Disentangling the control of tectonics, eustasy, trophic conditions and climate on shallow-marine carbonate production during the Aalenian–Oxfordian interval: From the western France platform to the western Tethyan domain

Simon Andrieu ⁎, Benjamin Brigaud a, Jocelyn Barbarand a, Eric Lasseur b, Thomas Saucède c

a GEOPS, Univ. Paris-Sud, CNRS, Université Paris-Saclay, Rue du Belvédère, Bât. 504, 91405 Orsay, France
b BRGM, 3 avenue Claude Guillemin, BP 36009, 45060 Orléans, France
c UMR 6282 Biogéosciences, Univ. Bourgogne Franche-Comté, CNRS, 6 Bd Gabriel, 21000 Dijon, France

Abstract

The objective of this work is to improve our understanding of the processes controlling changes in the architecture and facies of intracontinental carbonate platforms. We examined the facies and sequence stratigraphy of Aalenian to Oxfordian limestones of western France. Seventy-seven outcrop sections were studied and thirty-one sedimentary facies identified in five depositional environments ranging from lower offshore to backshore. Platform evolution was reconstructed along a 500 km cross-section. Twenty-two depositional sequences were identified on the entire western France platform and correlated with European third-order sequences at the biozone level, demonstrating that eustasy was the major factor controlling the cyclic trend of accommodation. The tectonic subsidence rate was computed from accommodation measurements from the Aalenian to the Oxfordian in key localities. Tectonism controlled the sedimentation rate and platform architecture at a longer time scale. Tectonic subsidence triggered the demise of carbonate production at the Bathonian/Callovian boundary while the uplift made possible the recovery of carbonate platform from Caen to Le Mans during the mid Oxfordian. Topography of the Paleozoic basement mainly controlled lateral variations of paleodepth within the western France platform until the mid Bathonian. A synthesis of carbonate production in the western Tethyan domain at that time was conducted. Stages of high carbonate production during the Bajocian/Bathonian and the middle to late Oxfordian are synchronous with low δ13C, high eccentricity intervals, and rather dry climate promoting (1) evaporation and carbonate supersaturation, and (2) oligotrophic conditions. Periods of low carbonate production during the Aalenian and from the middle Callovian to early Oxfordian correlate with high δ13C and low eccentricity intervals, characterized by wet climate and less oligotrophic conditions. Such conditions tend to diminish growth potential of carbonate platforms. This work highlights the importance of climate control on carbonate growth and demise at large scale in western Tethyan epicontinental seas.

© 2016 Elsevier B.V. All rights reserved.

Keywords:
Carbonate platform
Facies
Sequence stratigraphy
Tectonics
Jurassic

1. Introduction

In recent decades, numerous studies have been conducted on Jurassic carbonate platforms of the Western Tethys to characterize their architectures, facies, and discontinuities. They highlighted the importance of interactions among the multiple processes that control the evolution of carbonate depositional systems, that is: (1) eustasy, (2) tectonism in that it modifies accommodation, and (3) environmental conditions including seawater temperature, depth, salinity, chemistry, trophic conditions and productivity, with this last aspect determining the nature of carbonate producers and production rates (Pomar, 2001; Brigaud et al., 2009; Carpentier et al., 2010). Deciphering the precise influence of each factor on depositional sequences, on carbonate platform architecture, growth, and demise, and on facies distribution and composition is still a challenge (Mutti and Hallock, 2003; Léonide et al., 2007; Merino-Tomé et al., 2012), especially because accommodation and carbonate production are interdependent factors (Pomar, 2001). For instance, intracratonic basins were considered stable until recent studies showed that tectonics influenced sedimentation (Guillocheau et al., 2000; Carcel et al., 2010; Brigaud et al., 2014; Wright, 2014). Nevertheless, determining the specific control of tectonics on carbonate platform development is complex, especially because vertical movements are of low amplitude and detected over long wavelengths. Moreover, the control of the inherited basement structure over the architecture of carbonate
platform is rarely discussed (Allenbach, 2002); western Tethyan carbonate platforms underwent several stages of growth and demise during the Jurassic, which are particularly well-marked in the Paris Basin (Brigaud et al., 2014). However, no large-scale synthesis of such changes in sedimentation patterns and facies has been conducted, while it could provide new insights on the main factors controlling the type of carbonate producers and production rate.

From the Albian to the Oxfordian (about 17 My, Gradstein et al., 2012), a vast carbonate platform developed in western France. Several stratigraphic studies based on the distribution of ammonite and brachiopod species provide a precise and reliable biostratigraphic framework within which to study the evolution of this platform at the biozone level. However, such studies remain limited to specific geographic areas or stages (Caen to Alençon: Dugué, 1989; Rioult et al., 1991; Préat et al., 2000; Dugué, 2007; Poitiers region: Mourier, 1983; Lorenz, 1992; Gonnin et al., 1992, 1993, 1994; Lenoir, 2012; Saint-Maixent l’École to Montbron: Foucher, 1986; Faugeras, 1988; Branger, 1989). The western France platform has tremendous potential for improving our understanding of Jurassic intracratonic carbonate systems because (1) it offers a continuous record of 17 My that can be studied through numerous outcrops and boreholes with a good time control based on biostratigraphy, (2) it rests upon a highly structured Hercynian basement, and (3) it has never been studied as a whole.

The first objective of this work is to reconstruct changes in facies and architecture of the western France platform and to determine the respective influences of eustasy, tectonics, environmental conditions, and the topography of the Paleozoic basement on platform evolution. The second objective is to compare our results at broad scale with other platforms in western Tethyan epicontinental sea. Facies were described and classified so as to define depositional environments. A 500 km-long north–south cross-section from Deauville to Montbron is examined within a sequence stratigraphy framework to model the evolution of the platform from the Alençon to the Oxfordian. Four accommodation curves are established to quantify the amplitude of vertical movements and precisely identify tectonic control over the evolution of this intracratonic carbonate platform. A synthesis of sedimentation patterns in western Tethyan domain during the Middle Jurassic and the Oxfordian is conducted and a large-scale model for western Tethyan carbonate platforms evolution is provided.

2. Geological setting

The Paris and Aquitaine Basins are two intracratonic sedimentary basins filled with Triassic to Quaternary deposits. They developed respectively over a Cadomian and Variscan basement (Paris Basin, Guillou-Buccoux et al., 2000) and a Variscan basement (Aquitaine Basin, Biteau et al., 2006). The study area stretches from Deauville, in the northwestern Paris Basin, to Angoulême, in the northern Aquitaine Basin, along an approximately 500-km transect (Fig. 1). From Aalenian to Oxfordian times, western France was positioned at subtropical latitudes (20–30°N) and was covered by a shallow epicontinental sea. Sediments are predominantly composed of shallow marine carbonates deposited over a vast platform open to the Atlantic, Tethys, and Arctic oceans (Fig. 2A–B; Enay and Mangold, 1980; Thierry and Barrier, 2000).

Several studies of ammonite and brachiopod fauna have been conducted for Middle and Late Jurassic outcrops in areas around Caen (Rioult et al., 1991), Argental (Kuntz et al., 1989; Ménillet et al., 1997; Moguet et al., 1998), Poitiers (Gably et al., 1978; Mourier, 1983; Cariou and Joubert, 1989a; Lorenz, 1992), and between Saint-Maixent l’École and Montbron (Cariou et al., 1973; Foucher, 1986; Faugeras, 1988; Branger, 1989; Cariou et al., 2006). They have made it possible to define a precise and reliable biostratigraphical framework at the ammonite biozone level (Fig. 3). The stratigraphic framework is briefly described for the five areas of our study, displaying numerous outcrops and various facies: Caen, Argental/Alençon, Poitiers Saint-Maixent-l’École, and Montbron. Readers should refer to Mégnien and Mégnien (1980) for a detailed description of lithostratigraphic units of the study area.

The Caen area was more or less isolated from the Paris Basin during the Trias and at the onset of the Jurassic, and remained separated from Argental by immersive basement reliefs until the end of the Bathonian (Enay and Mangold, 1980; Dugué, 2007). Around Caen, the Bajocioc to Oxfordian succession includes two major episodes of carbonate platform growth, (1) from the late Bajocioc (parkinsoni Zone) to the late Bathonian (discus Zone) and (2) from the mid (densiplicatum Zone) to late Oxfordian (glosense Zone; Fig. 3, Rioult et al., 1991). From the late Bajocioc to the late Bathonian, bioclastic facies rich in echinoderms (Calcaires de Caen, Calcaires de Creully, Calcaires de Ranville Formations), bryozoans (Calcaires de Largrune Formation), and sponge bioconstructions (Caillolasses de Basse-Ecarde Formation) dominated. The middle to upper Oxfordian deposits display ooid- and coral-dominated facies (Calcaires d’Auberville, Calcaires oolithiques de Trouville, Coral Rag Formations; Fig. 3; Rioult et al., 1991). The Alençon to late Bajocioc (garantiana Zone) was, like the Toarcian, a period of low carbonate production, with highly condensed ferruginous levels (Oolithe ferrugineuse aulénienne, Couche verte, Conglomérat de Bayeux, Oolithe de Bayeux Formations, Rioult et al., 1991). A carbonate platform demise occurred from the very late Bathonian (discus Zone) to late early Oxfordian (cordatum Zone), a period characterized by clayey deposits (Argiles de Lion, Marnes d’Escoville, Marnes d’Argences, Marnes à Belemnopsis latesulcata, Marnes et calcaires sableux de Crevecœur, Marnes de Dives, Marnes de Villers, Oolithe ferrugineuse de Villers et Argiles à Lopha gregarea Formations; Fig. 3).

The area from Argental to Alençon formed a small basin, separated from Caen by immersive basement reliefs (Dugué, 2007). Detrital deposits formed during the Aalenian (Sables et graviers de Tessé Formation), displaying brachiopods and ammonites (Graphoceras sp., Graphoceras cornu) from the murchisonae and concavum Zones (Fig. 3; Juignet et al., 1984). During the Bajocioc to late Bajocioc (discus Zone), a bioclastic to oolitic carbonate platform grew (Oolithe à pentacrinés, Oolithe de Villaines-la-Carelle, Calcaires d’Écouché, Calcaires de Sarceaux and Calcaires d’Argental Formations; Fig. 3; Juignet et al., 1984; Kuntz et al., 1989). The very late Bathonian (discus Zone), Callovian, and early Oxfordian deposits are characterized by clays to sands (Marnes de Montcel, Marnes du Chevain, Oolithe de Séuré, Marnes et calcaires sableux, Oolithe ferrugineuse de Chemillé, Marnes de Montarlo, Sables de Saint-Fulgent, Marnes à Vernes, Sables de Mortagne Formations; Fig. 3; Ménillet et al., 1997). The mid to late Oxfordian limestone displays ooid and coral rich facies (Calcaires coralliens de Mortagne, Calcaires à Astrates; Fig. 3; Moguet et al., 1998).

Around Poitiers, the marl/limestone alternations (Calcaires argileux d’Airvault Formation) of the early Alençon pass upwards into bioclastic to oolitic limestones (Calcaires finement dolomitiques à silex, Calcaires roux bioclastiques à oncolithes, Calcaires dolomitiques à bioclastiques de Poitiers, Calcaires graveleux de Poitiers, and Pierre des Lourdines Formations; Fig. 3; Mourier et al., 1986; Cariou and Joubert, 1989a). There is a sedimentary gap between the late Callovian and the late early Oxfordian (athleta, lamberta and mariae Zones; Cariou and Joubert, 1989a). Middle and upper Oxfordian deposits are composed of crinoid limestones (Pierre grise de Bonnillet, Calcaires crinoidiques Formations) and clayey sedimentary sets (Banc de Pierre subolithographique, Calcaires argileux et glauconieux de Mirebeau Formation; Fig. 3; Cariou and Joubert, 1989a).

The Saint-Maixent l’École area was located on a relatively low-energy platform at the margin of the Golfe Charentais basin during the Mid Jurassic (Enay and Mangold, 1980). The Saint-Maixent l’École area is dominated by marl/limestone alternations overlying the Alençon (Marnes bleues Formation) and from the early Callovian to the late Oxfordian (macrocephalus to binnamatum zones; Calcaires argileux de Pamproux, Marnes de Velluire, Pierre chauffante, Marnes gris-bleues à ammonites pyriteuses, Marnes grises à spongies, Marnes Schistes and Calcaires blancs de Fors Formations; Cariou et al., 2006). The
Bajocian and Bathonian limestones are characterized by bioclast-rich facies (Calcaires ponctués de Saint-Maixent, Calcaires à silex Formations; Cariou et al., 2006).

In the Montbron area, in the western part of the high-energy Central Platform (Figs. 2A, 3), dolomite (Dolomies de Montbron Formation) and shallow carbonate facies (Calcaires de Combe Brune, Calcaires de...
Saint-Sauveur, Calcaires de Vilhonneur, Calcaires crayeux à stromatolithes de Montbron, Calcaires oolithiques et coralliens de Garat Formations; Foucher, 1986; Le Pochat et al., 1986) were deposited from the Bajocian to Oxfordian. The biostratigraphy of brachiopod faunas (Foucher, 1986) reveals a sedimentary gap during the early Callovian. The Aalenian deposits are characterized by clay and marl facies (Argiles et marnes grises Formation).

3. Material and methods

3.1. Sedimentology

This study is based on the detailed examination of 77 outcrop sections lying between the cities of Deauville and Montbron. Observations were completed with data from previous works carried out on 13 outcrop sections (Foucher, 1986; Dugué, 1989; Louail et al., 1989; Maire, 1983; Rioult et al., 1991) and 37 boreholes (Guyader et al., 1970; Bourgueil et al., 1971; Roger et al., 1979; Rioult and Doré, 1989; Ménillet et al., 1999) and from borehole data of the French Geological Survey available online [http://infoterre.brgm.fr/]. Outcrop sections were logged in detail: lithology, texture (Dunham, 1962; Embry and Klovan, 1971; Insalaco, 1998), allochem content, and sedimentary structures were reported. A total of 268 thin sections were examined to determine the microfacies using a polarizing microscope. The proportion of facies components (bioclasts and non-bioclastic elements), texture, granulometry, grain sorting, and early cement types were characterized in each thin section. We followed the definition (1997) of James for photozoan and heterozoan grain associations. Photozoan grain association is rich in hermatypic corals, chlorophytes, benthic foraminifera and non-bioclastic components as ooids, peloids and aggregates.

Fig. 2. A. Global palaeogeographical map of the Callovian simplified from R. Blakey's maps [http://cpgeozyystems.com]. B. Location of the study area on a palaeogeographical map of the Western Tethys during the Callovian. C. Location of the study area during the Bathonian in a paleogeographic reconstruction (modified from Brigaud et al., 2009; based on Enay and Mangold, 1980; Ziegler, 1988; Thierry and Barrier, 2000; Hendry, 2002). D. Location of the study area during the Oxfordian in a paleogeographic reconstruction (modified from Strasser et al., 2015; based on Enay and Mangold, 1980; Ziegler, 1988; Thierry and Barrier, 2000).
Heterozoan facies association displays abundant rhodophytes, molluscs, echinoderms and bryozoans. Transitional facies designates grain associations in which neither heterozoan nor photozoan producers are prevailing.

### 3.2. Sequence stratigraphy

Following Embry's (2009) definition, outcrop sections and boreholes were interpreted in terms of sequence stratigraphy to establish a stratigraphic cross-section. In this sequence stratigraphy model, units are bounded either by a subaerial unconformity (SU) when the surface was exposed, or by a maximum regressive surface (MRS) when the surface was not exposed. Maximum regressive surfaces, also called transgressive surfaces (Van Wagoner et al., 1988), or flooding surfaces (Homewood et al., 1992), coincide with shifts in stacking patterns between shallowing-upward and deepening-upward trends, and correspond to the shallowest depositional environment recorded within a sequence (Embry, 2009). Maximum flooding surfaces (MFSs) mark a shift between deepening-upward and shallowing-upward trends, and correspond to the deepest facies encountered within a sequence. Subaerial unconformities or maximum regressive surfaces form sequence boundaries (SBs). Depositional sequences are composed of transgressive and regressive systems tracts (TSTs and RSTs). The transgressive systems tract is characterized by prograding architectures, with more proximal facies upward.

### 3.3. Decompacted depth, accommodation, subsidence, and tectonic subsidence calculation

Decompacted depth, also called total subsidence, corresponds to the thickness of sediments after decompaction (Steckler and Watts, 1978; Allen and Allen, 2005). Goldhammer (1997) demonstrates that, in mud-supported carbonates at least, compaction mainly occurs during shallow burial with 100 to 400 m overburden. Considering that all the formations of the Middle and Upper Jurassic were buried beneath more than 500 m of sedimentary deposits, the following compaction factors were used for the entire sedimentary series: 1.2 for grainstones, 1.5 for packstones, 2 for wackestones, 2.5 for mudstones, and 3 for marls (Hillgärtner and Strasser, 2003).

Accommodation space can be defined as the sum of the decompacted thickness of sediments and of paleodepth variations (Robin et al., 2000). For each facies, paleodepths were estimated based on the examination of sedimentary structures and fossil fauna (especially corals, Lathuilière et al., 2005). The classic wave zonation is used, considering depths of 10–15 m for the fair-weather wave base and more than 40 m for the storm wave base (Walker and James, 1992; Sahagian et al., 1996). Subsidence corresponds to the increase in accommodation that is not due to eustasy, and so created by basement movements. It was calculated by subtracting eustasy variations from accommodation changes.
using the eustasy curve of Haq et al. (1987). The curve was placed in the European context of third-order sequences as defined by Hardenbol et al. (1998b). Although there is debate about this curve (Miller et al., 2005), it is the only eustasy curve precise enough to estimate eustatic variations at the time resolution used in this study, and so the only curve that can be used to reconstruct subsidence variations at the third-order sequence scale. Current criticisms of Haq’s charts claim that it overestimates the amplitude of eustatic variations. In Cretaceous and Cenozoic times, Rowley (2013), based on worldwide continental flooding, have identified large overestimations of the magnitude of variations and restored elevations of eustatic level. No reassessment of the kind has yet been published for the Jurassic, but this could be a major bias in estimates of subsidence amplitudes. However, as the eustasy values subtracted from the measured accommodation are all from the same chart (Haq et al., 1987), this misestimate may affect the amplitude but not the trend. We therefore focus mainly on the trend and variations between sites.

Tectonic subsidence is independent of the isostatic adjustment due to sediment deposition or paleodepth variations. It was calculated using the equation of Steckler and Watts (1978):

$$Y = S \left( \frac{\rho_m - \rho_b}{\rho_m - \rho_w} \right) - \Delta SL \left( \frac{\rho_m}{\rho_m - \rho_w} \right) + Wd$$

$Y$ is tectonic subsidence, $S$ is decompacted sediment thickness, $\rho_m$ is mantle density, $\rho_b$ is mean sediment density, $\rho_w$ is water density, $\Delta SL$ is eustatic variation, and $Wd$ is paleodepth variation. Mean densities of $3.3 \text{ g cm}^{-3}$, $2.6 \text{ g cm}^{-3}$, and $2.4 \text{ g cm}^{-3}$ were used for mantle rocks, carbonates, and marls respectively.

Decompacted sediment thickness, accommodation, subsidence, and tectonic subsidence values were calculated for each third-order sequence top, corresponding to maximum regressive surfaces, and with an average time step of 0.8 My (Gradstein et al., 2012), to reconstruct sequence evolution over the Aalenian to Oxfordian time interval.

Uncertainties in accommodation, subsidence, and tectonic subsidence calculation may be due to uncertainties in the estimates of (1) the decompacted sediment thickness, (2) paleodepth, and (3) eustasy (for subsidence and tectonic subsidence only). Uncertainties in the decompacted sediment thickness depend on the chosen compaction law (Hillgärtner and Strasser, 2003). Uncertainties in paleodepth estimates are $60 \text{ m} +/− 15 \text{ m}$ for the lower offshore, $30 \text{ m} +/− 10 \text{ m}$ for the upper offshore, $10 \text{ m} +/− 5 \text{ m}$ for the shoreface and the lagoon, and $0 \text{ m} +/− 10 \text{ m}$ above sea level for continental environments. A $+/−5 \text{ m}$ uncertainty was used to take into account reading precision of the curve by Hardenbol et al. (1998b). However, as explained above, the eustasy variations used (Haq et al., 1987) are probably overestimated and the subsidence and tectonic subsidence curve should not be regarded as absolute values but as trends.

4. Results

4.1. Sedimentary facies

Thirty-one different facies were identified in the limestones under study. The facies were grouped into five positions within the downdip profile (Fig. 4): (1) the lower offshore for facies deposited below the storm wave base, (2) upper offshore for facies deposited between the storm wave base and the fair-weather wave base, (3) shoal/barrier environments for wave- and tide-dominated facies deposited above the fair-weather wave base, (4) lagoon to intertidal environments for most mud-supported carbonate rocks deposited in calm and shallow environments, and (5) backshore for facies deposited in sebkha, brackish, or continental environments, above high tides. Observations and descriptions are summarized in Table 1. Five synthetic sedimentological logs synthesize the succession of Aalenian to Oxfordian deposits in the Caen, Alençon, Poitiers, Saint-Maixent-l’École, and Montbron areas (Fig. 5).

4.1.1. Lower offshore: facies F1a to F1f

Claystones and marls with micritic limestone alternations (F1a–b–c, Fig. 6A–B), ferruginous ooid wackestones to packstones (F1d, Fig. 6C–D), ferruginous oncid floatstones (F1e, Fig. 6C–E–F), and glauconitic to phosphatic bioclastic wackestones to floatstones (F1f, Fig. 6G) are present in the lower offshore. Fauna are composed of ammonites, belemnites, bivalves, brachiopods, benthic and planktonic foraminifers, sponge spicules, ostracods, echinoderms, dinoflagellates, serpulites, and gastropods. Bioturbation is frequent and trace fossils abundant (Chondrites, Thalassinoides, Walthonensis; Dugué, 1989). Ferruginous oncoids of facies F1e are composed of microstromatolitic iron mats dominated by filamentous bacteria and fungi living below the photic zone in dysoxic waters (Préat et al., 2000). The very fine grain size, the absence of sedimentary structures, the absence of scleractinian corals that are typical of the euphotic zone, the abundant bioturbation, and the presence of ammonites and belemnites indicate a very calm depositional environment, probably below the storm wave base (Pomar, 2001; Lathuilière et al., 2005).

4.1.2. Upper offshore: facies F2a to F2f

Mostly bioclastic limestones, intercalated with thin marl interbeds are present in the upper offshore environments (Fig. 7A). Six facies were identified: (1) alternations of marl and echinoderm/peloid packstones to grainstones (F2a, Fig. 7A–B), (2) alternating marls and bioclastic/bioclastic packstones to grainstones (F2b, Fig. 7C), (3) marl/peloid grainstone alternations (F2c), (4) sponge wackestones to floatstones (F2d, Fig. 7D–E), (5) sheetstone to platestone facies with lamellar corals, and (6) sponge bioclastite in marls (F2e–f, Fig. 7F). The dominant fauna is composed of echinoderms, bivalves, sponges, lamellar microsolenid corals, brachiopods, bryozoans, and foraminifers. It indicates normal oxygenation and salinity conditions. The dominance of lamellar microsolenid corals argues for an environment located within the photic zone. In the eastern Paris Basin, similar coral associations developed in paleodepths ranging between 20 m and 40 m during the Oxfordian (Lathuilière et al., 2005). The upper surface of the limestone beds is often bioturbated by vertical burrows filled in with the overlying clays. The accumulation of fragmented bioclastic shells forming shelf-graded layers (F2a–b), gutter casts (F2a–b–c, Fig. 7A), hummocky cross-stratification (HCS, F2b), and upturned sponges (F2c) suggests that sedimentation was under storm influence in the upper offshore between the storm wave base and the fair-weather wave base.

4.1.3. Shoreface: facies F3a to F3l

Twelve facies are distinguished in shoreface environments: (1) very fine peloidal grainstones (F3a, Fig. 8A), (2) superficial ooid grainstones (F3b), (3) bioclastic grainstones to rudstones (F3c, Fig. 8B), (4) bioclastic peloidal grainstones (F3d, Figs. 8C, 9A), (5) quartz bioclastic sands to sandstones (F3e, Fig. 9B), (6) sponge bioclastite in bioclastic grainstones (F3f, Fig. 9C), (7) mixstone to domestone coral bioclastite (Isostrea, Thecosmilia, Rioult et al., 1989) in oolitic grainstones (F3g), (8) echinoderm grainstones (F3h, Figs. 8D, 9D), (9) briozoan grainstones to rudstones (F3i, Fig. 8E), (10) ooid grainstones to sands (F3j, Figs. 8F, 9E–F–G), (11) lithoclast-ooid grainstones to rudstones (F3k, Figs. 8G, 9F), (12) lithoclast-oyster calcareous conglomerates (F3l, Figs. 8H, 9H). The presence of grainstone to rudstone textures, common low-angle large-scale clinobeds (Fig. 9A), tabular to trough cross bedding in dunes (Fig. 9B, G), erosive channels (Fig. 9A), wedges in clinobedded oolitic grainstones (Fig. 9F), and herringbone cross bedding suggests a high-energy wave- or tide-influenced environment located above the fair-weather wave base. This is congruent with the mixstone to domestone texture of coral bioclastites and with the occurrence of the coral genera Isostrea and Thecosmilia, which characterize paleodepths between 20 m and
0 m in the eastern Paris Basin during the Oxfordian (Lathuilière et al., 2005).

4.1.4. Lagoon and intertidal environments: facies F4a to F4d

Four facies are distinguished in: (1) coral bioconstructions (mixstones to domestones) in micritic limestones (F4a), (2) bioclastic/peloid/quartz mudstones (F4b, Fig. 10A), (3) pellet/peloid grainstone/packstones (F4c, Fig. 10B), and (4) oncoid-ooid rudstones to packstones associated with grapstones and lithoclasts (F4d, Fig. 10C–D–E). In these mud-dominated facies, the fauna is mainly composed of bivalves, corals, gastropods, and miliolids. Brachiopods, bryozoans, and echinoderms occur in low numbers. From Argentan to the Alençon area, these facies may be directly transgressive on the basement highs and alternate with oolitic shoal deposits (Fig. 10F). The dominance of peloid to oncoid facies and the presence of miliolids and well-preserved gastropods (especially nerineids) argue for protected lagoonal environments. However, washover deposits in facies F4b are characteristic of events of high-hydrodynamism (storms probably). Tangential cross bedding in dunes, and grain-supported to mud-supported textures in facies F4c and F4d indicate variable hydrodynamic conditions. Microbialites and microbial peloids are common in facies F4c and F4d (Fig. 10C). Microstalactitic and meniscus cements also occur in both facies (Fig. 10D); they are characteristic of cementation occurring in a vadose diagenetic environment. The local occurrence of birdseyes (Fig. 10E) and of planar-bedding in facies F4c and F4d indicate that they can form in intertidal environments.

4.1.5. Supratidal to continental environments: facies F5a to F5c

Three facies can be present in supratidal to continental environments: (1) dolomitic facies (F5a, Fig. 10G), (2) lignite-rich facies (F5b), and (3) paleosoil layers (F5c). The dolomitic facies (F5a, Fig. 10G), exclusively present at Montrhon, includes stromatolites (Le Pochat et al., 1986). It is characteristic of supratidal environments of high salinity or evaporation rates present in sabkha environments. The lignite facies (F5b) consists of organic matter rich deposits, of very fine grain size (silt), and displays characeae gyrogonites and ostracods (Foucher, 1986) suggesting a very calm and protected brackish depositional environment. Paleosoils (F5c) are described by Dugué (1989) and Ménillet et al. (1999) around Caen and Argentan. They are composed of silts, clays, and quartz grains of about 0.1 mm diameter. The occurrence of characeae gyrogonites and roots suggests environments of coastal plain deposits (Dugué, 1989).

4.2. Facies architecture and depositional sequences

The spatial distribution and temporal evolution of the facies defined above is described below. Twenty-two transgressive-regressive cycles were identified between the early Aalenian and the late Oxfordian, whose maximum flooding surfaces are underlined by clays (Figs. 5, 11). Considering a time range of about 17 My for the entire interval (Gradstein et al., 2012), a rough estimate of the average duration of each cycle comes to 0.8 My, which is close to the duration range of...
<table>
<thead>
<tr>
<th>Position within the down dip profile (hydrodynamics)</th>
<th>Lithofacies</th>
<th>Non-bioclastic components</th>
<th>Bioclastic components</th>
<th>Predominant grain association</th>
<th>Sedimentary and biogenic structures</th>
<th>Sorting and grain size</th>
<th>Energy and depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower offshore</td>
<td>F1a: claystone</td>
<td>Silts (R), wood fragments (R)</td>
<td>Ammonites (R), belemnites (R), bivalves (R), sponge spicules (R), foraminifers (R), ostracods (R)</td>
<td>Abundant bioturbation; Chondrites, Thallasinoideae, Walthomnensis</td>
<td>Very well sorted; &lt;4 μm</td>
<td>Very low energy, lower offshore</td>
<td></td>
</tr>
<tr>
<td></td>
<td>F1b: marls</td>
<td>Silts (C), wood fragments (R)</td>
<td>Ammonites (R), belemnites (R), bivalves (R), brachiopods (R), sponge spicules (R), foraminifers (R), ostracods (R), echinoderms (R), dinoflagellates (R)</td>
<td>Heterozoan</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>F1c: marl/micritic limestone alternations</td>
<td>Peloids (A), quartz (C), glauconite (C), intraclasts (R), Ferruginous ooids (A), intraclasts (R), quartz (R), glauconite (R)</td>
<td>Echinoderms (F), bivalves (F), foraminifers (R), sponges (R), foraminifers (R), brachiopods (R), gastropods (R), bryozoans (R)</td>
<td>Heterozoan</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>F1d: ferruginous ooids packstone to wackestone</td>
<td>Ferruginous oncoids (A), ferruginous ooids (F), peloids (R), quartz (R), glauconite (R)</td>
<td>Bivalves (F), echinoderms (C), sponges (R), foraminifers (R), brachiopods (R), gastropods (R), bryozoans (R)</td>
<td>Heterozoan</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>F1e: ferruginous oncoid-oid floatstone</td>
<td>Glaucnite (F), phosphate (C), quartz (R), glauconite (R)</td>
<td>Echinoderms (F), bivalves (F), belemnites (C), ammonites (C), brachiopods (R), foraminifers (R)</td>
<td>Heterozoan</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>F1f: Glauconitic to phosphate bioclastic wackestone to floatstone</td>
<td>Peloids (F), quartz (R), oncoinds (R), superficial ooids (R), lithoclasts (R), pellets (C), lumps (C), oncoinds (C), glauconite (R)</td>
<td>Bivalves (A), brachiopods (R), crinoids (C), sponges spicules (R), serpulites (R)</td>
<td>Heterozoan</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper offshore</td>
<td>F2a: marl/echinoderm-peloid packstone to grainstone alternations</td>
<td>Peloids (F), quartz (R), oncoinds (R), superficial ooids (R), lithoclasts (R), pellets (C), lumps (C), oncoinds (C), glauconite (R)</td>
<td>Echinoderms (A), bivalves (C), brachiopods (R), foraminifers (R), ostracods (R), echinoderms (R), dinoflagellates (R)</td>
<td>Heterozoan</td>
<td>Gutter casts, shell graded layers, bioturbation</td>
<td>Well sorted, 200 μm–500 μm</td>
<td>Low to moderate energy, upper offshore</td>
</tr>
<tr>
<td></td>
<td>F2b: marl/bivalve-brachiopod wackestone to packstone alternations</td>
<td>Oolites (F), radial ooids (C), superficial ooids (R), micritic ooids (R), quartz (R), pellets (R), intraclasts (R), graptolites (R)</td>
<td>Oocoids (C), echinoderms (R), bivalves (R), foraminifers (R), brachiopods (R), bryozoans (R)</td>
<td>Transitional</td>
<td>Gutter casts</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>F2c: marl/ooid grainstone alternations</td>
<td>Peloids (C), oncoinds (C), superficial ooids (R), micritic ooids (R), quartz (R), pellets (C), intraclasts (R), graptolites (R)</td>
<td>Sponges (A), sponge spicules (F), echinoderms (F), bivalves (C), bryozoans (R), foraminifers (R), ostracods (R), brachiopods (R), gastropods (R), serpulites (R)</td>
<td>Transitional</td>
<td>Bioturbation, reversed sponges</td>
<td>Very poorly sorted, 100 μm–10 cm</td>
<td>Low to moderate energy, upper offshore</td>
</tr>
<tr>
<td></td>
<td>F2d: sponge wackestone to floatstone</td>
<td>Peloids (C), oncoinds (C), superficial ooids (R), intraclasts (R)</td>
<td>Lamellar corals (A), bivalves (R), echinoderms (F), sponge spicules (R), serpulites (R)</td>
<td>Transitional</td>
<td>Biostroms (50 cm to 1 m thick and 2–15 m wide)</td>
<td></td>
<td>Upper offshore</td>
</tr>
<tr>
<td></td>
<td>F2e: coral bioconstructions (sheetstone to platestone) in marls</td>
<td>Peloids (R)</td>
<td>Sponges (A), bivalves (C)</td>
<td></td>
<td>Biostroms (50 cm thick and 2–5 m wide)</td>
<td></td>
<td>Upper offshore</td>
</tr>
<tr>
<td></td>
<td>F2f: sponge bioconstruction (boundstone) in marls</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Shoreface</td>
<td>F3a: very fine peloidal grainstone</td>
<td>Peloids (A), superficial ooids (R), intraclasts (R), quartz (R), intraclasts (R)</td>
<td>Echinoderms (C), bivalves (C), miliolids (R), foraminifers (R), brachiopods (R), gastropods (R)</td>
<td>Transitional</td>
<td></td>
<td></td>
<td>Moderate energy, wave dominated shoreface</td>
</tr>
<tr>
<td></td>
<td>F3b: superficial ooid grainstone</td>
<td>Superficial ooids (A), radial ooids (C), micritic ooids (C), oolites (C), peloids (C), graptolites (R), quartz (R), oncoinds (R), intraclasts (R)</td>
<td>Bivalves (R), echinoderms (R), miliolids (R), foraminifers (R), brachiopods (R), bryozoans (R)</td>
<td>Photozoan</td>
<td>SCS (swaley cross stratification), cross stratification, bioturbations</td>
<td>Well sorted, 50 μm–1 cm</td>
<td>Moderate energy, wave dominated shoreface</td>
</tr>
<tr>
<td></td>
<td>F3c: bivalve grainstone to rudstone</td>
<td>Quartz (R), peloids (R), ooids (R), oncoinds (R), intraclasts (R), graptolites (R), quartz (R), oncoinds (R), intraclasts (R)</td>
<td>Bivalves (A), echinoderms (C), brachiopods (C), miliolids (R), corals (R), bryozoans (R), gastropods (R), red algae (R), foraminifers (R)</td>
<td></td>
<td></td>
<td>Very poorly sorted, 300 μm–10 cm</td>
<td>Moderate to high energy, wave dominated shoreface</td>
</tr>
<tr>
<td></td>
<td>F3d: bioclastic peloidal grainstone</td>
<td>Peloids (F), intraclasts (C), micritic ooids, superficial ooids, superclial ooids</td>
<td>Bivalves (C), echinoderms (C), bryozoans (C), miliolids (R), foraminifers (R),</td>
<td></td>
<td></td>
<td></td>
<td>High energy, wave dominated shoreface</td>
</tr>
</tbody>
</table>

(continued on next page)
<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Non-bioclastic components</th>
<th>Bioclastic components</th>
<th>Predominant grain association</th>
<th>Sedimentary and biogenic structures</th>
<th>Sorting and grain size</th>
<th>Energy and depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>F3c: quartz/bioclastic sand to sandstone</td>
<td>Ooids, radial ooids, oolites (C), oncoids (R), feldspar (R), glauconite (R), iron oxides, intraclasts (R), ferruginous ooids (R)</td>
<td>Gastropods (R), brachiopods (R), corals (R)</td>
<td>Echinoderm (R), bivalves (R), gastropods (R), echinoderms (R), foraminifers (R)</td>
<td>Well sorted</td>
<td>200 μm - 500 μm</td>
<td>High energy, wave to tidally dominated shoreface</td>
</tr>
<tr>
<td>F3f: sponge bioclastic sandstone (boundstone) in bioclastic grainstones</td>
<td>Sponges (Plectocladus magnus) (A), bivalves (C), bryozoans (C), serpulites (R), echinoderms (R), brachiopods (R), gastropods (R), red algae (R), miliolids (R)</td>
<td>Heterozoan</td>
<td>Bioclastic components</td>
<td>Bioclastic components</td>
<td>Bioclastic components</td>
<td>Bioclastic components</td>
</tr>
<tr>
<td>F3h: echinoderm grainstone</td>
<td>Peloids (C), ooids (R), oolites (R), quartz (R)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>F3i: bryozoan grainstone to rudstone</td>
<td>Peloids (R), ooids (R), oncoids (R), intraclasts (R)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>F3j: ooid grainstone to sand</td>
<td>Oolites, micritic ooids, radial ooids, superlithic ooids (A), peloids (C), nubecularia oncoids (R), intraclasts (R), graptolites (R), quartz (R)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>F3k: lithoclast-ooid grainstone to rudstone</td>
<td>Lithoclasts (F), oolites, micritic ooids, superlithic ooids, radial ooids (F), peloids (C), graptolites (R)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>F3l: lithoclast-oyster calcareous conglomerate</td>
<td>Quartz (A), intraclasts (C), peloids (R), ferruginous ooids (R), glauconite (R), iron oxide (R), feldspar (R)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>Lagoon and intertidal environments</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>F4a: coral bioconstructions (mixstone) in micritic limestones</td>
<td>Branching corals (Thomomastix dendroidea)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>F4b: bioclastic/peloid/quartz muddstone</td>
<td>Peloids (F), quartz (F), lithoclasts (C), micritic ooids, peloids (C), oncoids (R)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>F4c: pellet/peloid grainstone</td>
<td>Pellets (A), peloids (F), lithoclasts (F) superlithic ooids (C), oncoids (R), aggregates (R), micritic ooids, radial ooids (R), quartz (R)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>F4d: oncoid-ooid rudstone to packstone with aggregates and lithoclasts</td>
<td>Oolites, micritic ooids, superlithic ooids (F), nubecularia oncoids (F), aggregates (F), lithoclasts (F), peloids (R), quartz (R)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>Supratidal to continental environments</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>F5a: dolomite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>F5b: lignite</td>
<td>Silts (F)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
<tr>
<td>F5c: paleosol</td>
<td>Clays (F), silts (F), quartz (F), iron hydroxides (C)</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
<td>Photozoan</td>
</tr>
</tbody>
</table>

R = rare: <10%; C = common: 10–20%; F = frequent: 20–40%; A = abundant: >40%
Fig. 5. Synthetic sedimentological log with geological stages, ammonite biozones, and biostratigraphic data (ammonites, brachiopods, and foraminifers), lithostratigraphic formations, carbonate textures, depositional environments, third-order sequence stratigraphy and stratigraphic locations of outcrops and samples. The parts of the logs not observed in the field have been completed using boreholes and descriptions from the literature. A. Synthetic sedimentological log of the Caen region, completed from Rioult et al. (1989, 1991) and Maurizot et al. (2000). B. Synthetic sedimentological log of the Argentan to Alençon region, completed from Kuntz et al. (1989) and Moguedet et al. (1998). C. Synthetic log of the Poitiers region completed from Mourier et al. (1986) and Cariou and Joubert (1989a). D. Synthetic log of the Saint-Maixent-l’École region, completed from Cariou et al. (2006). E. Synthetic log of the Montbron region, completed from Le Pochat et al. (1986).
third-order cycles as reported by several authors (Haq et al., 1987; Hardenbol et al., 1998a; Schlager, 2004). Moreover, it corresponds to a multiple of 0.4 My, which is the duration of Earth’s eccentricity cycle, often considered as the controlling factor for third-order sequences (e.g. Strasser et al., 2000).

These cycles are arranged into lower-order cycles from an Aalenian regressive hemicycle topped by the maximum regressive surface Bj1 (Caen) or Bj3 (Montbron); a Bajocian–Bathonian cycle topped by the maximum regressive surface Bt2 (Caen) of Bt4 (Montbron; maximum flooding surface zigzag); and a Callovian–mid Oxfordian cycle until the
maximum regressive surfaces Ox6 (maximum flooding surface mariae or tenuiserratum).

A constant characteristic of the western France platform throughout the Aalenian to Oxfordian interval is the presence of clayey lower offshore environments from La Flèche to Mirebeau, where sedimentation only occurred during sequences MJIV, MJVII, LJVI and LJVII (Figs. 11, 12).

4.2.1. Aalenian to early Bathonian: detrital to bioclastic ramp (sequences MJ to MJVIII)

This interval contains eight third-order sequences from Deauville to Montbron (Figs. 5, 11): MJI (Middle Jurassic I, opalinum Zone), MJII (opalinum to murchisonae Zones), MJIII (bradforesis to concavum zones), MJIV (discites and laeviuscula Zones), MJV (propiquans and early humphriesianum Zones), MJVI (late humphriesianum and niortense zones).
Fig. 8. A. Very fine-grained peloidal (Pe.) grainstone with superficial ooids (Su. oo.), foraminifers (Fo.), echinoderms (Ech.), and bivalves (Bl.) – F3a – Bonnillet quarry (55) — Pierre des Lourdines Formation, sequence LJv. B. Bivalve (Bl.) and brachiopod (Br.) grainstone – F3c – Villedieu-lès-Bailleul quarry (25) — Calcaires de Villedieu Formation, sequence MJIX. C. Bioclastic peloidal (Pe.) grainstone with crinoids (Cr.), undifferentiated echinoderms (Ech.), gastropods (Ga.), bivalves (Bl.), serpulites (Se.) and a well-developed isopachous early cement (Iso. Ce.) – F3d – Belle-Eau quarry (32) — Calcaires de Belle-Eau Formation, sequence MJX. D. Echinoderm (Ech.) grainstone with crinoids (Cr.) and bivalves (Bl.) – F3h – Vierville-sur-Mer (3) — Calcaires de Saint-Pierre-du-Mont Formation, sequence MJVIII. E. Bryozoan (Br.) rudstone – F3i – Luc-sur-Mer (10) — Calcaires de Langrune Formation, sequence MJXI. F. Ooid (Oo.) sand with crinoids (Cr.) – F3j – Fay quarry (41) — Oolithe de Vilaines-la-Carelle Formation, sequence MJV. G. Lithoclast (Li.) rudstone with bryozoans (Br.), echinoderms (Ech.), ooids (Oo.), peloids (Pe.), and microbial peloids (Mi. Pe.) – F3k – Les Aucrais quarry (21) — Calcaires de Bon-Mesnil Formation, sequence MJX. H. Detritic conglomerate with quartz (Qz), lithoclasts (Li.), feldspaths (Fe.), and echinoderms (Ech.) – F3l – Fay quarry (41) — Sables et graviers de Tessé Formation, sequence MJIII.
Zones), MJVII (parkinsoni Zone), and MJVIII (zigzag/progracilis Zones). The Aalenian to early Bathonian is characterized by ramp geometries that are either detritic (Aalenian) or bioclastic (echinoderms, bivalves, and sponge bioherms; Bajocian/early Bathonian; Figs. 11, 12A to D). The zigzag Zone maximum flooding surface is marked by an increase in paleodepth leading to large-scale retrogradations and maximum expansion of clayey facies in the study area (Figs. 5, 11, 12D).

Fig. 9. A. Channel and large-scale low-angle clinobeds – F3d – Belle-Eau quarry (32) — Calcaires de Belle-Eau Formation, sequences MJIX and MJX. B. Cross-bedding stratifications – F3e – Fay quarry (41) — Sables et graviers de Tessé Formation, sequence MJIII. C. Sponge bioherms on bryozoan rudstones – F3f and F3i – Saint-Aubin-sur-Mer (8) — Calcaires de Basse-Ecarde Formation, sequence MJXI. D. Trough cross-bedding stratifications with alternating fine and coarse levels in echinoderm grainstone – F3h – Creully quarry (12) — Calcaires de Creully Formation, sequence MJVIII. E–F. Wedge of clinobedded oolitic grainstone foresets – F3j and F3k – Combe Brune quarry (84) — Calcaires de Combe Brune Formation, sequences MJIX to MJXI. G. Trough cross-bedded oolitic grainstone – F3j – Cosses Noires quarry (85) — Calcaires Vilhonneur Formation, sequence MJX. H. Lithoclast–oyster rich conglomerate, probably storm deposit – F3l – Fay quarry (41) — Sables et graviers de Tessé Formation, sequence MJIII.
Fig. 10. A. Bivalve (Bi.) mudstone – F4b – Garat Quarry (86) — Calcaires coralliens et oolithiques de Garat Formation, sequence MJXII. B. Peloidal (Pe.) and oolitic (Oo.) packstone to grainstone with quartz (Qz) – F4c – Boitron quarry (39) — Calcaires de Valframbert Formation, sequence MJX. C. Lithoclast (Li.) and encoid (On.) mudstone with miliolids (Mi.), ooids (Oo.), echinoderms (Ech.), and microbial peloids (Pe.) – F4d – Vilhonneur quarry (83) — Calcaires de Vilhonneur Formation, sequence MJX. D. Lithoclast (Li.) grainstone with meniscus cements (M. Ce.) – F4d – Combre Brune quarry (84) — Calcaires de Combe Brune Formation, sequence MJXI. E. Birdseye in a lithoclast (Li.) grainstone with ooids (Oo.), peloids (Pe.), and foraminifers (Fo.) – F4d – Chez Trappe quarry (87) — Calcaires de Saint-Martial Formation, sequence MJX. F. Unconformity of Bathonian limestone on Brioverian flyschs – F4b, F4c and F3j – Boitron quarry (39) — Calcaires de Valframbert Formation, sequence MJX. G. Dolomite formed in a presumably supratidal protected environment, as evidenced by the presence of stromatolites (Le Pochat et al., 1986) – F5a – La Rochebertrier (82) — Dolomies de Montbron Formation, sequence MJV. H. Lithoclast (Li)/Ooid (Oo) grainstone with gastropods (Ga) displaying an erosive surface (Er. S.) with microbial encrusting (Mi. En.), fibrous (Fi. Ce.), microstalactitic (Mi. Ce.) and meniscus cements (M. Ce.) – F3h – Les Aucrais quarry (21) — Calcaires de Bon-Mesnil Formation, maximum regressive surface B2.
Fig. 11. Correlation diagram of the Middle Jurassic and the Oxfordian in western France on a Deauville–Montbron transect. This illustration is composed of 61 outcrop sections studied in this work completed by previous descriptions of 13 outcrop sections (Foucher, 1986—outcrops 72, 73, 88, and 90; Dugué, 1989—outcrops 16, 17, and 40; Louail et al., 1989—outcrop 47; Rioult et al., 1991—outcrops 1, 4, 8, 10, 13, and 15) and 37 boreholes (Guyader et al., 1970—borehole A; Bourgueil et al., 1971—borehole AB; Roger et al., 1979—borehole AK; Rioult and Doré, 1989—borehole B; Ménillet et al., 1999—borehole C; and http://infoterre.brgm.fr/—remaining boreholes). Correlations are based on biostratigraphy (ammonite, brachiopod and foraminifer associations) and 22 recognized stratigraphic cycles (MJI to LJVI) delimited by sequence boundaries (Aa1 to Ox7).
Offshore environments dominate from Argentan to Deauville, where a ramp deepens northward. Sequence MJ1 consists in 1 m of ferruginous oncocid facies (F1d, Oolithe ferrugineuse oaléenne Formation, Opalinum Zone) and is capped with marly to silty limestones of sequences MJII to MJIV (facies F1c, Marnes de Port-en-Bessin Formation, concavum Zones, Figs. 5, 11, 12A). Sequences MJV and MJVI are located in a condensed succession composed of glauconitic limestones (F1f, Couche verte Formation, propinquus Zone), oncocid ferruginous facies (F1e, Conglomérat de Bayeux Formation, humphriesianum zone), and ooid ferruginous facies (F1d, Oolithe ferrugineuse de Bayeux Formation, garantiana to niortense Zones, Figs. 5, 11, 12B). Sequence MJVIII is marked by a mid-ramp rich in sponges that progrades northward (F2d, Calcaires à spongies Formation, Figs. 5A, 11, 12C). Around Caen, the maximum regressive surface between sequences MJVII and MJVIII (Bj5) is characterized by upper offshore sponge facies sharply overlain by lower off-shore marl/limestone alternations of sequence MJVII (F1a–b, Marnes de Port-en-Bessin Formation, Figs. 5D, 11, 12D).

Over the Aalenian to early Bathonian, the shallowest environments were located on Ordovician basement highs, between Argentan and Le Mans. For the Aalenian, facies are clastic sands with rare ammonites (Graphoceras sp. and Graphoceras cornu) characterizing the concavum Zone (facies F3e–l, Sables et graviers de Tessé Formation, Figs. 5B, 11, 12A). Sequence MJIII is bounded at the top by a maximum regressive surface marked by a perforated hardground with encrusting bivalves, also characterizing a sedimentary gap that occurs during sequences IV to VI (Bj1, Figs. 5, 11, 12B). Over sequence MJVII, basement highs are covered with oolitic sands to grainstones (F3j, Oolithe de Vilaines-fa-Curelle Formation, Figs. 5B, 11, 12C) that prograde towards Saumur. These shallow oolitic facies were dated to the niortense and parkinsoni Zones based on ammonites Pseudogarantiana sp. and Parkinsonia cf. parkinsoni respectively. At the top of sequence MJVII, the maximum regressive surface Bt5 is underlined by (1) a lignite layer attesting to a mid Bathonian age (Subcontractus to Morissi Zones; sequence MJX; Rioult and Fily, 1975; Rioult et al., 1991). The Calcaire oolithe ferrugineuse à Montlivaltia Formation displays cephalopods (Parocreatostrus waageni, Grossouvraria bathonica, Homoepiolanulites sp., Cosmoceras contrarium, Opis lorieri, and Opis similis) and brachiopod (Rioultina triangularis and Avonothyris sp.) indicating the early late Bathonian, hodsoni Zone (sequence MJX; Juignet et al., 1989; Le Gall et al., 1998). The fauna can be used to date the flooding of the basement highs to the mid and late Bathonian, subcontractus to hodsoni Zones (sequence MJX and MJX). During sequence XI (Discus Zone), a flat and shallow bryozoan to bivalve-rich ramp stretched from Deauville to Le Mans (F3i–c; Figs. 11, 12F). Oolitic limestones prograde from Le Mans to La Flèche (Fig. 11).

For sequences MJX and MJXI, a ramp shallowing southward developed between Poitiers and Saumur. Facies alternate between echinoderm to sponge wacke/packstones (sequences MJX and MJXI; F2a–d; Calcaires à silex and Calcaires graveleux de Poitiers Formation) and prograding superficial ooid grainstones (sequence MJX and regressive systems tract of sequence MJXI; F3b; Calcaires graveleux de Poitiers Formation; Figs. 5C–D, 11, 12E). Another ramp profile deepened southwards from Poitiers (shoreface to upper offshore) to Saint-Maixent-l’École (upper to lower offshore; Figs. 11, 12F). Around Montbron, shallow lagoon environments rich in peloids and oncoids (F3b–c–d, Calcaires de Saint-Saureu and Calcaires de Vilhonneur Formations, Figs. 5E, 11, 12E–F) were edged by ooid/lithoclast wedges prograding to the northwest over the mid to late Bathonian (F3j–k, Calcaires de Combe Brune Formation).

4.2.2. Mid and late Bathonian: echinoderm/bryozoan ramp from Deauville to Le Mans, oolitic platform at Montbron (sequences MJX–XI)

The middle to late Bathonian is composed of three third-order sequences: MJX (subcontractus to early morrisi Zones), MJX (late morrisi to orbis Zones), and MJXI (early discus Zone). Maximum regressive surfaces (Bt2–3–4) are marked either by marine perforated surfaces encrusted with bivalves in upper offshore and shoreface environments, or by exposure surfaces with vadose cements, stromatolites, or lignite levels in shoal and lagoonal environments (Figs. 5A–B–E, 11). During sequences MJX and MJX1, an echinoderm-rich shallow ramp, prograding northwards, stretched from Argentan to Deauville (F3h, Calcaires de Saint-Pierre-du-Mont and Calcaires de Ranville Formations, Figs. 5A, 11, 12E). A protected lagoon composed of peloidal mudstones to packstones (F4b–c–d, Calcaires de Valfambert Formation, Figs. 11, 12E) developed between the basement highs from Argentan to Alençon. Basement highs were overlain by shallow shoreface and lagoonal facies during the Bathonian (Figs. 11, 12A; Calcaires de Villedieu, Calcaires de Valfambert, Calcaire et oolithe ferrugineuse à Montlivaltia Formations; Dassiat et al., 1982; Juignet et al., 1989, Ménillet et al., 1994, 1997). The brachiopod fauna in the Calcaires de Valfambert Formation is composed of Burmirmychina turgida, Rhynchonelloidea eleganattula, and Ephyris oxonica that indicate a mid Bathonian age (Subcontractus to Morissi Zones; sequence MJX; Rioult and Fily, 1975; Rioult et al., 1991). The Calcaire oolithe ferrugineuse à Montlivaltia Formation displays cephalopods (Parocreatostrus waageni, Grossouvraria bathonica, Homoepiolanulites sp., Cosmoceras contrarium, Opis lorieri, and Opis similis) and brachiopods (Rioultina triangularis and Avonothyris sp.) indicating the early late Bathonian, hodsoni Zone (sequence MJX; Juignet et al., 1989; Le Gall et al., 1998). The fauna can be used to date the flooding of the basement highs to the mid and late Bathonian, subcontractus to hodsoni Zones (sequence MJX and MJX). During sequence XI (Discus Zone), a flat and shallow bryozoan to bivalve-rich ramp stretched from Deauville to Le Mans (F3i–c; Figs. 11, 12F). Oolitic limestones prograde from Le Mans to La Flèche (Fig. 11).

For sequences MJX and MJXI, a ramp shallowing southward developed between Poitiers and Saumur. Facies alternate between echinoderm to sponge wacke/packstones (sequences MJX and MJXI; F2a–d; Calcaires à silex and Calcaires graveleux de Poitiers Formation) and prograding superficial ooid grainstones (sequence MJX and regressive systems tract of sequence MJXI; F3b; Calcaires graveleux de Poitiers Formation; Figs. 5C–D, 11, 12E). Another ramp profile deepened southwards from Poitiers (shoreface to upper offshore) to Saint-Maixent-l’École (upper to lower offshore; Figs. 11, 12F). Around Montbron, shallow lagoon environments rich in peloids and oncoids (F3b–c–d, Calcaires de Saint-Saureu and Calcaires de Vilhonneur Formations, Figs. 5E, 11, 12E–F) were edged by ooid/lithoclast wedges prograding to the northwest over the mid to late Bathonian (F3j–k, Calcaires de Combe Brune Formation).


The Callovian to early Oxfordian contains six third-order sequences: MJXII (herveyi and early koegini Zones), MJXIII (late koegini and calloviensis Zones), MJXIV (jason and coronatum Zones), MJXV (athleta and lamberti Zones), LIJ (mariae and early corodatum Zones), and LIIJ (Late corodatum Zone). Transgressive systems tracts are characterized by increasingly clayey sedimentation upward, with a gradual disappearance of carbonate beds. The upper reaches of regressive systems tracts are commonly marked by prograding sandy to carbonate deposits until maximum regressive surfaces, corresponding to marine surfaces (Call to Cal4, Oxl1), which can display encrusting bivalves and phosphate coatings (Cal4, Foucher, 1986), or exposure surfaces, with early meniscus cements (Oxl1, Sables ferrugineux de Varais Formation; Figs. 5, 11). From Deauville to Le Mans, the study area is dominated by a flat clayey ramp (F1a–b–c; Figs. 5A–B–C–D, 11, 12G). From Mirebeau to Poitiers, a ramp progressively shallowed to upper offshore environments dominated by echinoderm...
Fig. 12. Facies distribution on carbonate architecture morphology of western France in nine successive steps from Aalenian to Oxfordian in a third-order sequence stratigraphy framework. Potentially active faults, Paleozoic basement and Lias deposits are represented to illustrate the multiple influences on carbonate platform/ramp morphology. A. Ramp geometry marked by the dominance of silty to sandy facies — sequences MJII and MJIII, Aalenian. B. Ramp geometry with condensed ferruginous facies in Caen and bioclastic upper offshore facies from Poitiers to Montbron — sequence MJV, Bajocian. C. Predominance of a storm dominated ramp with sponges, although oolitic facies develop in shallow environments (Montbron, Le Mans) — sequence MJVII, Bajocian. D. Bioclastic ramp rich in echinoderms — sequence MJVIII, Bathonian. E. Shallow ramp with echinoderm facies from Deauville to Le Mans, oolitic platform in Poitiers and Montbron — sequence MJX, Bathonian. F. Shallow ramp with bryozoan facies and sponge bioherms from Deauville to Le Mans, upper offshore bioclastic ramp in Poitiers, oolitic shallow platform in Montbron — sequence MJX, Bathonian. G. Lower offshore clayey ramp from Deauville to Mirebeau, bioclastic upper offshore in Poitiers; lagoon protected by an oolitic shoal in Montbron — sequence MJXIV, Callovian. H. Shallow protected platforms in Argentan and Montbron, separated by clayey deposits in lower offshore environments — sequence LJVI – Oxfordian. I. Emerged platform with soil horizon in Argentan separated from a shallow platform in Montbron by clayey deposits in lower offshore environments – sequence LJVI – Oxfordian.
Fig. 12 (continued).
to bivalve wackestones/packstones during Callovian sequences MJXII to MJXIV (F2a–b, Pierre des Lourdines and Calcaires argileux de Pamproux Formations, Figs. 5C, 11, 12G). Sequences MJXV and LJ are missing (gap of the athletta to mariae Zones) and sequence LJII is condensed in a less than 1 m-thick carbonate level in Poitiers. A ramp gradually deepened from Poitiers to Ruffec, and the paleodepth then abruptly shallowed at Montbron, with the development of a shallow platform lagoon protected by an oolitic shoal over sequence MJXIV (F3j and F4b; Calcaires Crayeux à Stromatolithes de Montbron Formation; Figs. 5E, 11, 12G). Brachiopod-based biostratigraphy indicates a gap for sequences MJII and MJXIII (Foucher, 1986), and sequences MJXV to LJ might be missing as well (Figs. 5C–D–E, 11).

4.2.4. Middle to late Oxfordian: coral/oooid platforms development (sequences LJIII–IV–V–VI–VII)

The mid to late Oxfordian interval is composed of five third-order sequences: LJIII (Early plicatilis Zone), LJIV (Late plicatilis Zone), LJV (tenuserratum Zone), LJVI (glosense to early serratum Zones), and LJVII (late serratum to rosenkranzi Zones). Ooid/coral shallow platform geometries predominate, with very steep slopes (Figs. 11, 12H–I).

Transgressive systems tracts are characterized by major retrogradations from Deauville to Argentan, from La Flèche to Le Mans, and from Saint-Maixent l’École to Montbron. The maximum flooding surfaces are marked by the deposition of marl to mud-supported upper offshore facies over the entire study area. Regressive systems tracts are marked by major progradations of lagoonal to shelfal environments from Argentan to Deauville, from Le Mans to Saumur, and from Montbron to Saint-Maixent-l’École (e.g. Calcaire oolithique de Lisieux Formation, Fig. 11). Maximum regressive surfaces correspond to periods of maximum expansion of lagoonal and shoreface facies, and may be either marine (Okx, OX5 and Ox6) or immiscive, with the occurrence of meniscus cements (Ox4, Argentan to Alençon) or soils (Ox6, Caen to Le Mans, Figs. 11, 12D).

Sequence LJIII occurs only in the northern part of the study area, where it is composed of clayey offshore facies. Over sequence LJIV and LJV, a shallow platform stretches from Caen to Le Mans, where a 100 km-long lagoon developed (F4d and F4a; Calcaires coralliens de Mortagne and Calcaires coralliens de la Ferté-Bernard Formations; Figs. 5A–B, 11, 12H). An oolitic to coral barrier delimited the northern and southern lagoon (F3a–j, Figs. 11, 12H). During the regressive systems tract of sequence LJIV, cross-bedded sands prograded from Caen towards Deauville (Flè. Sables de Glos Formation, Fig. 11). They change abruptly to lower offshore clays around Deauville (Argiles de Villers Formation, Fig. 11). These are overlain by the lagoon, mud-supported facies of sequence LJVII (F5c; Calcaires à Astrates Formation; Figs. 5B, 11, 12D; Dugué, 1989; Ménillet et al., 1999).

During sequence LJV, facies changed suddenly from marls (Marnes à spongioïdes Formation) to fine peloid/echinoderm grainstones between Mirebeau and Poitiers (F3a–h; Pierre grise de Bonnillet and Calcaires crinoidiques Formations), and a ramp deepened from Poitiers to Ruffec, where lower offshore facies dominated. Over sequences LJVI and LJVII, lower offshore clayey environments predominate from Saumur to Ruffec (F1c; Banc de pierre subtholithographique, Calcaires argileux et glauconieux de Mirebeau, Marnes grises à spongioïdes Formations; Figs. 5C, 11, 12H–I). During the mid/late Oxfordian, the Montbron area saw the emplacement of an ooid/peloidal barrier with coral biostromes ahead of a calm lagoon where coral/peloidal mudstones accumulated (F3a–d–j, F4b, Calcaires oolithiques et coralliens de Garat Formation, Figs. 5E, 11, 12H–I).

4.3. Depocentered sedimentation, accommodation, subsidence, and tectonic subsidence

- Early to early miid Aalenian (sequences MJII and MJIII). This interval is characterized by stable accommodation and near-zero decomposed sediment thickness rates (Figs. 13, 14). Notable decomposed accumulation (13 m/My) and subsidence rates (5 m/My) are only present in the Poitiers area.

- Late mid Aalenian to early Bajocian (sequences MJIII to MJV). Accommodation and decomposed sediment thickness rates for the entire interval range from 0 m/My in Argentan to 15 m/My in Montbron (Fig. 13). A substantial accommodation rate is only recorded in Poitiers for sequence MJV (70 m/My, Fig. 14). This interval is marked by a tectonic uplift, synchronous over the entire western France platform (Fig. 13).

- Late Bajocian to late Bathonian (sequences MJVI to MJXI). In this time interval, decomposed sedimentation rates vary from 27 m/My in Poitiers to 46 m/My in Caen, and accommodation rates vary from 27 m/My in Poitiers to 34 m/My in Argentan (Fig. 13). The highest accommodation rates are recorded for sequence MJXI, with values of 83 m/My in Montbron, 130 m/My in Caen and Poitiers, and 240 m/My in Argentan. Continuous subsidence and tectonic subsidence are recorded in the Argentan, Poitiers, and Montbron areas.

- Early Callovian (sequence MJXII). The Caen and Argentan areas show high rates of decomposed sedimentation (respectively 76 m/My and 65 m/My) and accommodation (104 m/My, Figs. 13, 14), whereas Poitiers and Montbron are characterized by decomposed sedimentation and accommodation rates of about zero. High positive subsidence and tectonic subsidence rates are recorded for Caen and Argentan, whereas negative tectonic subsidence rates characterized the Poitiers and Montbron areas (Fig. 13).

- Late early Callovian to mid Oxfordian (sequences MJXIII to LJIV). Over this interval, high decomposed sedimentation rates (respectively 51 m/My and 58 m/My) and accommodation rates (41 m/My and 46 m/My) are recorded in Caen and Argentan. A negative accommodation rate (~17 m/My) coupled with a high negative tectonic subsidence rate mark an uplift in Caen during sequence LJIV, following a period of relative stability (Figs. 13, 14). Stage 5 is characterized by near-zero rates of decomposed sedimentation, accommodation, and subsidence (Figs. 13, 14). Subsidence and tectonic subsidence curves display negative rates in Poitiers and Montbron over the entire interval. Only sequence MJXIV displays a different pattern in Montbron, with high decomposed sedimentation (154 m/My) and accommodation rates (163 m/My).

- Late mid to late Oxfordian (sequences LJV to LJVII). This last phase is characterized in Caen by high rates of decomposed sedimentation (109 m/My) and accommodation (124 m/My), which correlate with major tectonic subsidence (Figs. 13, 14). In Argentan, stage 6 exhibits low decomposed sedimentation and accommodation rates of 27 m/My and 34 m/My, and steady tectonic subsidence. The Poitiers area recorded a very high accommodation rate over sequence LJV (192 m/My), contemporaneous with significant tectonic subsidence (Figs. 13, 14). During deposition of sequences LJVI and LJVII, a stable tectonic phase is recorded in Poitiers, where the decomposed sedimentation and accommodation rates are 34 m/My (Figs. 13, 14). Around Montbron, the decomposed sedimentation rate (67 m/My) and accommodation rate (67 m/My) might have been constant over the last stage, and the area was subsiding (Figs. 13, 14).

5. Discussion

5.1. Accommodation changes on the western France carbonate platform – global and regional trends

Platform architecture and sedimentary dynamics change with time from Deauville to Montbron over the Mid Jurassic and the Oxfordian, raising questions about the factors controlling this evolution. Sedimentary sequences, systems tracts, and platform architectures depend directly on the ratio between accommodation space (A) and sedimentation rate (S); (Jervey, 1988; Schlager, 1993, 2005; Catuneanu et al., 2009). The A/S ratio controls the progradational, aggradational, or retrogradational...
Fig. 13. Synthetic chronostratigraphic diagram showing western France sequences, eustasim curve (Hardenbol et al., 1998b), bathymetry, decompacted sediment thickness, accommodation, subsidence, and tectonic subsidence curves in Caen/Deauville, Argentan, Poitiers, and Montbron.
geometries of sediments. Retrograding structures and facies that are more distal upward mark the transgressive systems tract, when the A/S ratio is higher than 1. Prograding structures and upwardly more proximal facies characterize the regressive system tract, when the A/S is lower than 1. Accommodation space is generated by changes in (1) global eustasy and/or (2) subsidence, the latter being associated with (3) isostatic adjustment due to sediment deposition or paleodepth variation, and with (4) tectonics. At local scale, tectonic subsidence can be explained by syn-sedimentary active faults. Sedimentation rate relates to (1) terrigenous input or to (2) in-situ carbonate production and its redistribution, which is largely influenced by environmental conditions (e.g. trophic conditions and seawater temperature).

Overall eustasy variations in European basins have been modeled by Hardenbol et al. (1998b) and are presented in Figs. 13, 15. Changes in depth, decompacted sediment thickness (observed total subsidence, Steckler and Watts, 1978; Allen and Allen, 2005), accommodation (total subsidence +/− depth variation), subsidence, and tectonic subsidence (Steckler and Watts, 1978; Fig. 13) were reconstructed in four key localities to discuss the factors controlling sediment architectures (Caen, Argentan, Poitiers, and Montbron, Fig. 13). By reconstructing...

Fig. 14. Accommodation rate for each western France sequence in Caen/Deauville, Argentan, Poitiers, and Montbron for the Aalenian to Oxfordian interval.

Fig. 15. Synthetic chronostratigraphic diagram from the Aalenian to Oxfordian showing western France sequences, second- and third-order European cycles/sequences (Hardenbol et al., 1998b), δ13C isotopic curve (Martinez and Dera, 2015), and semi-quantitative estimation of decompacted carbonate production, carbonate producers, and depositional profile on the western France platform.
long-distance depositional sequences and dating basement flooding it was possible to reconstruct the ante-Aalenian topography and discuss its influence on platform architecture.

5.1.1. Eustasy as the major control on 3rd-order transgressive–regressive systems tracts

The overall correlation between European third-order sequences and the 22 depositional sequences along with their continuity over the entire western France platform suggests that eustasy is the main factor affecting accommodation variations at the scale of third-order sequences (Hardenbol et al., 1998a, Fig. 15). The maximum flooding surfaces identified were also found in other European basins of similar ages at the biozone scale (Hardenbol et al., 1998a). However, five third–order sequences described in other European basins were not identified in the study area: two within the Bathonian deposits (zigzag/prograculis Zones transition and discus subzone maximum flooding surfaces), two within the Callovian succession (koegini and lamberti Zones maximum flooding surfaces), and one within the late Oxfordian (bimmamatum Zone; Fig. 15; Hardenbol et al., 1998a).

The absence of several European basin sequences from western France can be explained by (1) non-deposition/preservation in proximal areas (hardgrounds and exposure surfaces), especially at the Bathonian/Callovian boundary (Figs. 5, 11), and by (2) condensation in deep clay-dominated environments, which is the case for Callovian and early Oxfordian sequences (koegini, lamberti and bimmamatum Zones; Fig. 11).

However, there are significant differences in second-order transgressive–regressive cycles between European basins and the western France platform, but also within the western France platform (Figs. 11, 15). Lateral dissimilarities in sedimentation rate and carbonate production, and the sharp local depth variations recorded over the western France platform cannot have been under global or eustatic control and must be related to local tectonics (Fig. 11).

5.1.2. Tectons contribution on sedimentation rate, carbonate platform architecture, growth, and demise

Several episodes of facies and platform architecture change are interpreted to be connected with tectonic activity over all or part of western France. The major tectonic uplift, associated with an accommodation rate of about zero, that occurred over the entire study area from the mid-Aalenian to the early Bajocian (sequences MJIII to MJV, Fig. 13) corresponds to the Mid-Cimmerian Unconformity, a subtle deformation event identified all around the Paris Basin (Guillocheau et al., 2000; De Graciansky and Jacquin, 2003) and in Western Europe related to North Sea doming (Underhill and Partington, 1993). Eustatic rise between sequences MJIII and MJV was not sufficient to offset the loss of accommodation due to the uplift (Haq et al., 1987). The consequences were the very low sedimentation rates (0 to 15 m/My) and overall shallowing (Figs. 11, 13). Around Montbron, lagoon environments expanded northwards during the humphriesianum Zone, marking the regressive maxima of a second-order cycle, and oolitic wedges formed in Poitiers over sequences MJIV and MJV, (Figs. 11, 15). In Caen, uplift was active until the end of sequence MJVII (negative accommodation rate of −17 m/My, Fig. 14) and may have induced the shallowing trend marked by upper offshore sponge facies prograding northwards.

During the late Bajocian and the Bathonian, about 150 m of accommodation space was created, despite a global sea-level fall (Figs. 13, 14; Hardenbol et al., 1998b). This indicates that the western France platform underwent a major phase of tectonic subsidence (estimated at about 100 m considering a sea-level fall of 45 m; Fig. 13; Hardenbol et al., 1998b). The accommodation generated by tectonics, coupled with the subsidence related to sediment loading, made it possible for 100 to 150 m of decompacted carbonates to accumulate (Fig. 13). Over sequences MJIX to MJXI, the Caen and Montbron shallow platforms prograded northwards, indicating that carbonate production was sufficient to fill the accommodation space generated by tectonic subsidence (Fig. 11). Significant variations in decompacted sedimentation/accommodation rate were observed within sequence MJXI between Falaise (decompacted sedimentation/accommodation rate about 36 m/My) and Argentan (about 240 m/My; Gradstein et al., 2012, Figs. 11, 14). This implies a local tectonic subsidence of about 40 m/My in the Argentan area, the remaining subsidence being controlled by sediment loading. Argentan is bordered by three faults affecting the Mesozoic series, which may have controlled this tectonic subsidence for the latest Bathonian: the Montabard–Gouffern fault to the north (Ménillet et al., 1997; Gigot et al., 1999), the Pommereux–Sentilly fault to the west, and the Sées fault to the southwest (Kuntz et al., 1989; Fig. 1). The Montabard–Gouffern fault displays a high subvertical dip and a vertical offset of between 30 and 40 m. At its northwestern extremity, Bajocian to Bathonian limestones were brought into contact with Triassic to Ordovician deposits, indicating Bathonian to post-Bathonian activity (Gigot et al., 1999).

The early Callovian is marked by a sudden carbonate demise that is commonly explained by a eustatic rise and a consequent extensive drowning event within the Paris Basin (Jacquin et al., 1992, 1998; Jacquin and de Graciansky, 1998; Brigaud et al., 2014). The sudden disappearance of carbonates in favor of clay sedimentation occurred from Deauville to Le Mans during deposition of the early transgressive systems tract of sequence MJXII (Figs. 11, 12G, 13). The progressive rise in the overall eustatic level, estimated at about 15 m, was not sufficient to explain sudden shallowing of at least 40 m (Hardenbol et al., 1998b; Figs. 11, 13). From Deauville to Le Mans, the steep deepening resulting in carbonate demise for the earliest Callovian was triggered by prominent tectonic subsidence of an amplitude between at least 35 m in Argentan and 55 m in Caen (Fig. 13). The increased accommodation was then maintained through the Callovian and into the early Oxfordian due to a eustatic rise (Hardenbol et al., 1998b). Several reactivated basement faults affecting the Mesozoic series from Falaise to Alençon may have caused this tectonic subsidence: the Cordéy-Ronai, Montabard-Gouffern, Merlerault, Courtomier, and Fresnay-sur-Sarthe faults (Fig. 1). The Merlerault and Courtomer faults display a vertical offset varying between 50 m and 130 m in the Mesozoic succession (Ménillet et al., 1997, 1998). Northwards, the Cordéy-Ronai and Montabard-Gouffern faults affect the Middle Jurassic series with a vertical offset of about 40 m (Ménillet et al., 1998).

In Montbron, the accommodation rate was zero for sequences MJXII and MJXIII over a period of eustatic rise (about +20 m, Figs. 13, 14, Hardenbol et al., 1998b). This implies an uplift, the amplitude of which is estimated at about 40 m, preceding emersion, which occurred during a second-order transgression in European basins, and accounting for the absence of sequences MJXII and MJXIII at Montbron (Fig. 11; Hardenbol et al., 1998a). This local tectonics implies that the sequence boundary between the two complete second-order cycles in western France corresponds to Bt4 in Montbron whereas it corresponds to Bt2 in Caen and Argentan (Figs. 11, 15). In Poitiers and Montbron, the uplift phase remained constant until the end of sequence LJIV (Middle Oxfordian, Fig. 13). The postulated lack of deposition of sequences MJXV to LJIV in Montbron was certainly due to emersion due to uplift. Over sequences LJII to LJIV, uplift in the Poitiers area controlled sedimentation despite global eustatic rise, leading to the non-deposition of sequences LJII and LJIII and to the shallowing-upward trend marked by the deposition of cross-bedded bioclastic grainstones during sequence MJXIV (Figs. 11, 13). From Poitiers to Montbron, the maximum flooding surface of the Callovian–Oxfordian cycle cannot be positioned in the mariae Zone as for the Caen and Argentan areas or for European second-order cycles, but is located in the termiseraturn Zone (Figs. 11, 15).

In Argentan, the accommodation rate was zero for sequences LJII and LJIII over a period of eustatic rise (about +15 m, Figs. 13, 14, Hardenbol et al., 1998b). This implies an uplift of about 30 m, leading to the progradation of shelf-edge sands topped by an exposure surface (Roussier de Gacé et Sables ferrugineux de Vairais Formations, Ox1, Figs. 11, 13). In Caen, drastic shallowing from lower offshore to
shoreface between sequences LJI and LJV during a period of eustatic rise (about + 10 m, Hardenbol et al., 1998b) implies a major uplift, of about 40 m (Fig. 13). This early/mid Oxfordian uplift made possible the growth of a shallow carbonate platform prograding northwards during sequence LJV. This carbonate platform persisted until the end of the Oxfordian from Caen to Le Mans. However, the sharp carbonate demise resulting in the deposition of lower offshore clays and in retrograding environments southwards cannot have been the consequence of the eustatic rise alone (about 12 m, Hardenbol et al., 1998b) and must also have been due to intense tectonic subsidence occurring in Deauville over sequences LJV and LVI (Figs. 11, 13). Deauville is separated from Caen by the Villier- Reux fault which displays a vertical offset of about 55 m and brought Jurassic and Cretaceous series into contact (Fig. 1, Pareyn, 1970). This fault may have been reactivated during the Late Jurassic, causing the major tectonic subsidence recorded in Deauville. The deepening from shoreface to lower offshore around Poitiers during sequence LJV, in a context of relative eustatic stability (+3 m Hardenbol et al., 1998b), and leading to carbonate demise, was the consequence of sudden and intense tectonic subsidence. This event may have been due to the late Oxfordian reactivation of the Lusignan fault: a 100 km-long basement structure oriented NW–SE and affecting Jurassic series to the north of Poitiers (Cariou and Joubert, 1989b). In Montbron, the increased accommodation due to gradual and slow tectonic subsidence coupled with a global eustatic rise over sequences LJV to LVI (Hardenbol et al., 1998b, Fig. 13) accelerated sedimentation and led to the deposition of about 180 m of decompacted carbonates (Figs. 11, 13).

5.2. Basement topography as a major control of paleodepth on the western France carbonate platform

The Jurassic platform of western France developed over a pre-structured bottom surface. This includes basement rocks that are mostly Cambrian and Ordovician quartz sandstones, displaying major paleotopographic changes that can be directly observed in the field (Juignet et al., 1989; Ménillet et al., 1994; Cariou et al., 2006). Post-Variscan relief of the basement has been partly lowered by the deposition of Liassic series, for example in the Montbronn area that is filled by about 250 m of Triassic to Liassic deposits (Roger et al., 1979). The topography of the basement was reconstructed for the time interval under study from borehole data with paleobathymetric interpretations (Fig. 12).

Between Argentan and Alençon, sequences MJII to MJVIII are not found overlying the basement, indicating an emerged area from the early Aalenian to the mid Bathonian (Figs. 11, 12A to D). These areas were flooded over sequences MJI to MJX, during which shallow environments were essentially located on basement highs: from Argenton to Alençon and in Poitiers (Fig. 12A to D). From sequence MJII to MJX, deep platform environments (lower offshore) were located in basement depressions in Caen and from Le Mans to Saumur (Fig. 12A to D).

Over the Aalenian to early Bathonian, the basement topography was the major factor determining lateral variations of paleodepth, as shallow environments were located on basement highs and deeper environments in basement depressions (Fig. 12A to E). From the late Bathonian to late Oxfordian, the basement depressions were filled by sediments and flattened, and the influence of the basement topography on bathymetries became negligible, except from Le Mans to Saumur, where lower offshore environments persisted (Fig. 12F to I). Local tectonic subsidence and uplifts changed basement topography through time as described in Section 4.1.2. However, tectonic movements were rather homogenous across western France through the Aalenian to early Bathonian and the basement topography remained approximately the same (Figs. 12A to E, 13). From the late Bathonian to late Oxfordian, local tectonic movements greatly modified the basement topography (Figs. 12F to I, 13).
Table 2
Literature used to reconstruct the evolution of sedimentation during the Middle Jurassic and the Oxfordian in western Tethyan platforms. For each reference, different characteristics of the studied sedimentary rocks are provided: country, precise location, age, depositional environment, and lithology/carbonate producers.

<table>
<thead>
<tr>
<th>Country</th>
<th>Location</th>
<th>Age</th>
<th>Depositional environment</th>
<th>Lithology/carbonate producers</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>France</td>
<td>Eastern Paris Basin</td>
<td>Aalenian to Oxfordian</td>
<td>Lower offshore to shoreface</td>
<td>Ooid/coral limestones (photozoan), crinoid/coral limestones (transitional) and marls</td>
<td>Brigaud et al. (2014)</td>
</tr>
<tr>
<td>France</td>
<td>Southeastern France</td>
<td>Aalenian to early Bajocian</td>
<td>Shallow marine</td>
<td>Crinoid limestones (heterozoan)</td>
<td>Roussele and Dromart (1996)</td>
</tr>
<tr>
<td>France</td>
<td>Quercy (southern France)</td>
<td>Aalenian to Oxfordian</td>
<td>Shallow platform</td>
<td>Bioclastic (mixed grain association, Aalenian and Callovian) to ooid (photozoan, Bajocian, Bathonian, Oxfordian)</td>
<td>Cubaynes et al. (1989)</td>
</tr>
<tr>
<td>Germany</td>
<td>Oberrhein (southwestern Germany)</td>
<td>Early Bajocian</td>
<td>Offshore</td>
<td>Marl-limestone alternations</td>
<td>Ohmert (1994)</td>
</tr>
<tr>
<td>Germany</td>
<td>Southwestern Germany</td>
<td>Aalenian to Oxfordian</td>
<td>Shallow marine</td>
<td>Marl-limestone alternations</td>
<td>Maxwell et al. (2012)</td>
</tr>
<tr>
<td>Germany</td>
<td>Süntel Mountains (northern Germany)</td>
<td>Callovian-Oxfordian</td>
<td>Lower offshore (Callovian and early Oxfordian) to shallow marine (middle–late Oxfordian)</td>
<td>Marl to sandstone (Callovian and early Oxfordian) and oolitic limestone (photozoan), middle–late Oxfordian</td>
<td>Helm and Schülke (1998)</td>
</tr>
<tr>
<td>Germany</td>
<td>North German Basin</td>
<td>Middle Jurassic (Aalenian-Callovian)</td>
<td>Offshore to shallow-marine (Wave dominated)</td>
<td>Claystone to sandstone</td>
<td>Zimmermann et al. (2015)</td>
</tr>
<tr>
<td>Italy</td>
<td>Umbria-Marche Basin, Central Italy</td>
<td>Aalenian to Oxfordian</td>
<td>Offshore/shallow marine</td>
<td>Marl-limestone alternations</td>
<td>Bartolini et al. (1996)</td>
</tr>
<tr>
<td>Italy</td>
<td>Umbria-Marche Basin, Central Italy</td>
<td>Aalenian to Oxfordian</td>
<td>Offshore</td>
<td>Marl-limestone alternations/Heterozoan limestones</td>
<td>Moretti et al. (2002)</td>
</tr>
<tr>
<td>Italy</td>
<td>Latium-Abruzzi Platform, Central Italy</td>
<td>Aalenian to Oxfordian</td>
<td>Shallow marine</td>
<td>Heterozoan (Aalenian-early Oxfordian) to photozoan (middle–late Oxfordian)</td>
<td>Bartolini et al. (1996)</td>
</tr>
<tr>
<td>Italy</td>
<td>Apulia platform, southern Italy</td>
<td>Callovian to Oxfordian</td>
<td>Oolitic shoals and lagoon</td>
<td>Peloid mudstones to packstones/oolitic grainstones (photozoan)</td>
<td>Bosellini et al. (1999)</td>
</tr>
<tr>
<td>Italy</td>
<td>Apennines platform, southern Italy, Sorrento Peninsula</td>
<td>Aalenian to Oxfordian</td>
<td>Very shallow</td>
<td>Ooid limestone with Chlorophyte</td>
<td>Iannace et al. (2011)</td>
</tr>
<tr>
<td>Morocco</td>
<td>High Atlas</td>
<td>Aalenian</td>
<td>Offshore</td>
<td>Marl-limestone alternations/Heterozoan limestones</td>
<td>Prêt et al. (2006)</td>
</tr>
<tr>
<td>Morocco</td>
<td>High Atlas</td>
<td>Aalenian to Bathonian</td>
<td>Inner to mid-ramp</td>
<td>Marl-limestone alternations/Oolitic limestones (photozoan)</td>
<td>Pierre et al. (2010)</td>
</tr>
<tr>
<td>Morocco</td>
<td>High Atlas</td>
<td>Late Bajocian</td>
<td>Inner ramp (Aalenian, late Bajocian, Bathonian) to lower offshore (early Bajocian)</td>
<td>Oolitic/bioclastic limestones (photozoan, Aalenian, late Bajocian), marls (early Bajocian) and sandstones (Bathonian)</td>
<td>Tomas et al. (2013); Alt Addi and Chaffik (2013)</td>
</tr>
<tr>
<td>Poland</td>
<td>Pieniny Kippen Basin, Carpathians (southern Poland)</td>
<td>Aalenian to early Bajocian</td>
<td>Lower offshore</td>
<td>Organic-rich claystones</td>
<td>Christ et al. (2012)</td>
</tr>
<tr>
<td>Poland</td>
<td>Polish Jura (south-central Poland)</td>
<td>Late Bajocian</td>
<td>Lower offshore</td>
<td>Clay</td>
<td>Tyszka (1994); Tyszka and Kaminski (1995)</td>
</tr>
<tr>
<td>Poland</td>
<td>Cracow Area, southern Poland</td>
<td>Bajocian-Bathonian</td>
<td>Lower offshore (Callovian and early Oxfordian) to shallow marine (middle–late Oxfordian)</td>
<td>Marl to sandstone (Callovian and early Oxfordian) and oolitic-rich limestone (transitional, middle–late Oxfordian)</td>
<td>Matyszewsowicz and Felisiak (1992)</td>
</tr>
<tr>
<td>Poland</td>
<td>Tatra Mountains, southern Poland</td>
<td>Aalenian to Callovian</td>
<td>Offshore to shallow-marine</td>
<td>Crinoidal limestones (heterozoan), ferruginous condensed, Sponge limestone (transitional)</td>
<td>Luczynski (2002)</td>
</tr>
<tr>
<td>Poland</td>
<td>Polish Jura chain (southwestern Poland)</td>
<td>Late Oxfordian</td>
<td>Shallow marine</td>
<td>Marl-limestone alternations, condensed ferruginous, heterozoan/photozoan limestones, sandstones</td>
<td>Weizbrowski (2015); Hesselbo (2008)</td>
</tr>
<tr>
<td>Spain</td>
<td>Asturias, Basque-Cantabrian, Pyrenean and Iberian basins; northern Spain</td>
<td>Callovian-Oxfordian</td>
<td>Offshore to shallow marine</td>
<td>Marl-limestone alternations</td>
<td>Ait et al. (2003)</td>
</tr>
<tr>
<td>Spain</td>
<td>Betic Cordillera, southern Spain</td>
<td>Middle to late Oxfordian</td>
<td>Lower offshore</td>
<td>Marl-limestone alternations</td>
<td>Pittet et al. (2000)</td>
</tr>
<tr>
<td>Switzerland</td>
<td>Swiss Jura</td>
<td>Late Bajocian</td>
<td>Shallow water ramp (shoreface)</td>
<td>Ooid/coral grainstones to boundstones (photozoan)</td>
<td>Pittet and Strasser (1997)</td>
</tr>
<tr>
<td>Switzerland</td>
<td>Swiss Jura</td>
<td>Late Bajocian</td>
<td>Shallow water ramp (shoreface)</td>
<td>Oolitic limestone (photozoan)</td>
<td>Gonzalez and Wetzler (1996)</td>
</tr>
<tr>
<td>Switzerland</td>
<td>Swiss Jura</td>
<td>Late Bajocian</td>
<td>Shallow water ramp (shoreface)</td>
<td>Oolitic limestone (photozoan)</td>
<td>Wetzel et al. (2013)</td>
</tr>
</tbody>
</table>
climate disturbed by short episodes of intensive rainfalls and storms. Although these monsoonal episodes intensified the net transfer of nutrient, organic carbon and carbonates ions to the ocean, organic carbon burial was prevented because of the prevailing dry climate and the efficient oxidizing conditions in seawater (Fig. 17B; Martinez and Dera, 2015). This lead to (1) an enrichment of seawater in $^{12}$C and a
consequent decrease of $\delta^{13}$C values in the oceanic reservoir, and to (2) carbonate supersaturation in epicontinental seas due to high evaporation (Martinez and Dera, 2015). This model is consistent with the high carbonate production rate recorded in western Tethyan epicontinental seas during the Bajocian/Bathonian and middle/late Oxfordian intervals. The ratio between photozoan and heterozoan associations is mainly controlled by seawater temperature, trophic conditions, and bathymetry (Fig. 17A; James et al., 1997, Mutti and Hallock, 2003).

Photozoan association development is favored by high seawater temperatures (subtropical to tropical, >18 °C) and low trophic resources (oligotrophic to mesotrophic), but can also be found in temperate waters under oligotrophic conditions (Fig. 17A; Mutti and Hallock, 2003). Heterozoan grain association dominates in warm waters (>18 °C) under eutrophic conditions or in cool waters (<18 °C) whatever the trophic conditions. In western Tethyan platforms, during the Bajocian/Bathonian and middle/late Oxfordian, photozoan producers were...
distributed in shoreface environments, indicating prevailing oligotrophic conditions, whereas heterozoan producers are mainly present in offshore environments (Figs. 16, 17B). Nevertheless, heterozoan producers can also be found locally in shallow wave to tide-dominated environments, as in northwestern France during the Bathonian (Fig. 16). This could be induced by local changes in environmental conditions as an increase of trophic inputs or a decrease of seawater temperatures (e.g. Lécuyer et al., 2003).

5.4.3. Carbonate production during low-eccentricity intervals

During the Jurassic, low eccentricity intervals induced wet conditions, which promoted high weathering rates, nutrient inputs, productivity levels, and organic burial in the ocean, and led to an increase of δ13C values (Martinez and Dera, 2015). This happened during the Aalenian and from the middle Callovian to early Oxfordian (Fig. 17B; Martinez and Dera, 2015). The high kaolinite ratio in the Paris Basin (>20% of the clay mineralogical association; Pellenaard and Deconinck, 2006; Brigaud et al., 2009), and vascular plant biomarkers used as a proxy for palaeoflora (Hautevelle et al., 2006) support enhanced weathering conditions under a humid climate during the middle Callovian to early Oxfordian. Moreover, organic-rich layers are found in Central Atlantic (Dromart et al., 2003) and in the western Tethyan domain (England, Kenig et al., 1994; Saudi Arabia, Carrigan et al., 1995) during the middle Callovian. The Aalenian clay mineralogical association is also dominated by kaolinite in the southern Paris Basin (Delavanne et al., 1989) and in England (Sellwood and Sladen, 1981). The presence of anoxic conditions at the sea bottom is evidenced in southern Poland by the presence of agglutinated foraminifera, but the association with organic-rich deposits is not obvious (Tyszka, 1994; Tyszka and Kaminski, 1995). It is likely that organic carbon was either (1) locked up but disseminated in shelf mudstones and therefore, is not readily identified, (2) stored in deep marine areas or (3) stored in terrestrial settings (Price, 2010). Clay contents and palaeoflora data, as well as evidences of prevailing anoxic conditions at the sea-floor during the Aalenian and from the middle Callovian to early Oxfordian are consistent with the hypothesis of a wet climate during low-eccentricity intervals leading to high nutrient inputs. Such an eutrophication of sea waters is noxious for photozoan carbonate producers and tends to diminish the growth potential of carbonate platforms (Weissert and Mohr, 1996; Bartolini et al., 1996; Bartolini and Cecca, 1999; Mutti and Hallock, 2003), promoting siliclastic sedimentation (Figs. 15, 16, 17B). On the western France platform, decompacted carbonate accumulation rate was over 60 m/My during the Bathonian and middle/late Oxfordian but fell down under 20 m/My at the Aalenian and during the middle Callovian/early Oxfordian interval. The mean biological carbonate content in marls varies from 35% to 20% in the Callovian to early Oxfordian, the lowest values being recorded in the marias Zone (Dugué, 1989), which correlates with the δ13C maxima (Martinez and Dera, 2015). Proximal deposits are sandstone or heterozoan to transitional grain associations, whose development is favored by eutrophic conditions (Figs. 16, 17B). Photozoan facies can develop locally, probably due to local changes in trophic conditions or seawater temperatures (e.g. London platform during the Aalenian). Low sedimentation rates (mainly between 0 and 10 m/My), the abundance of condensed ferruginous deposits and gaps during the Aalenian are the consequence of a decrease of accommodation space related to a global uplift in Western Europe. This event corresponds to the Mid-Cimmerian unconformity (Figs. 16, 17B; Underhill and Partington, 1993; Guilloucheau et al., 2000; De Graciánski and Jacquin, 2003). High clayey sedimentation rates during the middle Callovian to early Oxfordian, generally between 20 and 50 m/My, was allowed by an increase in accommodation space both due to (1) the global eustatic sea-level rise (Hardenbol et al., 1998b) and (2) increase of subsidence rate (Fig. 17B). In western France, the demise of the carbonate platform initiated at the Bathonian/Callovian boundary because of a sharp, local tectonic subsidence, but it persisted in deeper and shallower environments until the middle Oxfordian due to (1) the eutrophic conditions of neritic environments and to (2) the global eustatic sea level rise.

6. Conclusions

Thirty-one facies were characterized in Aalenian to Oxfordian formations of western France, deposited in lower offshore to backshore settings. Changes in platform geometry and facies were identified over a time interval of ~17 Ma (Aalenian to Oxfordian) and across a long distance (500 km). A high-resolution correlation scheme was realized at the ammonite biozone scale in a sequence stratigraphy framework. Twenty-two third-order depositional sequences have been defined. They are bounded by maximum regressive surfaces that are either marine (e.g. perforated hardgrounds encrusted with bivalves) or immersive (microstalactitic, meniscus, and dogtooth cements, lignite, paleosols). Depositional sequences of the western France platform correlate at the ammonite biozone scale with the third-order sequences of European basins, indicating that eustasy is the major factor controlling third-order transgressive–regressive systems tracts (Hardenbol et al., 1998a). Topography of the Paleozoic basement controlled lateral depth variations until the mid Bathonian, resulting in the development of (1) shoreface environments from Argenton to Le Mans and in Poitiers and to (2) lower offshore environments in Caen and from La Flèche to Mirebeau. Evolution of tectonic subsidence was reconstructed in four key areas of western France: Caen, Argenton, Poitiers, and Montbron. A major uplift was identified to have occurred over the middle/late Aalenian and the early Bajocian. It corresponds to the Mid-Cimmerian Unconformity. Throughout the Middle Jurassic and Oxfordian, tectonic controlled sedimentation rate and platform architecture; it also triggered two major phases of carbonate growth and demise. Uplifts favored low sedimentation rates (i.e. late Aalenian to early Bajocian), the occurrence of exposure surfaces and prograding systems, as observed around Montbron during the early to mid Callovian. Tectonic subsidence promoted high sedimentation rates, for example during the Bathonian over the entire western France platform (decompacted sedimentation rate about 70 m/My) or during the mid/late Oxfordian around Montbron (decompacted sedimentation rate about 70 m/My). The major carbonate production demise at the Bathonian/Callovian boundary was triggered by sharp tectonic subsidence of about 35 m to 55 m, causing an increase in paleodepths. In contrast, the development of the prograding mid-Oxfordian carbonate platform in north-western France was made possible by the tectonic uplift that generated shallow-water and favorable conditions for carbonate growth. A synthesis of sedimentation on western Tethyan platforms was conducted. Two periods of high carbonate production during the Bajocian/Bathonian and middle/late Oxfordian, with prevailing photozoan producers in shallow-marine environments, are synchronous with high eccentricity intervals marked by low δ13C values (Martinez and Dera, 2015). Carbonate production was promoted by dry climates disturbed by short episodes of intensive rainfalls and storms leading to (1) high evaporation and carbonate supersaturation and (2) low trophic conditions. Both periods of low carbonate production during the Aalenian and from the middle Callovian to early Oxfordian are synchronous with low eccentricity intervals and high δ13C values. This was marked by a wet climate and the eutrophication of epicontinental seas that tend to diminish growth potential of western Tethyan carbonate platforms. The uplift of the Aalenian (corresponding to the Mid-Cimmerian unconformity) lead to the formation of sedimentation hiatus and condensed levels, whereas the global eustatic sea-level rise of the Callovian–early Oxfordian generated an important accommodation space and favored high sedimentation rates.

The present work shows that the development of carbonate platforms in intracontinental settings was influenced by multiple factors, each one exerting a precise control on sedimentation. In particular, we highlighted the control of basement topography on lateral depth variations and the influence of local and regional tectonics on carbonate
platform architecture, even in intracratonic basins. Finally, climate evolution in western Tethys was controlled by long-term eccentricity variations and can be related to the major stages of carbonate platform growth and demise and largely influenced the producer types (photozoan versus heterozoan).

Acknowledgments

This work is the result of collaborative project no PO4990 between Paris-Sud University and the BRGM. This project was funded by the BRGM and by a doctoral PhD from the French Ministry of Research and Higher Education (2013–134). We are grateful to Philippe Blanc (Lithologie Bourgogne) for the high-quality thin-sections and to Christophe Dutel and Philippe Courtivolle for the stratigraphy of Aalenian deposits of Orne.

The authors thank the anonymous reviewer and associate editor Brian Jones for their constructive comments.

References


