order sea-level rise through changes in ocean-basin volume (Ziegler 1988). A global sea-level rise is documented in many regions during the Late Berriasian and Valanginian (refer to chapter 6.7). Initially, at the transition from the Late Jurassic to Earliest Cretaceous, the Kimmerian tectonic phase led to major tectonic uplift related to thermal doming in large areas of the North Atlantic and Northern Europe. This is documented by widespread erosional unconformities in Northern Europe (Ziegler 1988) and matched regionally in the Jura domain by shallow-marine to marginal-marine sediments of the Purbeckian and major unconformities marked by erosion and subaerial exposure at the base of the studied interval (SB Be3 and SB Be4, Strasser & Hillgärtner 1998, refer also to Fig. 6.13). This emphasizes the fact that global plate-tectonic dynamics control intra-plate stress patterns that have strong effects on regional and local tectonic activity and, thus, subsidence rates and patterns of the entire platform or parts of it (Cloetingh 1986, 1988). Passive margins, even mature ones, are tectonically active and display "jerky" movements along normal faults, thus developing rather a stepped subsidence curve than a smooth one (Satterley 1996, Turcotte et al. 1977).

The observed phases of tectonic activity corroborate such regionally effective episodic movements but also are correlatable over large distances (between the Jura platform, the Vocontian Trough, and the Moroccan Atlantic margin; Fig. 6.10). This suggests that changing stress regimes synchronously affected the regional and local tectonic activity in the Atlantic and northern Tethyan domains (chapter 6.7).

The transition to a generally more humid and seasonal climate towards the Valanginian points to higher levels of greenhouse gases in the atmosphere in the case of CO_2 probably related to higher volcanic activity during the phase of accelerated rifting and spreading (Berner 1990, Lini et al. 1992). The combination of such long-term climatic change and regionally effective tectonic movements in the hinterland of the Jura domain and on the Jura platform itself favored high input levels of siliciclastics and organic matter during the Late Berriasian and Early Valanginian (chapter 8).

Local subsidence rates are one of the most important factors influencing platform morphology, and thus bathymetry, distribution of depositional environments, and energy conditions. Consequently, thickness variations and sedimentological expression and type of bounding discontinuities of individual small-scale sequences are strongly controlled by local and regional tectonic activity.

10.2.2. Eustasy

Eustatic sea-level variations play an important role in the formation of depositional sequences at different scales. Their stacking pattern is the result of superimposition of different orders of sea-level change, subsidence patterns, and rates of carbonate production/sediment supply.

Long-term sea-level variations

Sea-level changes on a long term play an important role in the general evolution of depositional environments. Together with regional subsidence patterns, their influence on bathymetry and hydrodynamic conditions intimately controls ecological succession, carbonate production, nutrient availability and, thus, large-scale stratigraphic architecture (progradation and retrogradation) (e.g., Homewood 1996).

In all studied domains, a general long-term transgressive trend from the late Middle Berriasian to the late Early Valanginian is evident with an opening and deepening of the sedimentary systems, incipient platform drowning, and overall platform retrogradation. However, phases of platform progradation during Late Berriasian and Earliest Valanginian times suggest a slow-down of relative sea-level rise. To what extent this relates to long-term changes in eustatic sea level and/or changes in regional subsidence pattern cannot be clearly defined. The relative sea-level trend in Morocco, however, displays an opposite pattern, which suggests a strong tectonic control in either domain. A third factor, namely high sediment production during bathymetrically and climatically favorable conditions and advantageous platform morphology, probably caused high sediment accumulation rates and filling of available space after an initial lag phase on the Jura platform (catch up). This may have contributed to platform progradation during continuing relative sea-level rise on the long term. The Essaouira platform, in contrast, was continuously stressed by siliciclastic input and never developed such a growth potential. Similarly, accelerated differential subsidence on the Jura platform and less favorable conditions for carbonate production contribute significantly to the incipient drowning of the platform at the transition to the Late Valanginian.

Short-term sea-level variations

High-frequency eustatic sea-level changes have a much more direct influence on the formation of elementary to medium scale depositional sequences. Regionally effective tectonic activity commonly acts at longer time scales. Interference of intrinsic dynamics of sedimentary systems with the orbitally induced eustatic sea-level changes, however, determine which frequencies can be recorded and preserved. Long-term trends in accommodation change have an influence on the type of depositional sequence and bounding surfaces formed in phase with high-frequency sea-level change (chapter 5).

In platform environments of the Jura, depositional sequences related to the 1st and 2nd eccentricity cycles are well recorded and correlatable across the platform. In the Vocontian basin, however, precessional cycles primarily related to climatically controlled productivity changes are the dominant stratigraphic control. Medium-scale depositional sequences (related to the 2nd eccentricity cycle of 400 ky) correlate between all studied domains and are the fundamental stratigraphic building blocks. Zones of well-marked exposure repetitively occur in time intervals resembling 3rd-order sequences (sensu Vail et al. 1991, chapter 6.7). However, it appears that the interference between sea-level changes in phase with the 2nd eccentricity cycle and larger-scale accommodation trends (related to tectonic activity and 2nd-order eustasy), as well as climatically influenced changes in carbonate productivity led to these stratigraphic patterns rather than purely eustatic "3rd-order" sea-level variations (chapter 6.7, see also below).

10.2.3. Climate

Climate influences sedimentary systems through changes in temperature, humidity, and seasonality on different scales.

Changes in temperature directly affect sea-level by means of water-mass expansion on the long and the short term but may be the main controlling factor for the latter under a greenhouse climate (chapter 8). Changes in water temperature also have ecological effects and can influence facies distribution and modes of carbonate production (e.g., coral-ooid mode or photozoan assemblage vs.

bioclast-peloid mode or heterozoan assemblage indicating higher, respectively lower temperatures; Pittet 1996 and James 1997, respectively).

Elevated humidity leads to intensified production of terrestrial organic matter, paleosol formation, and higher siliciclastic and nutrient run off. Abundance and composition of siliciclastics and organic matter on the carbonate platform therefore have a direct relation to changes in humidity. Salinity changes through fresh-water influx in coastal regions have significant ecological and diagenetic effects (chapter 8). High nutrient influx from the continents reduces water transparency and favors the thriving of suspension-feeding animals, many of them being bioeroders. Associated higher concentrations of metals and phosphorous in bottom- and pore waters can have toxic effects for carbonate producers (Kuhn 1996). All this can seriously affect carbonate productivity and cementation (Hallock & Schlager 1986, James 1997). Such environmental changes occur on a small scale but are well recorded also by major facies changes on a longer term-scale.

Abundance of tempestites or sedimentary structures related to storm events (HCS) on the platform can be a measure for high seasonality. However, sedimentological evidence of storms in shallow platform environments with low accumulation and high intensity of bioturbation is prone to be destroyed (Davis 1995). Most of the corresponding features are therefore preserved in locations with elevated accommodation rates (Guiers Member, Upper Chambotte Formation). Abundance of smectite, a clay mineral formed predominantly in paleosols under highly seasonal climate, gives further evidence for changes in seasonality (chapter 8). During times of elevated seasonality, atmospheric and oceanic circulation is stronger. This has an effect on nutrient supply in open-marine environments and, consequently, affects planktonic productivity and autochthonous carbonate production. Stronger currents may also influence platform environments and play a role in shaping their topography. Transport of open-marine fauna to internal platform positions suggests the presence of persistent and strong tidal currents in the upper Vions Formation.

10.3. The interplay of tectonics, eustasy, and climate through time

10.3.1. Late Middle Berriasian to early Late Berriasian

This time interval comprises the *Privasensis* to *Paramimounum* ammonite zones (Fig 10.1) and has a cyclostratigraphically deduced duration of 1.6 My (Fig. 7.2). The potential for carbonate production and, consequently, platform growth was high due to a combination of semi-arid climate, warm waters, low detrital input, and enhanced water circulation on the platform, all favoring oligotrophic conditions.

Accelerated differential subsidence and sea-level rise on the long term are interpreted to have initiated the change from a flat-topped platform morphology (during the Purbeckian, stage 1 in Fig. 10.1) to a distally steepened ramp. It was the main factor enhancing water circulation in this time interval. Lateral extent of observed surface zones supports this hypothesis. Exposure commonly is restricted to proximal positions, whereas accommodation change in distal platform positions exceeded sedimentation rate, and smaller-scale sea-level drops could not lead to exposure. Here, condensation surfaces are dominant (refer also to Fig. 6.2).

After flooding of the flat-topped platform and a start-up phase in platform growth (stage 2 in Fig 10.1) high carbonate production progressively outpaced accommodation and led to progradation of the platform with shoaling in more distal positions (stage 3 in Fig 10.1). Intertidal to protected shallow-marine conditions began to prevail in most areas of the platform.

Intense plate-tectonic activity probably triggered regional tectonic movements and increase in global $p\mathrm{CO}_2$. This in turn induced a gradual rise in humidity and precipitation, which is evidenced by the first influx of detrital material in proximal platform positions (paleogeography stage 2 in Fig 10.1).

Basinal environments were characterized by low sedimentation rates and very slowly increasing marly sedimentation towards the Upper Berriasian. Gravity flows occasionally imported platform material. This points to generally low autochthonous carbonate production and low influx of siliciclastics, all indicating sluggish atmospheric/oceanic circulation (low seasonality) and elevated aridity on the long term.

10.3.2. Middle Late Berriasian

This time interval comprises the upper *Paramimounum* and *Picteti* ammonite zones (Fig 10.2). The cyclostratigraphically deduced duration is approximately 1.2 My (Fig. 7.2). The interval is characterized by reduced carbonate productivity and platform growth potential in the studied domain. This is mainly due to high detrital input and attenuated water circulation on the platform, which locally caused mesotrophic conditions.

Widespread exposure and shallow karstification of platform carbonates in the middle of the *Paramimounum* zone (SBZ Be5) suggest that not only platform progradation was advanced and platform morphology was

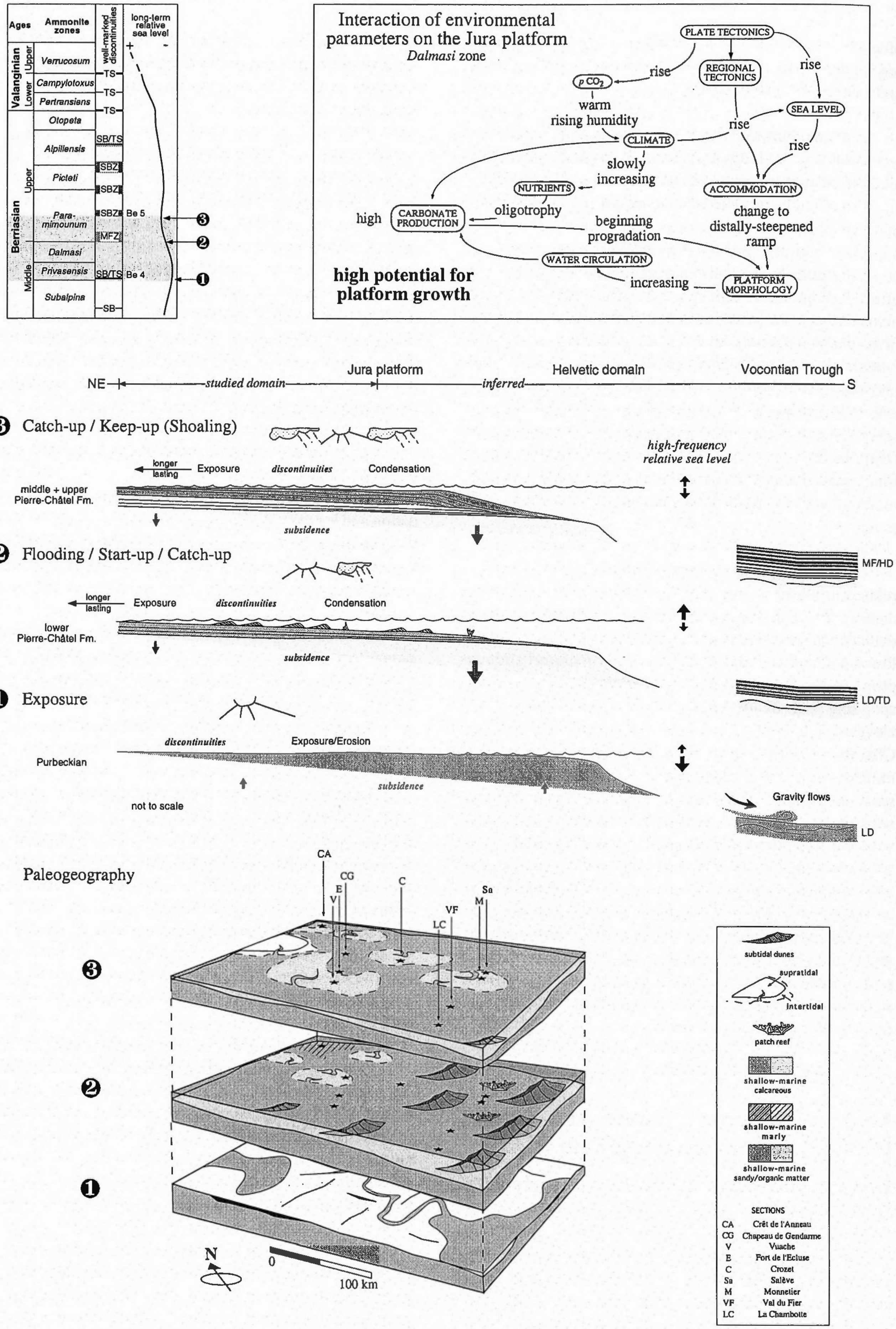


Fig. 10.1. Evolution of the Jura platform during the late Middle Berriasian and early Late Berriasian.

that of a flat-topped platform, but also that relative sea-level fell at least for a few meters (stage 1 in Fig. 10.2). This implies that long-term relative sea-level rise slowed, and climatically-induced smaller-scale sea-level falls were sufficient to expose large areas of the platform. It remains speculative to what extent tectonic uplift and/or slow-down in eustatic sea-level rise contributed to the relative sea-level fall on the longer term. Differential subsidence certainly played a role in the subsequent interval, as revealed by laterally changing accommodation rates (stage 2 in Fig. 10.2). However, no synsedimentary faults were observed in this study and faults shown in Figure 10.2 remain hypothetical. Generally, low accommodation rates on the platform were easily outpaced by sedimentation rates although carbonate production was stressed by high detrital input. The consequence was a continuing platform progradation that accentuated the flat-topped platform morphology. Short-term sea-level falls caused repetitive, well marked, and laterally extensive exposure features. Locally, correlative condensation surfaces and small-scale erosion surfaces may either indicate restricted topographic lows where subaerial exposure was not attained, or, more probably, point to erosion of all features attesting for exposure during the subsequent marine flooding (ravinement surface) (refer also to Fig. 6.2).

Platform progradation was, however, mainly fed by carbonates produced close to the platform rim. Here, conditions for carbonate production were much better due to retention of siliciclastics on the inner platform and higher-energy conditions on the outer platform, which favored water circulation. Continuous oligotrophic conditions and favorable water quality are indicated by small reefs and dominant ooid sedimentation in outer platform positions now located in the Helvetic nappes (Detraz & Mojon 1989, Pasquier 1995). Maximum progradation is attained at the end of the *Picteti* ammonite zone and indicated by tidal-flat deposits in distal platform positions. Stacking pattern of small-scale and medium-scale sequences reveals a hiatus of 100 ky in the transition of *Picteti* and *Alpillensis* ammonite zones (chapter 7). It is most probably condensed at the point of maximum progradation and lowest accommodation potential at the end of the *Picteti* zone.

Further increasing levels of pCO_2 led to intensified greenhouse conditions during the concerned time interval, with high humidity, elevated precipitation, and rising seasonality. This favored karstification, paleosol formation and high influx of detrital and organic material onto the platform.

Basinal deposits document intense slumping in dominantly calcareous sediments and only occasional input of platform material. This may also point to tectonic instability, since most slumps and gravity flows are of intrabasinal origin. Low content of marls during times of elevated humidity and precipitation corroborates a retention of siliciclastics on the surrounding platforms, at least on a regional scale.

10.3.3. Latest Berriasian

This time interval comprises the *Alpillensis* and *Otopeta* ammonite zones (Fig 10.3) and has a duration of 1.6 My on the basis of cyclostratigraphic analysis (Fig 7.2). The interaction of environmental parameters in the *Alpillensis* zone again induced more oligotrophic conditions and increasing potential for carbonate production. The high burial rate of organic matter, the intensified weathering and beginning carbonate production probably reinforced an external lowering of pCO_2, possibly already induced by lowered volcanic degassing. This resulted in reduced humidity, diminishing precipitation, and decreasing input of detritus and nutrients on the platform.

A variation in pCO_2 in the latest Berriasian is not figured on the global curve of Berner (1990), but this certainly is due to the different scale of observation. The effect of variations in the carbon cycling on climate on higher frequencies has not yet been evaluated in detail for the Mesozoic, but potentially it may be similar to the one described for the Late Jurassic or the Early Cretaceous on the long term (Weissert & Mohr 1996, Föllmi et al. 1994). It is known that the climate system has more than one equilibrium state, and that perturbations can easily trigger overturns (e.g., Broecker et al. 1985). Not only absolute pCO_2, but also its rate of change can have significant influence on thermohaline circulation and, thus, can significantly influence the entire climate system (e.g., Stocker & Schmittner 1997). In any case, the acceleration of the carbon cycle during the Early Cretaceous with its maximum in the Aptian was irregular and episodic (e.g., Föllmi et al. 1994). This suggests that extrinsic CO_2 input (volcanism) was episodic, and/or that internal feedback mechanisms and the self-regulating capacity of the ecosphere (see chapter 8) were active on many scales.

The base of the *Alpillensis* ammonite zone is characterized by a phase of accelerated differential subsidence, probably related to differential movements of individual blocks on the platform (stage 1 in Fig. 10.3). The lateral correlation of exposure surfaces in more distal positions with condensation surfaces in parts of the platform interior and the outer platform suggests the existence of a morphologic barrier (refer to Fig. 6.2). Accommodation in platform positions indicating condensation is comparable or higher than accommodation in positions that give evidence for exposure. Consequently, a major truncation surface that eroded an interval with signs of subaerial exposure, in positions where it is not observed today, seems improbable to explain this surface distribution.

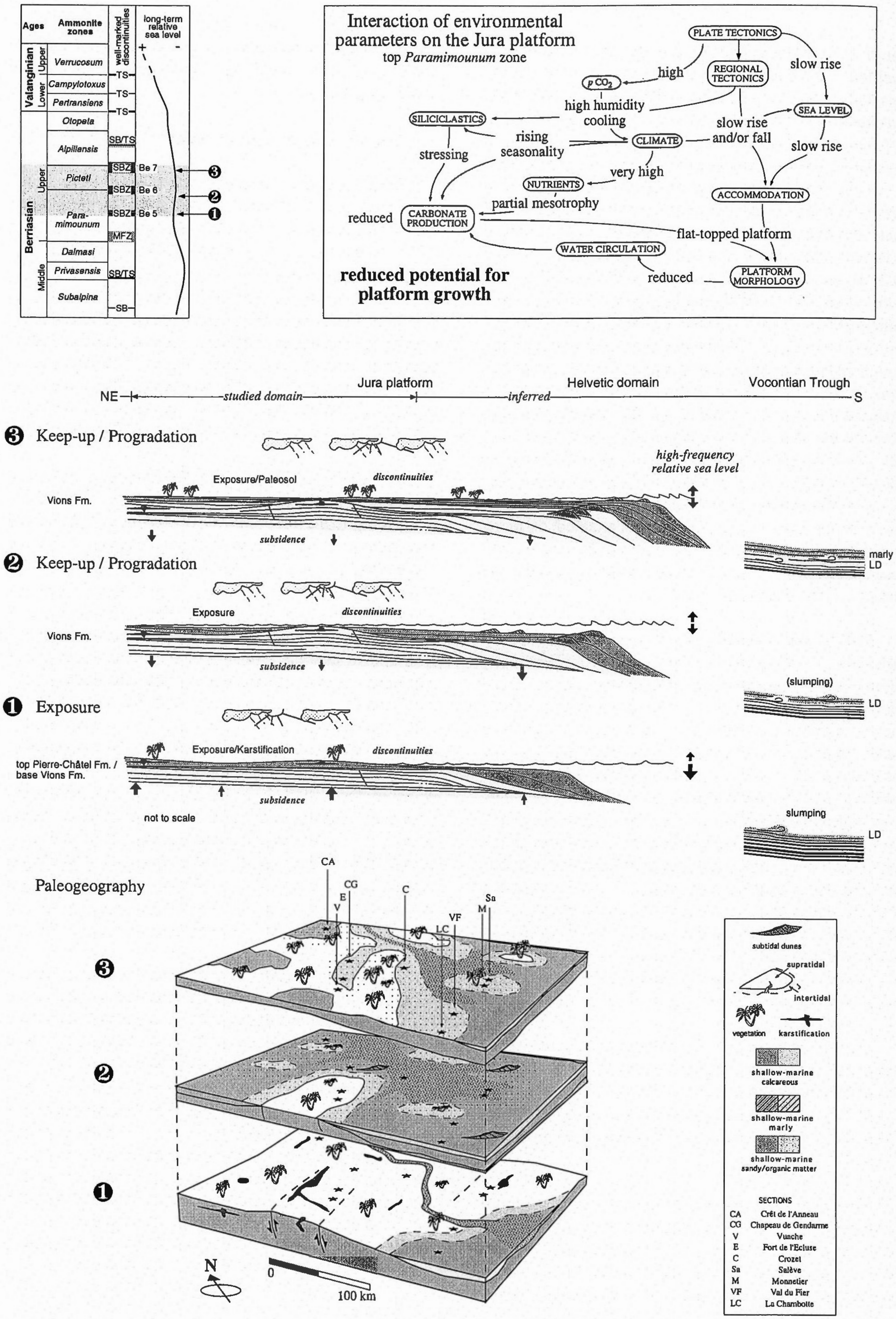


Fig. 10.2. Evolution of the Jura platform during the middle Late Berriasian.

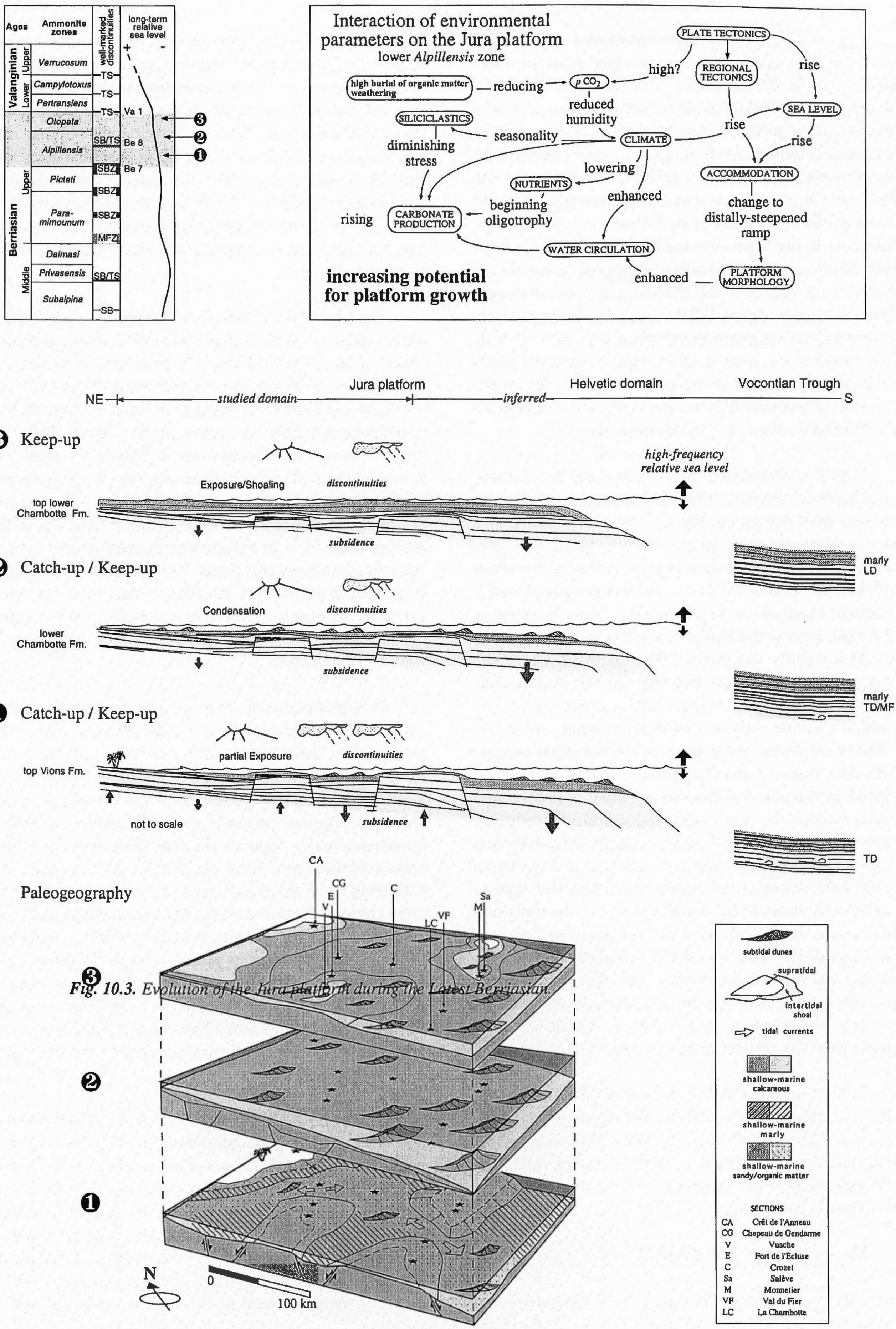


Fig. 10.3. Evolution of the Jura platform during the Latest Berriasian.

Fig. 10.3. Evolution of the Jura platform during the Latest Berriasian.

Enhanced morphology on the platform favored strong tidal currents recorded even in proximal platform positions (Chapeau de Gendarme section). Significant tidal influence is also evident in the Helvetics, where tidal delta deposits occur in the Upper Berriasian (Detraz 1989). General accommodation trends and facies evolution indicate a continued relative sea-level rise on the long term (chapter 9). High accommodation rates in topographic lows on the platform are marked by laterally continuous condensation during superimposed high-frequency sea-level rises. Short-lived rapid sea-level drops led to partial exposure of the platform but affected only areas that were morphologically elevated (SBZ Be8). Proximal platform positions lived long-lasting subaerial exposure (Crêt de l'Anneau section), whereas thick deposits of oolite dunes are characteristic for distal platform positions in the Helvetics (Pasquier 1995). Both underline the importance of differential subsidence in this time interval.

With the termination of block faulting but continuing differential subsidence in proximal and distal regions, the Jura platform again attained the morphology of a distally steepened ramp (stage 2 in Fig. 10.3). The internal platform consequently was exposed to open-marine, high-energy conditions. The quite abrupt disappearance of detrital material can be attributed to better winnowing of the platform and passing of a climatic threshold that caused a rapidly diminished detrital input. Since the Chambotte Formation and laterally equivalent units characteristically are poor in siliciclastics in the entire Jura domain, a simple shift of a siliciclastic point source to a different region cannot explain this lithological change. With progressively increasing water quality through increased circulation and diminishing input of detrital material, carbonate producers encountered gradually improving conditions. Sediment thus aggraded quickly and outpaced relative sea-level rise (stage 3 in Fig. 10.3). Small-scale sea-level variations could therefore lead to localized exposure (SBZ Va1), whereas at the transition between *Alpillensis* and *Otopeta* zones (stage 2 in Fig. 10.3) condensation during small-scale sea-level rises had predominated. According to data presented in other studies, however, the platform did not prograde as far or further into the basin as in the *Paramimounum/Picteti* ammonite zones (Detraz & Mojon 1989, Blanc 1996).

Basinal deposits display a general increase in clay content, pointing to more effective winnowing of the platform and transport paths into the basin. However, calcareous sediments are still dominant, probably reflecting reduced primary input of detrital material due to the more arid climatic conditions.

10.3.4. Early Valanginian

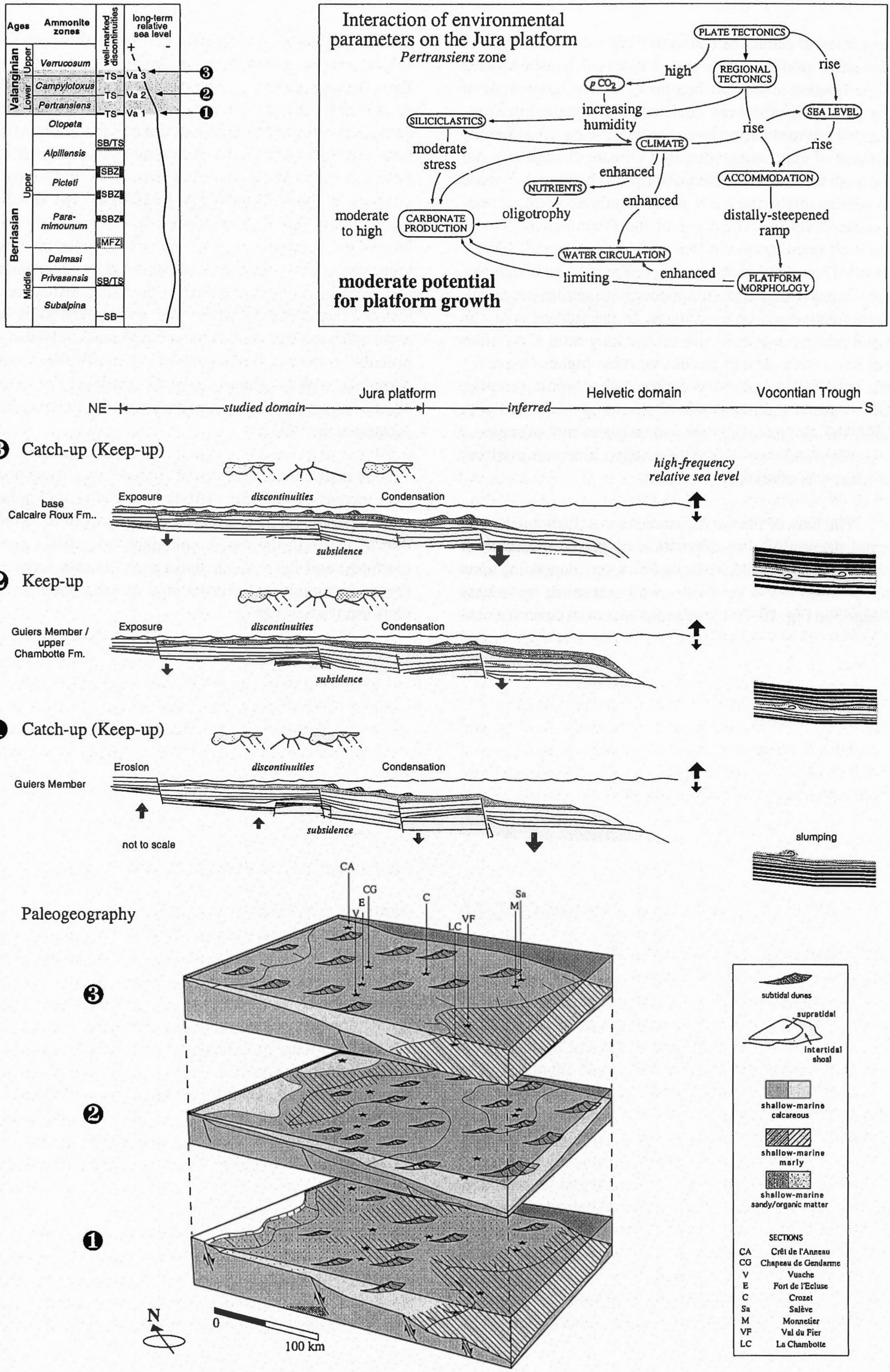
This time interval comprises the *Pertransiens* and *Campylotoxus* ammonite zones (Fig 10.4). The cyclostratigraphically deduced duration amounts 1.6 My (Fig. 7.2). Oligotrophic conditions prevailed in this time interval due to good water circulation on the platform (increased seasonality, general sea-level rise, ramp morphology), although intense tectonic activity on a regional and probably global scale caused uplift in proximal areas and erosion of older rocks rich in siliciclastics, and intensified greenhouse conditions. Carbonate production therefore resumed after incipient drowning at moderate to high levels, but rapid tectonic subsidence was not entirely compensated.

The base of the *Pertransiens* ammonite zone is characterized by an important tectonic event, which is recognized in large areas of the Jura platform, including the eastern Jura platform and the Helvetics (Mohr & Funk 1995, Detraz 1989). It is marked by tilting of large blocks and significant differential subsidence (Detraz 1989, this study). However, no synsedimentary faults are observed, but rapid lateral facies changes suggest a strong tectonic control on facies distribution. The overall gradient of differential subsidence across the platform created a distally-steepened ramp, with a slope significantly steeper than in the preceding scenarios (stage 1 in Fig. 10.4). This tilting of the entire platform in the Early Valanginian has been related to distensive movements successively affecting the Subalpine, Pre-Subalpine, and Jura domains (Detraz & Steinhauser 1988).

Proximal platform areas were exposed, and erosion along cliffs caused reworking of paleosols and older rocks rich in siliciclastics (Blanc 1996). In distal platform positions, in contrast, open-marine conditions below swell wave base and locally below storm wave base prevailed. Initially, carbonate production ceased, locally leading to condensed marly deposits (Marnes d'Arzier) and widespread formation of firm- and hardgrounds. During the following catch-up phase, rapid sedimentation and sufficient content of organic matter led to reducing conditions in the sediment and formation of "black facies" (chapter 2). Discontinuity surfaces related to high-frequency sea-level variations typically are of the condensation type. A hiatus of 400 to 800 ky in more proximal platform positions, mainly due to a lack of accommodation, can be inferred on the basis of cyclostratigraphic analysis (chapter 7).

A slow-down in the rate of differential subsidence allowed platform aggradation and progradation, and accumulation of open-lagoonal sediments in distal platform positions (Upper Chambotte Fm. s.str., stage 2 in Fig. 10.4.) Synchronously, in more proximal positions, sea-level rise induced flooding of formerly emergent areas and deposition of high-energy oolitic and bioclastic shoals (Calcaire Roux Fm. s.l.). To what extent climatic conditions influenced the reduced input of siliciclastics observed in this

Fig. 10.4. Evolution of the Jura platform during the Early Valanginian. →


Ages
Ammonite zones
well-marked discontinuities
long-term relative sea level
Valanginian
Lower
Upper
Berriasian
Middle
Verrucosum
Campylotoxus
Pertransiens
Otopeta
Alpillensis
Picteti
Para-mimounum
Dalmasi
Privasensis
Subalpina
TS
Va 3
Va 2
Va 1
SB/TS
SBZ
MFZ
SB
Interaction of environmental parameters on the Jura platform
Pertransiens zone
PLATE TECTONICS
REGIONAL TECTONICS
high
rise
p CO_2
increasing humidity
SEA LEVEL
SILICICLASTICS
rise
CLIMATE
moderate stress
enhanced
ACCOMMODATION
NUTRIENTS
enhanced
oligotrophy
distally-steepened ramp
moderate to high
CARBONATE PRODUCTION
WATER CIRCULATION
limiting
enhanced
PLATFORM MORPHOLOGY
moderate potential for platform growth
Jura platform
Helvetic domain
Vocontian Trough
NE
studied domain
inferred
S
Catch-up (Keep-up)
high-frequency relative sea level
Exposure
discontinuities
Condensation
base Calcaire Roux Fm..
subsidence
Keep-up
Guiers Member / upper Chambotte Fm.
Catch-up (Keep-up)
Erosion
Guiers Member
not to scale
slumping
Paleogeography
CA
CG
E
V
C
Sa
M
VF
LC
N
0
100 km
subtidal dunes
supratidal
intertidal shoal
shallow-marine calcareous
shallow-marine marly
shallow-marine sandy/organic matter
SECTIONS
CA Crêt de l'Anneau
CG Chapeau de Gendarme
V Vuache
E Fort de l'Ecluse
C Crozet
Sa Salève
M Monnetier
VF Val du Fier
LC La Chambotte

time interval cannot be elaborated precisely. Possibly, the less active uplift in coastal areas caused siliciclastic sources to be flooded and/or pushed back, so that large areas of the outer platform were depleted in continental detritus. Basinal environments, however, received a considerable amount of clay, suggesting that climate change was not the main factor but that the platform was bypassed. A phase of presumably slowed and subsequently accelerated relative sea-level rise at the top of the *Pertransiens* zone is not well documented in the studied domain (SBZ Va2). This SBZ probably is better marked in regions with lower subsidence rates and environments more sensitive to small-scale relative sea-level changes. In the studied area, the rapid relative sea-level rise on the long term (2nd-order eustasy + subsidence) attenuated most higher-frequency sea-level falls, even those on the 2nd eccentricity cycle. Only basinal sediments record slumping associated with SBZ Va2, but this is interpreted rather as sign of regional tectonic instability than pure eustasy, since no platform detritus was observed.

The base of the *Verrucosum* zone is distinguished by rapid deepening and deposition of marls and crinoid-bryozoan sands (Calcaire Roux Fm. s.str.), depending upon the position below or above swell and storm wave base (stage 3 in Fig. 10.4). These sediments even cover the central and northern Jura (Neuchâtel/Biel area), which were subject to exposure and erosion during at least parts of the Early Valanginian (e.g., Steinhauser & Charollais 1971, Adatte 1988, Adatte & Rumley 1989). The abrupt deepening is manifested by an important condensation surface followed by condensed marly sediments in distal platform positions, indicating incipient drowning and/or backstepping of the productive platform areas. The flooding discontinuity (TS Va3) resembles the discontinuity at the base of the Valanginian (TS Va1) and may also be of tectonic origin. This hypothesis is reinforced by major slumping occurring at the same time in the basin, testifying for tectonic instability. However, no areas undergoing tectonic uplift are known, and the relative sea-level rise most probably is the result of a general rise in subsidence rates combined with accelerated eustatic sea-level rise, which attained its regional maximum during the Hauterivian (e.g., Jacquin et al. 1991).

Climate in the concerned interval was transitional with increasing humidity, strong seasonality, and moderate temperatures, as monitored by increasing accumulation of marls in the basin, abundant tempestites on the platform, and heterozoan fossil assemblages (crinoids, bryozoans) pointing to cooler (and deeper) waters on the platform (James 1997).

11 - Conclusions

Nine sections on the Jura platform and two sections each in the Vocontian Trough and on the Essaouira platform were studied with the aim to unravel the evolution of the sedimentary environments and to understand the role of controlling factors, as well as extent and timing of environmental changes.

Data are obtained through detailed analysis of sedimentological, stratigraphical, geochemical, and biostratigraphical characteristics. Benthic foraminifers and calpionellids are the main tools to establish the large scale time- and correlation framework.

The great facies variability and rapid transitions between depositional environments suggest environmental change was heavily influenced by factors extrinsic to the sedimentary systems such as tectonics, sea level, and climate.

11.1. Facies and discontinuity analysis

Reactions of the sedimentary system to rapid and drastic environmental changes are represented rather by discontinuity surfaces that delimit homogeneous strata and/or strata with a gradual facies change than by the sediment itself. The discontinuity surfaces actually record times of the highest dynamics in environmental change. Reactions of sedimentary systems to such changes are expressed by subaqueous erosion, exposure (including any erosion in a subaerial setting), subaqueous omission, and facies changes. Changes in energy regime, relative sea level, accumulation rate, and sediment type are the principal factors leading to the formation of discontinuity surfaces.

In the Lower Cretaceous of the French and Swiss Jura, all observed small-scale discontinuities are classified according to the type of environmental change they express: subaerial exposure, subtidal condensation, subtidal erosion, and/or facies changes. Exposure surfaces and condensation surfaces form zones of repetitive occurrence in all sections. On the basis of a biostratigraphic framework, exposure and condensation zones can be correlated, although single surfaces cannot be traced in the same way. Correlation based on zones of surfaces rather than single surfaces is more realistic in shallow-water settings because such zones span a much larger time interval and therefore are less sensitive to local variations of depositional systems and lateral facies changes. Variation in the lateral continuity of surface zones and the coeval occurrence of exposure and condensation zones can therefore indicate morphological structuring of the platform and areas of differential subsidence.

The careful investigation of all discontinuity surfaces present in a sedimentary succession can reveal information about depositional processes not necessarily indicated by the sedimentary facies alone, and the results can serve as an additional tool for correlation and interpretation of platform evolution.

11.2. Sequence analysis

Integration of all sedimentological, geochemical and biostratigraphic data reveals a hierarchical organization of depositional sequences on the Lower Cretaceous Jura platform. Depositional sequences are interpreted to have been formed mainly in response to relative sea-level changes on the basis of their facies evolution and the lateral correlatability of the stacking pattern. Sea-level changes on a shallow marine platform under a greenhouse climate are a complex superimposition of high-frequency, low-amplitude, climatic eustasy in tune with orbital frequencies, of larger-scale tectono-eustasy, and of local to regional subsidence rates. Different types of depositional sequences are distinguished on the basis of bathymetric trends and types of bounding surfaces. In terms of sea-level change they are defined as TS-, MF-, and SB-sequences, reflecting the superimposition and intensification of sea-level change on different scales. They form through enhancement of initial flooding, maximum flooding and relative sea-level fall. Accordingly, stacking pattern and arrangement of the different sequence types reflect long-term evolution of relative sea level.

The advantage of this perception of sequences over the classical sequence-stratigraphic concept (sensu Vail et al. 1977, 1991) is that it integrates stratigraphic signatures in response to relative sea-level changes on all scales. It also opens a link (or closes a gap, depending on the point of view) between cyclostratigraphy and classical sequence stratigraphy.

Four hierarchies of depositional sequences are identified. Their stacking pattern on the Jura platform, in the Vocontian Basin, and on the Essaouira platform in Morocco within a common biostratigraphic framework suggest that elementary sequences on the scale of a bed reflect environmental changes in phase with the Earth's precessional frequency (20 ky). This relationship is especially well developed in basinal environments. Small-scale sequences and medium-scale sequences correspond to the 1st eccentricity cycle (100 ky) and the 2nd eccentricity cycle (400 ky), respectively. The cyclostratigraphically deduced time span of 6 My for the studied interval compares well with the radiometric data of 5.5 My (Hardenbol et al. 1998).

Sequences attributed to the 100 and 400 ky eccentricity cycles display a good correlatability over long distances from platform to basin. The 400 ky cycle can even be correlated between platforms on two different continents bordering different oceans. This underlines their potential to serve as a high-resolution chronostratigraphic tool. It has to be clearly noted, however, that the correlation of these sequences between different sections does not imply that their bounding discontinuities and/or the deposits themselves are physically correlatable. Both rather reflect environmental changes in phase with the same external forcing factor but governed by locally different environmental conditions. Consequently, their correlation is a best-fit solution and the closest deterministic approach possible to an inherently chaotic system.

A fourth hierarchy of depositional sequences can be distinguished mainly on the basis of well-expressed discontinuity surfaces or laterally correlatable clusters of discontinuities on the Jura platform. They correspond well to the 3rd-order sequence stratigraphic-framework outlined by Hardenbol et al. (1999). The duration of these sequences determined on the basis of cyclostratigraphic analysis varies between 800 ky and 1.6 My, with one sequence being identical to a 400 ky cycle. Hence, they are not governed by a regular external forcing factor. Lateral facies changes and analysis of local accommodation changes point to the importance of episodic tectonic activity in many of the time intervals displaying well-expressed discontinuities. Most of the tectonic events are correlatable between all studied domains (Jura, Vocontian basin, Moroccan shelf) and imply a common plate-tectonic influence affecting intra-plate stress patterns on a regional scale.

Depositional sequences on the Jura platform that can be correlated to the 3rd-order sequence stratigraphic framework of Hardenbol et al. (1999) are interpreted as the result of a combination of: 1. eustatic sea-level rise on a long-term, 2nd-order scale extending from the Middle Berriasian to the Hauterivian with a slow-down in the rate of rise during the early Late Berriasian; 2. high-frequency eustasy in tune with the 2nd eccentricity cycle (400 ky), which episodically attenuated or enhanced the general trend of relative sea-level change and caused the development of well-marked discontinuities; 3. episodic, tectonic activity causing local and regional changes in subsidence pattern; 4. sediment input and carbonate productivity intimately related to climate change and tectonics and controlling platform growth potential.

Consequently, sequences and discontinuities on the Jura and Essaouira platforms, which can be attributed to "3rd- order cycles" in terms of duration, thickness and stratigraphic patterns are clearly not reflecting eustasy alone. 3rd-order sea-level falls with an amplitude of more than 100 meters during Late Berriasian and Early Valanginian times, as indicated by Haq et al. (1987), cannot be corroborated.

On the basis of cyclostratigraphic analysis it can be stated that sedimentation on the Jura platform was highly discontinuous on the small scale, but it appears that no major hiatuses occurred on a platform-wide basis. High-frequency sea-level rises of a few meters amplitude always provided enough accommodation to leave a sedimentary record, except for local areas that were subject to tectonic uplift and/or strongly reduced subsidence rates. This stands in contrast to other studies that inferred considerable hiatuses over wide areas of the platform at single "3rd-order" sequence boundaries (e.g., Blanc 1996).

11.3. Platform evolution

Upper Berriasian and Lower Valanginian strata of the southern Jura platform display obvious, large-scale facies changes in the so-called "Trilogy". It is the ensemble of the lightly-colored, calcareous Pierre-Châtel Formation, the brownish Vions Formation rich in siliciclastics and organic matter, and the whitish, calcareous Lower Chambotte Formation. A similar change in facies is repeated with the more marly Guiers Member and the more calcareous Upper Chambotte Formation in distal platform positions.

The Pierre-Châtel Formation was deposited during 1.2 My on a low-angle, distally-steepened ramp mainly controlled by a large-scale eustatic sea-level rise, regionally effective differential subsidence, and high carbonate productivity. The top of the Formation is characterized by a flat-topped platform morphology attained through progradation during slowing sea-level rise and lower rates of differential subsidence. The climate evolved from semi-arid to more humid and seasonal.

The abrupt facies change to the Vions Formation can be attributed to a gradual evolution towards a more humid climate that caused weathering and formation of paleosols in the hinterland, and to tectonic activity suddenly creating pathways for continental detritus onto the platform. During approximately 1.0 - 1.2 My the platform continued to prograde, and small-scale sea-level changes in tune with the Milankovitch frequencies caused repeated exposure of wide platform areas.

The top of the Vions Formation is characterized by flooding, which mainly was the effect of increased differential subsidence and block-faulting. The rapid burial of organic matter and accelerating carbonate production served as carbon sinks and probably initiated and/or enhanced lowering of atmospheric pCO_2. This led to weakened greenhouse conditions and a climate change to more arid conditions in the Lower Chambotte Formation.

Good conditions for carbonate-producing organisms brought about a rapid platform aggradation and subsequent progradation that outpaced relative sea-level rise of probably mixed tectonic and eustatic origin.

Incipient drowning of the distal platform witnessed by the Guiers Member in distal platform positions was the effect of rapidly accelerated differential subsidence. This tectonic event is recognized widely in the Atlantic and Tethyan domains. It is interpreted as sign of enhanced plate-tectonic activity that contributed also to renewed rise in pCO_2. More humid and seasonal but cooler climate prevailed and, together with enhanced erosion of newly exposed sedimentary rocks in the hinterland, caused the revived input of siliciclastics and organic matter. The Guiers Member can, however, not be compared to the Vions Formation insofar as platform morphology, bathymetric conditions, and climate were significantly different.

The same is valid for the Upper and Lower Chambotte Formations, although their resemblance, at least in the type locality La Chambotte, is remarkable. The Upper Chambotte Formation documents regional depletion in detrital material and waning tectonic activity, which again provoked favorable environmental conditions and platform aggradation.

This evolution was suddenly stopped by renewed acceleration of tectonic subsidence and rise of relative sea level on the long term. During the rapid and platform-wide transgression recorded by the Calcaire Roux Formation, distal platform positions were subject to incipient drowning.

Environmental changes on the short term throughout the studied interval are dominated by high-frequency sea-level changes in response to insolation variations and associated changes in temperature and seasonality. These short-term global changes interact with the long-term local and regional environmental conditions outlined above. Thus, their sedimentological expression can vary considerably over short distances.

11.4. Implications and perspectives

The development of the global ecosphere through billions of years was governed by the interaction of the biosphere and its geophysical environment as defined by insolation, plate tectonics, atmosphere-hydrosphere system, and biogeochemical cycles. The present study documents an example of the intimate interactions and interdependancies of some of the variables that control environmental change.

The fundamental system properties of the ecosphere such as feedback mechanisms, regulative ability, and adaptability to changing external forcings depend not only on the functions of its subsystems, i.e., atmosphere, hydrosphere, terrestrial and marine biosphere, pedosphere, cryosphere and lithosphere. These subsystems are linked to each-other in such a way that the entire system behaves as a dynamic entity consisting of strongly interacting processes of high complexity (Schellnhuber 1997). Comprehensive sedimentological and stratigraphical analyses as demonstrated here help to unravel the history of biogeochemical cycles as the main linkage between the different subsystems of the ecosphere, and to identify conditions specifically sensible to environmental change. A high-resolution time and correlation framework is important for the quantification of short-term and long-term biogeochemical fluxes.

The results of this work open various perspectives for the future. Investigations with a similar approach and detail in different regions of the world could further refine, justify, or reject interpretations on the type, extent, and importance of environmental change. Semi-quantitative analysis of biogeochemical fluxes can be refined within the established high-resolution time framework. This opens ways to investigate the short-term self-regulating capacity of the ecosphere (homeostasis) in the geological past. Here, different methods including systems analysis as applied in modern global-change scenarios and relatively new methods such as systems analysis on the basis of non-deterministic variables (Fuzzy Logic) hold a good promise. A holistic perception of the geologic record that does justice to the complexity in ecosphere dynamics demands in any case a multidisciplinary approach.

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Plates

PLATE 1 - MICROFACIES

1. Bioclastic grainstone (bio-pelsparite) with abundant echinoderm debris and lamination. Facies L1/B3, open-marine, high-energy lagoon to bioclastic bar, sample F240.

2. Bioclastic grain-dominated packstone(bio-pelmicrite) with crinoid- and bryozoan fragments, corals (upper left corner), and intra-clasts. Facies L1b, open-marine, external lagoon, sample F131.

3. Packstone(biomicrite) with abundant sponge fragments. Facies R1, proximal ramp below storm wave base, sample F128.

4. Oncoid-rich, grain-dominated packstone (onco-bio-pelmicrite). Facies L9, open-marine lagoon with reduced sedimentation rate, sample LC13.

5. Grain-dominated packstone(bio-onco-pelmicrite) with coral debris encrusted by Lithocodium and high diversity of benthic foraminifers (textulariids, cyclamminids). Facies L5/L6, open-marine, high-energy lagoon, sample Sa138.

6. Wackestone (bio-pelmicrite) with abundant serpulids. Facies L10, open-marine, protected lagoon with reduced sedimentation rate, sample Sa0.1.

Remark: Images of microfacies are taken from thin sections in normal light analysis. Facies code refers to the classification presented in Fig. 2.5. Sample numbers indicate the respective section by their initial (e.g., Sa for Salève section).

Ramp & Open lagoon

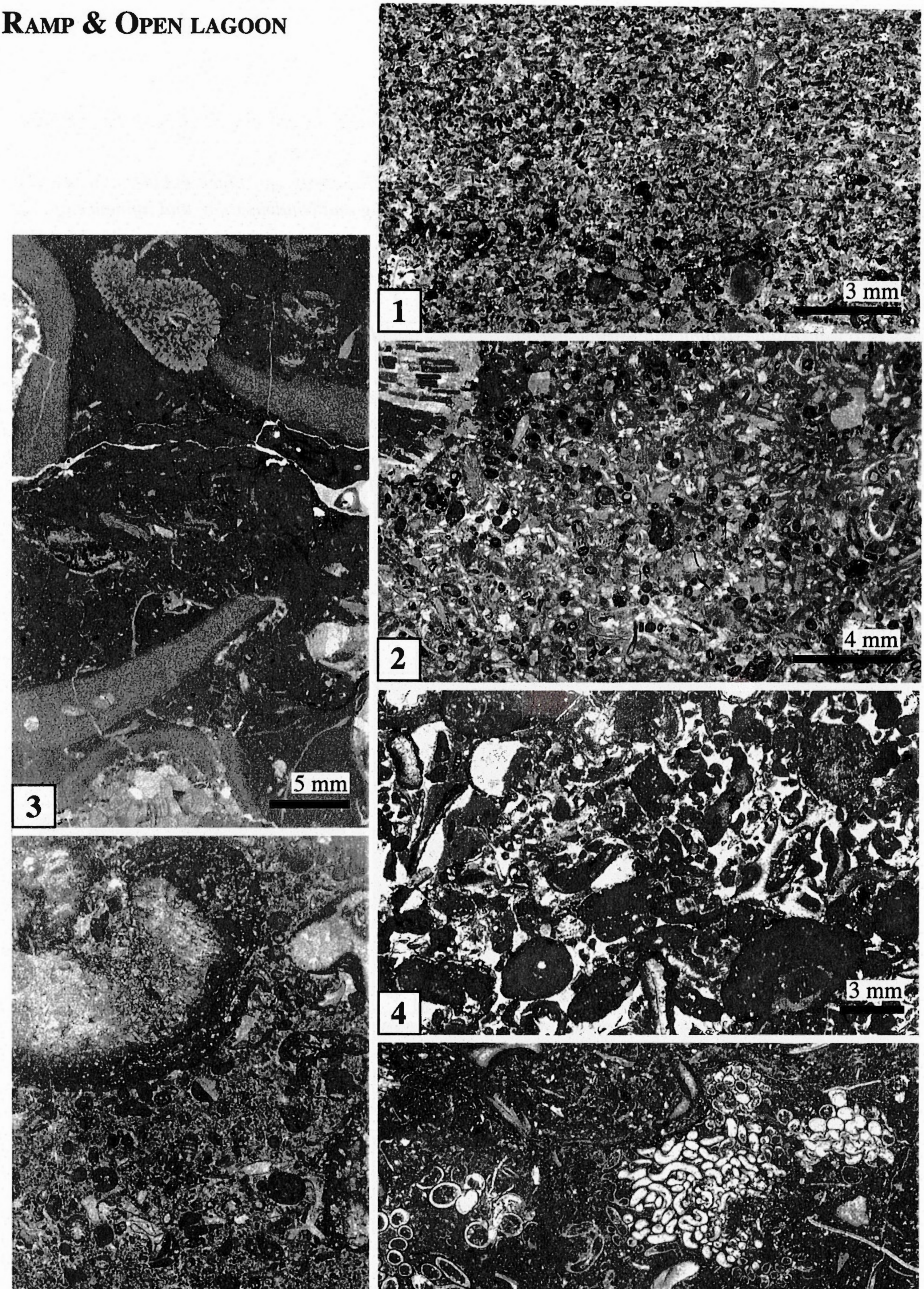


Plate 2 - Microfacies

1. Wackestone (bio-pelmicrite) with abundant dasycladales and blackened peloids. Facies L11, open-marine, protected lagoon with low-energy conditions, sample Sa21.

2. Grainstone (bio-intra-pelsparite) with abundant echinoderm fragments and black pebbles. Facies L14, shallow internal lagoon with proximal detrital influence, reworking and condensation, and influence of tidal currents, sample C67.

3. Grain-dominated packstone (oo-bio-pelmicrite) with blackened ooids (type 2a) and peloids. Facies L12, internal to open lagoon, rich in organic matter, sample R42.

4. Mud- to wackestone (biomicrtite) with abundant molluscs and detrital quartz. Facies L15, protected internal lagoon, sample R11.

5. Very well sorted grainstone to grain-dominated packstone (bio-pelsparite/micrite) with intense bioturbation. Facies L13, shallow internal to open lagoon with varying energy conditions, sample LC130.

6. Brecciated mud- to wackestone (pelmicrite) with stromatolitic microbial mats, silty, peloidal laminae, and fenestrae (arrows for all three features). Facies T1, microbial tidal flat, sample LC103.

Lagoon & Tidal flat

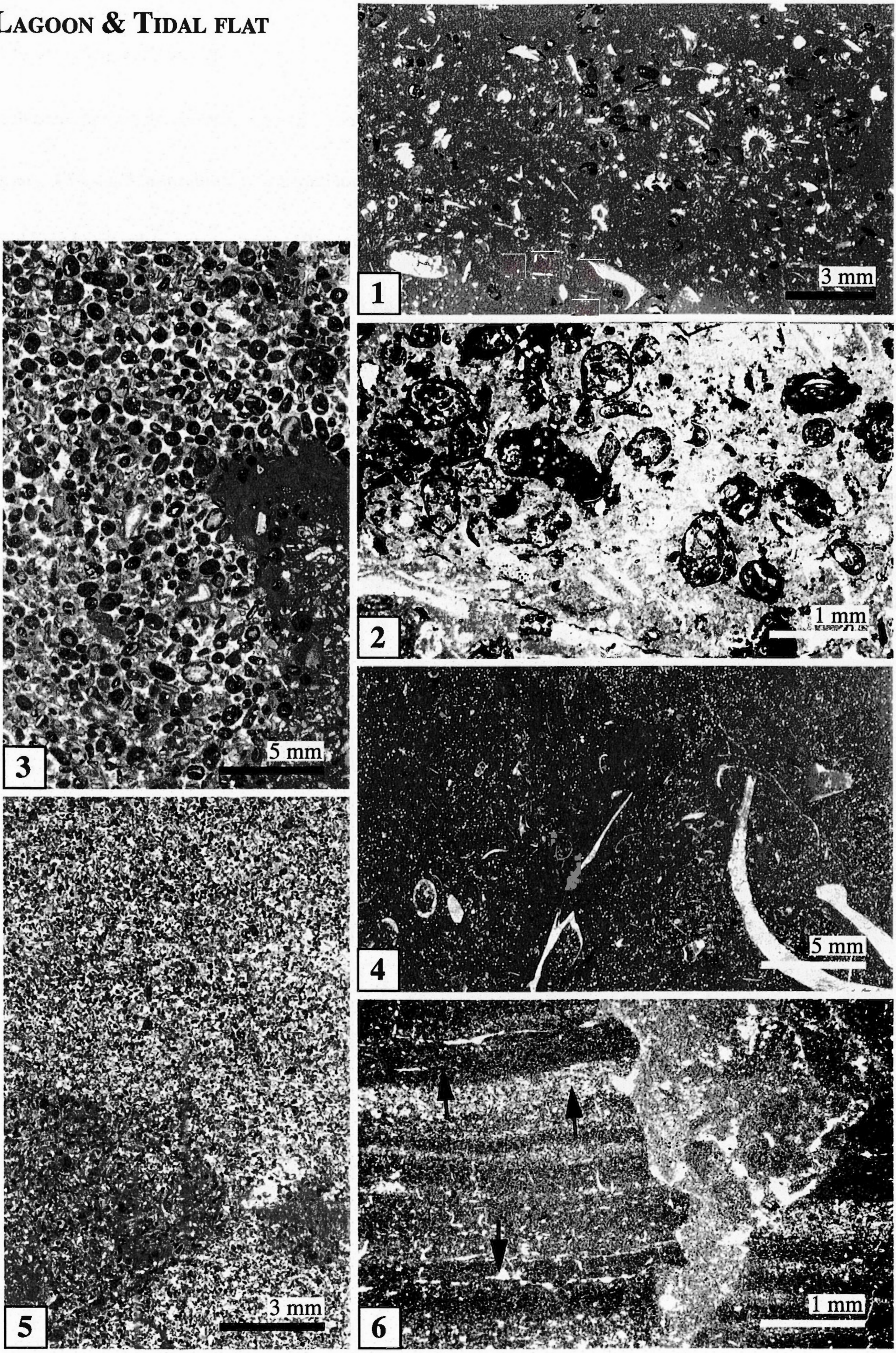


Plate 3 - Microfacies

1. Mudstone (biomicrite) with fenestrae and desiccation cracks. Facies T2, alteration A1, (microbial) mud flat, sample LC102.

2. Grain-dominated packstone (bio-pelmicrite) with abundant detrital quartz and lamination. Facies T4, sandy mud flat to sand flat, sample Sa6.5.

3. Very well sorted grainstone (oo-pelsparite, ooid types 1a, 1b). Facies B1, open-marine, submerged ooid bar, sample Sa165.

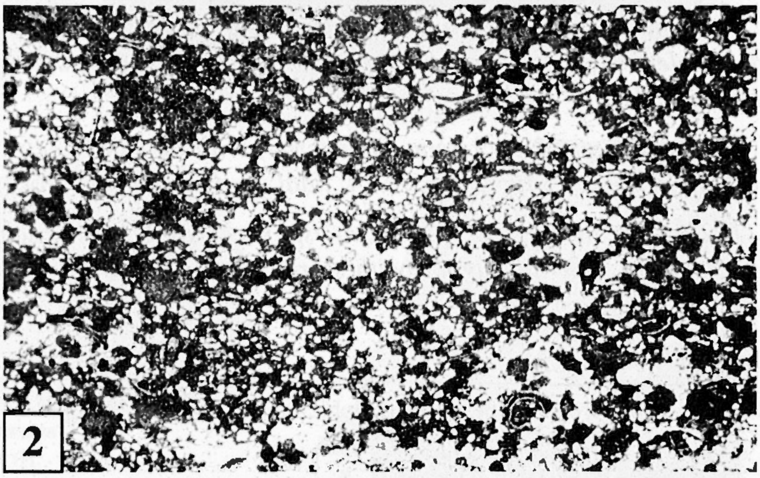
4. Bio-intra-pelmicrite with charophytes and blackened rip-up clasts. Facies T6, semi-restricted intertidal lagoon, sample Sa0.0.

5. Wackestone (bio-(intra)-micrite) with abundant debris of bryozoans, oysters, and crinoids. Facies B5, open-marine, external bioclastic bar, sample Sa171.

6. Root molds with early diagenetic pendant cements (arrow) overprinting restricted lagoonal facies (L16/L17). Alteration A2, sample E26.

TIDAL DOMAIN & HIGH-ENERGY BARS


1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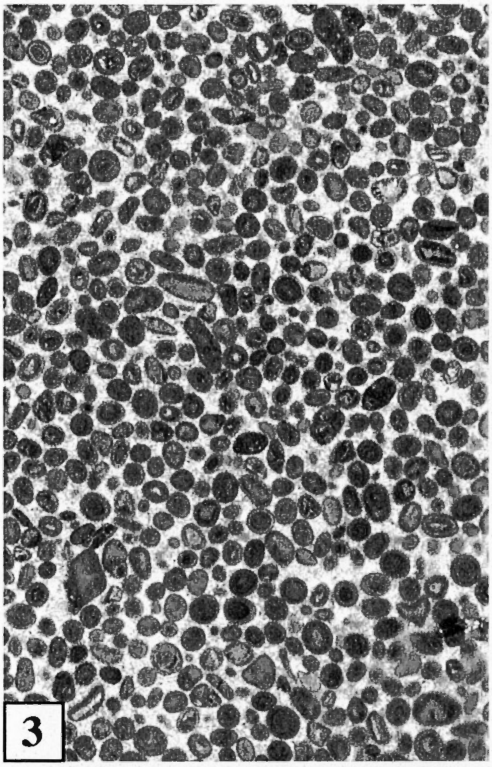
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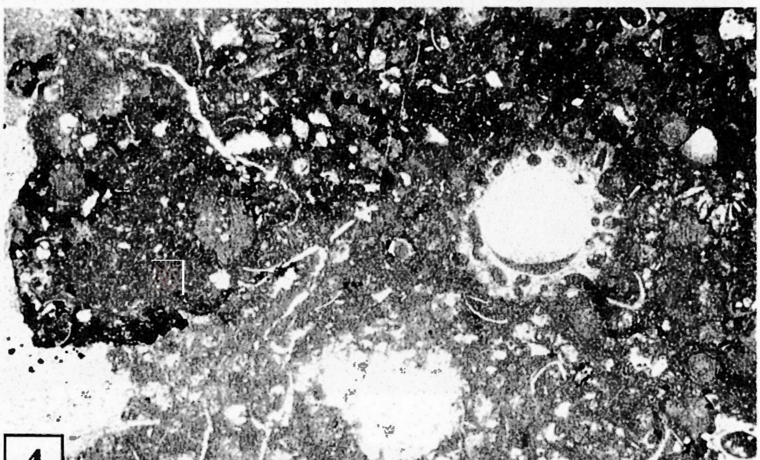
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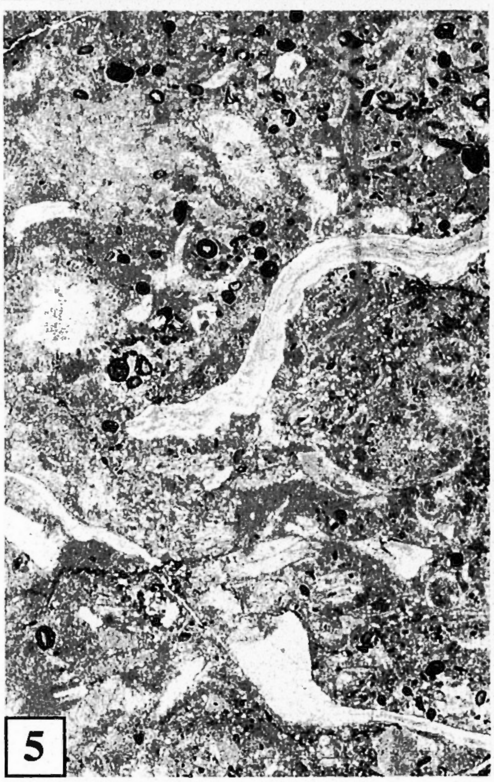
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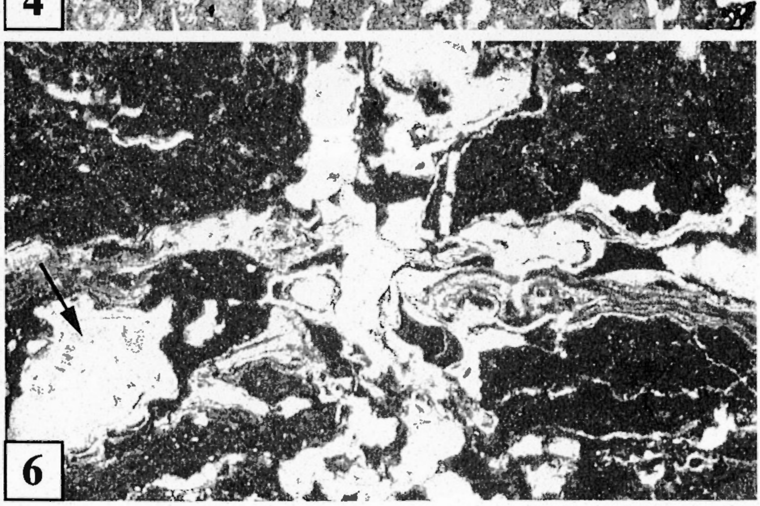
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Plate 4 - Facies

1. Coarse, laminated grainstone (bio-pelsparite) with keystone vugs (lower half of the sample). Beachfacies Bch1, sample Sa125.

2. Well-sorted grainstone (oo-pelsparite) with lamination and intense black staining (pyrite). Facies D1, small ooid dunes near shoaling in internal lagoon, sample R41.

3. Colony of encrusting bryozoans in coarse grainstone (bio-pelsparite). Facies L1b/L6, open-marine, high-energy lagoon, sample V76.

4. Rudist-framestone. Facies Bh1, small rudist bioherm in a shallow, open-marine lagoon, sample Sa91.2.

5. Pack- to grainstone / coral-floatstone (bio-oo-pelsparite/micrite) with Stylinids (arrow, upper left) and Microsolenids (arrow, center). Facies L5, open-marine lagoon probably in the vicinity of small patchreefs, sample M11.

6. Grain-dominated packstone (biomicrite) with abundant detrital quartz and ripple marks (white arrows point to faint lamination). Facies T4, tidal sand flat, sample R60.

7. Pack- to grainstone (bio-pelmicrite/sparite) with large intraclasts (outlined by white arrows). Facies L2, open-marine, external lagoon with storm influence. Intraclasts are derived from an incipient hardground, sample Sa163.1.

Remark: black arrows point to stratigraphic top.

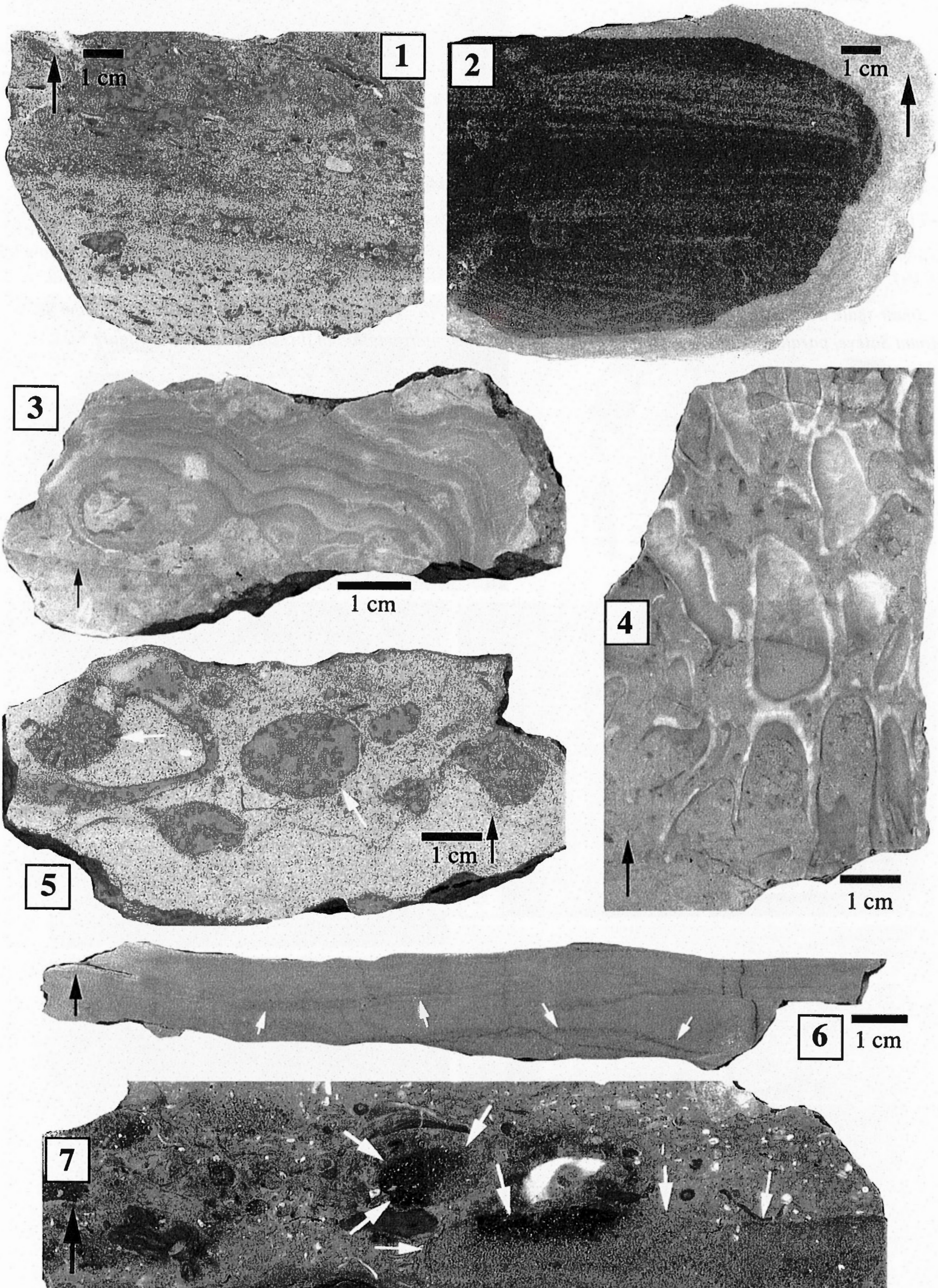
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Plate 5 - Facies & Stratigraphy

1. Karst pit in lagoonal limestone which is infilled with overlying marly limestone, SBZ Be5, Monnetier section at 29 meters.

2. Desiccation (circumgranular) cracks in mudflat facies (facies T2) immediately below a paleosol, sample Sa3.9.

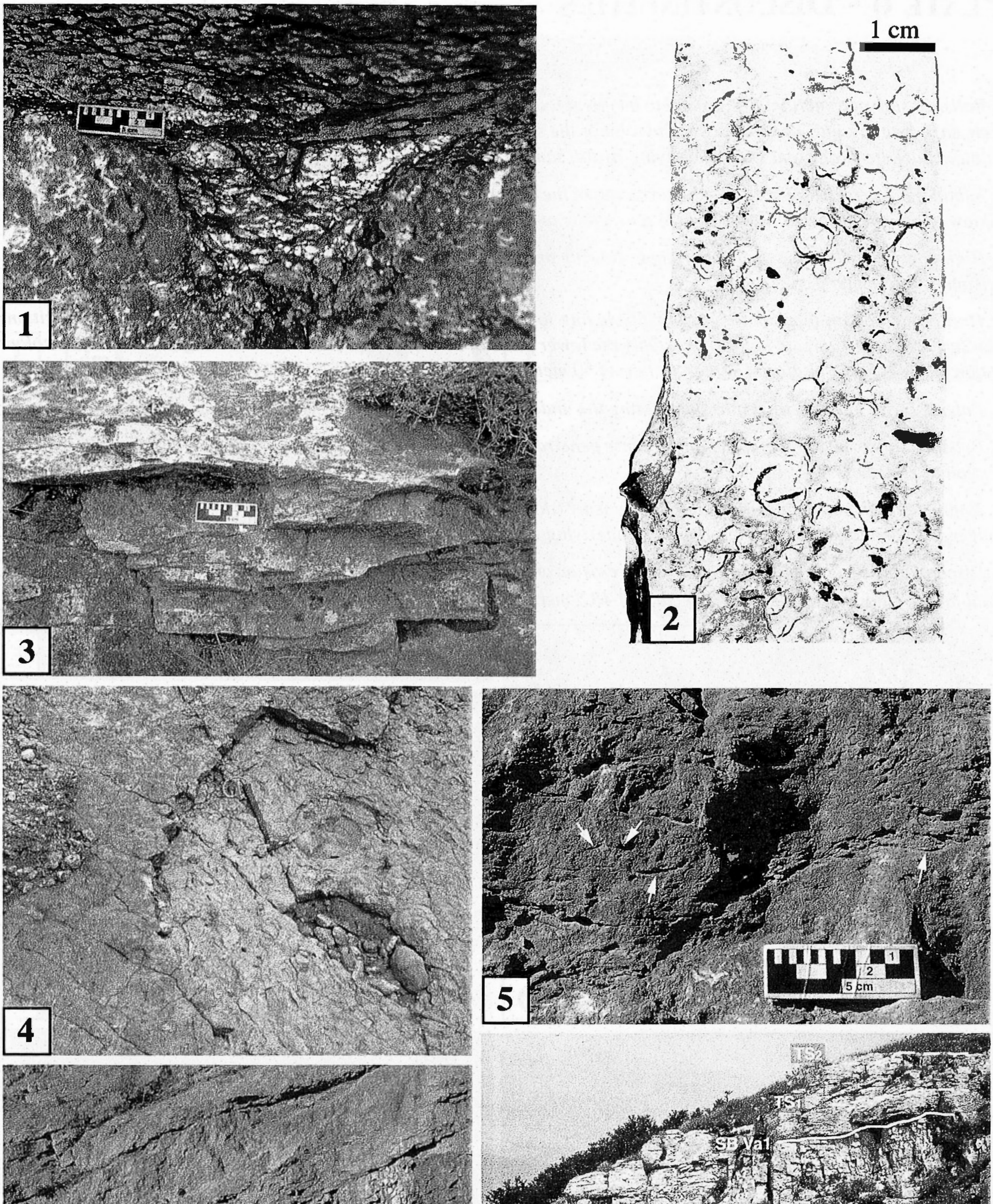
3. Hummocky cross-stratification. External lagoon/bioclastic bar, Val du Fier section at 85 meters.

4. Stained condensation surface with imprint of indeterminant ammonite. MFS Be4 in Mradma section, Morocco.

5. Flaser bedding, pointing to tidal influence in an internal lagoon (facies L14). Crozet section at 16 meters.

6. Sigmoidal foresets (white arrows) in oo-biosparites, pointing to tidal currents and shallowing tendency just below SB/TS Va1. Crozet section at 40 meters.

7. Small-scale SB-sequences with beach facies (black arrows). Top of lower Chambotte Formation just below SB Va1. Mount Salève, parallel to Salève section. For a detailed facies interpretation of this interval refer to Figure 5.23.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small-scale sequence 36
TS
SB
7

Plate 6 - Discontinuities

1. Polished slab of a subtidal firmground to incipient hardground. The sharp borders of the burrows (Thalassinoides) and their dark staining indicate early consolidation of the sediment and impregnation by Fe-oxides. Burrows were filled by peloidal pack- to grainstone of the overlying strata, sample V26.

2. Subtidal hardground with intense bioturbation of the underlying sediment and boring bivalves penetrating the surface (arrow to the left of scale). The surface is marked by an irregular morphology, Salève section at 56.5 meters.

3. High-energy hardground on an ooid shoal showing perforation by lithophages (1) and encrusting by oysters (2), Crozet section at 46.5 meters.

4. Hardground on an intertidal mudstone displaying low density and low diversity of borings with rare superpositions. The surface is very flat and knife-sharp. Only the lower parts of borings are preserved, indicating abrasion after colonization on a wave-cut platform, Salève section at 50 meters, scale in millimeters.

5. Paleosol with massive rhizoliths penetrating the underlying rock, Val du Fier section at 51 meters.

6. Polished slab of a microkarst showing intense penetrative staining by Fe-oxides and microrelief Chapeau de Gendarme section at 26 meters (SBZ Be5).

7. Encrusted bioclast displaying stalactite cements (thin section, scale bar equals 0.5 mm). Such indicators of vadose diagenesis commonly are found in the rock underlying diagenetic discontinuities, sample Sa127.

8. Discontinuity surface indicating small-scale erosion and facies change from lagoonal facies to siliciclastically influenced tidal facies, Chapeau de Gendarme section at 48.5 meters. Lens cap is 5 cm in diameter.

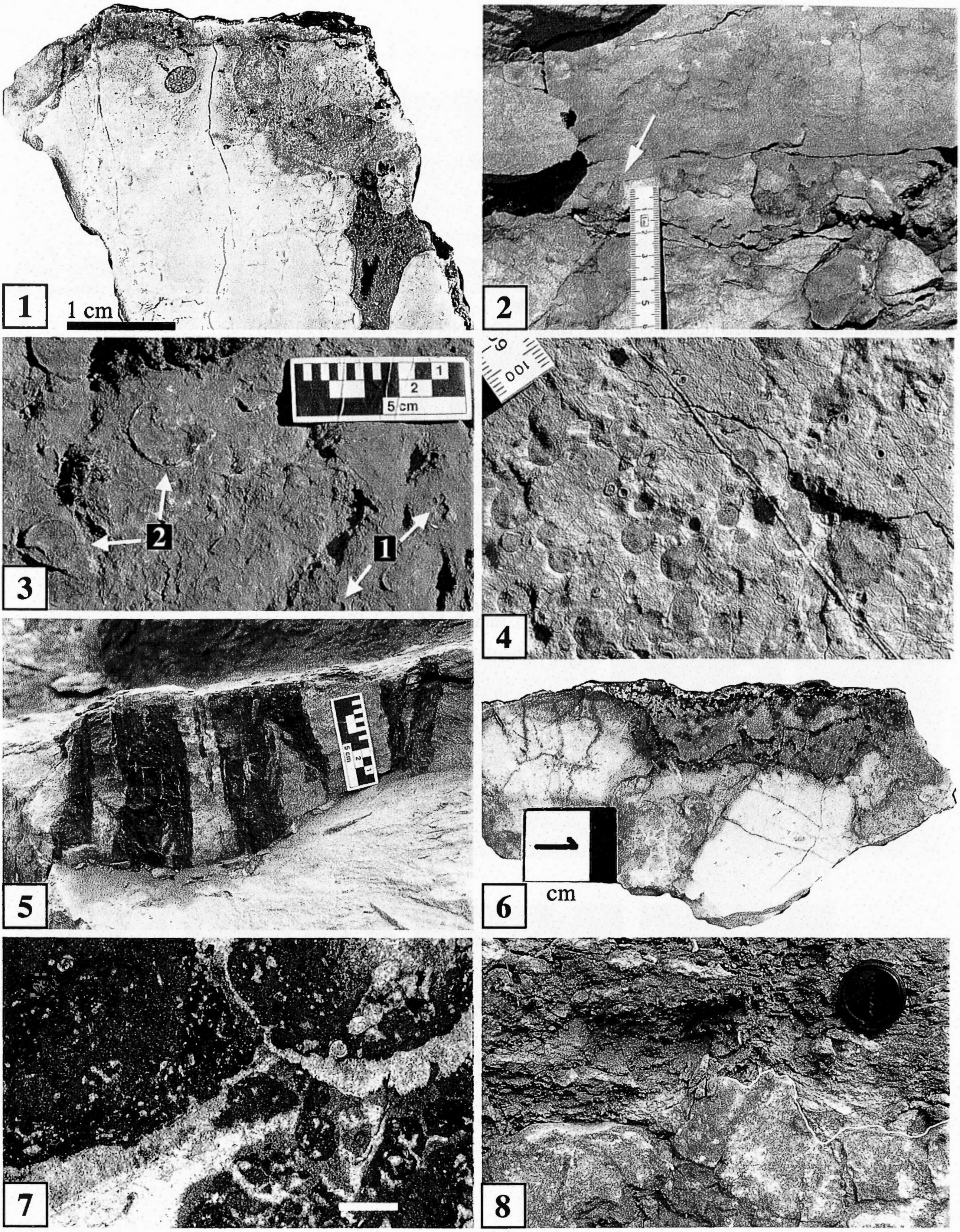
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Plate 7- Biostratigraphy

1. *Pavlovecina allobrogensis* (STEINHAUSER, BRÖNNIMANN & KOEHN ZANINETTI 1969) tangential section, sample Sa3.6.

2 + 3. *Pseudotextulariella courtionensis* (BRÖNNIMANN & KONRAD 1966)
2. axial section, sample Sa 5.3,
3. transversal section, sample LC33.

4 + 5. *Pfenderina neocomensis* (PFENDER 1938)
4. oblique section, sample C126,
5. oblique section , sample C128.

6 + 7. *Montsalevia salevensis* (CHAROLLAIS, BRÖNNIMANN & ZANINETTI 1966)
6. axial section, sample Sa151,
7. transversal section, sample R84.

8. *Calpionella alpina* (LORENZ 1902), sample Angles 124.

9. *Tintinnopsella carpathica* (MURGEANU & FILIPESCU 1933), sample Monctlus 17a.

10. *? Remaniella borzai* (POP 1994), sample Monctlus 17a.

11. *? Tintinnopsella longa* (COLOM 1939), sample Monctlus 20a.

12. *Calpionellopsis ? simplex* (COLOM 1939), sample Monctlus 16a.

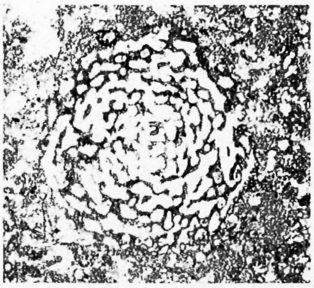
13. *Calpionellopsis oblonga* (CADISCH 1932), sample Monctlus 18b.

14. *Lorenziella hungarica* (KNAUER & NAGY 1961), sample Monctlus 18a.

15. *Calpionellites darderi* (COLOM 1934), sample Monctlus 20b.

16. *Calpionellites major* (COLOM 1948), sample Monctlus 21a.

FORAMINIFERS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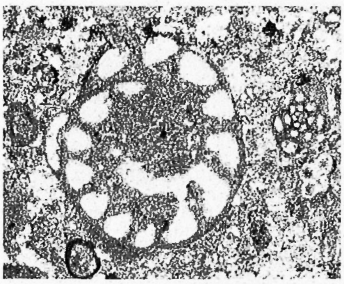
3 150 μm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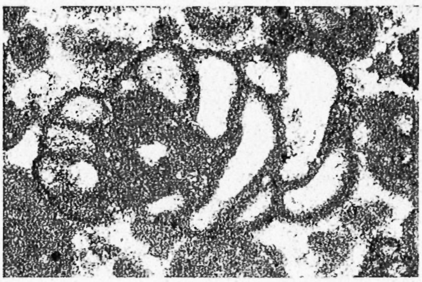
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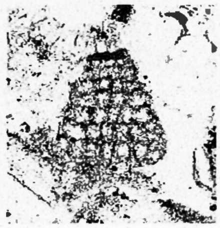
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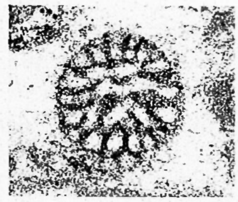
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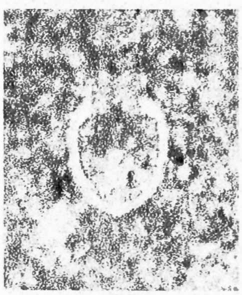
CALPIONELLIDS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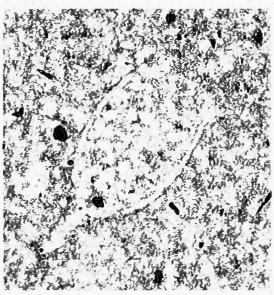
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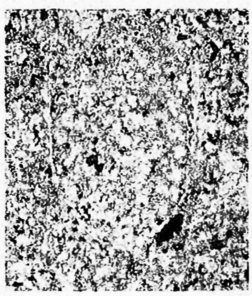
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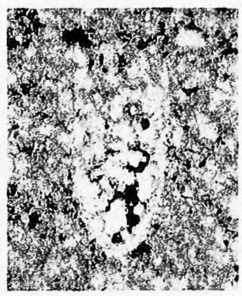
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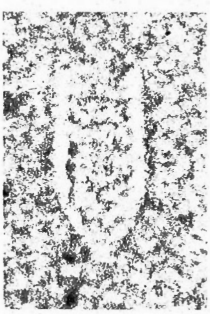
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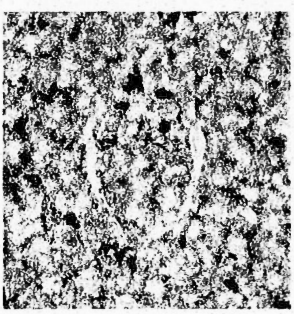
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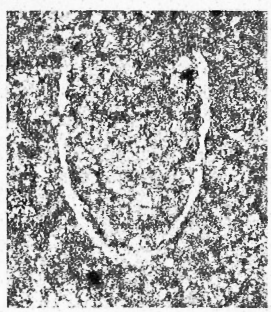
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Plate 8 - Stratigraphy

1. *Complete stratigraphic succession of the studied interval. The outcrop is located on the northern flank of the Val du Fier (eastern end of the valley). The Val du Fier section was partly logged along this outcrop (sectionline 2). The easily weathering, slightly marly facies of the Vions Formation and the Guiers Member are preferentially vegetated, whereas the massive limestones of the Piere-Châtel and Chambotte Formations form the cliffs.*

2. *Western face of Mount Salève with well-exposed Vions and Chambotte Formations. The identified medium-scale depositional sequences and large-scale sequence-boundary zones are pointed out on the photograph. Numbers of depositional sequences correspond to the section in Figure 5.22.*

3. *Id ou Belaid section in the Essaouira basin, Morocco. The morphological ridge formed by the carbonate-rich, shallow-ramp deposits of the Late Berriasian is well visible. Maximum flooding surfaces Be4 and Be5 are correlatable with condensation surfaces in Mradma section (Fig. 6 on this Plate).*

4. *Basinal sediments in the Monclus section (Vocontian Trough). The change in stacking pattern from thin-bedded, limestone-dominated strata to well-developed marl-limestone couplets at the limit between LD/TD Be8 and HD Be8 is interpreted as MF. SB Va1 is inferred at the base of a slightly thicker marly interval with a thin, wavy limestone bed introducing a general change to thicker marly interbeds.*

5. *Small-scale stacking pattern and discontinuity surfaces in the Monnetier section. For a detailed interpretation refer to Figure 5.8.*

6. *Mradma section in Morocco. Carbonate-rich TD of sequence Be4 in the foreground with subsequent, more marly MF and distinct condensation surfaces (MFS Be4, MFS Be5).*

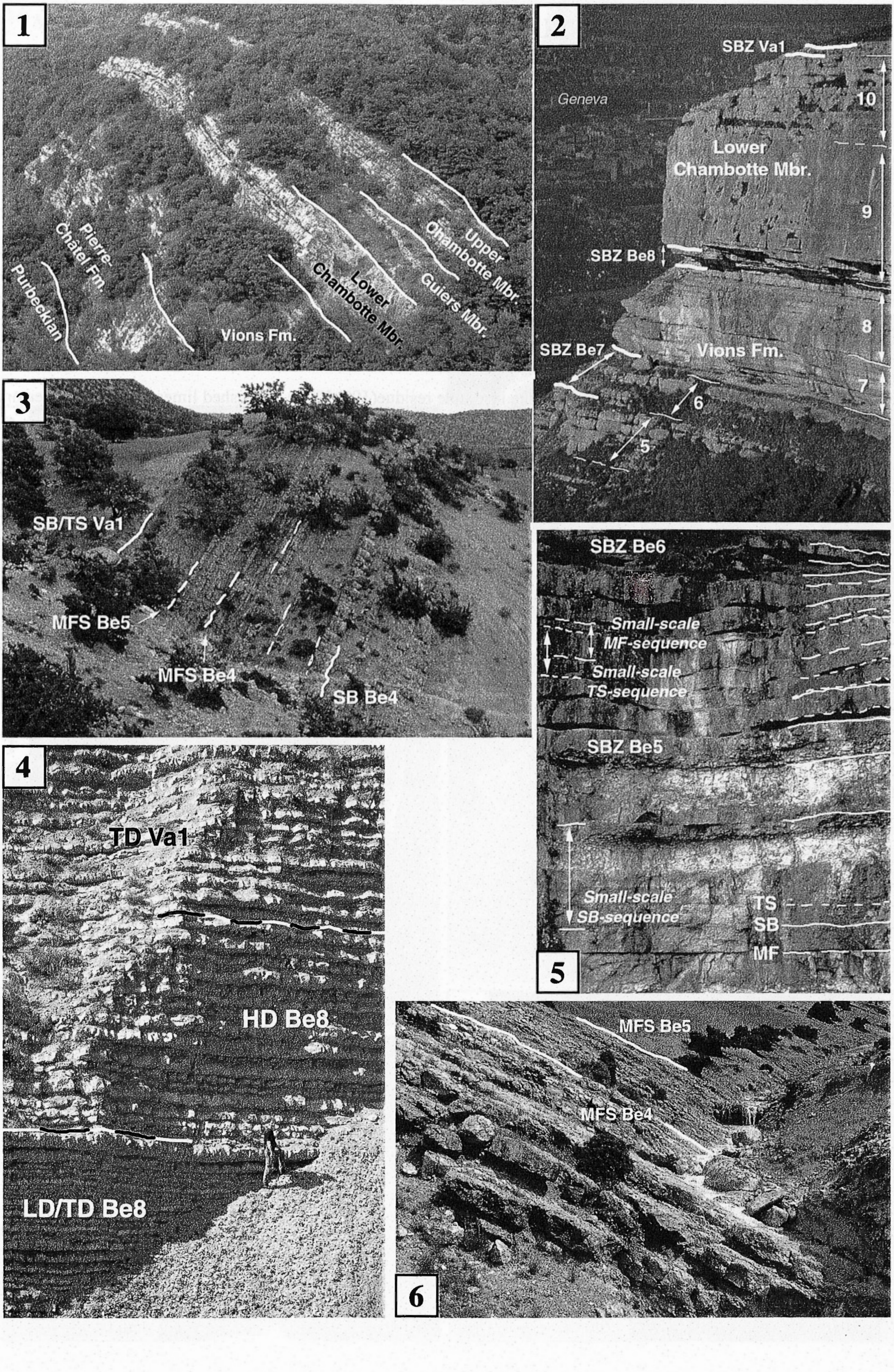

1
Pierre-Châtel Fm.
Purbeckian
Vions Fm.
Lower Chambotte Mbr.
Guiers Mbr.
Upper Chambotte Mbr.
2
SBZ Va1
Geneva
10
Lower Chambotte Mbr.
9
SBZ Be8
8
SBZ Be7
Vions Fm.
7
6
5
3
SB/TS Va1
MFS Be5
MFS Be4
SB Be4
4
TD Va1
HD Be8
LD/TD Be8
5
SBZ Be6
Small-scale MF-sequence
Small-scale TS-sequence
SBZ Be5
Small-scale SB-sequence
TS
SB
MF
6
MFS Be5
MFS Be4

PLATE 9 - DETRITAL INPUT

1. SEM image of detrital quartz grains. Angular, multifacetted shapes point to fluvial transport, sample Sa8.5.

2. SEM image of a well-rounded quartz grain with slightly pitted surface. Transport may initially havebeen aeolian, sample Sa8.5.

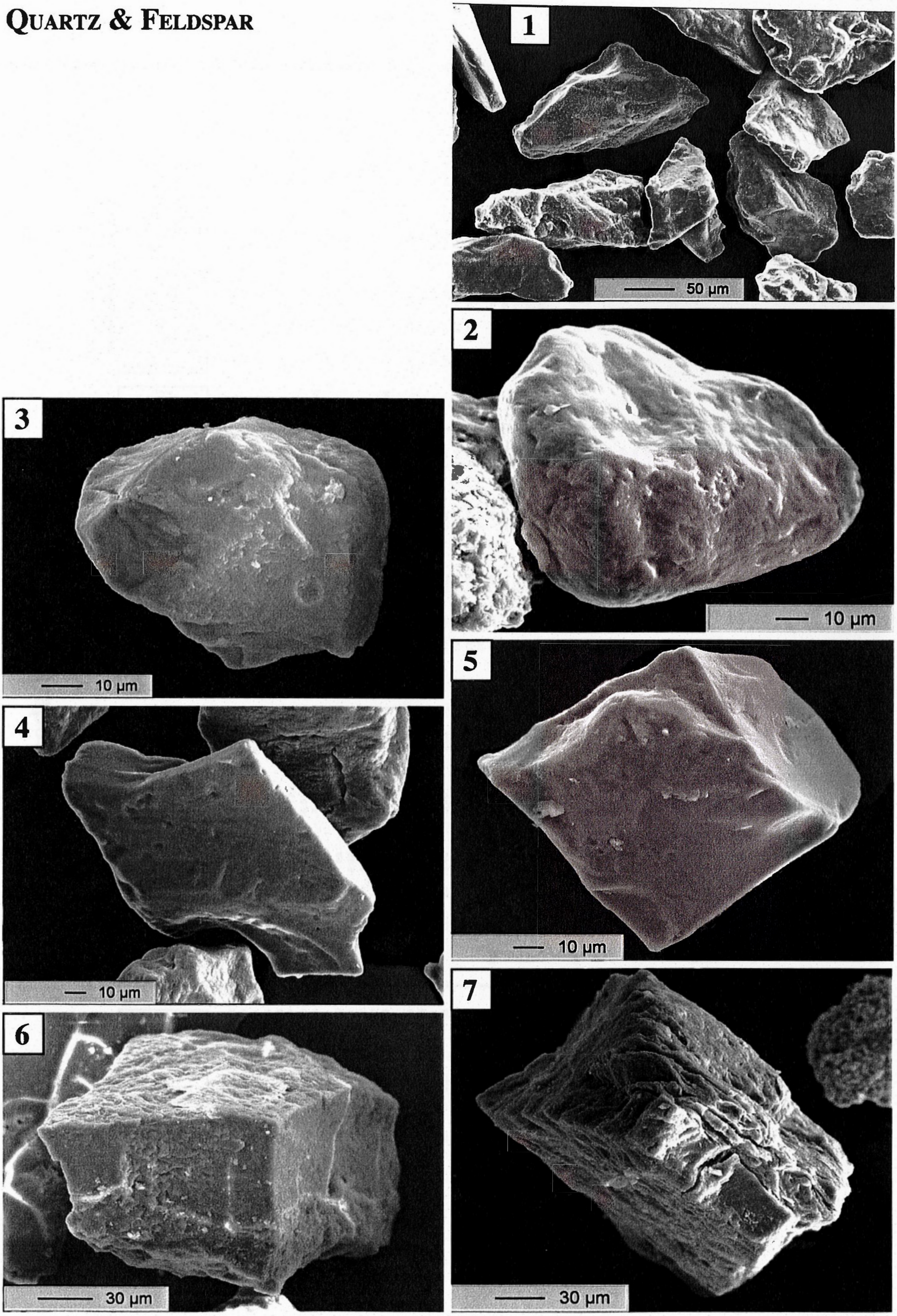
3. SEM image of a well-rounded, detrital quartz grain (fluvial transport), sample Sa8.5.

4 + 5. SEM images of quartz grains with a rhombohedral shape.This is interpreted as result of an authigenic pseudomorphosis after dolomite, sample Sa3.1.

6 + 7. SEM images offeldspars. The almost euhedral shape of the grain in Figure 6 suggests an authigenic origin, sample Sa 7.4.

Remark: The figured detrital grains are found in the insoluble residue(10% HCL) of crushed limestone/marly-limestone samples.

QUARTZ & FELDSPAR



Appendix 1

Accommodation analysis

Monnetier section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
	1	4	4.00	2	6.00		3.04
	2	5	3.54	3	6.54		3.58
	3	6	4.77	3	7.77		4.81
	4	7	4.92	3	7.92		4.96
	5	8	6.00		6.00		3.04
	6	9	3.23		3.23		0.27
	7	10	2.00		2.00		-0.96
	8	11	0.92		0.92		-2.04
Be5							
	9	12	1.08		1.08		-1.88
	10	13	2.15		2.15		-0.81
	11	14	1.08		1.08		-1.88
	12	15	1.69		1.69		-1.27
	13	16	0.77		0.77		-2.19
Be6							
	14	17	1.23		1.23		-1.73
	15	18	2.92		2.92		-0.04
	16	19/20	2.85		2.85		-0.11
	17	21	1.38		1.38		-1.58
	18	22	1.69		1.69		-1.27
	19	23	1.85		1.85		-1.11
Be7							
	20	24	1.08		1.08		-1.88
	21	25	2.62		2.62		-0.34
	22	26	2.15		2.15		-0.81
	23	27	3.38	1	4.38		1.42
	24	28	2.00		2.00		-0.96
	25	29	0.92		0.92		-2.04
Be8							
	26	30	0.92		0.92		-2.04
	27	31	1.85	1	2.85		-0.11
	28	32	2.46		2.46		-0.50
	29	33	4.00	2	6.00		3.04
	30	34	3.08	2	5.08		2.12
	31	35	3.08	2	5.08		2.12
			total thickness		94.62	2.96	

Salève section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
Be4							
	1	1	2.31	2	4.31		1.19
	2	2	3.38	2	5.38		2.26
	3	3	3.85	2	5.85		2.73
	4	4	4.77	2	6.77		3.65
	5	5	4.31	3	7.31		4.19
	6	6	2.77	4	6.77		3.65
	7	7	4.31	4	8.31		5.19
	8	8	3.08	1	4.08		0.96
	9	9	2.15	1	3.15		0.03
Be5	10-12		4.62		4.62		-4.74
	13	13	2.62		2.62		-0.50
	14	14	1.69		1.69		-1.43
	15	15	1.54		1.54		-1.58
	16	16	0.77		0.77		-2.35
Be6							
	17	17	0.77		0.77		-2.35
	18	18	3.85		3.85		0.73
	19	19	2.46		2.46		-0.66
	20	20	0.46		0.46		-2.66
	21	21	1.85		1.85		-1.27
	22	22	1.54	1	2.54		-0.58
	23	23	1.38		1.38		-1.74
Be7							
	24	24	1.54		1.54		-1.58
	25	25	1.85		1.85		-1.27
	26	26	2.00		2.00		-1.12
	27	27	3.08		3.08		-0.04
	28	28	2.46		2.46		-0.66
	29	29	0.92		0.92		-2.20
Be8							
	30	30	1.38		1.38		-1.74
	31	31	4.46		4.46		1.34
	32	32	2.77		2.77		-0.35
	33	33	5.23	2	7.23		4.11
	34	34	2.46	2	4.46		1.34
	35	35	1.85		1.85		-1.27
	36	36	2.00		2.00		-1.12
	37	37	3.08		3.08		-0.04
			total thickness		115.56	3.12	
Va1							
		38	1.54	5	6.54		3.075
		39	2.62	10	12.62		9.155
		40	2.46	10	12.46		8.99
		41	1.85	5	6.85		3.385
Va2							
		45/51	7.69	15	22.69		1.899
Va3			total thickness		176.72	3.47	

Val du Fier section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
Be4							
	1	1	2.46	2	4.46		0.56
	2	2	2.46	3	5.46		1.56
	3	3	2.46	3	5.46		1.56
	4	4	4.15	3	7.15		3.25
	5	5	4.62	3	7.62		3.72
	6	6	3.85	3	6.85		2.95
	7	7	2.62	3	5.62		1.72
	8	8	4.15		4.15		0.25
	9	9	1.85		1.85		-2.05
	10	10	1.85		1.85		-2.05
	11	11	1.23		1.23		-2.67
Be5							
	12	12	1.23		1.23		-2.67
	13	13	3.08	1	4.08		0.18
	14	14	1.69		1.69		-2.21
	15	15	0.92		0.92		-2.98
	16	16	0.31		0.31		-3.59
Be6							
	17	17	0.92		0.92		-2.98
	18	18	3.85		3.85		-0.05
	19	19	2.15		2.15		-1.75
	20	20	1.85		1.85		-2.05
	21	21	1.85		1.85		-2.05
	22	22	0.92		0.92		-2.98
	23	23	0.62		0.62		-3.28
Be7							
	24	24	0.77		0.77		-3.13
	25	25	2.31		2.31		-1.59
	26	26	2.85	1	3.85		-0.05
	27	27	1.85	2	3.85		-0.05
	28	28	1.54	2	3.54		-0.36
	29	29	1.54	2	3.54		-0.36
Be8							
	30	30	1.85	2	3.85		-0.05
	31	31	1.54		1.54		-2.36
	32	32	3.08		3.08		-0.82
	33	33	2.62	2	4.62		0.72
	34	34	1.38	2	3.38		-0.52
	35	35	1.69	2	3.69		-0.21
	36	36	2.77	1	3.77		-0.13
	37	37	1.85		1.85		-2.05
Va1							-3.90
	38	38	0.46	2	2.46		-1.44
	39	39	3.38	10	13.38		9.48
	40	40	2.00	15	17.00		13.10
	41	41	1.54	10	11.54		7.64
			total thickness		160.08	3.90	

La Chambotte section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
	2	5	2.74		2.74		-2.75
	3	6	3.55	3	6.55		1.06
	4	7	2.74	2	4.74		-0.75
	5	8	4.03	2	6.03		0.54
	6	9	1.45	1	2.45		-3.04
		10	0		0		-5.49
	7	11	0.97	1	1.97		-3.52
Be5							
	8	12	1.45	1	2.45		-3.04
	9	13	2.26	1	3.26		-2.23
	10	14	2.90		2.90		-2.59
	11	15	1.61		1.61		-3.88
	12	16	1.53		1.53		-3.96
Be6							
	13	17	1.13		1.13		-4.36
	14	18	2.90		2.90		-2.59
	15	19	2.42		2.42		-3.07
	16	20	0.32		0.32		-5.17
	17	21	2.42		2.42		-3.07
	18	22	1.61		1.61		-3.88
	19	23	0.97	1	1.97		-3.52
Be7							
	20	24	1.61		1.61		-3.88
	21	25	2.10		2.10		-3.39
	22	26	1.61		1.61		-3.88
	23	27	2.74		2.74		-2.75
	24	28	0.97		0.97		-4.52
	25	29	1.13	2	3.13		-2.36
Be8							
	26	30	1.77	2	3.77		-1.72
	27	31	3.39	3	6.39		0.90
	28	32	3.55	3	6.55		1.06
	29	33	2.26	3	5.26		-0.23
	30	34	2.74	2	4.74		-0.75
	31	35	3.87	2	5.87		0.38
	32	36	1.94	2	3.94		-1.55
Va1							
	33	37/38	3.23	4	7.23		-3.75
	34	39	2.42	10	12.42		6.93
	35	40	1.77	10	11.77		6.28
	36	41	1.77	10	11.77		6.28
	37	42	2.42	15	17.42		11.93
	38	43	2.42	10	12.42		6.93
	39	44	2.90	10	12.90		7.41
Va2							
	40	45	2.26	10	12.26		6.77
	41	46	2.74	5	7.74		2.25
	42	47	2.42	5	7.42		1.93
	43	48	2.42	5	7.42		1.93
	44	49	2.74	5	7.74		2.25
	45	50	2.90	5	7.90		2.41
Va3							
	46	51	3.23	15	18.23		12.74
			total thickness		252.34	5.49	

Accommodation analysis

Crêt de l'Anneau section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
	1	3	1.83		1.83		0.12
	2	4	2.17		2.17		0.46
	3	5	3.17		3.17		1.46
	4	6	2.17		2.17		0.46
	5	7	2.00	3	5.00		3.29
	6	8	1.33	3	4.33		2.62
	7	9	1.67		1.67		-0.04
	8	10	1.83		1.83		0.12
Be5		11	0.00		0.00		-1.71
	9	12	0.83		0.83		-0.88
	10	13	2.00		2.00		0.29
	11	14	3.08		3.08		1.37
	12	15	0.80		0.80		-0.91
Be6		16	0.00		0.00		-1.71
	13	17	1.00		1.00		-1.71
	14	18/19	2.17		2.17		0.46
	15	20	0.42		0.42		-1.29
	16	21	0.33		0.33		-1.38
	17	22	0.33	1	1.33		-0.38
			total thickness		34.13	1.71	
Be7							
Be8							
		30/38	4.67	5	9.67	1.07	-0.03
Va1							
		39/44	3.50		3.5	0.5	-0.60
Va2							
		45/51	1.83	5	6.83	0.98	-0.13
Va3					54.13	1.10	

Chapeau de Gendarme section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
Be4							
	1	1	2.31		2.31		-0.05
	2	2	1.85		1.85		-0.51
	3	3	2.46		2.46		0.10
	4	4	3.38		3.38		1.02
	5	5	2.77		2.77		0.41
	6	6	3.08		3.08		0.72
	7	7	3.23	3	6.23		3.87
	8/9	8	5.54		5.54		3.18
	10	9	1.08		1.08		-1.28
Be5		10/11	0.00		0.00		-4.72
	11	12	0.77		0.77		-1.59
	12	13	2.15		2.15		-0.21
	13	14	1.85	1	2.85		0.49
	14	15	1.54		1.54		-0.82
Be6		16	0.00		0.00		-2.36
	15	17	1.54		1.54		-0.82
	16	18/19	3.38		3.38		-1.34
	17	20	0.62		0.62		-1.74
	18	21	1.54		1.54		-0.82
	19	22	1.31		1.31		-1.05
Be7		23	0.00		0.00		-2.36
	20	24	0.54		0.54		-1.82
	21	25	2.77		2.77		0.41
	22	26	1.85	2	3.85		1.49
	23	27	2.77	2	4.77		2.41
	24	28	1.54	2	3.54		1.18
	25	29	2.15		2.15		-0.21
Be8							
	26	30	2.31	2	4.31		1.95
	27	31	3.23	2	5.23		2.87
	28	32	4.31		4.31		1.95
	29	33	1.54		1.54		-0.82
	30	34	2.69		2.69		0.33
	31	35	2.46		2.46		0.10
			total thickness		82.54	2.36	

Vuache section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
Be4							
	1	1	1.46	1	2.46		0.45
	2	2	2.77		2.77		0.76
	3	3	2.00		2.00		-0.01
	4	4	2.46		2.46		0.45
	5	5	3.85		3.85		1.84
	6	6	4.31		4.31		2.30
	7	7	5.38	2	7.38		5.37
	8	8	5.23	1	6.23		4.22
	9	9	1.54		1.54		-0.47
Be5							
	10	10/16	0.46		0.46		-11.60
Be6							
	11	17	1.08		1.08		-0.93
	12	18	1.54		1.54		-0.47
	13	19	2.15	1	3.15		1.14
	14	20	1.08		1.08		-0.93
	15	21/22	2.31		2.31		-1.71
	16	23	1.31		1.31		-0.70
Be7							
	17	24	0.77		0.77		-1.24
	18	25	2.31		2.31		0.30
	19	26	2.92		2.92		0.91
	20	27	1.85		1.85		-0.16
	21	28	1.54		1.54		-0.47
Be8		29	0.00		0.00		-2.01
	22	30	1.69	2	3.69		1.68
	23	31	3.23	2	5.23		3.22
			total thickness		62.23	2.01	

Ecluse section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
	8	8	5.85		5.85		3.24
	9	9	0.92		0.92		-1.69
Be5							
	10	10/16	0.46		0.46		-15.20
	11	17	1.38		1.38		-1.23
	12	18	1.08		1.08		-1.53
	13	19	2.00	1	3.00		0.39
	14	20	1.23		1.23		-1.38
	15	21/22	1.38		1.38		-3.84
	16	23	2.15		2.15		-0.46
Be7							
	17	24	0.62		0.62		-1.99
	18	25	2.46		2.46		-0.15
	19	26	3.00		3.00		0.39
	20	27	2.31	1	3.31		0.70
	21	28	1.54	1	2.54		-0.07
Be8		29	0.00		0.00		-2.61
	22	30	2.31	1	3.31		0.70
	23	31	2.62	2	4.62		2.01
	24	32	2.77	4	6.77		4.16
	25	33	1.69	4	5.69		3.08
	26	34	1.85	4	5.85		3.24
	27	35	2.31	2	4.31		1.70
	28	36	3.23	2	5.23		2.62
	29	37	2.46	2	4.46		1.85
Va1							
	30	38	2.15	5	7.15		4.54
	31	39	1.69	5	6.69		4.08
			total thickness		83.46	2.61	

Crozet section

position of large-scale SB in section	# of small-scale dep.seq. in section	# of small-scale dep.seq. in correlation	thickness of small-scale dep.seq. in m	correction for bathymetry at inferred SB in m	accommodation not corrected for compaction	mean accommodation per 100 ky	deviation of mean accommodation in m
	1	22	2.15		2.15		-1.39
Be7							
	2	23	1.69		1.69		-1.85
	3	24	1.85		1.85		-1.69
	4	25	2.23	2	4.23		0.69
	5	26	2.77		2.77		-0.77
	6	27	6.30	2	8.3		4.76
	7	28	3.38		3.38		-0.16
	8	29	3.23		3.23		-0.31
	9	30	2.00		2		-1.54
Be8							
	10	31	2.62	2	4.62		1.08
	11	32	2.77	2	4.77		1.23
			total thickness		38.99	3.54	
		33/37	8.46	2	10.46		3.21
Va1							
		38	0.69	2	2.69		1.24
		39	1.46	3	4.46		3.01
		40	1.08	3	4.08		2.63
		41	1.38	3	4.38		2.93
Va2							
		45/50	2.46	5	7.46		-1.24
Va3			total thickness		72.52	1.45	

Accommodation analysis

# of small-scale dep.seq. in correlation	accommodation not corrected for compaction									mean accommodation across the platform	mean accommodation per 100 ky	non-cumulative deviation of mean accommodation in m	cumulative deviation of mean accommodation in m
	Crêt de l' Anneau	Chapeau de Gendarme	Vuache	Fort de l'Ecluse	Crozet	Monnetier	Salève	Val du Fier	La Chambotte				
													0
1		2.31	2.46				4.31	4.46		3.39		0.09	0.09
2		1.85	2.77				5.38	5.46		3.87		0.57	0.65
3	1.83	2.46	2				5.85	5.46		3.52		0.22	0.87
4	2.17	3.38	2.46			6	6.77	7.15		4.66		1.36	2.23
5	3.17	2.77	3.85			6.54	7.31	7.62	2.74	5.21		1.91	4.14
6	2.17	3.08	4.31			7.77	6.77	6.85	6.55	4.81		1.51	5.65
7	5.00	6.23	7.38			7.92	8.31	5.62	4.74	6.72		3.42	9.06
8	4.33	5.54	6.23	5.85		6	4.08	4.15	6.03	5.12		1.82	10.88
9	1.67	1.08	1.54	0.92		3.23	3.15	1.85	2.45	2.43		-0.87	10.01
10	1.83	0	0.46	0.46		2	1.54	1.85	0	1.32		-1.98	8.04
11	0	0	0	0		0.92	1.54	1.23	1.97	0.46		-2.84	5.20
12	0.83	0.77	0	0		1.08	1.54	1.23	2.45	0.93		-2.37	2.82
13	2.00	2.15	0	0		2.15	2.62	4.08	3.26	1.93		-1.37	1.46
14	3.08	2.85	0	0		1.08	1.69	1.69	2.9	1.71		-1.59	-0.14
15	0.8	1.54	0	0		1.69	1.54	0.92	1.61	1.17		-2.13	-2.26
16	0	0	0	0		0.77	0.77	0.31	1.53	0.43		-2.87	-5.13
17	1.00	1.54	1.08	1.38		1.23	0.77	0.92	1.13	1.18		-2.12	-7.25
18	1.08	1.69	1.54	1.08		2.92	3.85	3.85	2.9	2.14		-1.16	-8.41
19	1.08	1.69	3.15	3		1.43	2.46	2.15	2.42	2.23		-1.07	-9.48
20	0.42	0.62	1.08	1.23		1.43	0.46	1.85	0.32	1.19		-2.11	-11.59
21	0.33	1.54	1.16	0.69		1.38	1.85	1.85	2.42	1.14		-2.16	-13.75
22	1.33	1.31	1.16	0.69	2.15	1.69	2.54	0.92	1.61	1.58		-1.72	-15.47
23	0	0	1.31	2.15	1.69	1.85	1.38	0.62	1.97	1.18		-2.12	-17.59
24	0	0.54	0.77	0.62	1.85	1.08	1.54	0.77	1.61	1.02		-2.28	-19.87
25	0	2.77	2.31	2.46	4.23	2.62	1.85	2.31	2.1	2.24		-1.06	-20.93
26	0	3.85	2.92	3	2.77	2.15	2	3.85	1.61	2.52		-0.78	-21.72
27	0	4.77	1.85	3.31	8.3	4.38	3.08	3.85	2.74	3.46		0.16	-21.56
28	0	3.54	1.54	2.54	3.38	2	2.46	3.54	0.97	2.42		-0.88	-22.44
29	0	2.15	0	0	3.23	0.92	0.92	3.54	3.13	1.30		-2.00	-24.44
30	1.07	4.31	3.69	3.31	2	0.92	1.38	3.85	3.77	2.63		-0.67	-25.11
31	1.07	5.23	5.23	4.62	4.62	2.85	4.46	1.54	6.39	3.71		0.41	-24.70
32	1.07	4.31		6.77	4.77	2.46	2.77	3.08	6.55	3.95		0.65	-24.05
33	1.07	1.54		5.69	2.09	6	7.23	4.62	5.26	4.35		1.05	-23.00
34	1.07	2.69		5.85	2.09	5.08	4.46	3.38	4.74	3.74		0.44	-22.56
35	1.07	2.46		4.31	2.09	5.08	1.85	3.69	5.87	3.16		-0.14	-22.70
36	1.07			5.23	2.09		2	3.77	3.94	3.79		0.49	-22.21
37	1.07			4.46	2.09		3.08	1.85	3.62	2.88		-0.43	-22.63
38	1.07			7.15	2.69		6.54	2.46	3.62	4.11		0.81	-21.83
39	0.7			6.69	4.46		12.62	13.38	12.42	7.32		4.02	-17.81
40	0.7				4.08		12.46	17	11.77	9.63		6.33	-11.48
41	0.7				4.38		6.85	11.54	11.77	7.91		4.61	-6.87
42	0.7				0		0		17.42	4.03		0.73	-6.15
43	0.7				0		0		12.42	4.77		1.47	-4.67
44	0.7				0		0		12.9	3.56		0.26	-4.41
45	1.14				1.24		3.78		12.26	4.48		1.18	-3.23
46	1.14				1.24		3.78		7.74	4.53		1.23	-2.00
47	1.14				1.24		3.78		7.42	3.69		0.39	-1.61
48	1.14				1.24		3.78		7.42	3.88		0.58	-1.03
49	1.14				1.24		3.78		7.74	3.31		0.01	-1.02
50	1.14				1.24		3.78		7.9	3.19		-0.11	-1.13
51	1.14						3.78		18.23	4.23		0.93	-0.21
								total thickness of (lateral) mean accommodation		168.09	3.30		

Appendix 2

Calpionellids in Montclus (Mo) and Angles (An) sections (Vocontian Trough)

SAMPLE Mo	14a	14b	15a	15b	16a	16b	17a	17b	18a	18b	19a	19b	20a	20b	21a	21b	22a	22b	23a	23b	24a	24b	25a	25b	26a	26b	An	124	123
CALP. SPEC.																													
Calpionella gr. elliptica			x?	x																								x	
Calpionella minuta			x?	x																									
Calpionella gr. alpina			x?	x																								x	
Lorenziella hungarica									x									x	x?										
Tintinnopsella carpath.	x	x	x	x	x	x			x	x			x	x	x	x	x		x						x	x			x
Tintinnopsella longa									x?	x	x	x	x	x	x	x	x	x											
Calpionellopsis oblonga		x				x?	x	x	x	x																			
Calpionellopsis simplex	x	x	x	x	x	x?	x?	x	x			x?																	x
Remaniella borzai		x?					x		x?																				
Praecalpionellites sirin.															x														
Calpionellites darderi													x	x	x				x?	x		x							
Calpionellites major															x?			x	x?		x?								

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Curriculum vitae

I was born in Wiesbaden, Germany, on the 18. January 1965 to Günter and Beate Hillgärtner (née Linz). Returning to Germany after a 6 year long stay in Burma (Myanmar) I attended primary school in Johannisberg (Rheingau). In 1984 I obtained the Abitur degree from the Rheingaugymnasium Geisenheim.

After two years of military service I enrolled in the University of Freiburg i. Brsg. with a major in geology. Having obtained the pre-diploma degree in 1988, I continued my studies in geology at Brock University, St. Catherines (Canada), where I received the bachelors degree with distinction in 1989. The Diploma degree in geology was issued in 1994 at the University of Freiburg. The diploma thesis carried out under the guidance of Prof. Thilo Bechstädt was entitled "Geologie im Bereich Sass da Putia - Lungiaru (Nördliche Dolomiten, Italien): Faziesanalyse und paläogeographische Interpretation permischer und triadischer Sedimente".

In Oktober 1994 I began to work on my doctoral thesis on sedimentology, sequence- and cyclostratigraphy at the Institute of Geology and Paleontology of the University of Fribourg (Switzerland) under the direction of Prof. André Strasser. I occupied the position of a research and teaching assistant kindly financed by the Swiss National Science Foundation (Projects 20-37386.93 and 20-46625.96).

Fribourg 30. October 1998