A new sedimentary and biostratigraphic framework for the Callovian-Oxfordian transition on the Atlantic margin of Morocco

Duval-Arnould, Aude; Bulot, Luc; Masrour, Moussa; Simmons, Mike; Bonnot, Alain; Charton, Rémi; Redfern, Jonathan; Schröder, Stefan

Citation (APA)
Duval-Arnould, A., Bulot, L., Masrour, M., Simmons, M., Bonnot, A., Charton, R., Redfern, J., & Schröder, S. (2024). A new sedimentary and biostratigraphic framework for the Callovian-Oxfordian transition on the Atlantic margin of Morocco. *Journal of African Earth Sciences*, *210*, Article 105164. https://doi.org/10.1016/j.jafrearsci.2023.105164
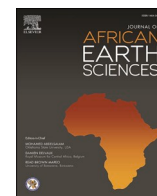
Important note
To cite this publication, please use the final published version (if applicable).
Please check the document version above.

A new sedimentary and biostratigraphic framework for the Callovian-Oxfordian transition on the Atlantic margin of Morocco

Aude Duval-Arnould [a], Luc Bulot [a], Moussa Masrour [b], Mike Simmons [c], Alain Bonnot [d], Rémi Charton [e], Jonathan Redfern [a], Stefan Schröder [a,*]

[a] Basin Studies, Department of Earth and Environmental Sciences, University of Manchester, UK
[b] Laboratoire de Géologie Appliquée et Géo-Environnement, Université Ibn Zohr, Agadir, Morocco
[c] Neftex-Halliburton, Abingdon, UK
[d] Biogéosciences, UMR 6282 CNRS, Université de Bourgogne, Dijon, France
[e] Department of Geoscience and Engineering, Applied Geology Section, TU Delft, Netherlands

ARTICLE INFO

ABSTRACT

A major global marine transgression occurred during the Callovian to Early Kimmeridgian, which was interrupted locally by a hiatus during the Late Callovian to Early Oxfordian. The transgression may have been a major driver for extensive coral buildup development in the Oxfordian. The depositional hiatus may be related to a combination of eustasy, local tectonic activity and hinterland movements, highlighting the potential influence of tectonism on sedimentary evolution in sedimentary basins. Whilst a regional Late Callovian-Early Oxfordian hiatus has been recorded extensively in the Tethys realm, this study has improved the biostratigraphic and sedimentary record from Morocco, and for the first time documents the Late Callovian-Early Oxfordian depositional hiatus in the Atlantic realm. Detailed sedimentary facies analysis of Callovian to Oxfordian carbonates and siliciclastics (Ouanamane Formation) demonstrates a major transgression from continental and coastal sediments, through oolitic mid-inner ramp deposits, brachiopod-rich middle ramp deposits, and ultimately to outer ramp marls, sharply overlain by Middle Oxfordian coral buildups across the basin. Repeated hard- and firmgrounds and bioturbated bed tops highlight transgressive surfaces traceable basinwide, which suggests low sedimentation rates in the upper Ouanamane Formation. Localized siliciclastic input is evidence for episodes of erosion in the same interval. Collection of new fossil specimens (ammonites, brachiopods, echinoderms, foraminifera) and revision of existing fossil material suggest this interval corresponds to a hiatus or condensed section in the Callovian-Oxfordian transition. The uppermost part of the Ouanamane Formation is of early Middle Oxfordian age and is directly overlain by Middle Oxfordian coral buildups. Onset of buildup construction is considered synchronous based on the new biostratigraphic data. Observations in Moroco compare with the sedimentary evolution around the Tethys, and in particular with the Arabian Plate. The Callovian-Oxfordian depositional hiatus corresponds to eustatic sea level changes, which were possibly driven by global cooling, and was likely overprinted by local tectonics.

1. Introduction

The Callovian to Lower Kimmeridgian succession records a global transgressive interval (Haq, 2018), which is well recognized along the NW African Atlantic margin (Ambroggi, 1963; Adams, 1979; Davison, 2005). In the northern Neo-Tethys (Ruban, 2012), the Lusitanian Basin (Azerêdo et al., 2002), and the Adriatic platform (Husinec et al., 2022), this transgression was punctuated during the Late Callovian to late Early Oxfordian by a prominent depositional hiatus, followed by a subsequent

Oxfordian transgression. A similar depositional pattern is seen on the Arabian Plate (Al-Mojel et al., 2020; Al-Mojel and Razin, 2022). In Morocco, source-to-sink and low-temperature geochronology evidence points to important Callovian tectonic movements in the hinterland and an associated sedimentary provenance shift from the Anti-Atlas to the Meseta (Charton et al., 2021), which could relate to such a depositional hiatus. At a global scale, relative sea level fall recorded by the hiatus might reflect a postulated minor glacial period (Dromart et al., 2003a; Cecca et al., 2005). Identification of this hiatus in Morocco would test

whether tectonics influenced sedimentation on what is generally considered a tectonically quiet passive margin, and whether the hiatus extended beyond the Tethys into the Atlantic realm.

The peak transgression is associated with a prominent interval of coral buildups resting directly above the Callovian-Oxfordian interval studied here (Weissenberger et al., 2006; Martin-Garin et al., 2007; Olivier et al., 2012; Duval-Arnould et al., this volume). Buildups form subsurface hydrocarbon reservoir targets on the conjugate margins of Morocco and Nova Scotia. In Canada, the top-Callovian maximum flooding surface is one of the main seismically mappable horizons (Weston et al., 2012), and thus an updated chronostratigraphic framework would allow better prediction of this important interval.

The above geological events require precise dating, for which an updated biostratigraphy is essential. The biostratigraphy of the Callovian-Oxfordian interval in Morocco has been essentially based on brachiopods and microfossils (Roch, 1930; Ager, 1974; Adams, 1979; Adams et al., 1980; Bouaouda et al., 2004, 2009), which give only wide age ranges. Ammonites, which are the standard for chronostratigraphy in the Jurassic because of their rapid evolution allowing identification down to the biohorizon, have been reported (Gentil and Lemoine, 1906; Roch, 1930; Ambroggi, 1963), but are rarely used due to poor general preservation and scarcity of specimens (for an exception see Olivier et al., 2012). In addition, this fauna needs revision and calibration with more recent biostratigraphic scales (Cariou and Hantzpergue, 1997). There is therefore a need for biostratigraphy refinement in Morocco, using ammonites where possible.

The Atlantic margin of Morocco offers the best outcrops to improve the age constraints and to identify vertical and lateral facies variations of the Callovian-Oxfordian interval. This study has collected ammonites, brachiopods, echinoderms and foraminifera within a detailed sedimentary context from the Essaouira-Agadir Basin (EAB) in Morocco and has revised the identification of existing specimens. The work presented here aims to refine the biostratigraphic framework of the Callovian-Oxfordian succession in Morocco for better age constraints, enabling the correlation of biostratigraphic intervals across the conjugate margins. In combination with a detailed reconstruction of depositional processes and environments, this allows an assessment whether the Callovian-Oxfordian hiatus is widespread beyond the Tethys realm, and to discuss the age and synchroneity of the onset of coral buildup growth at the scale of the basin.

2. Geological setting

2.1. Structural context

The EAB is the southwestern extension of the Western High Atlas, located at the intersection of the Mesozoic Atlas and the Central Atlantic rifts (Fig. 1). The N–S to NNE-SSW trending syn-rift structures of the Atlantic margin are linked to an E-W transfer fault system inherited from older Hercynian faults (Medina, 1995; Bouatmani et al., 2003; Laville et al., 2004; Hafid et al., 2006). In the EAB, most of the Jurassic formations outcrop along E-W and NE-SW trending anticlines (Fig. 1). These anticlines are linked to salt tectonics triggered by transtentional and transpressional movement in the Essaouira segment, and fault reactivation during the Tertiary inversion in the Agadir segment (Piqué et al., 1998; Fernández-Blanco et al., 2020). The anticlines are deeply eroded and expose the entire Jurassic stratigraphy.

2.2. Upper Triassic and Lower-Middle Jurassic stratigraphy

Central Atlantic rifting started in Permo-Triassic times, forming extensional half-grabens filled by continental deposits (Mader and Redfern, 2011; Leleu et al., 2016). Extensive salt deposition occurred in many of the syn-rift grabens, with the maximum extension of salt deposition corresponding to the latest syn-rift stage (Tari and Jabour, 2013). The Triassic-Jurassic boundary is marked by the last phase of rifting, associated with the Central Atlantic Magmatic Province (CAMP) basaltic event (Marzoli et al., 1999). Final break-up of the Central
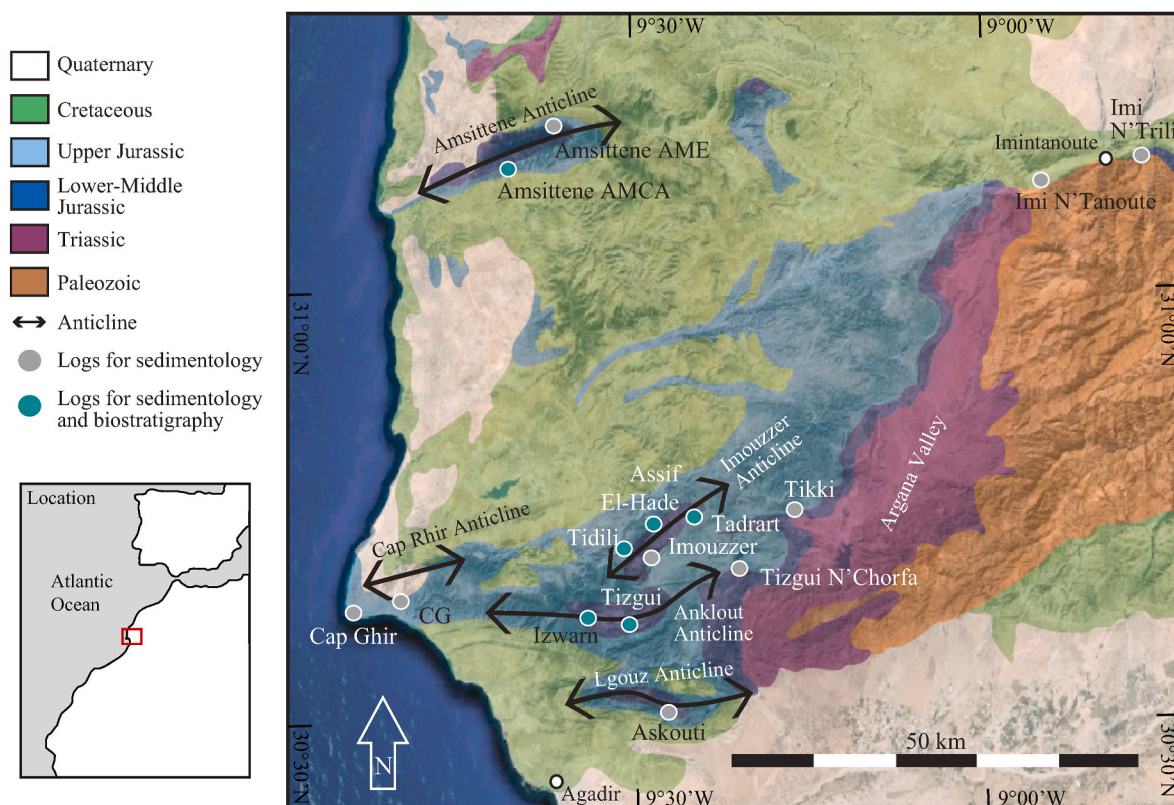

Fig. 1. Locality map of the study area and sections logged. Biostratigraphically relevant sections are indicated in blue.

Atlantic and the start of the post-rift stage occurred in the Early Jurassic (Hafid, 2000; Le Roy and Piqué, 2001; Laville et al., 2004; Tari and Jabour, 2013).

The Lower Jurassic stratigraphy in the EAB is dominated by marine carbonates (Roch, 1930; Ambroggi, 1963; Adams et al., 1980; Peybernès et al., 1987; Du Dresnay, 1988; Bouaouda, 2004; Duval-Arnould et al., 2021). During most of the Middle Jurassic (Aalenian-Bathonian), the onshore Agadir sub-basin was dominated by siliciclastic shallow-marine to fluvial deposits (Ameskhoud Fm.; Fig. 2A and 3), whereas the northern Agadir sub-basin (Jbel Amsittène), the Essaouira sub-basin, and offshore areas saw deposition of shallow-marine dolostones and limestones (Ambroggi, 1963; Adams et al., 1980; Peybernès et al., 1987; Du Dresnay, 1988; Bouaouda, 2004; Duval-Arnould et al., 2021).

2.3. The Callovian-Oxfordian transition

Transgression during the Callovian to Oxfordian re-established carbonate platforms along the conjugate Central Atlantic margins (Davison, 2005). This transgression corresponds to the Middle to Late Jurassic 2nd order global transgression (Jacquin et al., 1998; Haq, 2018). In the western High Atlas of Morocco, these carbonates belong to the Ouanamane, Lalla Oujja and Iggui El Behar formations (Fig. 2A and 3). According to Adams (1979, 1980), the transgression started in the Lower Callovian with the deposition of oolite shoals in the lower part of the Ouanamane Fm. The transgression reached its maximum extent with the deposition of the argillaceous limestones of the upper Ouanamane Fm. (assigned to Middle Oxfordian by Adams, 1980) and the development of reefal systems in the lower part of the Lalla Oujja Fm. (Fig. 2B and 3; Duval-Arnould et al., this volume). With minor changes, Adam's views, mostly based on the study of the Imouzzer area, were subsequently confirmed by Peybernès et al. (1987) and Bouaouda (2004).

3. Methods

3.1. Facies analysis

The Middle-Upper Jurassic of the EAB was studied along E-W trending anticlines and along the Argana Valley in 11 logged sections, 2 in the Essaouira sub-basin and 9 in the Agadir sub-basin (Fig. 1). Detailed sedimentological analyses were conducted in the field, including description of bed geometries, lithologies, sedimentary structures, fossil content and bioturbation features. In addition, 50 petrographic thin sections were studied for microfacies characterization (texture, allochems, sedimentary structures, bioturbation). For carbonates, the classification by Dunham (1962) with the extension by Embry and Klovan (1971), was used. The siliciclastic descriptions were based on grain size, sedimentary structure, following the Wentworth (1922) grain size scale. Facies were grouped into facies associations, and their vertical and lateral architecture was analyzed along two local and one regional stratigraphic transects combining depositional sequences and biostratigraphy.

3.2. Biostratigraphy

The existing biostratigraphic framework for the study interval was low-resolution and heterogeneous. In order to build a high-resolution biostratigraphic framework, bed by bed collection of ammonites, brachiopods and echinoids was conducted at 6 field sections (Fig. 1). Despite extensive efforts only 21 identifiable ammonite specimens were collected. This work showed that the limited number of ammonites reported in the literature reflects the scarcity of this fossil group at the scale of the basin. Therefore, an extensive search for biostratigraphically significant brachiopods (~500 specimens) was carried out, completed by a survey of benthic foraminifera markers in thin sections. The combination of all available markers was used to refine the biostratigraphic framework in the Essaouira sub-basin and establish a high-resolution

stratigraphic transect for the Agadir sub-basin. Unless otherwise mentioned, the age assignments are derived from the Mediterranean and western European biostratigraphic scales established by the French Jurassic Group (Cariou and Hantzpergue, 1997). Since ammonites are the prime markers for Jurassic biostratigraphy, the biostratigraphic interpretation of brachiopods, echinoids and benthic foraminiferal assemblages can only be considered reliable when direct calibration to ammonite scales is available. The limitations of those groups were discussed by Bassoullet (1997).

In addition to collection of new data, the material collected by Ambroggi (1963) from the Anklout anticline (Izwarn area) was reinvestigated since it complements new findings at the same locality. Besides, Ambroggi's collection that is kept in the Palaeontological Museum of the Ministère de l'Energie et des Mines (MEM) of Rabat (Morocco) was reinvestigated and key specimens were revised.

4. Stratigraphic results

4.1. Lithostratigraphic subdivision

The lithostratigraphic framework retained herein is based on the scheme established by Adams (1979) and Adams et al. (1980), with additions by Peybernès et al. (1987) and Bouaouda (2007) (Fig. 2A). The Ouanamane Fm. was originally subdivided into four members by Adams et al. (1980). The lowermost member (Transition Dolomite Mb.) that marks the passage between the Ameskhoud and Ouanamane formations was only recognized in the Imouzzer anticline. We consider that it cannot be used at the scale of the EAB and therefore a threefold subdivision of the Ouanamane Fm. was retained herein (Fig. 2B and 3).

- Unit 1 (Iggui-n'Tarhazout Oolite Mb.) is dominated by oolitic grainstone and bioclastic packstone. Its base is marked by a firmground that caps the Transition Dolomite Mb., or rests directly on the fluvial siliciclastics of the Ameskhoud Fm.
- Unit 2 (*Somalirhynchia* Limestone Mb.) is composed of marly and more indurated brachiopod-rich floatstones and rudstones regularly topped by very bioturbated levels or by firmgrounds, alternating with some mudstones.
- Unit 3 (Marl Mb.) is dominated by thick marls with carbonate mudstones and floatstones.

These three units are homogeneous in the center of the EAB. They are associated with iron rich and very bioturbated firmgrounds and hardgrounds. The main hardgrounds coincide with the boundaries of the units and are associated to a major change in sedimentation (Fig. 2B).

4.2. Review of the biostratigraphic framework

Over the past 25 years, Bouaouda and collaborators have built a micropalaeontological data set that was integrated with ammonite and brachiopod distributions in order to produce a biochronologic framework of the Middle and Upper Jurassic of the EAB (Bouaouda et al., 2004, 2009). The data set used and presented by these authors is largely heterogeneous, as will be shown below. Consequently, the established and new framework based on this study are reviewed in this section.

4.2.1. Ammonites

4.2.1.1. Review of the existing framework. The ammonite record was interpreted from a limited number of specimens listed in the literature (Gentil and Lemoine, 1906; Roch, 1930; Ambroggi, 1963), most of which were never formally documented and illustrated. The systematic assignment of this historical material needs revision. In addition, brachiopods were illustrated and described by Gentil and Lemoine (1906), but the only modern study was performed by Alméras et al. (1991). A
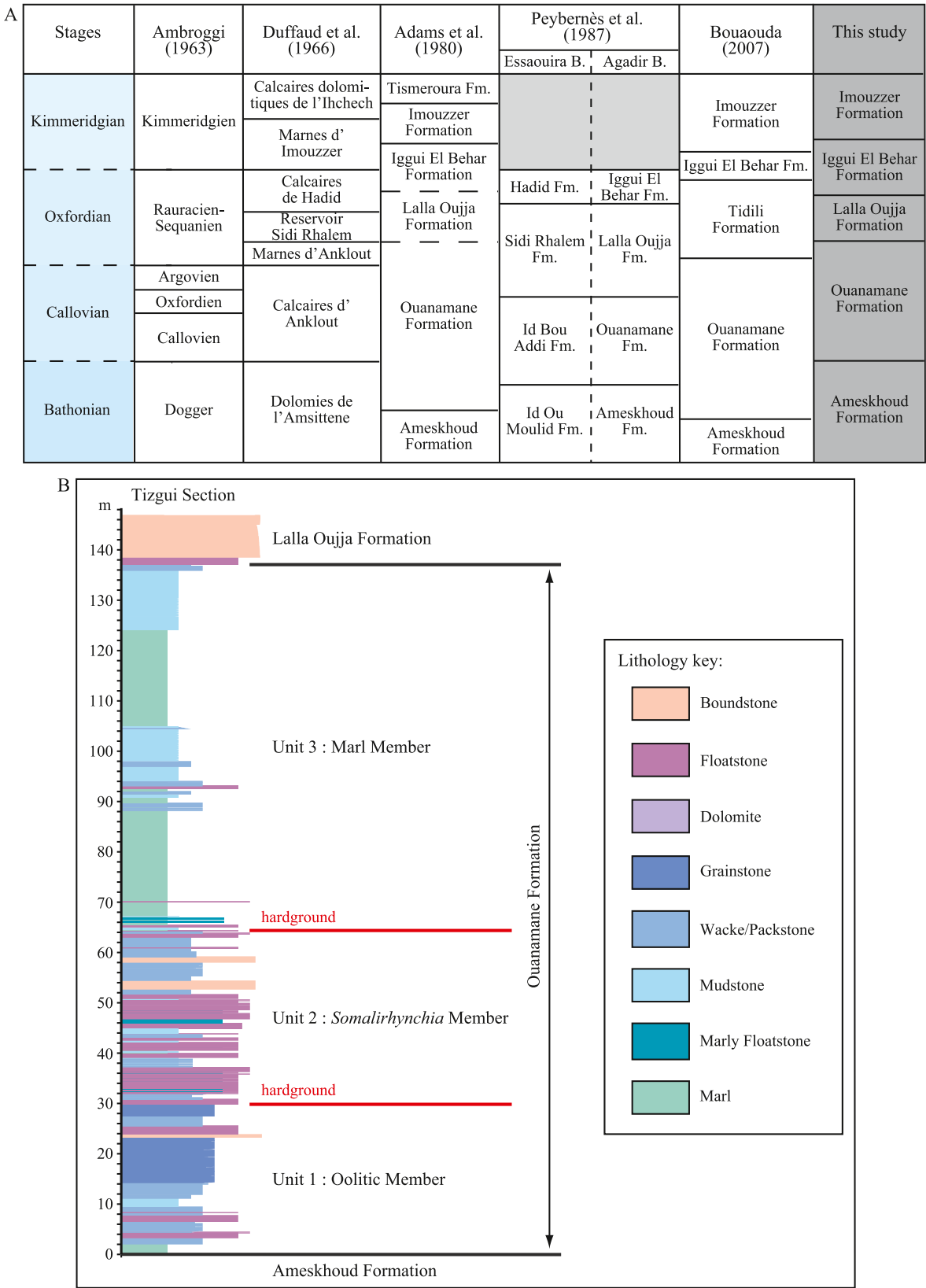
A

Stages	Ambroggi (1963)	Duffaud et al. (1966)	Adams et al. (1980)	Peybernès et al. (1987) Essaouira B.	Peybernès et al. (1987) Agadir B.	Bouaouda (2007)	This study
Kimmeridgian	Kimmeridgien	Calcaires dolomitiques de l'Ihchech	Tismeroura Fm.			Imouzzer Formation	Imouzzer Formation
Kimmeridgian	Kimmeridgien	Marnes d' Imouzzer	Imouzzer Formation			Imouzzer Formation	Imouzzer Formation
Kimmeridgian	Kimmeridgien	Marnes d' Imouzzer	Iggui El Behar Formation			Iggui El Behar Fm.	Iggui El Behar Formation
Oxfordian	Rauracien-Sequanien	Calcaires de Hadid	Lalla Oujja Formation	Hadid Fm.	Iggui El Behar Fm.	Tidili Formation	Lalla Oujja Formation
Oxfordian	Rauracien-Sequanien	Reservoir Sidi Rhalem	Lalla Oujja Formation	Sidi Rhalem Fm.	Lalla Oujja Fm.	Tidili Formation	Lalla Oujja Formation
Oxfordian	Rauracien-Sequanien	Marnes d'Anklout	Lalla Oujja Formation	Sidi Rhalem Fm.	Lalla Oujja Fm.	Tidili Formation	Lalla Oujja Formation
Callovian	Argovien	Calcaires d' Anklout	Ouanamane Formation	Id Bou Addi Fm.	Ouanamane Fm.	Ouanamane Formation	Ouanamane Formation
Callovian	Oxfordien	Calcaires d' Anklout	Ouanamane Formation	Id Bou Addi Fm.	Ouanamane Fm.	Ouanamane Formation	Ouanamane Formation
Callovian	Callovien	Calcaires d' Anklout	Ouanamane Formation	Id Bou Addi Fm.	Ouanamane Fm.	Ouanamane Formation	Ouanamane Formation
Bathonian	Dogger	Dolomies de l'Amsittene	Ouanamane Formation	Id Ou Moulid Fm.	Ameskhoud Fm.	Ouanamane Formation	Ameskhoud Formation
Bathonian	Dogger	Dolomies de l'Amsittene	Ameskhoud Formation	Id Ou Moulid Fm.	Ameskhoud Fm.	Ameskhoud Formation	Ameskhoud Formation

B

Fig. 2. (A) Middle to Upper Jurassic lithostratigraphic subdivisions of the Essaouira-Agadir Basin according to various authors, and as followed in this study. (B) Simplified section of the Ouanamane Fm. to illustrate the subdivision into 3 members.
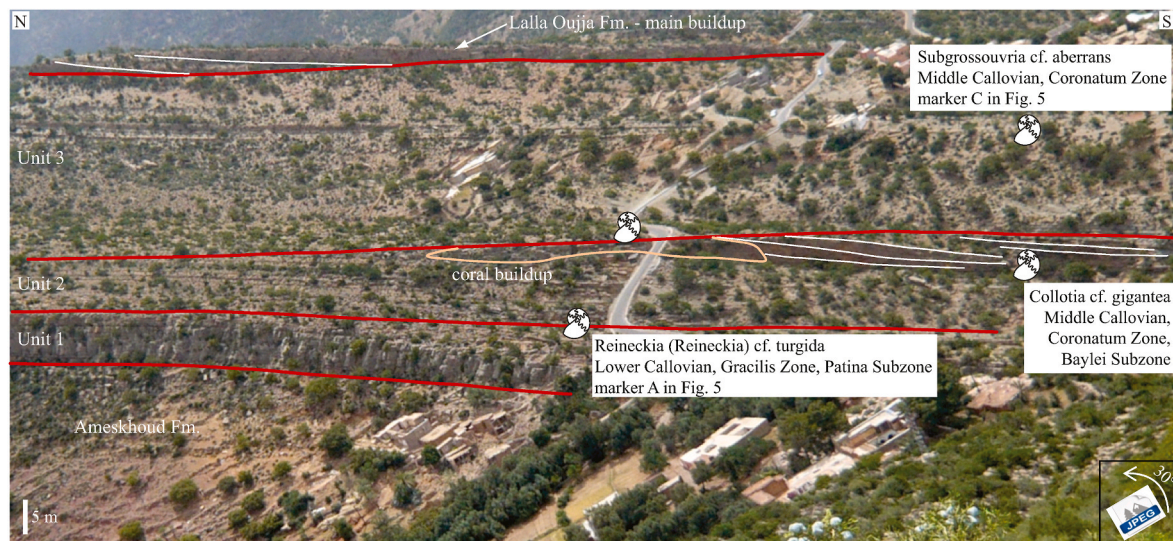
Fig. 3. Photomosaic of the Tizgui transect with associated ammonite fauna, illustrating the subdivision of the Ouaname Fm. into 3 units identifiable in outcrop. A small coral buildup with clinoforms developed near the top of Unit 2, whereas progradational clinoforms formed at the base of the Lalla Oujja Fm. main buildup.

selection of the most representative material was illustrated by Peybernès et al. (1987), but no formal description of these fauna is available. As a consequence, the resolution of the framework proposed by Bouaouda and collaborators is largely handicapped by uncertainties regarding the systematics and precise distribution of the biostratigraphic markers.

A good example of these uncertainties is given by the single Jurassic ammonite ever illustrated from the EAB, i.e. *Perisphictes chavattensis* de Loriol, 1903 in Gentil and Lemoine (1906, pl. 4, n° 6). This specimen was later re-identified as *Perisphinctes* cf. *furcula* Neumayr, 1871 by Roch (1930, p. 203–205) and *Arisphinctes vorda* Arkell, 1939 by Adams et al. (1980, p. 72). The quality of the illustration is fairly poor and the description minimal. Re-examination of the original specimen would be crucial for a proper interpretation of its biostratigraphic value. Our efforts to locate it failed, despite the fact that most brachiopods described in the same publication by Gentil and Lemoine (1906) are still preserved in the collections of Université Pierre et Marie Curie (Paris). The original data certainly do not allow the exact position of the specimen in the stratigraphic succession to be pinpointed. Nevertheless, a level close to the base of the Lalla Oujja Fm. was suggested by Ambroggi (1963).

Bouaouda et al. (2004) report three identifications of the same ammonite from two different stratigraphic levels, as if they were three different specimens. Moreover, the biostratigraphic interpretations of Bouaouda et al. (2004) derived from those erroneous occurrences are not supported by the data presented.

- *Perisphinctes furcula* is said to mark the base of the Lower Oxfordian, while it is well established that this species is a Lower Callovian *Homeoplanulites* (Mangold, 1970; Mangold et al., 1996).
- *Arisphinctes vorda*, which has been transferred to the subgenus *Kranaosphinctes* in recent literature, is a marker of the lower Middle Oxfordian (Głowniak, 2002), and certainly not indicative of the Upper Oxfordian.
- *Perisphinctes chavattensis* lacks recent taxonomic reassessment and the precise stratigraphic position of the species in the Oxfordian remains unknown.

A similar problem arises with the faunal list given by Ambroggi (1963) from what is now regarded as the upper part of the Ouanamane Fm. After a nomenclatural update, the species reported by this author include *Euaspidoceras (E.) phoenicium* Gemellaro, 1877, *Pseudaspidoceras choffati* de Loriol, 1903, *Idoceras hodiernae* Gemmellaro, 1878,

Perisphinctes (Kranaosphinctes) maximus Young and Bird, 1828 and *Perisphinctes (Kranaosphinctes) pickeringius* Young and Bird, 1822.

Based on the I of *E. (E.) phoenicium* alone, a Middle Oxfordian age (Transversarium Zone) was accepted by Olivier et al. (2012). The same authors attributed an Upper Oxfordian age (Bimammatum and Planula zones) to the top of the Ouanamane Fm. from which they report *Subdiscosphinctes* sp. and *Orthosphinctes* sp., based on two specimens that were not illustrated. These ages are far less established than it might seem at first sight. Indeed, most species listed by Ambroggi (1963) indicate a Middle Oxfordian age, but the fauna includes elements of the Plicatilis and Transversarium zones. If confirmed, the occurrence of *Idoceras hodiernae* would indicate a Kimmeridgian age (Olóriz, 2002). Moreover, isolated specimens of *Subdiscosphinctes* sp. and *Orthosphinctes* sp. do not allow precise biostratigraphic control. *Subdiscosphinctes* is known to range through the Middle and Upper Oxfordian, while the taxonomic content and stratigraphic distribution of *Orthosphinctes* s.s. still remains unclear (e.g. Atrops and Meléndez, 1993; Meléndez et al., 2006).

4.2.1.2. New ammonite data. The material studied mainly originates from the Jbel Amsittène and Izwarn anticlines (Fig. 1). Both areas proved to be more ammonite-rich than the other localities investigated. The bulk of the ammonite fauna was obtained from the Ouanamane Fm. (Units 2 and 3), and locally, from the base of the Lalla Oujja Fm. The stratigraphic distribution of the fauna is summarized in Fig. 4, and the most important specimens are illustrated on plates 1 to 3.

The oldest ammonite occurrences lie in the lower third of Unit 2 of the Ouanamane Fm. at Tizgui and Jbel Amsittène. The fauna includes *Indosphinctes patina* Neumayr, 1870 (AMCA-P27A, pl. 1, n° 5), *Indosphinctes* aff. *Choffati* Parona and Bonarelli, 1897 (AMCA-P26C, pl. 1, n° 4). *Reneckeia (R.)* cf. *turgida* Cariou and Meléndez, 1990 (TZG-P-10B). This assemblage indicates an upper Lower Callovian age (Gracilis Zone, Patina Subzone).

At Jbel Amsittène, the upper part of the *Somalirhynchia* Limestone Mb. (Unit 2) is fairly rich and includes: *Reineckeia (R.)* gr. *anceps* Reinecke, 1818 (AMCA-P27C – macroconch, pl. 1, n° 6), *Choffatia* gr. *subbalinensis* Siemiradzki, 1894 (AMCA-P27A – macroconch), *Reineckeia (R.) stuebeli* Steinmann, 1881 (AMCA–II–25 – macroconch, pl. 1, n° 1–2 and AMCA II 26 – microconch, pl. 1, n° 3) and *Rehmania (Loczyceras)* gr. *richei* Flamand, 1911 (AMCA P29 – macroconch). The assemblage indicates the lower part of the Middle Callovian (Anceps Zone). *R. (R.) stuebeli* is the index species of the Stuebeli Subzone.
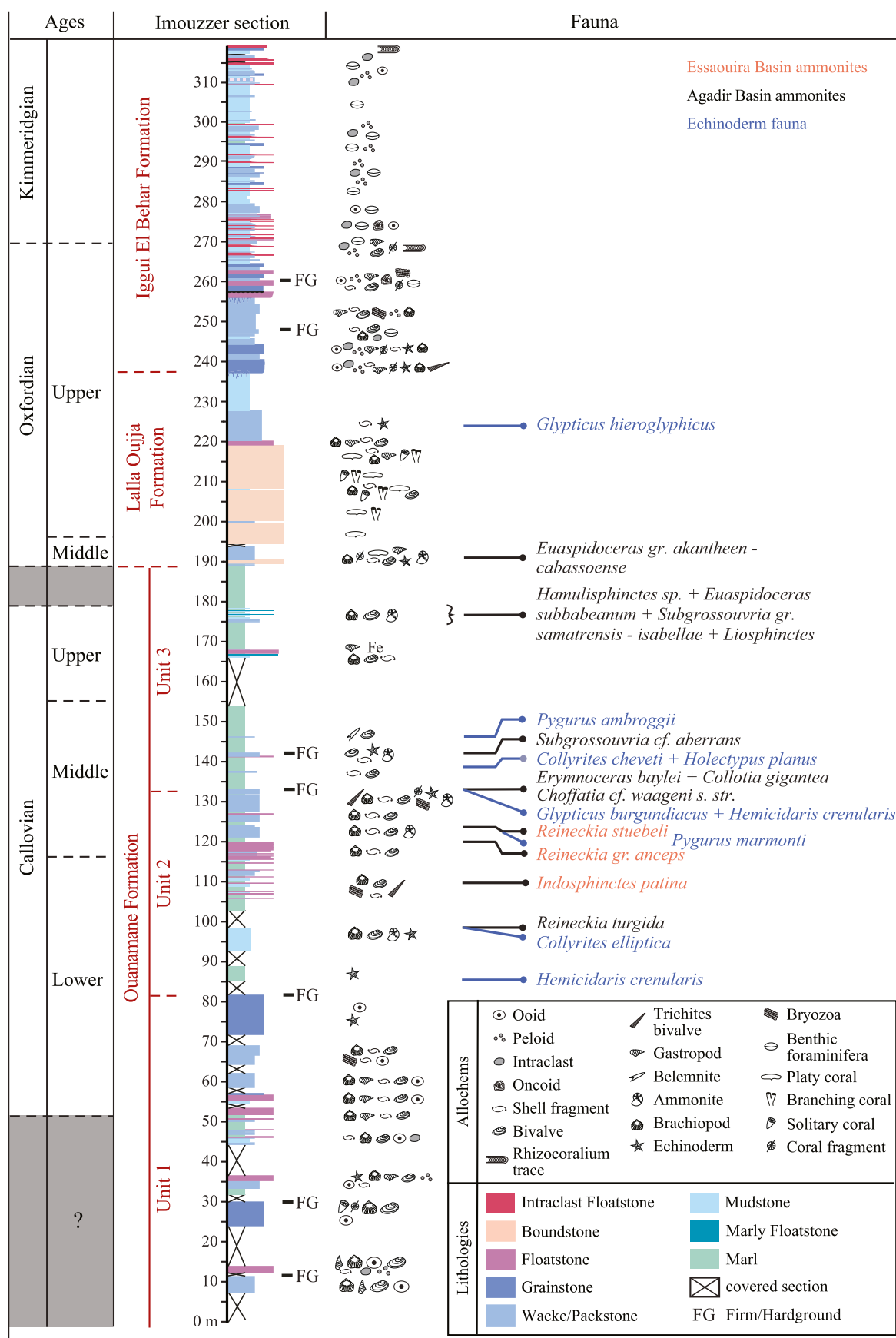
Fig. 4. Callovian and Oxfordian type section with distribution of ammonites and echinoderms in the Essaouira-Agadir Basin.

At Izwarn, an interesting assemblage was reported by Ambroggi (1963) and its stratigraphic position at the top of Unit 2 has been confirmed. Two macroconchs of *Erymnoceras baylei* Jeannet *sensu* Gill et al. (1985) (MEM-AM-955, pl. 2, n° 5–6 and MEM-AM-1425, pl. 2, n° 11–13) indicate the lower part of the Middle Callovian Coronatum Zone (Baylei Subzone). Equivalent levels at Tizgui yielded a very large macroconch of *Collotia* cf. *gigantean* Bourquin, 1968 (TZG-P29. SE) that also indicates the lower part of the Coronatum Zone. At both localities, the fauna is dominated by macroconchs of *Choffatia waageni* Teysseire, 1889 (IZW-A1-P1004, pl. 1, n° 7, IZW-A1-P1005, pl. 1, n° 11, MEM-AM-999b-c, pl. 2, n° 7–8). Ambroggi's collection from Izwarn also includes *Reineckeia (R.)* cf. *rugosa* Cariou and Meléndez, 1990 (MEM-AM-1427) and *Rehmannia (Loczyceras)* sp. Fragmentary body chambers of a form that recalls *Subgrossouvria aberrans* Waagen, 1875 were collected from the lower part of Unit 3 (TDE-P-1, pl. 1, n° 9–10). According to Bardhan et al. (2012), *S. aberrans* characterizes the uppermost part of the Middle Callovian in India. Juvenile specimens of *Reineckeia (R.)* gr. *reissi* Steinmann, 1881 from Imouzzer (IM3) and Assif El Hade (AS-P69) also support an upper Middle Callovian age for the lower part of Unit 3.

The specimens collected by Ambroggi (1963) from the upper part of the Ouanamane Fm. at Izwarn have been revised. Four specimens provide relevant biostratigraphic information.

- MEM-AM-993 (pl. 2, n° 4) (= *Perisphinctes hodiernae* in Ambroggi, 1963) is an *Hamulisphinctes* of the *hamulatus* Buckman, 1921 group that indicates the lower Upper Callovian and most likely the lower part of the Athleta Zone.
- MEM-AM-1001 (pl. 2, n° 9) (= *Arisphinctes maximus* in Ambroggi, 1963) is reinterpreted as a *Subgrossouvria* that recalls *S. isabellae* Bonnot et al., 2008 and *S. samatrensis* Spath, 1931. These species are known to occur in the lower Upper Callovian (Athleta Zone, see discussion in Bonnot et al., 2014).
- MEM-AM-1007 (pl. 2, n° 10) (= *Aspidoceras choffati* in Ambroggi, 1963), is re-identified as *Euaspidoceras subbabeanum* (Sinzow) *sensu* Jeannet (1951) and Bonnot (1995). The species indicates the Uppermost Callovian (Lamberti Zone) (Pellenard et al., 2014a).
- MEM-AM-997 (= *Arisphinctes pickeringius* in Ambroggi, 1963) (pl. 2, n° 3) belongs to the group of *Arisphinctes plicatilis* Sowerby, 1817 that was transferred to *Liosphinctes* by Głowniak (2002). Those forms indicate the lower to uppermost lower Middle Oxfordian (Głowniak, 2002).

The specimen originally identified as *Aspidoceras phoenicium* in Ambroggi (1963) is too worn to allow identification at the genus or species level. Even though the number of specimens is limited, three of them point to an Upper Callovian age (Athleta and Lamberti zones), while the fourth indicates a lower to uppermost lower Middle Oxfordian age. Ambroggi (1963) gave no details about the precise position of those specimens. Preservation and matrix suggest that they may come from different beds within the 3 m interval from which they are reported and there is no evidence that all material was found *in situ*. This part of the Izwarn succession was unfortunately not accessible at the time of field work.

New elements also allow to re-evaluate the age of the base of the Lalla Oujja Fm. A large specimen of *Euaspidoceras* (TIDILI-2, pl. 3, n° 1–5) was collected just above the base of the formation at Tidili. It is closely allied to *E. akantheen* Buckman, 1928 and *E. cabassoense* Spath, 1931. Both species indicates a Middle Oxfordian age (Plicatilis and Transversarium zones). Finally, a fragmentary ammonite from Cap Ghir is identified as *Dichotomosphinctes* sp. (CG-1, pl. 1, n° 8). Although preservation prevents identification at the species level, the specimen matches the sub-adult ornamental stage of *D. antecedens* Salfeld, 1914. This species is a marker of the upper part of the Middle Oxfordian Plicatilis Zone in Southern Europe (Cariou and Meléndez, 1990).

4.2.2. Brachiopods

4.2.2.1. Review of the existing framework. The first account on the Jurassic brachiopods from the EAB was published by Gentil and Lemoine (1906). This material was discussed by Roch (1930) who added personal observations. Ager (1974), Adams (1979) and Adams et al. (1980) discussed the occurrence of *Somalirhynchia* and *Bihenithyris* and their palaeobiogeographic affinities with the fauna of the Jurassic Ethiopian Province (Ager, 1971; Hallam, 1977; Manceñido, 2002).

The occurrence of *Bihenithyris weiri* Muir-Wood (1935), *B. barringtoni* Muir-Wood (1935), *Somalirhynchia africana* Weir (1925), *S. africana ampla* Douvillé, 1885, *Flabellothyris dichotoma* Kitchin, 1900, *Sphaeroidothyris? browni* Muir-Wood (1935), *Lophrothyris? euryptycha* Kitchin, 1900, *Septaliphoria orbignyana* Oppel, 1856, *Burmirhynchia gregoryi* Weir (1929), *Kutchirhynchia indica* d'Orbigny, 1849, *Ornithella calloviensis* Douglas and Arkell, 1928, and "*Rhynchonella*" *marocanina* Gentil and Lemoine (1906), led Peybernès et al. (1987) to suggest a Lower to Upper Callovian age for the Ouanamane Fm. The same authors outlined that *Kutchithyris acutiplicata* Kitchin, 1900, *K. aurata* Kitchin, 1900 and *K. planiconvexa* Kitchin, 1900 may indicate an Upper Bathonian age for the lower part of the formation. Eleven of the 15 taxa listed were illustrated by those authors, but their detailed distribution was not documented. A synthetic log with the stratigraphic distribution of some key species was subsequently published by Bouaouda et al. (2004) who modified the original list of Peybernès et al. (1987) and recorded a new taxon, *Bihenithyris bihenensis* morph *superstes* Douvillé, 1916.

The understanding of the stratigraphic distribution of most species listed above has been largely improved in recent years by the study of the extensive collections made in the Middle East (Cooper, 1989; Feldman et al., 1991, 2001; Alméras et al., 2010) and India (Mukherjee et al., 2003; Mukherjee, 2007). These recent revisions challenge some of the biostratigraphic interpretations of Bouaouda et al. (2004, Fig. 5). For example, the association of *Kutchithyris acutiplicata*, *K. aurata* (a junior synonym of *K. propinqua* Kitchin, 1900) and *K. planiconvexa* indicates a Middle to Upper Bathonian age according to Mukherjee et al. (2003) and Mukherjee (2007). This is not compatible with the well-established Middle Callovian age (Alméras et al., 2010; Alméras and Cougnon, 2013) indicated by *S. orbignyana* and *B. barringtoni* from the same assemblages.

4.2.2.2. New brachiopod data. The stratigraphic distribution of *A. bihinensis* morph *superstes*, *B. barringtoni, B. weiri, S. africana, S. orbignyana*, and *K. indica* in provide new age constraints on the Ouanamane Fm. *A. bihinensis* morph *superstes* was recently revised by Alméras et al. (2010). In Saudi Arabia, this taxon first occurs in the Upper Bathonian and becomes extinct in the middle part of the Middle Callovian (Alméras and Cougnon, 2013). The type specimen from Sinai is assumed to be Bathonian even though there is no solid evidence for this age.

Based on well-constrained collections made in Israel, Algeria, Tunisia, Saudi Arabia, the lower to lowermost middle Upper Callovian age of *B. barringtoni* is now clearly established (Alméras et al., 2010). The same is true for *B. weiri* that is known to range from the upper Middle Callovian into the Upper Callovian (Alméras et al., 2010, p. 69–70). It should be noted that the Callovian age of the East African type material is still poorly constrained.

S. africana is most often reported from the Callovian and Oxfordian stages (Alméras and Cougnon, 2013), even though according to Feldman et al. (1991) it already occurs in the Lower Bajocian of Egypt. This early occurrence was documented by Feldman et al. (1991) and excluded from the synonymy of *S. africana* by Alméras and Cougnon (2013). The Middle to Upper Callovian age of *S. africana* is well-documented in Ladakh (Alméras et al., 1991), Egypt (Feldman et al., 1991), Israel (Feldman et al., 2001) and Tunisia (Ben Ismail et al., 1989). It should also be noted that even if the Saudi Arabia occurrence is clearly dated to
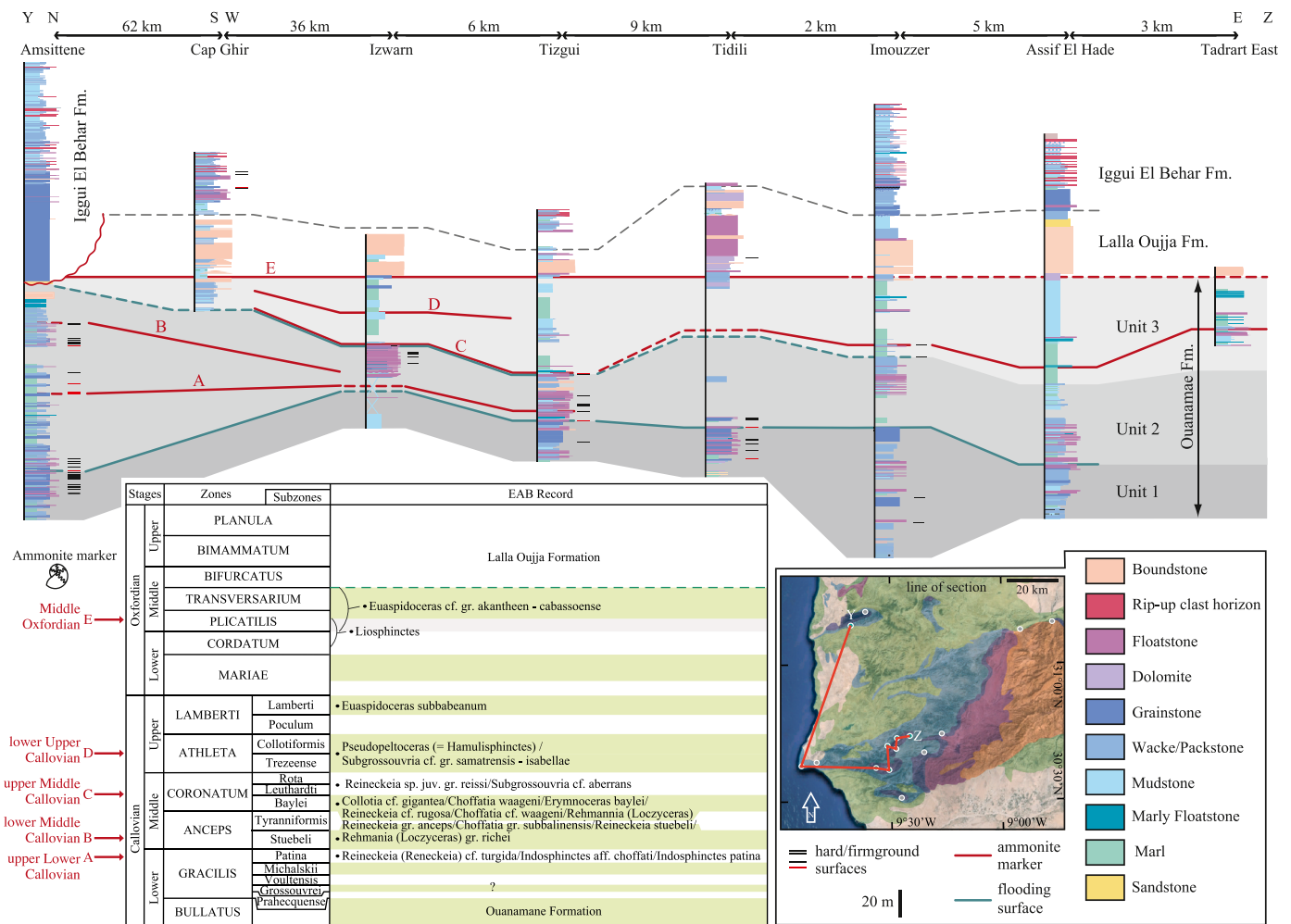
Fig. 5. Correlation of logged sections between the east-central and northwestern part of the Essaouira-Agadir Basin (see inset map for location). The vertical and lateral architecture of facies associations is shown. Red markers indicate biostratigraphic correlations based on ammonites. The lower left inset shows the synthetic ammonite biostratigraphic scale for the Callovian and Oxfordian of the Essaouira-Agadir Basin. Key sequence stratigraphic surfaces are shown in blue and green. Horizontal distances are not to scale.

the Oxfordian (Cooper, 1989), the age of the East African type material is still poorly constrained (Weir, 1925, 1929; Muir-Wood, 1935). Material from the EAB is fairly well documented by Gentil and Lemoine (1906, pl. 5, n° 1–2 as *Rynchonella ampla* Douvillé, 1885), Peybernès et al. (1987, pl. 1, n° 1–2) and Alméras and Cougnon (2013, pl. 14, n° 8). Our new collections clearly show that in the EAB, the first occurrence of *S. africana* predates the upper Lower Callovian, and that the species ranges throughout the Middle Callovian.

S. orbignyana is a useful marker that ranges from the uppermost Lower Callovian (Gracilis Zone, Patina Subzone) to the lower Upper Callovian (Athleta Zone) in Western Europe (France, UK; Laurin, 1984; Alméras and Cougnon, 2013).

The Callovian age of *K. indica* is also well established. Middle Eastern and Tunisian occurrences are of Upper Callovian age, but the species occurs already in the Lower Callovian of India, Pamir and Madagascar (Alméras et al., 2010).

The *Kutchithyris* species that were reported by Peybernès et al. (1987) from the lower part of the *Somalirhynchia* Limestone Mb. Indicates an Upper Bathonian age. This is not consistent with the age obtained from the rest of the brachiopod fauna and associated ammonites. According to Mukherjee (written communication 2018), the Morroccan specimens are likely misidentified and rather match *Kutchithyris mitra* Mukherjee (2007) (= *K. acutiplicata in* Peybernès et al., 1987) and *K. dhosaensis* Kitchin, 1900 (= *K. aurata in* Peybernès et al., 1987), both of which are

present from the Lower and Middle Callovian of Kachch (Mukherjee, 2007).

4.2.3. Benthic foraminifera

Benthic foraminifera can be observed in thin sections from the Ouanamane, Lalla Oujja and Iggui El Behar formations. Peybernès et al. (1987), Bouaouda (2002, 2006, 2009) and Bouaouda et al. (2004, 2009) have highlighted their potential to contribute to biostratigraphic studies, although the ranges of many taxa present are either quite long, and/or require more research to establish precise chronostratigraphic calibration (Septfontaine, 1988; Kuznetsova and Sorkina, 1996; Bassoullet, 1997; Velić, 2007; Hughes, 2018). In ascending order through the Ouanamane Fm. and overlying (Oxfordian) units, the following broad assemblages were recognized (Bouaouda et al., 2004; numbering not in the original paper).

1) *Archaeosepta platierensis* Wernli (1970)*, Andersenolina palastiniensis* Henson (1948)*, Praekurnubia crusei* Redmond, 1964*, Pseudoeggerella elongata* Septfontaine (1988)*, Pseudocyclammina maynci* Hottinger (1967). This assemblage occurs in grainy carbonates towards the base of the Ouanamane Fm. (oolitic Unit 1) and was assigned an upper Bathonian age, although this can be disputed.

2) *Everticyclammina* n. sp., *Ammobaculites* spp., *Verneuilinoides mauritii* Terquem, 1866, *Ophthalmidium strumosum* Gümbel, 1862, *Paleopfenderina gracilis* Redmond, 1964, *P. maynci.* This assemblage occurs in marly units and shales throughout the Ouanamane Fm. It was assigned a Lower Callovian age in the lower part, but in the upper two thirds mostly contains *Everticyclammina* n. sp. only, and was given a Middle and Upper Callovian to Lower Oxfordian age, which is broadly plausible.

3a) *Alveosepta jaccardi* Schrodt, 1894, *Nautiloculina oolithica* Mohler, 1938, *Rectocyclammina* cf. *chouberti* Hottinger (1967). This assemblage occurs in the carbonates of the Lalla Ouija Fm. and was assigned a Middle Oxfordian age, which is possible, but may be younger.

3b) This assemblage contains all elements from 3a, and in addition *Everticyclammina virguliana* Koechlin, 1942. It occurs in the lower carbonates of the Iggui El Behar Fm. The assigned Middle and Upper Oxfordian age is possible, but it may be younger.

The age interpretation of these assemblages can be critically reviewed based on published understanding of the stratigraphic ranges of the taxa present alongside consideration of the illustrations provided by Bouaouda (2002, 2006, 2009) and Bouaouda et al. (2004, 2009). The Upper Bathonian age implied for assemblage 1 in the lower Ouanamane Fm. can be questioned as follows.

1) *Archaeosepta platierensis* was originally described from the Jura Mountains and subsequently from other parts of the northern Tethyan margin (Wernli, 1970; Septfontaine, 1978; Bassoullet and Colchen, 1987; Zagorchev et al., 2000; Haas et al., 2006, 2013; Wernli and Görög, 2007). The record of Chiocchini and Mancinelli (1996) extends the occurrence onto the southern Tethyan margin. The species has been regarded as Upper Bajocian to Upper Bathonian (Septfontaine et al., 1991; Bassoullet, 1997; Chiocchini et al., 2008) based on the original report of Wernli (1970). Nonetheless, the precise range of the species is questioned by Septfontaine (1978), especially since it can be confused with younger taxa such as *Protopeneroplis striata* Weynschenk, 1950. Although potentially biostratigraphically important (presumably a key justification for the Upper Bathonian age assignment of the lower Ouanamane Fm.), it was not illustrated thus precluding confirmation of its occurrence.

2) *Andersenolina palastiniensis* was originally described as *Trocholina palastiniensis* by Henson (1948) from the Kurnub Anticline of the Levant region in strata regarded as Late Callovian (Hudson, 1958; Maync, 1966). Subsequently it was widely described from Jurassic strata in Arabia/Neotethys (usually within the genus *Trocholina*) and considered long-ranging in the Middle and Upper Jurassic (Haas et al., 2006, 2013). Rigaud et al. (2013) have argued that correct generic assignment for this species is *Coscinoconus* (the senior synonym of *Andersenolina*, and distinct from *Trocholina*). The specimens of *A. palastiniensis* illustrated by Bouaouda et al. (2004) are poorly preserved, but comparable with the type material. It was also recorded in the current study from a similar stratigraphic level. Essentially this is something of a "bucket" species for mid-Jurassic trocholinids and given the poor quality of the types, probably requires taxonomic revision to establish its identity and the nature of the species it represents. This may explain why it is assigned a relatively long stratigraphic range. Nonetheless, a *Trocholina palastiniensis* zone is recognized in the Tuwaiq Mountain Limestone of Oman of Late Bathonian–Callovian age (Forbes et al., 2010), whilst Grossowicz et al. (2000) considered *T. palastiniensis* to be very common in the Bathonian–Callovian of Israel. Énay et al. (1987) and Manivit et al. (1990) note that *T. palastiniensis* is common in the Tuwaiq Mountain Limestone of Saudi Arabia, a formation that is undoubtedly Middle–Late Callovian in age (Al-Mojel and Razin, 2022). In summary, whilst the species can occur in Bathonian strata, it is often found in those of Callovian age.

3) *Praekurnubia crusei* is considered ancestral to *Kurnubia* (e.g. Septfontaine, 1988) which arose close to the Callovian/Oxfordian boundary. Thus, the presence of *Praekurnubia* and absence of *Kurnubia* can be a useful key to distinguishing Middle Jurassic strata from Upper Jurassic strata. It occurs in Callovian (and younger) as well as Bathonian strata (Bassoullet, 1997; de Matos, 1997; Grossowicz et al., 2000; Velić, 2007; Chiocchini et al., 2008). The species was not illustrated by Bouaouda et al. (2004), but identifications by Bouaouda (2002) and Bouaouda et al. (2009) seem plausible.

4) *Pseudoeggerella elongata* was originally described from the Bathonian of Switzerland (Septfontaine, 1988). It was regarded as a probable synonym of *Riyadhella regularis* Redmond 1964 by Banner et al. (1991). The type specimens of that species are from the basal member of the Upper Dhruma Formation of Saudi Arabia (Callovian; Al-Mojel and Razin, 2022). Youssef and El-Sorogy (2015) record this species from the overlying Callovian Tuwaiq Mountain Limestone. This places the range of the species within the Bathonian–Callovian. *Pseudoeggerella elongata* was not illustrated by Bouaouda et al. (2004).

5) *Pseudocyclammina maynci* was considered of a broad Middle Jurassic to Oxfordian stratigraphic range by Hottinger (1967). A recent revision of Moroccan stratigraphy regarded the species as an index for the Upper Bajocian–Middle Bathonian (Medina et al., 2011). Bassoullet (1997) considered this species as tentatively no older than Uppermost Bathonian and grading into forms better described as *Pseudocyclammina lituus* Yokoyama, 1890 in the Upper Jurassic. Illustrations by Bouaouda (2002) and Bouaouda et al. (2009) are plausible.

In summary, the evidence for an Upper Bathonian age of the lower part of the Ouanamane Fm. based on foraminiferal assemblage 1 is contestable. The remainder of the Ouanamane Fm. contains foraminiferal assemblage 2 with an assigned Callovian–Lower Oxfordian age. None of the taxa available provide strong support for this assertion, although they do not conflict with it. *Ammobaculites* is a long ranging genus (Boudagher-Fadel, 2018) whilst *Verneuilinoides mauritii* and *Ophthalmidium strumosum* are poorly known taxa that are difficult to recognise in thin section. *V. mauritii* was originally described from the Lower Jurassic of France but is known from the Middle and Upper Jurassic (Zagorchev et al., 2000). *O. strumosum* has a broad Middle–Upper Jurassic range (Tudoran, 1997; Zagorchev et al., 2000; Vuks, 2012). Age calibration of "*Everticyclammina* n. sp." is not possible since it is a new taxon, but it appears to represent a form in the plexus *Everticyclammina praevirguliana* – *Everticyclammina virguliana*. This plexus occurs throughout the Jurassic (Fugagnoli, 2000; Velić, 2007; Schlagintweit and Velić, 2011). *Everticyclammina* was reported from throughout the mid-Jurassic of the UAE (De Matos, 1997). *Paleopfenderina gracilis* is a poorly known taxon that may be a synonym of *Paleopfenderina salernitana* Sartoni and Crescenti, 1962. This species is considered a marker for Bathonian–Callovian strata by Septfontaine (1988) and Tasli et al. (2008). On the other hand, Bassoullet (1997) considered it largely restricted to the Callovian, whilst Velić (2007) considered it restricted to the Bathonian on the Adriatic Platform. The illustrations by Bouaouda (2002) do not confirm this attribution.

Foraminiferal assemblage 3a occurs in the Lalla Oujja Fm. and has an assigned Middle Oxfordian age. The taxa present do not conflict with this, although they could be younger. *Alveosepta jaccardi* is a typical Middle Oxfordian–Lower Kimmeridgian species (Banner and Whittaker, 1991; Bassoullet, 1997), and Mahboubi et al. (2023) have described it from the Middle Oxfordian of the Algerian Central Saharan Atlas (age constrained by brachiopods). The illustration by Bouaouda (2002) and Bouaouda et al. (2004, 2009) is plausible. *Nautiloculina oolithica* is a long-ranging species throughout most of the Jurassic (Bassoullet, 1997; Velić, 2007), whereas *Rectocyclammina chouberti* is typically a Kimmeridgian–Tithonian species (Hottinger, 1967; Bassoullet, 1997). Note that Bouaouda (2006) reinterprets some of the occurrences of

"*Rectocyclammina* cf. *chouberti*" as *Sievoides kocyigiti* Farinacci and Ekmekci (2004)*,* a species described from the Kimmeridgian of Turkey (Farinacci and Ekmekci, 2004). Schlagintweit et al. (2005) described "*Rectocyclammina* sp." from the Kimmeridgian of Austria and considered the specimen of *R.* cf. *chouberti* illustrated by Bouaouda et al. (2004) to be their taxon which is separate from *R. chouberti*.

Foraminiferal assemblage 3b occurs in the lower carbonates of the Iggui El Behar Fm. and is assigned an Upper Oxfordian age. In addition to the taxa present in assemblage 3a, *Everticyclammina virguliana* is present (illustrated by Bouaouda et al., 2004). The timing of the appearance of true *E. virguliana* from its ancestor, *E. praevirguliana* is difficult to place because of transitional forms, but may occur around the Oxfordian/Kimmeridgian boundary (Darga and Schlagintweit, 1991; Fugagnoli, 2000), giving some support to the Upper Oxfordian age assignment of this assemblage.

Table 1
Micro- and macrofacies of the Ouanamane Formation.

Facies	Main components	Groundmass	Sedimentary structures and bioturbation	Bedding
F1 Claystone	Red and grey claystones	N/A	Homogeneous; Rare paleosol horizons	0.2–7 m thick beds, laterally continuous
F2 Sandstone	Very fine to medium quartz, well sorted	Calcite cement	Massive; Horizontal laminations	1–3 m thick beds, laterally continuous
F3 Sandy/silty Dolostone	Silt to medium-grained quartz, well sorted 20–50%; Shell fragments 0–5%; Serpulids; Occasional wood fragments	Euhedral and anhedral dolomite; calcite veins	Massive; Vugs and recrystallization (100–500 μm diameter) < 10%; Alternating with marls	0.2–2 m thick beds, laterally continuous
F4 Sandy Grain-/ Floatstone	Ooids and coated grains (100–500 μm) 10–40%; Medium to granule sized quartz (coated and uncoated) 2–40%; Peloids 0–20%; Coral fragments 2–10%; Gastropods 2–10%; Shell fragments 5–25%; Echinoderm fragments 2–8%; Dasycladacean algae 0–3%	Drusy and blocky calcite sparite 20–30%	Cross-bedding; Low-angle planar cross-bedding; Wave ripples; Planar bedding; Truncated sets; Iron-rich sandy intervals	Thick units (2–35 m thick), composed of sets 15–25 cm thick
F5 Oolitic/ bioclastic Wacke-/Pack-stone	Ooids (200–800 μm) 20–40%; Intraclasts 1–30%; Peloids 10–15%; Superficial ooids 0–10%; Shell fragments (<150 μm diameter) 0–10%; Echinoderms (urchin plates, spines, crinoids) 1–25%; Foraminifera <1%; Bryozoa <1%; Coral fragments <1%; Annelids <1%	Micrite; Locally sparite; Euhedral dolomite <30%	Poorly sorted bioclasts (50–1000 μm diameter); *Thalassinoides*, often better developed on bed tops	0.5–2 m thick beds, laterally continuous
F6 Rud/ Floatstone	Coral fragments 20–35%; Ooids and coated grains (100–500 μm) 0–25%; Gastropods 5–10%; Peloids 15–25%; Shell fragments 5–20%; Bivalves (*Trichites*) 0–10%; Echinoderms 5–10%	Grainstone (peloids, shell fragments, coral fragments)	Cross-bedding; Low-angle cross-bedding; *Skolithos* and *Conichnus*	Units 2 m thick, composed of sets 10–30 cm thick
F7 Oncoidal Floatstone	Oncoids (200–1000 μm) 0–40%; Peloids 15–25%; Coral fragments 0–20%; Gastropods 5–10%; Shell fragments 5–20%; Bivalves 0–10%; Echinoderms 5–10%	Wacke- to packstone (peloids, shell fragments, coral fragments)	*Thalassinoides*	Up to 2 m thick beds
F8 Branching coral Boundstone	Branching corals 5–25%; Solitary corals 0–5%; Bivalves (mainly *Trichites*) 10–15%; Coral fragments 10–30%; Shell fragments 10–20%; Gastropods 10–15%; Echinoderm fragments 5–10%; Peloids 10–20%	Packstone and grainstone	*Thalassinoides* on bed tops Common firmgrounds at bed tops	1.5–5 m thick beds, laterally continuous;
Facies	**Main components**	**Groundmass**	**Sedimentary structures and bioturbation**	**Bedding**
F9 Oolitic Grainstone	Ooids (100–500 μm) 40%; Aggregate grains 10%; Peloids 5%; Coral fragments 0–5%; Echinoderm fragments 2–5%; Foraminifera <1%	Drusy calcite cement; Ferroan calcite sparite 35%	*Thalassinoides* on bed tops; Well sorted ooids; Firm-/hardgrounds on some bed tops	Laterally continuous to meter- and decameter-scale mounded geobodies
F10 Dolostone	Phantoms of ooids (200–400 μm diameter) 30–40%, of peloids (200–400 μm) 20–30%, and of shell fragments 5–15%	Anhedral and subhedral dolomite 60–80%; Calcite cement 15%	*Thalassinoides*; Vugs (dolomite dissolution, 100–1000 μm diameter) < 20%	0.2–3 m thick beds, laterally discontinuous
F11 Bioclastic Wacke-/ Packstone	Shell fragments 10–25%; Quartz (20–100 μm) 10–40%; Glauconite <1%; Echinoderms (plates, spines, crinoids) 8%; Peloids 2–5%; Brachiopods 5–10%; Bivalves 2–5%; Annelids (*Thartharella*?) 1%; Bryozoa <1%	Micrite; Calcite sparite; Ferroan calcite microspar; Euhedral dolomite 5–65%	Poorly sorted bioclasts (20–2000 μm diameter); Sharp base; Occasional wave ripples on bed tops	0.5–2 m thick beds, laterally continuous
F12 Brachiopod Float-/ Rudstone	Brachiopods (*Somalirhynchia* assemblage) 30–50%; Bivalves (*Ostrea gregaria, Pecten, Trichites, Pholadomya*) 20–40%; Shell fragments 5–10%; Echinoderms (mainly *Pygurus*) 0–20%	F5 – Bioclastic wacke- or packstone	Not to very bioturbated; *Thalassinoides* on bed tops Common firmgrounds and perforations at bed tops; Sharp base with fossil accumulations; Graded	0.2–2.5 m thick beds, laterally continuous
F13 Brachiopod Float-/ Rudstone	Brachiopods (*Rhynchonella trilobata* assemblage) 30–50%; Bivalves (mainly *Pholadomya*) 20–40%; Shell fragments 5–10%; Echinoderms 0–20%; Bryozoa 0–2%; Flat corals 0–1%; Sponges 0–5%; Coral fragments 0–2%; Ammonites 0–2%	F5 – Bioclastic wacke- or packstone	Not to very bioturbated; *Thalassinoides* on bed tops Common firmgrounds and perforations at bed tops; Sharp base	0.2–2.5 m thick beds, laterally continuous; Occasional low-angle clinoforms <100 m long
F14 Muddy Boundstone	Platy corals 10–40%; Sponges 0–10%; Thrombolites 0–20%; Coral fragments 0–8%; Echinoderm fragments (mainly spicules) 2–5%; Calcispheres 1–2%; Brachiopod fragments 1–2%; Bryozoa 0–1%; Small shell fragments 0–1%; Foraminifera <1%	Micrite	Locally sponges or thrombolites encrusting platy corals	Mounds 0.5–4 m thick and 6–100 m wide
F15 Marl	Mostly unfossiliferous grey and blue marls; Shell fragments <5% in nodular horizons	N/A	Thin horizons to homogeneous units, Nodular to massive; Very friable	0.1–30 m thick units
F16 Marly lime Mudstone	Dark grey mud; Silt 10–40%	Micrite	*Chondrites*	1–5 cm thick beds, very thin bedding

4.3. Biostratigraphic interpretations

Somalirhynchia africana brachiopods already occur a few metres above the base of Unit 1. All available data suggest that the species is unknown from strata older than the Callovian. Besides, the associated *Kutchithyris* belong to taxa that are known to characterize the Lower and Middle Callovian of India. Similarly, none of the benthic foraminifera reported by Bouaouda et al. (2004) from Unit 1 unequivocally support an Upper Bathonian age for this part of the Ouanamane Fm. As a consequence, we retain a Lower Callovian age for the lower part of the Ouanamane Fm. (Unit 1; Fig. 4), although a slightly older age, i.e. Uppermost Bathonian, cannot be excluded for the base of the formation.

Ammonite faunas clearly indicate an upper Lower Callovian (Gracilis Zone, Patina Subzone) to Middle Callovian (Coronatum Zone, Baylei Subzone) age for the middle and upper part of the *Somalirhynchia* Limestone Mb. (Unit 2; Figs. 4 and 5, marker A and B). This finding improves on the Lower Callovian age accepted by Bouaouda et al. (2004, their Fig. 5) for this part of the succession.

The lower part of the Marl Mb. (Unit 3) is unequivocally dated to the upper Middle Callovian (upper part of the Coronatum Zone) by ammonites (Fig. 5, marker C). Its base is marked by a fairly distinct turnover of the brachiopod faunas that are dominated by *Kutchirhynchia indica* and "*Rynchonella*" *marocanina*. The biostratigraphic value significance of this bioevent remains to be tested.

Based on revision of Ambroggi's collection, Upper Callovian ammonites (Athleta and Lamberti zones) are identified from a 3 m thick interval in the upper part of the Ouanamane Fm. (Fig. 5, marker D). From the same beds a single lower to uppermost lower Middle Oxfordian ammonite is reported. Even if this specimen was found loose, this ammonite assemblage strongly suggests that the Callovian-Oxfordian is affected by condensation and/or hiatuses that involve the Uppermost Callovian and most of the Lower Oxfordian. Finally, when dated by ammonites, the onset of coral buildup development at the base of the Lalla Oujja Fm. falls within the Middle Oxfordian (Fig. 4; Fig. 5, marker E).

5. Facies analysis

5.1. Macro and microfacies

Outcrop and thin section observations allowed definition of 16 facies summarized in Table 1. All facies, except Facies 1, 2 and 15, are interbedded to variable degrees with marl beds ≤1 m thick, containing scarce shell fragments and rare silt. The facies identified in the Ouaname Fm. group together into 5 facies associations (FA), which represent distinct environments from a continental/marine transition to the outer ramp. In addition, key sedimentary surfaces aid in reconstruction of sedimentary environments and basinwide correlations.

5.2. Key surfaces

Across most facies, tops of carbonate beds commonly show evidence for synsedimentary modification, lithification and the formation of firmgrounds and hardgrounds. Two types of modified bedding surfaces were observed: (1) *Thalassinoides* burrows are very common and can reach several decimeters downwards. They form a dense boxwork of tubular burrows with ≤2 cm diameter (Fig. 6A). (2) Beds without bioturbation and beds with *Thalassinoides* boxworks are capped by a sharp surface covered in iron-rich black crusts (Fig. 6B). The crusts are usually encrusted by serpulids, sponges, oysters and other bivalves. The most prominent of these surfaces further have perforations millimeters to centimeters in diameter and filled by overlying sediment, some with a boring bivalve still in them (Fig. 6C). Both surface types tend to be overlain by fossil-poor marls.

Thalassinoides burrows are formed by dwellers and deposit feeders under well-oxygenated shallow-water conditions in muddy-sandy
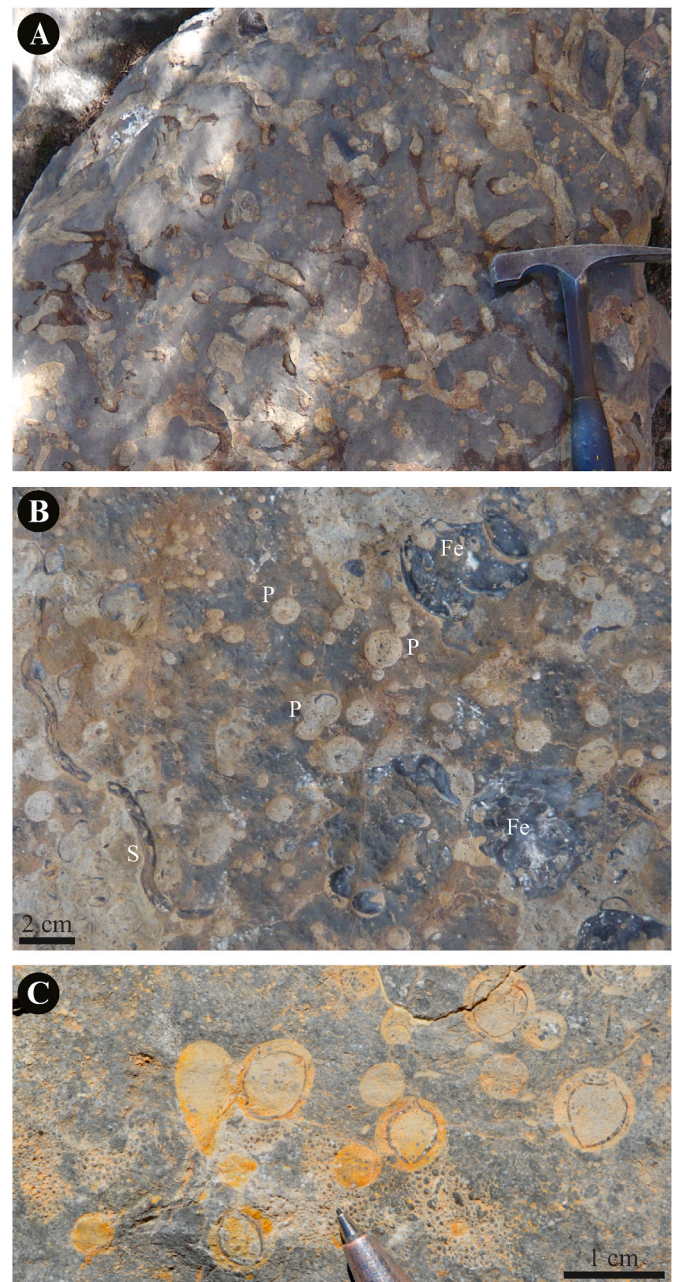
Fig. 6. Firm- and hardground surfaces. (A) Top-down view of a *Thalassinoides*-burrowed firmground bedding surface. Hammer head is 16 cm wide. (B) Top-down view of a hardground surface perforated by bivalves (P), and encrusted by iron (Fe, partly oyster shells?) and possible serpulids (S). (C) Detail of the hardground perforations with bivalve shells inside.

relatively stable substrates (Ekdale et al., 1984; MacEachern et al., 2006). The observed examples can represent both softgrounds and firmgrounds with incipient lithification (Pemberton et al., 2004; MacEachern et al., 2006), and may resemble *Thalassinoides*-dominated *Glossifungites* ichnofacies (Frey and Seilacher, 1980), although evidence for marine ravinement surfaces was not observed. The second type of surface shows clear evidence for iron accumulation and lithification to allow encrustations and borings (Bromley and Ekdale, 1984; Fürsich et al., 1992; Christ et al., 2015), and represents hardgrounds. Both types of surfaces are omission surfaces of unknown duration. As they are overlain by fossil-poor marls, they are interpreted to reflect episodic deepening of the environment.

5.3. Facies associations

5.3.1. Facies association 1: coastal transition (facies 1, 2, 3, 4)

This facies association includes red/grey claystones (Facies 1; Fig. 7A), carbonate-cemented unfossiliferous sandstones (Facies 2), sandy grain- and floatstones (Facies 4), as well as sandy-silty dolostones and dolomitic packstones (Facies 3; Fig. 7B) (Table 1). The sandstones are massive, horizontal-laminated (Facies 2), or trough cross-laminated with some low-angle parallel laminations and wave ripples (Facies 4; Fig. 7C). Paleosol horizons occur in Facies 1. Quartz is present throughout the coarser-grained facies (Fig. 7B), and sandstones are well sorted. Carbonate-coated quartz, ooids, peloids and bioclastic elements are common in the carbonate facies, and wood fragments occur locally in dolostone beds. The associated fauna consists mostly of gastropods and unidentified bivalves and is thus of relatively low diversity, although a more diverse marine fauna (coral fragments, echinoderm spicules, brachiopods, gastropods, dasycladacean algae) appears locally along discrete horizons in Facies 4.

The abundance of detrital quartz and presence of wood suggest proximity to a source of continental siliciclastic input. This detrital input was deposited by persistent unidirectional and oscillating traction currents, as indicated by cross and horizontal lamination. Absence of fossils in Facies 1 and 2, and occurrence of paleosols in Facies 1 indicate continental conditions with subaerial exposure. The low faunal diversity in Facies 3 suggests continental to brackish conditions in coastal lakes (Hudson et al., 1995), or near-coastal restricted marine conditions.

Green algae indicate shallow-water marine conditions with abundant light for Facies 4 (Flügel, 2010). Facies 4 is here interpreted as beach strandplain deposits (McCubbin, 1982; Dominguez and Wanless, 1991) due to the more open-marine fauna, the good sorting, and the sedimentary structures typical of beach deposits (Thompson, 1937). Collectively, this facies association represents transitional environments between continental and nearshore marine conditions, ranging from beach shoreface to coastal lakes or restricted lagoons.

5.3.2. Facies association 2: backshoal (facies 5, 6, 7, 8)

Bioturbated wackestones and packstones (Facies 5) dominate this facies association. They are interbedded with fossiliferous floatstones and rudstones (Facies 6), oncoidal floatstones (Facies 7), and locally associated with coral boundstones (Facies 8). Peloids, ooids (often only as superficial ooids), and a moderately diverse fauna of unbroken echinoderm and bivalve bioclasts occur in the wacke- and packstones (Fig. 8A). In contrast, float- and rudstones (Facies 6) contain a relatively diverse fauna of gastropods, echinoderms, *Trichites*, other bivalves, as well as branching and massive coral fragments. Horizontal bedding and trough cross-bedding occur in floatstones to rudstones. In Facies 7, lobate and concentric oncoids grow mainly on coral fragments and other bioclasts and occur in wackestone to packstone matrix of Facies 5 (Fig. 8B). Boundstones (Facies 8) consist of *in situ* branching coral colonies in m-thick beds (Fig. 8C). The interbedding of Facies 5 and 6 is often distinctly cyclic as m-scale to dm-scale fining-upward bedsets with gastropods and *Trichites* bivalves at the base (Facies 6), passing upwards
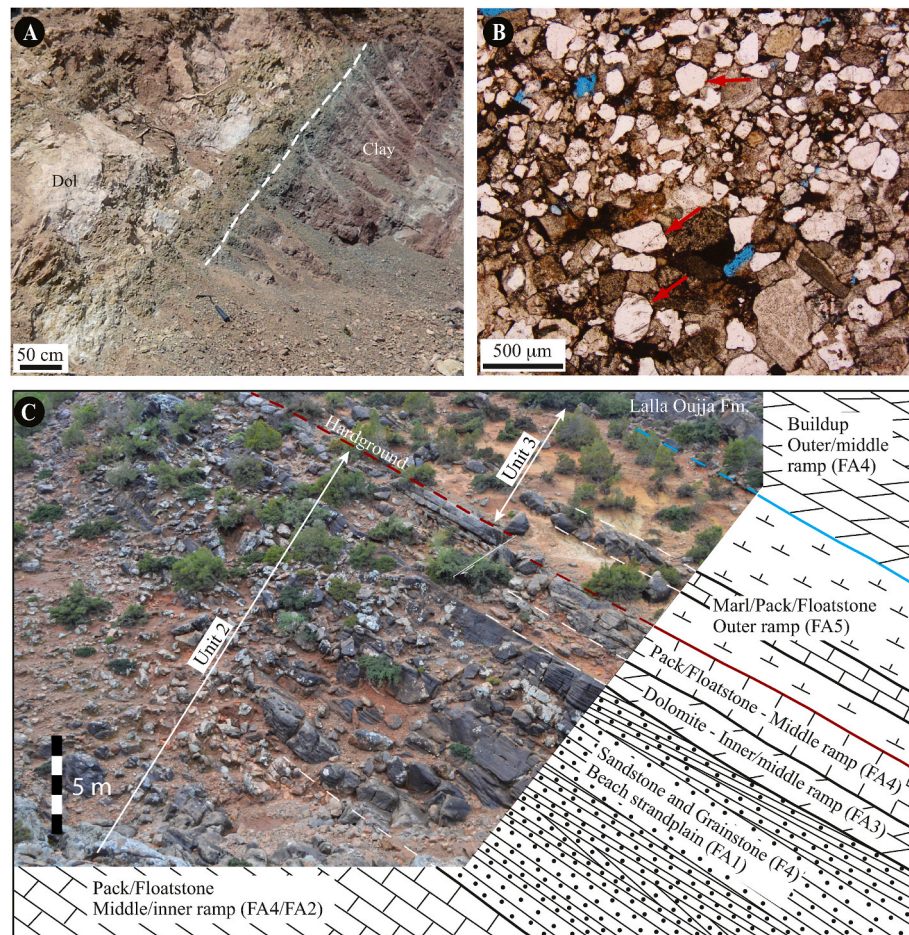

Fig. 7. Facies association 1. (A) Contact between the red claystones (Clay, Facies 1) of the Ameskhoud Fm. and grey-tan sandy dolomites (Dol, Facies 3) of the basal Ouanamane Fm. (Transition Mb.). (B) Sandy dolomite (Facies 3) microfacies. Arrows indicate angular quartz grains. (C) Succession of Ouaname Fm. Units 2 and 3, overlain by coral buildups of the Lalla Oujja Fm., in the south of the Amsittene anticline (section AME). Sandstones and grainstones with low angle cross bedding (Facies 4) are intercalated near the top of Unit 2.
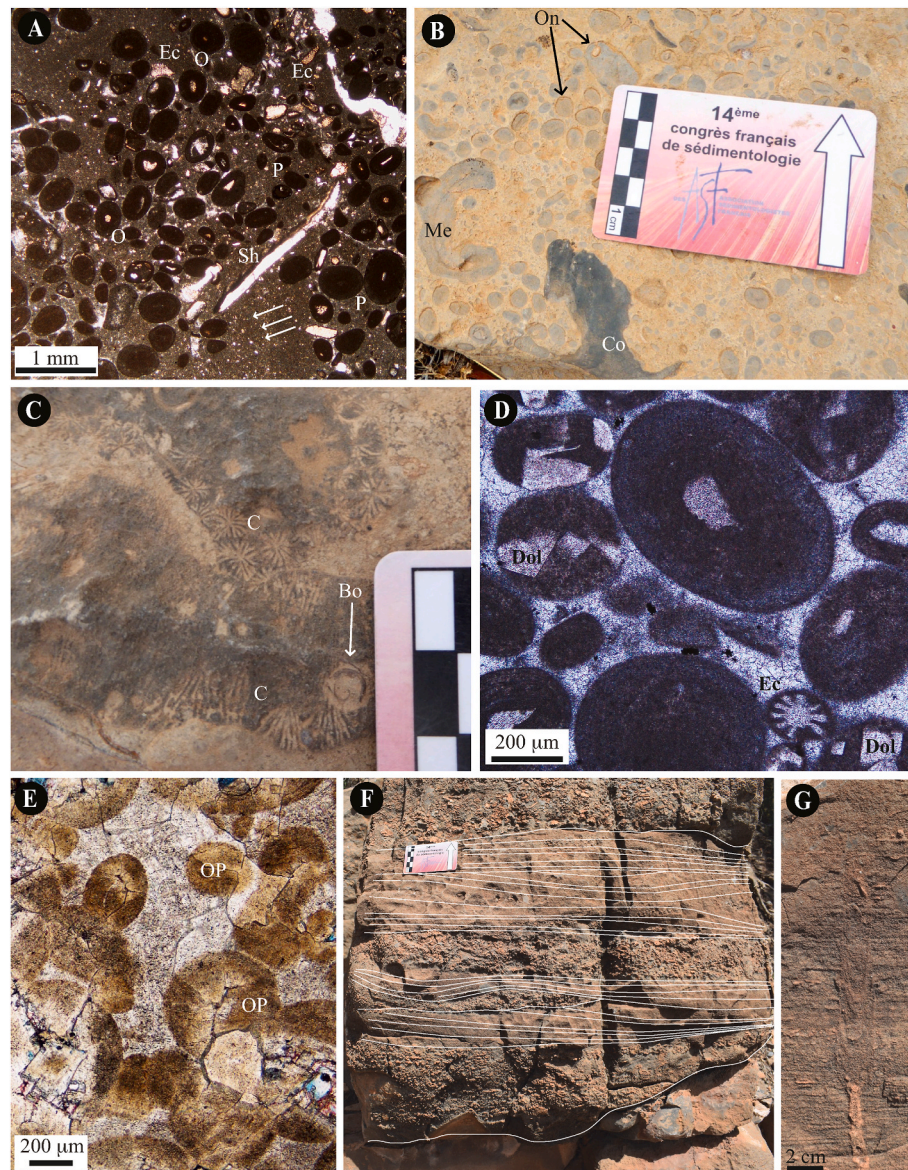
Fig. 8. Facies associations 2 and 3. (A) Microfacies of bioclastic-oolitic wacke-/packstone (Facies 5). Most allochems are peloids (P) and ooids (O), with bivalve shell fragments (Sh) and echinoderms (Ec). Arrows indicate silt-size quartz grains. (B) Oncoidal floatstones (Facies 7), dominated by oncoids (On), coral fragments (Co) and microbially coated shell (Me). Scale in centimeters. (C) Scleratinian coral in boundstone (Facies 8), bored by bivalve (Bo). Scale in centimeters. (D) Oolitic grainstone (Facies 9). Euhedral dolomite (Dol) crystals partially replace the ooids and pseudopeloids. Ec: micrite-coated echinoderm spine. (E) Dolostone (Facies 10) with ooid or peloid phantoms (OP). (F) Alternation of coral-rich rudstone (Facies 6) and oolitic-bioclastic grainstone (Facies 9), with trough cross-stratification. Scale in centimeters. (G) *Conichnus* trace fossil in Facies 6.

to wacke- and packstones (Facies 5). In other cases float- and rudstones form bed tops of wacke- and packstones. Cycle tops are often burrowed by *Thalassinoides*.

Extensive bioturbation by dwelling and deposit feeders (*Thalassinoides*), abundance of oncoids and in-situ coral boundstones representing small patch reefs all demonstrate a shallow and well oxygenated environment (Ekdale et al., 1984; MacEachern et al., 2006; Palma et al., 2007; Flügel, 2010; Nasiri et al., 2020). Facies 5 contains moderate faunal diversity, abundant micrite and largely unbroken shells, which reflects deposition under low hydrodynamism and likely somewhat restricted conditions (Strasser, 1986; Burchette and Wright, 1992). The presence of *Thalassinoides* trace fossils (*Cruziana* ichnofacies; Frey and Seilacher, 1980) and firmgrounds further support reduced hydrodynamism and sedimentation rate on relatively stable substrates (MacEachern et al., 2006; Fürsich et al., 1992). In contrast, higher-energy reworking is indicated for the float- and rudstones (Facies 6) by the prevalence of granular textures, traction current sedimentary structures and fragmented bioclasts. Thick *Trichites* shells were adapted to energetic peri-reefal environments (Hallam, 1976). The allochem composition in Facies 6 (more diverse fauna, ooids) suggests this material may have been transported at least in part from a neighboring high-energy belt (see section 5.3.3 below) (Palma et al., 2007; Olivier et al., 2008). The distinctly cyclic and fining-upward sedimentary architecture of Facies 6 and 5 suggests alternating higher and lower hydrodynamism, likely derived from waning traction and/or storm currents, although no direct evidence for storm deposits was found in this facies association. The evidence for partially protected conditions (moderate faunal diversity in Facies 5, oncoids in Facies 7; Flügel, 2010) points to a backshoal environment where traction currents and/or storms repeatedly moved allochems from a high-energy belt into an area that would have been partially protected by the same belt. Although the *Cruziana* ichnofacies observed here typifies mostly middle ramp settings,

it can extend into relatively low-energy more proximal environments (MacEachern et al., 2006). The backshoal may thus have been located in the lower part of the inner ramp.

5.3.3. Facies association 3: high-energy belt and shoals (facies 5, 6, 9, 10)

This facies association includes bioclastic-oolitic grainstones (Facies 9; Fig. 8D), coral- and gastropod-dominated rudstones (Facies 6), porous dolostones (Facies 10; Fig. 8E), as well as oolitic and bioclastic packstones (Facies 5). Beds of Facies 9 form continuous sheets or meter-scale to decameter-scale mounded geobodies, whereas dolostone beds (Facies 10) always consist of sharp-based ungraded carbonate sand bodies. Facies 6 occurs in this facies association as erosively-based beds with planar bedding, trough cross bedding (Fig. 8F), as well as *Conichnus* and *Skolithos* trace fossils (Fig. 8G). Dominant grains in Facies 9 are pseudopeloids, micritized tangential ooids, and ooid grain aggregates. Ooid nuclei comprise benthic foraminifera, echinoderm spines and shell fragments. Sorting of ooids is good. Some horizons also contain larger skeletal elements such as bivalve and brachiopod shells. The crystalline dolomite in Facies 10 almost always contains relics of ooids or peloids (Fig. 8E). The rudstones (Facies 6) have bioclasts of gastropods, echinoderms, bivalves, brachiopods, and corals. *Thalassinoides* trace fossils are developed on bed tops in Facies 5 and 9.

The thick micrite coatings and good sorting of the ooids, as well as evidence for persistent traction currents indicate a high-energy environment for Facies 9 (Strasser, 1986). This is consistent with *Skolithos* and *Conichnus* trace fossils (Facies 6), which were created by suspension feeders in a high-energy environment with unstable substrates (*Skolithos* ichnofacies; MacEachern and Pemberton, 1992; MacEachern et al., 2006; Gerard and Bromley, 2008). The high degree of bioturbation suggests shallow oxygenated conditions (Ekdale et al., 1984). The restricted lateral extent of many beds in Facies 9 and 10 is a good indicator for deposition within shoals or bars. Intermittent periods of reduced hydrodynamic energy, lower sedimentation rate and more

stable substrates are indicated by firmground development (Fürsich et al., 1992), aggregate grains and occurrence of oolitic-bioclastic packstones (Facies 5) (Strasser, 1986; Steinhoff and Strohmenger, 1996), *Thalassinoides* trace fossils (MacEachern et al., 2006), and association of *Skolithos* and *Cruziana* ichnofacies (Pemberton and MacEachern, 1997; Nasiri et al., 2020). These observations indicate that high-energy shoals were located at the transition from inner to middle ramp. Facies association 3 represents oolitic shoals and carbonate sand sheets (Burchette et al., 1990) that can be mapped out with a lateral extent of >10's of kilometers across the basin.

5.3.4. Facies association 4: middle ramp (facies 11, 12, 13, 14)

Highly bioturbated fossil-rich wackestones and packstones (Facies 11), brachiopod floatstones and rudstones (Facies 12 and 13), and muddy boundstones (Facies 14) constitute this association. The macrofauna in float- and rudstones is dominated by brachiopods (including *Rhynchonella* and *Terebratula*; Fig. 9A), associated with bivalves (*Pholadomya*, *Lopha*, *Trigonia* and *Trichites*). Gastropods, benthic foraminifera, regular echinoderms (large *Collyrites* and *Pygurus* echinoids), serpulids and coral debris are also common. Ammonites and bryozoans are rare. The distinction between Facies 12 and 13 is based on their dominant brachiopod assemblage (Facies 12: *Somalirhynchia* assemblage, Facies 13: *Rhynchonella trilobata* assemblage; Table 1). Finer-grained beds and the groundmass between large skeletal allochems are dominated by a bioclastic wackestone and packstone (Facies 11, Fig. 9B). The bioclasts are mainly subrounded to subangular fragments of brachiopods, bivalves and echinoderm plates derived from the same macro-elements present in the floatstone and rudstone units (Fig. 9B). Silt- and sand-sized quartz is common. Many bed tops were reworked by *Thalassinoides* boxworks. Hardgrounds with iron-stained, encrusted (oysters, serpulids, sponges) and bored surfaces are common (Fig. 6B and C). Facies 14 locally forms mounded geobodies (10–40 m wide) composed of microsolenid (*Dimorpharaea*) platy coral boundstones,
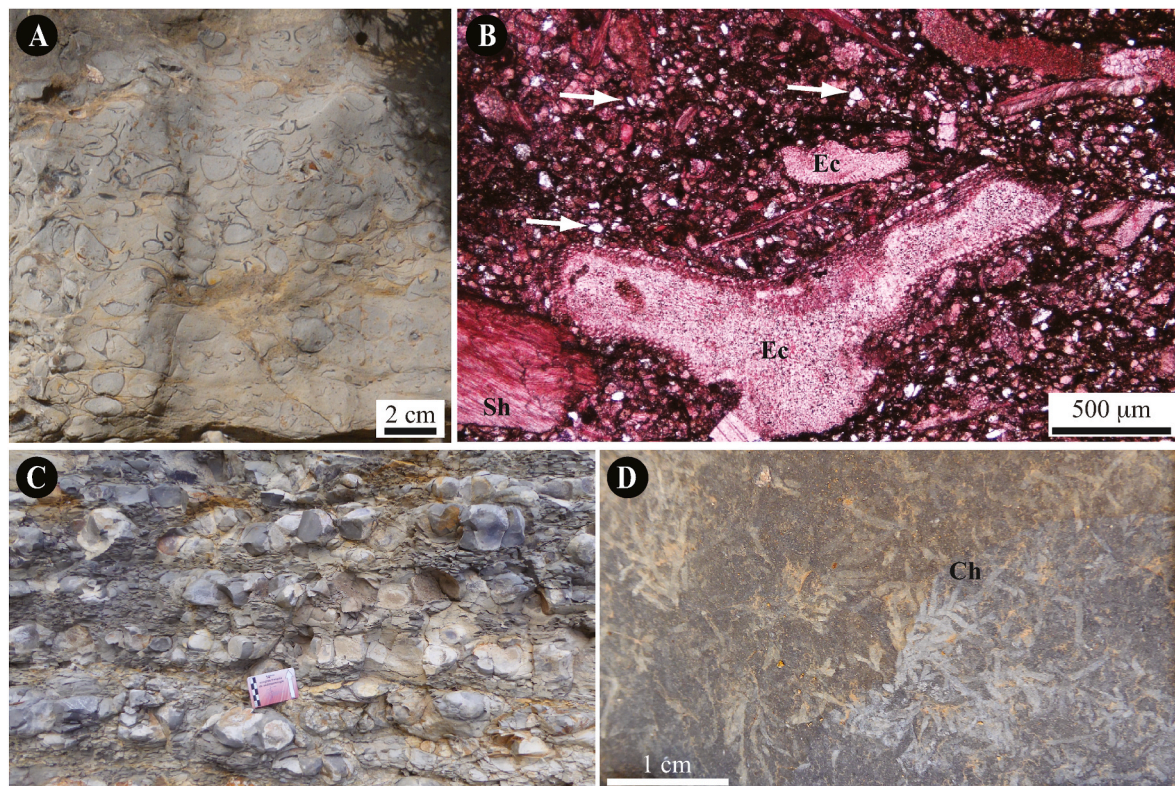

Fig. 9. Facies associations 4 and 5. (A) Brachiopod rudstone with *Terabratula* and *Rhynchonella* (Facies 13). (B) Bioclastic packstone with silt-sized quartz grains (arrows), echinoderms (Ech), and shell fragments (Sh) in a micrite matrix (Facies 11). (C) Grey nodular marl (Facies 15). Scale in centimeters. (D) Silty marl with *Chondrites* traces (Ch) (Facies 16).

associated with shell fragments, echinoderm spicules, and bryozoan fragments. Some buildups have low-angle clinoforms prograding off the buildup (Fig. 3). Associated background facies are wackestones (Facies 11).

The high amount of mud in the system and the many unbroken macrofossils in life position suggest the environment was relatively quiet, but still subject to intermittent reworking as shell fragments are present. Finer bioclasts (Facies 11) match the large allochems in Facies 12 and 13, and are subrounded to subangular, suggesting the bioclasts were not transported far, possibly by storms (Fürsich et al., 2003). There is no evidence for persistent traction or oscillatory currents. Sedimentation rate was intermittently slow, allowing widespread bioturbation of bed tops, but equally favouring the lithification of bed tops and hardground formation (Bromley and Ekdale, 1984; Fürsich et al., 1992; Christ et al., 2015). The high faunal diversity and moderate to low hydrodynamic energy are also consistent with the extensive development of *Cruziana* ichnofacies (presence of *Thalassinoides*), typical of environments between the fair weather and storm wave base with cohesive soft substrates (Frey and Seilacher, 1980; MacEachern et al., 2006). The flat plate-like habit of microsolenid corals in the buildups (Facies 14) is a common response to low light, low energy and low background sedimentation rates (Insalaco, 1996; Olivier et al., 2012). The background sedimentation to the buildups, with the first occurrence of ammonites and bryozoans, represents the deepest water within this facies

association, and is transitional to facies association 5. Overall, facies association 4 had low energy but sufficient light for a diverse macrofauna. Intermittent reworking and a small fraction of detrital input are indicated. This facies association developed in middle ramp settings.

5.3.5. Facies association 5: outer ramp (facies 12, 13, 15, 16)

Facies association 5 is dominated by marls and nodular calcareous marls (Facies 15; Fig. 9C), with intercalations of indurated brachiopod-rich floatstones and rudstones (Facies 12, 13). Dark grey lime mudstones with abundant *Chondrites* burrows occur in a few sections (Facies 16; Fig. 9D). The marls are normally devoid of fauna, except for occasional terebratulid brachiopods in the nodular marls. In contrast, floatstones and rudstones have a rich fauna often in life position, which is dominated by brachiopods, large gastropods, echinoderms, shell fragments, ammonites, and horizontal horizons of platy corals (Fig. 10A–D). Unbroken *Hemicidaris* regular echinoids often retain their spicules (Fig. 10A). Some of these horizons pinch out over <10's of meters, while other horizons can be followed for kilometres (Fig. 10E and F). The base of these beds is often sharp and erosive, and the floatstones become marlier towards the top, or they are terminated by firm- or hardgrounds, which are dominated by boring bivalves (Fig. 10B). The lime mudstones (Facies 16) contain up to 40% silt-sized quartz.

For the marls (Facies 15), the scarce fauna of terebratulids and accumulation of mud indicate a low-energy environment (Ager, 1965).
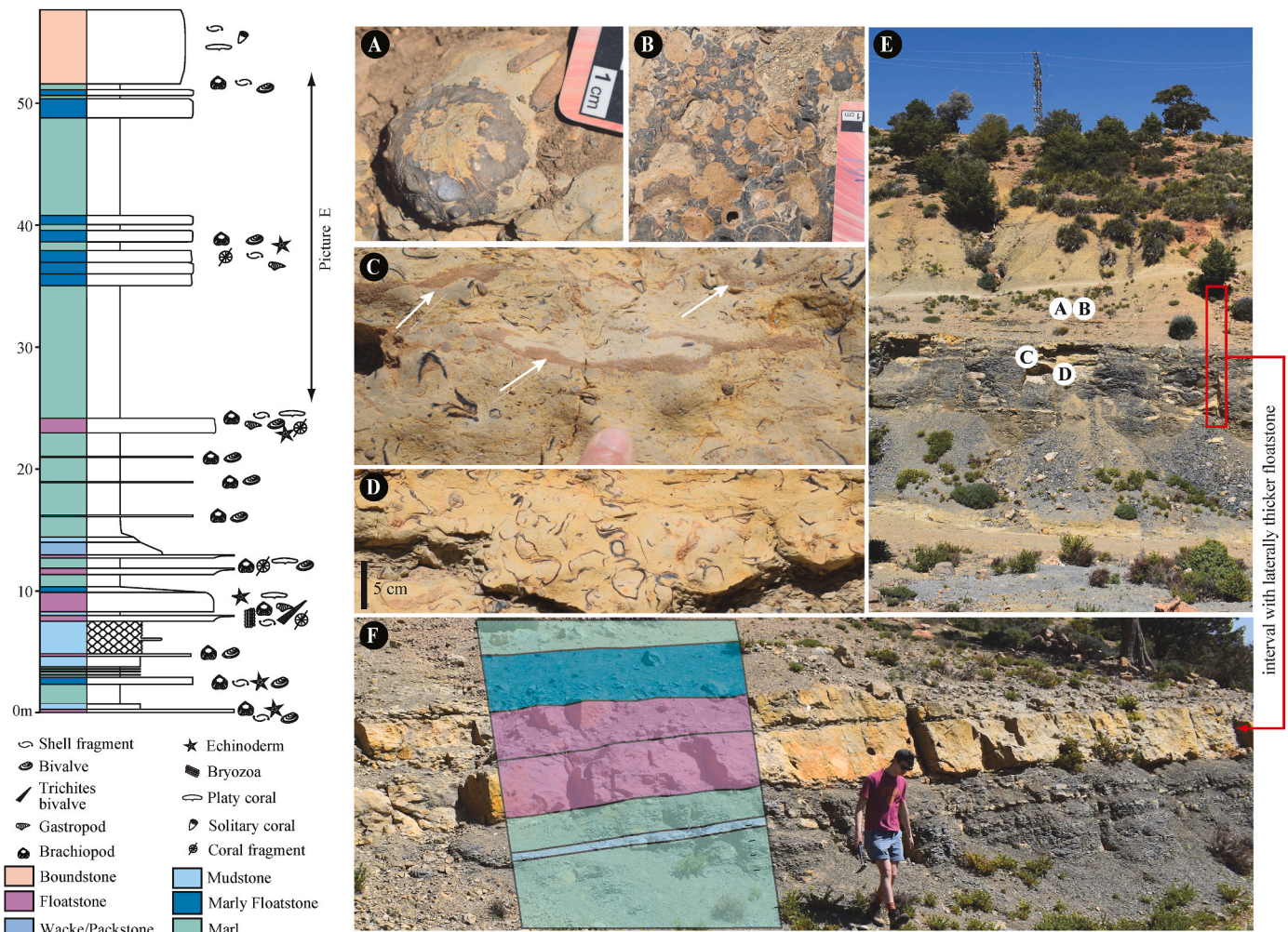

Fig. 10. Composite section (left) of the Tadrart East locality and key features. (A) Sea urchin with spicules in life position. (B) Hardground showing hardened and encrusted surface with bivalves. (C) Floatstone/boundstone (Facies 14); arrows point to platy corals in life position. (D) Brachiopod floatstone (Facies 12/13). (E) Overview of the upper part of the section with positions of photos A-D. (F) Close-up view showing how the indurated horizon thickens 200m northward from picture E.

When firm- or hardgrounds developed during episodes of reduced sedimentation rate, encrusting organisms were less common relative to facies association 4, suggesting a reduction of light and nutrients in deeper settings. In contrast, the richer fauna found in life position in the float- and rudstones (Facies 12, 13) is evidence for more favourable growth conditions, likely slightly shallower and/or with elevated nutrient levels. In-situ reworking by short-lived events of higher energy, possibly storms, is locally indicated by erosive bases (Fig. 10D; Fürsich et al., 2003). As these beds are often capped by firm- and hardgrounds, small-scale variations in relative sea level cannot be excluded. *Chondrites* trace fossils found in Facies 16 are a marine indicator, and the presence of such a monospecific trace fauna is an indicator of restricted
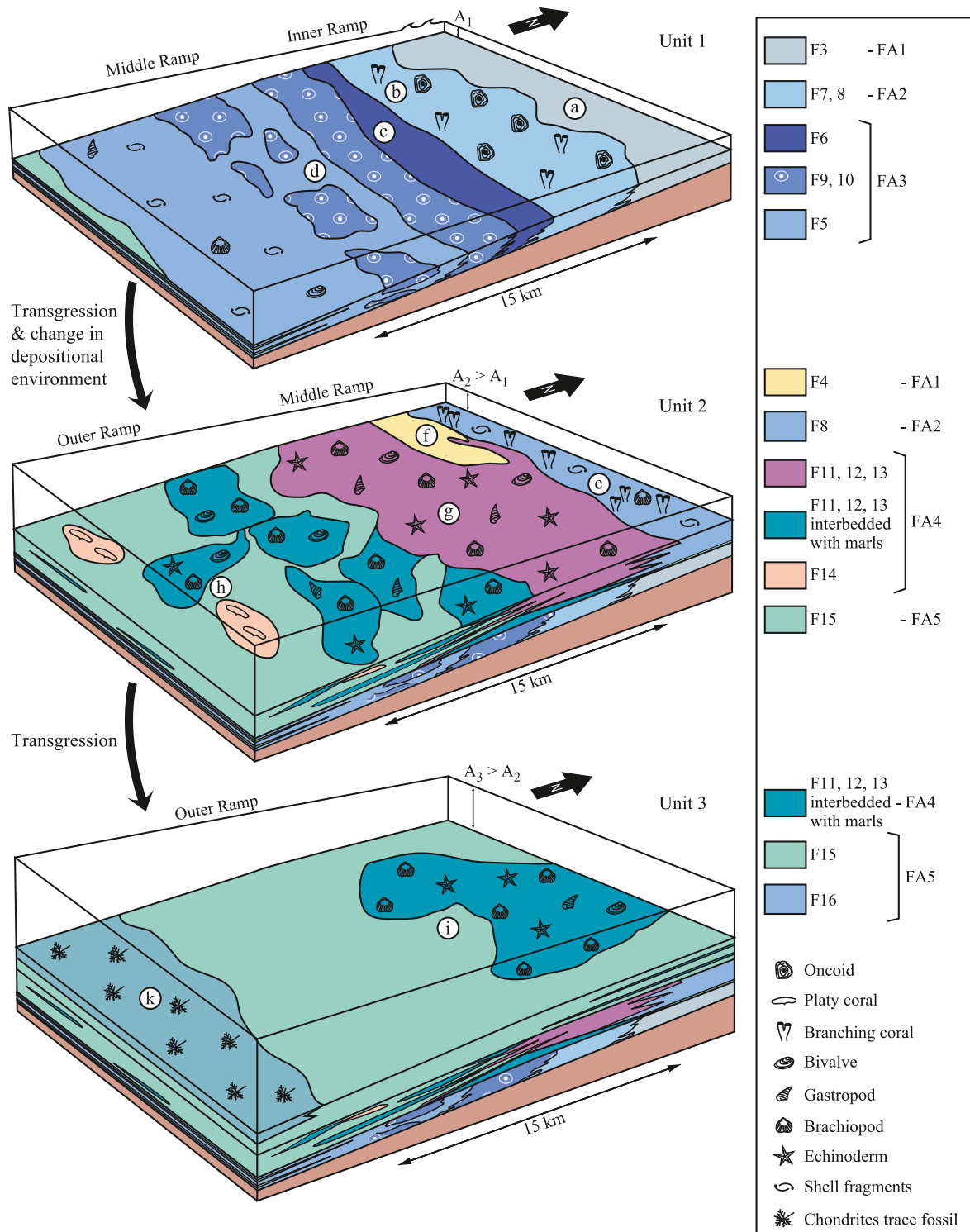

Fig. 11. Depositional environments of the Ouanamane Fm. for Units 1–3, with distribution of facies associations and key facies. Unit 1 depositional environments are based on observations from Tizgui N'Chorfa (a), Amsittène AMCA (b), Tikki (c), and facies associations from the center of the basin (d). Unit 2 environments are based on observations from Tikki and Tizgui N'Chorfa (e), Amsittène AME (f) and facies associations from the center of the basin (g, h). Unit 3 environments are based on Assif El Hade (k) and facies associations in the center of the basin (i).

dysaerobic conditions (Bromley and Ekdale, 1984; Gerard and Bromley, 2008). The organism responsible for *Chondrites* was systematically searching for nutrient-rich laminae and is very common in outer shelf to upper bathyal environments (Chamberlain, 1975). In summary, facies association 5 is characteristic of quiet and low-light conditions and was deposited in an outer ramp environment.

6. Vertical and lateral facies architecture

The three main stratigraphic units of the Ouanamane Fm. have been correlated around the basin, with boundaries defined by major hardground surfaces (Fig. 5). The detailed description of the stratigraphic succession and facies associations, together with analysis of facies variations across the basin, has allowed three depositional models to be established for each unit of the Ouanamane Fm. (Fig. 11). The vertical development is best understood in the center of the basin (Assif El Hade, Imouzzer, Tidili, Tizgui and Izwarn localities), where sections are most complete. Variations occur in the eastern and north-western sectors of the basin (Amsittène, Tikki, Tizgui N'Chorfa localities).

6.1. Unit 1: Iggui-n'Tarhazout oolitic member

The passage from the fluvial Ameskhoud Fm. to Unit 1 of the marine Ouanamane Fm. is transitional via the vertical superposition of Facies 1 to 3 (FA1; Fig. 7A). The succession was defined as the Transition Mb. by Adams et al. (1980) and is part of a general transgressive trend with the progressive establishment of fully marine conditions across the basin. It was only observed in the Tidili and Imouzzer localities; elsewhere the marine Ouanamane Fm. lies directly on the continental Ameskhoud Fm. (Fig. 5). A burrowed and iron-enriched firmground at the top of the last sandy dolomite bed marks the lithostratigraphic base of the Ouanamane Fm.

Unit 1 is dominated in the basin center (e.g. Tidili and Tizgui localities, Fig. 5) by oolitic and bioclastic deposits of facies association 3 that were deposited in the transition between inner and middle ramp (Fig. 5; Fig. 11, locations c and d). Further east (Assif El Hade locality, Fig. 5), inner ramp facies, partially protected by these shoals, are dominated by bioclastic limestones and small patch reefs (FA2; Fig. 11, location b). In the most proximal locations studied (e.g. Tikki, Fig. 1), the inner ramp passed to coastal environments represented by facies association 1 (Facies 3; Fig. 11, location a). Towards the west and northwest, fine-grained lithologies of facies associations 4 and 5 developed in Unit 1 (Amsittene and Izwarn localities, Fig. 5). On the southern flank of the Amsittène Anticline, these pass upwards rapidly to oolitic (FA3) and coral-rich inner ramp facies (FA2) (Amsittene locality, Fig. 5). In the northern section of the Amsittène Anticline (Fig. 1), oolitic-bioclastic grainstones (Facies 9) are overlain by a mixed siliciclastic-carbonate shoreface unit assigned to Facies 4 (FA1). The siliciclastic content decreases rapidly upwards, below the transition to Unit 2.

6.2. Unit 2: Somalirhychia limestone member

The transition from Unit 1 to Unit 2 is marked by the disappearance of ooids, a shift to brachiopod-rich carbonates and, in most localities, the development of a prominent hardground surface resting on highly bioturbated carbonates (*Thalassinoides* burrows). The surface itself is rich in iron nodules, encrusting fauna, perforated by small bivalves (Fig. 6B and C), and contains the first occurrence of a faunal association of *Pygurus* echinoids, serpulids, gastropods and coral fragments. The hardground provides a very good marker for basinwide correlations (Fig. 5). It represents a period of non-deposition that is characteristic of sea-level rise (Kendall and Schlager, 1981; Fürsich et al., 1992; Christ et al., 2015).

In the basin center, outer to middle ramp, brachiopod-rich carbonates (FA4) rest on the basal transgressive interval (Izwarn and Tizgui localities, Fig. 5; Fig. 11, location g). The faunal association is initially dominated by suspension-feeders and epifaunal grazers (*Collyrites* and *Pygurus* echinoids, gastropods, corals). Towards the upper part of the unit, with the disappearance of gastropods and echinoderms, the suspension-feeders become the only visible macrofauna. Facies association 4 passes downdip to marls alternating with beds rich in brachiopods, bivalves, rare ammonites and local buildups of platy coral boundstone (FA4 to 5; Amsittene locality, Fig. 5; Fig. 11, location h). Eastward, shallower conditions prevailed as evidenced by the interbedding of facies association 4 with facies association 2, a reduction in marls, and the development of branching coral patch reefs (Imouzzer and Assif El Hade localities, Fig. 5; Fig. 11, location e). Although deepening is indicated in the northwest of the basin (Amsittene locality, Fig. 5), beach strandplain deposits also record localized siliciclastic input around the Amsittène Anticline (locality AME; Facies 4; Fig. 7C; Fig. 11, location f).

6.3. Unit 3: marl and shale member

The transition from Unit 2 to Unit 3 is marked by firmgrounds and hardgrounds comparable to the transition between Units 1 and 2. The interval includes a very diverse fauna (bryozoans, crinoids, echinoids, corals, sponges, bivalves, brachiopods, gastropods and ammonites) that reflects more open marine influence. These deposits mark a strong change in the sedimentation from a fossiliferous unit to a marly unit.

Unit 3 is dominated by mud and marl (FA 5) and records a sensibly deeper environment than fossiliferous Unit 2 (Fig. 5; Fig. 11, location i). The deepest part of the ramp is characterized by marly lime-mudstones with *Chondrites* burrows indicating restricted, possibly dysaerobic conditions (Fig. 11, location k). Unit 3 is very uniform across the study area with little lateral variations (Fig. 5). It is however absent in the southern locality of the Amsittène Anticline, where ammonites collected from the underlying Unit 2 do not extend beyond the Middle Callovian, Anceps Zone (Stuebeli Subzone) (Fig. 5). In this locality, Unit 2 is overlain by a mixed siliciclastic-carbonate shoreface facies (Facies 4) that evolves upwards rapidly to the shallow-marine carbonates of the Iggui El Behar Fm. (Adams, 1980). The reefal Lalla Oujja Fm. (Duval-Arnould et al., this volume) is absent. The mixed siliciclastic-carbonate unit could thus be the lateral equivalent of Unit 3 and the Lalla Oujja Fm. on the margin of the basin, or Unit 3 and the Lalla Oujja Fm were eroded and these beach deposits belong to the basal Iggui El Behar Fm.

6.4. Depositional model and sequence stratigraphy

Vertical and lateral facies distribution represent the evolution of an open ramp system with gradual facies changes from high energy to low energy in a westward direction, and no evidence of break in slope (Fig. 11). Inner ramp facies association 2 may have been partly protected by high-energy facies in facies association 3. Facies variability in the proximal, eastern locations, is highest for Unit 1, whereas Units 2 and 3 are more uniform across the basin (Fig. 5). For Units 2 and 3, no eastern shoreline was identified in the studied sections, but the Amsittène anticline acted as a localized source area for siliciclastics throughout deposition of the Ouanamane Fm.

The vertical transition from the continental Ameskhoud Fm. through the 3 lithostratigraphic units of the Ouanamane Fm. reflects an overall transgression (Figs. 4 and 5). In this transgressive context, the depositional environment changed from oolite-dominated Unit 1 to macrofauna-dominated Unit 2, whereas the transition from Unit 2 to Unit 3 is mainly a change in bathymetry or nutrient availability. Lithostratigraphic boundaries have been drawn at hardground surfaces correlating between sections, and which represent major transgressive surfaces (Fig. 5). At basin scale, marls of facies association 5 (Unit 3) represent the deepest-water facies recorded, and a maximum flooding surface can be placed near the base of coral reefs of the overlying Lalla Oujja Fm. (Fig. 5). Reefs initiate in marls and lime-mudstones/wackestones of facies association 5. Internal reef facies develop from

microsolenid buildups (Facies 13) to branching coral buildups equivalent to Facies 14 (Duval-Arnould et al., this volume). These reefal buildups thus represent the start of basin-scale regression (Duval-l-Arnould et al., this volume). The Ouanamane Fm. thus represents a single transgressive systems tract, equivalent to the 2nd order Bathonian-Oxfordian transgression (Jacquin et al., 1998; Haq, 2018). Higher-order transgressive/regressive packages are likely present in Units 1 and 2 (Fig. 5), but these are difficult to correlate across the basin.

The mixed siliciclastic-carbonate facies observed on top of the Ouanamane Fm. in the Amsittène anticline (Fig. 5) represents a localized sequence boundary where shoreface sediments shifted downwards onto outer ramp facies. This is however spatially restricted to the anticline, whereas everywhere else, the top of the Ouanamane Fm. is a maximum flooding surface (Fig. 5).

The deposits of Unit 1 are similar to the ramp system of the Lower to Middle Jurassic Amellago transect in the Central High Atlas (Pierre et al., 2010), and to the Middle Jurassic carbonate ramp of the Lusitanian Basin (Azerêdo, 1998). Both systems present inner ramp deposits, with large oncoids, ooids and peloidal packstones, laterally evolving to inner/middle ramp deposits dominated by oolitic packstones and grainstone shoals. Units 2 and 3 present a facies distribution comparable to the Kimmeridgian of the Iberian Basin (Aurell et al., 1998). However, tempestite deposits are not as common in Morocco, and the coral buildups are better developed in the Iberian Basin.

7. Discussion

7.1. The Callovian-Oxfordian transition in the Atlantic and Tethyan realms

The ammonite data, supplemented by re-assessment of the brachiopod and foraminifera biostratigraphy, strongly suggest non-deposition for the Uppermost Callovian (Lamberti Zone pro parte) and Lower Oxfordian (Mariae and Cordatum zones pro parte). Sedimentologically, this study has provided evidence for low sedimentation rates (common development of hard- and firmgrounds and bioturbated bed tops) and localized erosion (siliciclastic input along the Amsittène Anticline) in the Ouanamane Fm. (Fig. 5), superimposed on a longer-term transgressive trend from Ameskhoud Fm. continental siliciclastics to outer ramp marls of the Ouanamane Fm. The Uppermost Callovian to Lower Oxfordian interval can therefore be considered as a condensed interval, possibly related to regression-transgression. Regression is supported by the sequence boundary along the Amsittène Anticline (Fig. 5).

On the conjugate continental margin in Canada, the shale-dominated Misaine Mb. of the Abenaki Fm. records a transgressive peak of Callovian age that can be identified as a seismically mappable maximum flooding surface (Fig. 12; Weston et al., 2012; Weissenberger et al., 2006). The lithostratigraphic units in Canada are bounded by flooding surfaces that could correlate to the transgressive surfaces identified in this study (Fig. 12). Although the Callovian transgression is therefore a major regional feature on both sides of the Central Atlantic Margin, there is currently no evidence for a hiatus in Canada due to lack of sufficiently resolved biostratigraphic data.

Episodes of non-deposition and hiatuses are however a common feature along Atlantic and mainly Tethyan margins during the Callovian to Oxfordian boundary interval (Fig. 13). Although continuous sedimentation occurred in the center of the Subalpine Basin (Fortwengler et al., 2013; Pellenard et al., 2014b), its western and southern margins do not record any Upper Callovian strata (Elmi, 1990; see Jacquin et al., 1998 for an overview). In Northern Switzerland, the Uppermost Callovian Henrici Subzone (= Poculum Subzone) is affected by a hiatus, and the Lower Oxfordian Mariae Zone is locally absent in most of the distal part of the basin (Gygi et al., 1998). In the Moray Firth rift system of the North Sea, the Lower Oxfordian is encompassed in a hiatus, which locally (on paleohighs) extends into the Middle Oxfordian (Davies et al., 1996). In the Iberian Basin (Northern Spain), the Lamberti Zone to Cordatum Zone pro parte is affected by a stratigraphic gap (Aurell et al., 1994; Meléndez et al., 2005). The Upper Bathonian-Lower Oxfordian record in the Subbetics of southern Spain is condensed (O'Dogherty et al., 2018). In the Lusitanian Basin (Azerêdo et al., 2002), the Upper Callovian (Lamberti Zone absent) to Middle Oxfordian (Mariae and Cordatum zones not recognized) unconformity is erosive. Although the Callovian–Oxfordian interval has been long interpreted as a global sea-level rise (Haq et al., 1988; Hallam, 2001; Wierzbowski et al., 2009), the examples above, from the Atlantic and Tethyan margins (Fig. 13), record smaller-scale intervals of sediment missing within that trend. These gaps have been linked to higher-order sea-level falls and more locally to tectonic movements. In the Lusitanian Basin, which is geographically closest to the EAB, the Middle to Late Jurassic discontinuity is associated to a forced regression (Azerêdo et al., 2002). In the Prebetics, weathering of shelf deposits was induced by a relative sea-level fall during the Middle to Late Jurassic (Reolid and Abad, 2018). Episodes of shallowing to emergence were also recorded the Iberian Basin (Aurell et al., 1994; Ramajo and Aurell, 2008), and in North Africa

Scotian Basin			Essaouira-Agadir Basin			Substages	Stages
Formation	Member	Lithology	Lithology	Member	Formation		
Abenaki	Baccaro				Iggui El Behar	Upper	Oxfordian
					Lalla Oujja	Middle	
						?	
						Upper	Callovian
	Misaine			Marl (Unit 3)	Ouanamane		
						Middle	
	Scatarie			Somalirhynchia (Unit 2)			
						Lower	
				Oolite (Unit 1)			

Fig. 12. Stratigraphic chart comparing simplified lithologies of the Scotian Basin and the Essaouira-Agadir Basin for the Callovian-Oxfordian interval. Scotian Basin stratigraphy adapted from Weston et al. (2012).
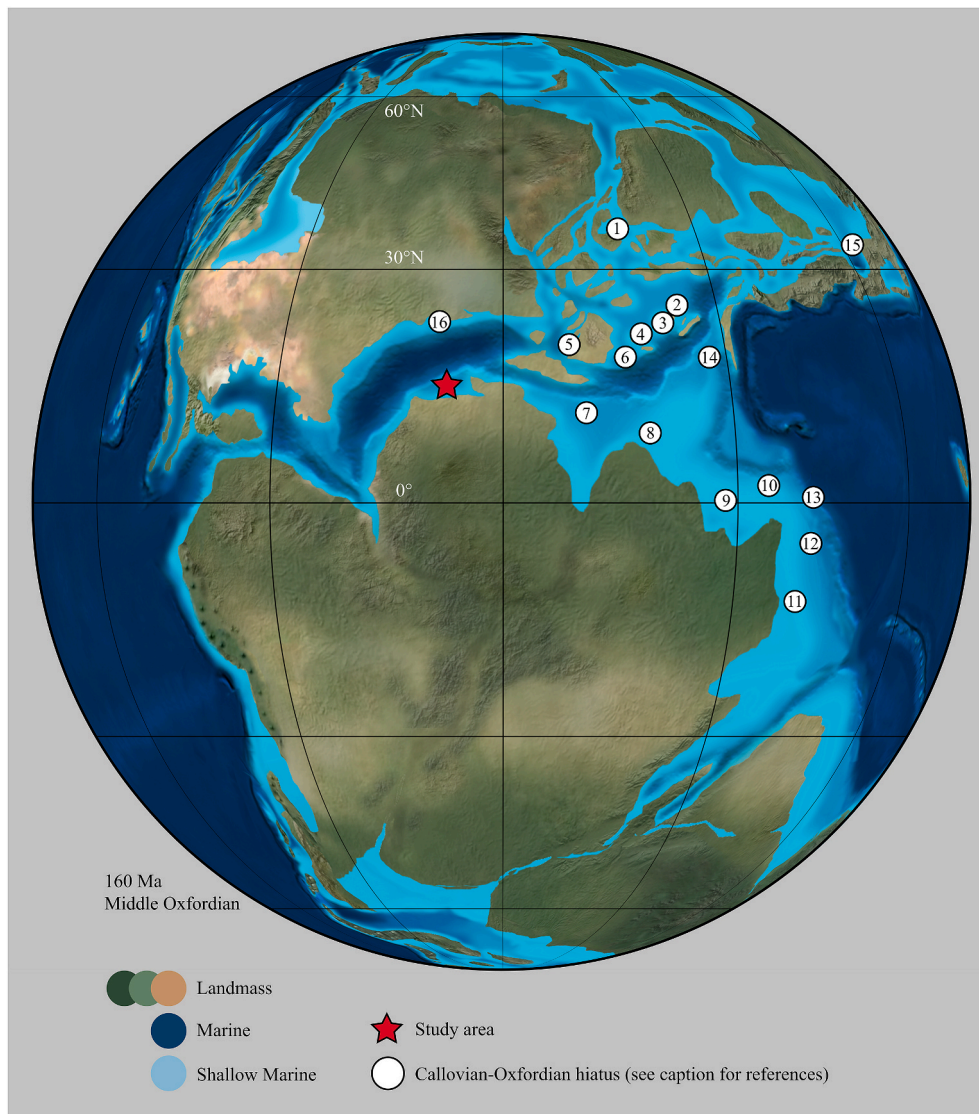
Fig. 13. Paleogeographic plate reconstruction at 160 Ma (Middle Oxfordian) from Colorado Plateau Geosystems Inc. Deep Time Maps™ (www.deeptimemaps.com). Locations of a documented Callovian-Oxfordian hiatus are indicated by the star (study area) and the numbered circles – 1: North Sea (Davies et al., 1996); 2: Switzerland (Gygi et al., 1998); 3: borders of the Subalpine Basin (Elmi, 1990; Jacquin et al., 1998; Fortwengler et al., 2013); 4: Iberian Basin (Aurell et al., 1994; Meléndez et al., 2005). 5: Lusitanian Basin (Azerêdo et al., 2002). 6: Subbetics (Reolid and Abad, 2018; O'Dogherty et al., 2018); 7: Algeria (Carr, 2003); 8: Tunisia/Libya (Walley, 1985; Mette, 1997); 9: Egypt (Carr, 2003); 10: Levant (Gardosh et al., 2011); 11: Saudi Arabia (Manivit et al., 1990; Al-Mojel et al., 2020); 12: Kuwait (Kadar et al., 2015); 13: NW Iran (Bayet-Goll et al., 2022); 14: Adriatic platform (Husinec et al., 2022); 15: Amu Darya Basin (Carmeille et al., 2020). The Scotian Basin is at number 16 (Weston et al. (2012).

(Walley, 1985; Mette, 1997; Carr, 2003).

A particularly good comparison is possible with the well-studied Callovian–Oxfordian succession on the Arabian Plate. The relative tectonic quiescence of Arabia during deposition means that sequences can be regionally correlated and may have a eustatic driver (Simmons et al., 2020; Al-Mojel and Razin, 2022). On the other hand, the formation of Middle and Late Jurassic intrashelf basins (Murris, 1980) may have been at least partially controlled by tectonic events (Bayet-Goll et al., 2022), although differential accumulation rates appear to be the primary driver (Gravestock et al., 2020). Due to the cyclic nature of Arabian Jurassic stratigraphy, sequence stratigraphic interpretation provides a valuable means of correlation and determining the spatial and temporal location of key surfaces and facies trends, and thus constraining the underlying driving mechanisms (Fig. 14).

Across much of Arabia, the Middle Bathonian, and possibly the Late Bathonian to Early Callovian is represented by a hiatus associated with the J40 sequence boundary (Hirsch et al., 1998; Kadar et al., 2015;

Al-Mojel and Razin, 2022; Al-Mojel et al., 2020; Bayet-Goll et al., 2022). A major Callovian transgression re-established a carbonate platform across much of the Arabian Plate. This is represented by the Uweinat Mb. of the Araej Fm. in many parts of the southern Gulf and can be correlated with confidence to the outcrops of Saudi Arabia, where the equivalent carbonates are called the Atash Mb. of the Dhruma Fm. (Al-Mojel and Razin, 2022). Transgression continued into the Tuwaiq Mountain Fm. and its equivalents (Fig. 14). The J40 maximum flooding surface is best picked near the base of the limestone unit T3, which is supported by both the regional stratigraphic architecture and faunal content of the limestone (Manivit et al., 1990; Al-Mojel and Razin, 2022). This maximum flooding surface is associated with ammonites of the Middle Callovian Coronatum Zone (Fig. 14; Énay et al., 2009; Al-Mojel and Razin, 2022), and can be traced into the Levant margin and Yemen (Gill et al., 1985; Gardosh et al., 2011; Howarth and Morris, 1998). It likely corresponds to the flooding surface separating Units 2 and 3 of the Ouananame Fm., constrained by Coronatum Zone ammonites (Fig. 5, marker D).
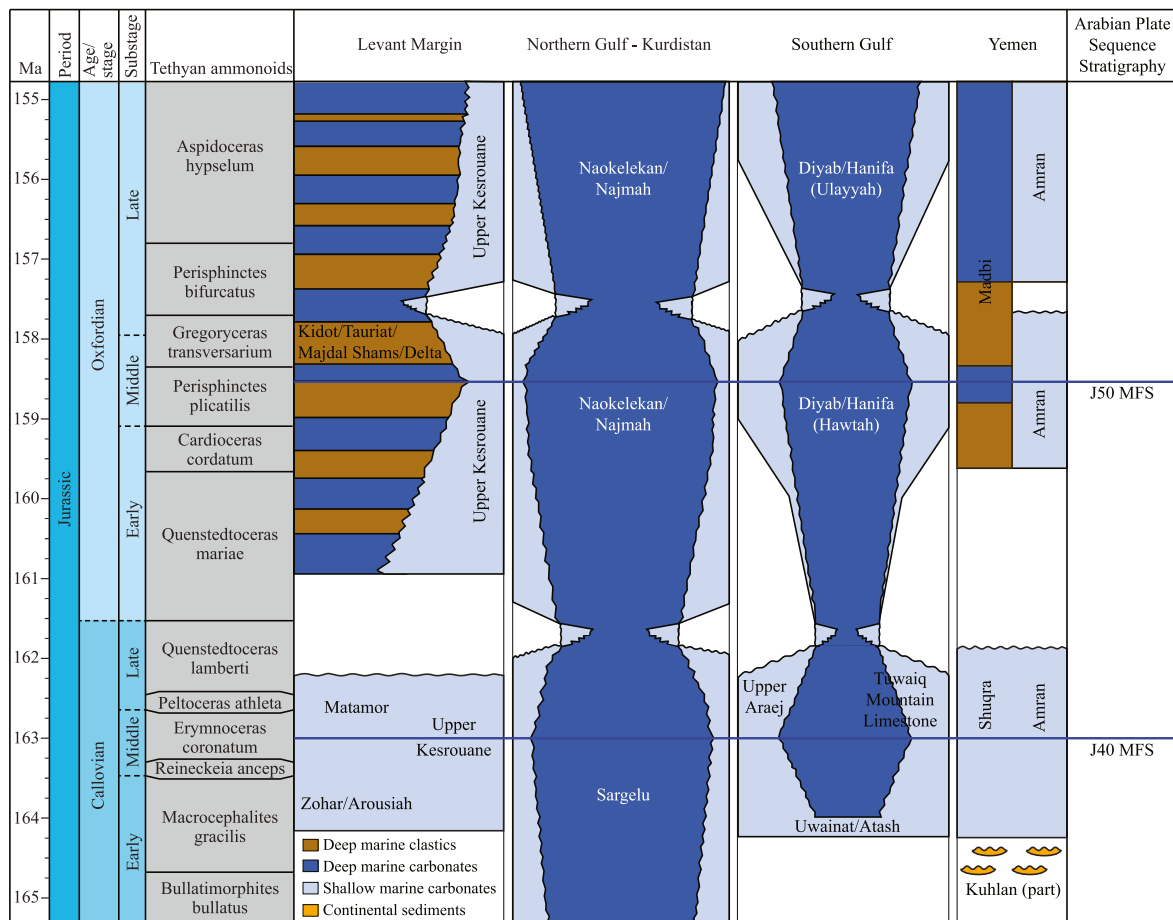
Fig. 14. Chronostratigraphic chart summarising the sequence stratigraphy and facies distribution of the Callovian–Oxfordian succession of the Arabian Plate. Geological timescale after Gradstein et al. (2020) and sequence stratigraphy modified after Sharland et al. (2001). See text for details of age control vs lithostratigraphy. Lowstand deposits are symbolically represented.

The top of the Tuwaiq Mountain Limestone Fm. in proximal locations (e.g. at outcrop in Saudi Arabia) has long been regarded as an important sequence boundary (J50 sequence boundary; Manivit et al., 1990; Sharland et al., 2001). Late Callovian and much of Early Oxfordian strata (Mariae Zone) can be absent (Fig. 14), although a correlative unconformity extends downdip into the intrashelf basin, which includes sections with Late Callovian (Athleta and Lamberti zones) ammonites (Al-Mojel et al., 2020; Al-Mojel and Razin, 2022) (Fig. 14). This correlative conformity can also be picked at the base of calciturbidites of the Najmah Fm. in Kuwait (Kadar et al., 2015), and at the boundary between the Sargelu and Najmah formations in the Zagros (Bayet-Goll et al., 2022) (Fig. 14). On the Levant margin, an exposure surface at the top of the Zohar Fm. probably relates to the J50 sequence boundary (Picard and Hirsch, 1987; Gardosh et al., 2011). Following deposition and subsequent exposure of the Tuwaiq Mountain Limestone and its equivalents over the central parts of the Arabian Plate, the Hanifa Fm. represents a new cycle of deposition. A maximum flooding surface can be located within open marine carbonates of the Hawtah Mb. at outcrop in Saudi Arabia. Ammonite and nannofossil biostratigraphy indicate that this maximum flooding surface lies within the Plicatilis Zone of the Middle Oxfordian (Fig. 14; Énay et al., 1987). Plicatilis Zone ammonites were also recorded in the transgressive (although associated with rift-driven subsidence) Madbi Shale of Yemen (Howarth and Morris, 1998). On the Levant margin it probably lies within the open marine shales and carbonates of the Kidod/Tauriat, Majdal Shams and Delta formations (Fig. 14; Hirsch et al., 1998; Gardosh et al., 2011). This surface corresponds to the base of the Lalla Oujja Fm. buildups in Morocco (Fig. 5, marker E), which reflects maximum transgression in the EAB.

The Callovian-Oxfordian transgressive and regressive events on the Arabian Plate, especially in the reasonably well dated stratigraphy outside of the intrashelf basins, thus closely match the record in the EAB. In particular, the Middle Callovian transgression is followed by a hiatus that can span much of the Late Callovian and Early Oxfordian, before renewed transgression leads to a Middle Oxfordian sea-level maximum (Fig. 14). This is also reflected in the latest revised global sea-level curve (Haq, 2018).

7.2. Global environmental change in the Callovian-Oxfordian transition

The widespread nature of these Callovian-Oxfordian depositional trends and their correlation along the Gondwanan margin of Neotethys and to the Atlantic realm (Fig. 13) suggest eustasy as a significant driver. Isotopic thermometric studies indicate that the Late Callovian to Early Oxfordian was subject to global cooling (Dromart et al., 2003a, 2003b; Cecca et al., 2005; Pellenard et al., 2014b; Andrieu et al., 2016; Price et al., 2023). Limited growth of polar ice caps may have caused the observed regression, and lower pCO_2 during this cooling episode would have led to reduced carbonate production. A period of global cooling is also supported by clay mineral data from the Paris Basin, where the proportion of illite and mixed illite/smectite layers increased at the expense of kaolinite during the Upper Callovian and Lower Oxfordian (Pellenard and Deconinck, 2006). Neodymium isotopes further support a change in seawater temperature in the Callovian–Oxfordian transition (Dera et al., 2015). The observed negative Neodymium isotope excursion has been interpreted to reflect circulation of Arctic waters into the northwestern Tethys, and it is possible that these waters also reached the
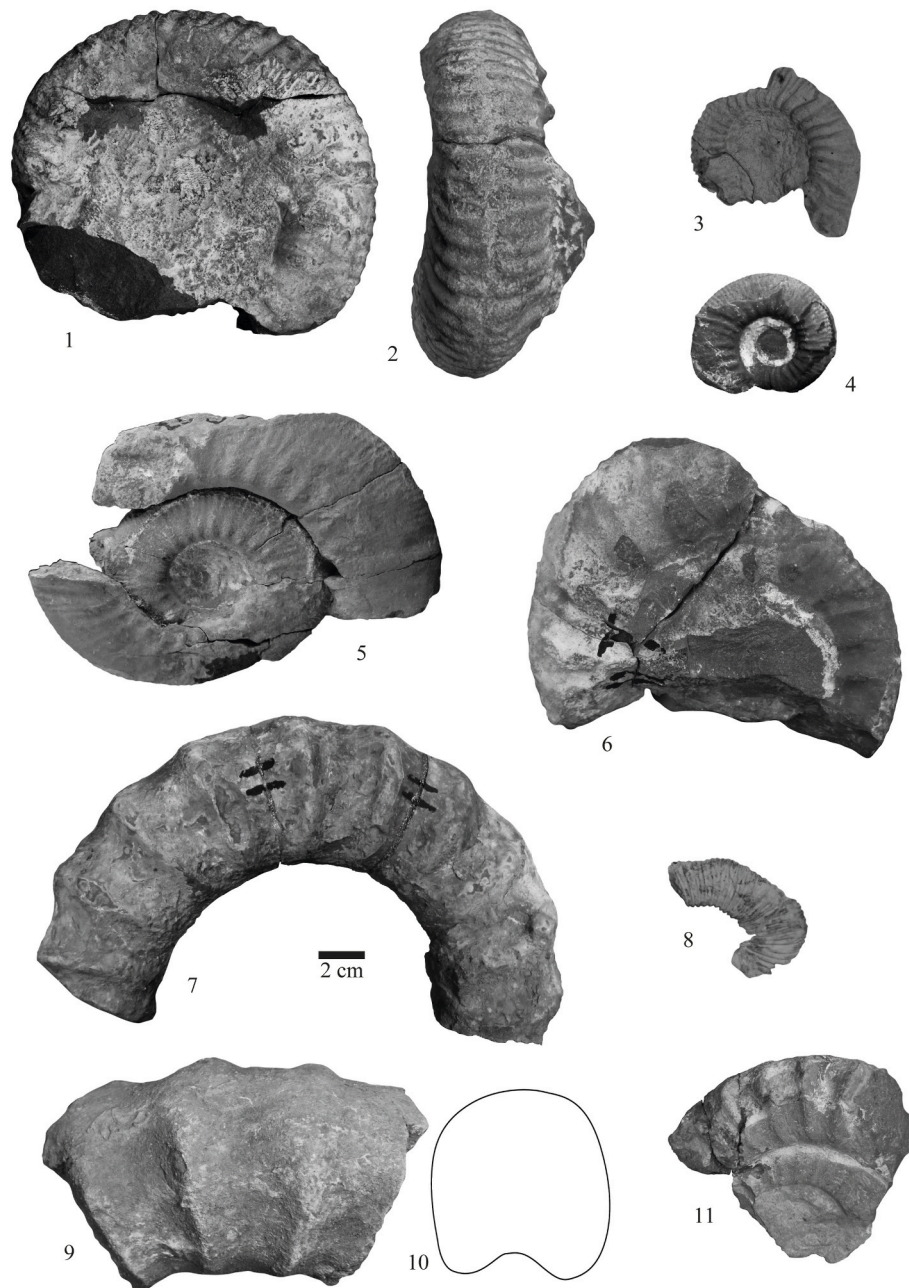
Plate 1. 1. *Reineckeia (R.) stuebeli* Steinmann, 1881; AMCA–II–25 **2**. *Reineckeia (R.) stuebeli* Steinmann, 1881; AMCA–II–25 - macroconch **3**. *Reineckeia (R.) stuebeli* Steinmann, 1881; AMCA II 26 – microconch **4**. *Indosphinctes* aff. *choffati* Parona and Bonarelli, 1897; AMCA-P26C **5**. *Indosphinctes patina* Neumayr, 1870; AMCA-P27A **6**. *Reineckeia (R.)* gr. *anceps* Reinecke, 1818; AMCA-P27C **7**. C*hoffatia waageni* Teysseire, 1889; IZW-A1-P1004, **8**. *Dichotomosphinctes* sp.; CG-1, **9**. *Subgrossouvria aberrans* Waagen, 1875; TDE-P-1 **10**. *Subgrossouvria aberrans* Waagen, 1875; TDE-P-1 - section **11**. *Choffatia waageni* Teysseire, 1889; IZW-A1-P1005.

Central Atlantic. Following the cooling, temperatures recovered in the Middle Oxfordian (Dromart et al., 2003b; Cecca et al., 2005). Palynological data from the Boreal province support a transition to a warmer and more arid climate during the Middle Oxfordian (Abbink et al., 2001), and a positive Neodymium isotope excursion in the Middle Oxfordian has been linked to a widening of the Hispanic corridor and potential ocean current destabilization by warmer bottom waters (Dera et al., 2015).

The Callovian-Oxfordian interval equally saw significant perturbations in the carbon cycle. The cooling recorded at the Middle-Late Jurassic boundary was preceded by a positive excursion of δ^{13}C in the Middle Callovian (Coronatum Zone), which correlates with a short warming episode and the accumulation of organic-rich facies worldwide (Dromart et al., 2003a, 2003b; Pellenard et al., 2014b; Carmeille et al.,

2020; Price et al., 2023). In Morocco, dysoxic conditions as indicated by *Chondrites*-rich dark grey limestones of Facies 16, are constrained between the Coronatum zone and the Middle Oxfordian, and thus possibly correspond to this event.

7.3. Timing of Oxfordian buildup development

The biostratigraphic analysis has consistently shown a Middle Oxfordian fauna underlying the Lalla Oujja Fm. coral buildups, indicating a Middle Oxfordian age for the base of the buildups. This applies to the center of the basin, between Tadrart and Izwarn, but equally to the Cap Ghir Anticline, where the base of the first thick coral buildups has been dated to the Middle Oxfordian. Although small platy coral buildups already appeared during the Callovian (Cap Ghir and Tizgui localities,
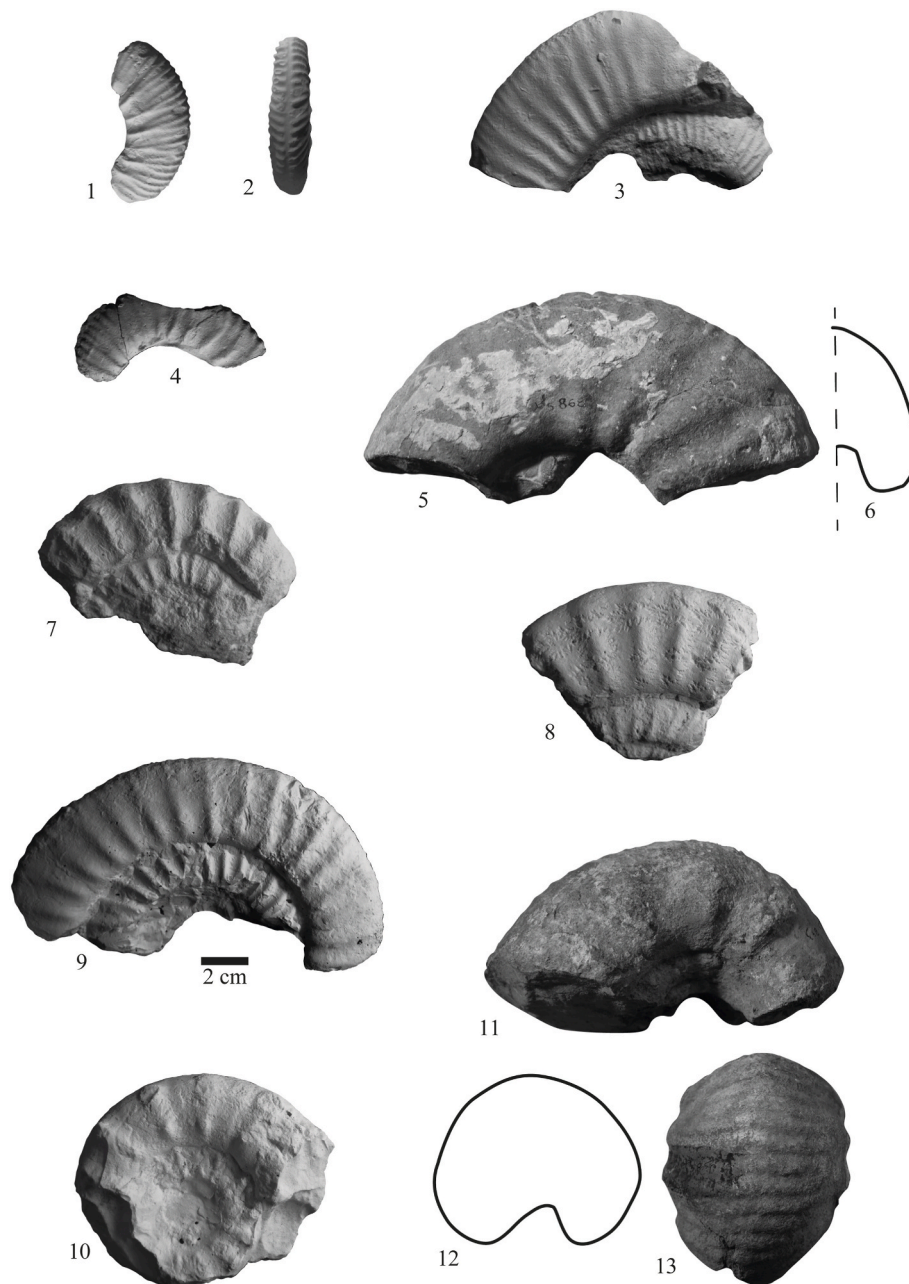
Plate 2. 1. *Reineckeia (R.) stuebeli* Steinmann, 1881; AMCA–II–25 – macroconch 2. *Reineckeia (R.) stuebeli* Steinmann, 1881; AMCA–II–25 – macroconch 3. *Arisphinctes plicatilis* Sowerby, 1817 transferred to *Liosphinctes* by Głowniak (2002); MEM-AM-997 4. *Hamulisphinctes* of the *hamulatus* group Buckman, 1921; MEM-AM-993 5. *Erymnoceras baylei* Jeannet *sensu* Gill et al. (1985); MEM-AM-955- macroconch 6. *Erymnoceras baylei* Jeannet *sensu* Gill et al. (1985); MEM-AM-955-section 7. *Choffatia waageni* Teysseire, 1889; MEM-AM-999b-c 8. *Choffatia waageni* Teysseire, 1889; MEM-AM-999b-c 9. *Subgrossouvria* that recalls *S. isabellae* Bonnot et al., 2008 and S. samatrensis Spath, 1931; MEM-AM-1001 10. Euaspidoceras subbabeanum (Sinzow) *sensu* Jeannet (1951) and Bonnot (1995); MEM-AM-1007 11. *Erymnoceras baylei* Jeannet *sensu* Gill et al. (1985); MEM-AM-1425 - macroconch 12. *Erymnoceras baylei* Jeannet *sensu* Gill et al. (1985); MEM-AM-1425 – section 13. *Erymnoceras baylei* Jeannet *sensu* Gill et al. (1985); MEM-AM-1425 - macroconch.

Fig. 3), they never reached thicknesses of more than a few tens of meters, which could be explained by the likely lower seawater temperatures in the Callovian (Dromart et al., 2003a, 2003b) favoring more temperate or cooler faunal assemblages.

Equivalent Oxfordian buildups were identified in the Cap Juby oil field of the offshore Tarfaya Basin (Morabet et al., 1998; Davison, 2005), and in the Deep Panuke gas field offshore Nova Scotia (Ellis et al., 1985; Weissenberger et al., 2006). Stratigraphically, the Deep Panuke reservoir belongs to the Baccaro Mb. of the Abenaki Fm., which is generally dated Kimmeridgian to Tithonian. The revised biostratigraphy in Morocco would indicate that the base of the Baccaro Mb. buildups is

Middle Oxfordian, slightly older than previously thought (Fig. 12).

The basal Middle Oxfordian buildups, which are synchronous across the outcrops in the basin, are normally followed by younger buildup generations (Martin-Garin et al., 2007; Olivier et al., 2012; Duval-Arnould et al., this volume). This implies a rapid development of the buildups across the basin and a kilometer-scale lateral extent of the buildup unit. Buildup development likely related to basin-scale changes in the environmental conditions, albeit modulated by local accommodation changes (Duval-Arnould et al., this volume).

The most recent global sea-level curves (Haq, 2018) indicate the Middle Oxfordian as the onset of the Upper Jurassic transgression. As

Plate 3. 1. *Euaspidoceras*; TIDILI-2 2. *Euaspidoceras*; TIDILI-2 3. *Euaspidoceras*; TIDILI-2 4. *Euaspidoceras*; TIDILI-2 5. *Euaspidoceras*; TIDILI-2.

already mentioned, the Latest Callovian to Early Oxfordian interval was marked by decreasing temperatures, pCO$_2$, and sea level (Dromart et al., 2003a, 2003b; Cecca et al., 2005; Pellenard et al., 2014b; Price et al., 2023). A renewed rise of these global factors in the Middle Oxfordian (Dromart et al., 2003a, 2003b) likely played a major role in driving coral development across Europe, North Africa, Middle East and Central Asia (Cecca et al., 2005). Global coral buildup recovery was relatively rapid in the Middle and Upper Oxfordian (Cecca et al., 2005).

8. Conclusions

This study provides a refined stratigraphic framework for the late Middle to early Upper Jurassic carbonate platform of the Moroccan Atlantic basin and identifies regional facies variations and controls on the evolution of the depositional environment. It presents the following

new observations and interpretations of the stratigraphy and faunal assemblages in the Essaouira-Agadir Basin.

(1) Integrated biostratigraphic analysis indicates an Early Callovian age for the base of the Ouanamane Fm., reassessing and discounting a Late Bathonian age proposed previously. The age of the three members of the Ouanamane Fm. has been refined. An Early Callovian age is indicated for the Iggui n'Tarhazout Oolite Member (Unit 1). The *Somalirhynchia* Limestone Mb. (Unit 2) is bracketed between the Early Callovian Gracilis Zone (Patina Subzone) and the Middle Callovian Coronatum Zone (Baylei Subzone). The Marl Mb. (Unit 3) is inferred to range from the Middle Callovian (Coronatum Zone) to the late Early to Middle Oxfordian (upper Cordatum to lower Plicatilis zones), with a

hiatus/condensation interval that spans the Latest Callovian and most of the Lower Oxfordian.

(2) The new high-resolution biostratigraphic correlation allows reconstruction of basin-scale variations in depositional environments. The facies distribution indicates deposition on a west-dipping ramp. Overall, the Ouanamane Fm. reflects progressive deepening of environments during the Callovian to Oxfordian global sea level rise. Distinct flooding surfaces (hard- and firm-grounds) form the boundaries between the three members of the Ouanamane Fm.

(3) The Iggui-n'Tarhazout Oolite Mb. is dominated by oolitic and bioclastic inner to middle ramp deposits across the center of the basin. It is absent in the more proximal eastern part, where it passes to sandy carbonates and shoreface deposits. Towards the north of the basin, the inner ramp deposits are platform patch reefs and oncoidal deposits partly protected by oolitic shoals. The overlying *Somalirhynchia* Limestone Mb. is dominated by brachiopod-rich outer to middle ramp deposits at the scale of the basin. Finally, the Marl Mb. consists of outer ramp marls with rare brachiopod-rich floatstones.

(4) Where dating by ammonites is possible, the base of the overlying Lalla Oujja Fm. coral buildups is always of Middle Oxfordian age, suggesting a synchronous establishment of coral facies across the EAB.

(5) The Middle to Upper Jurassic succession of the EAB records the major mid-Callovian and mid-Oxfordian transgressive pulses that are recognized at the global scale. There is no evidence of a late Callovian global sea-level rise, but instead the EAB succession records a major regressive event close to the Callovian–Oxfordian boundary, consistent with other sections in the Atlantic and Tethyan realms that may reflect a global cooling event associated to the Middle-Late Jurassic transition.

CRediT authorship contribution statement

Aude Duval-Arnould: Writing – original draft, Investigation, Conceptualization. **Luc Bulot:** Writing – original draft, Supervision, Investigation, Conceptualization. **Moussa Masrour:** Investigation. **Mike Simmons:** Writing – review & editing, Writing – original draft, Investigation. **Alain Bonnot:** Writing – review & editing, Investigation. **Rémi Charton:** Writing – review & editing, Investigation. **Jonathan Redfern:** Writing – review & editing, Supervision, Investigation, Funding acquisition. **Stefan Schröder:** Writing – review & editing, Supervision, Investigation, Conceptualization.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

Data will be made available on request.

Acknowledgements

References

Abbink, O., Targarona, J., Brinkhuis, H., Visscher, H., 2001. Late jurassic to earliest cretaceous palaeoclimatic evolution of the southern North Sea. Global Planet. Change 30, 231–256.

Adams, A.E., 1979. Sedimentary environments and palaeogeography of the western high Atlas, Morocco, during the middle and late jurassic. Palaeogeogr. Palaeoclimatol. Palaeoecol. 28, 185–196.

Adams, A.E., 1980. The stratigraphy and sedimentology of a Jurassic marine transgression, Western High Atlas, Morocco. Geol. Mediterr. 7, 223–231.

Adams, A.E., Ager, D.V., Harding, A.G., 1980. Géologie de la région d'Imouzzer des Ida-ou-Tanane (Haut-Atlas Occidental). Notes et Mémoires du Service géologique du Maroc 41, 59–80.

Ager, D.V., 1965. The adaptation of Mesozoic brachiopods to different environments. Palaeogeogr. Palaeoclimatol. Palaeoecol. 1, 143–172.

Ager, D.V., 1971. Space and time in brachiopod history. In: Middlemiss, F.A., Rawson, P. F., Newall, G. (Eds.), Faunal Provinces in Space and Time, vol. 4. Geological Journal Special Issue, pp. 95–110.

Ager, D.V., 1974. Storm deposits in the jurassic of the Moroccan high Atlas. Palaeogeogr. Palaeoclimatol. Palaeoecol. 15, 83–93.

Alméras, Y., Benigni, C., Tintori, A., 1991. New brachiopod faunas from the middle jurassic of zanskar (Ladakh, India). Riv. Ital. Paleontol. Stratigr. 96, 427–460.

Alméras, Y., Cougnon, M., 2013. Les Brachiopodes jurassiques (Spiriferida et Rhynchonellida). Principaux genres et leur évolution. Les espèces, extensions verticales et répartitions géographiques. Doc. Lab. Geol. Lyon 170, 1–227.

Alméras, Y., Cougnon, M., Enay, R., Mangold, C., 2010. Brachiopodes du Jurassique inférieur et moyen d'Arabie Saoudite centrale. Paléontologie, Biostratigraphie et Paléoenvironnements. Échelles chronostratigraphiques. Doc. Lab. Geol. Lyon 168, 1–247.

Al-Mojel, A., Razin, P., 2022. Sequence stratigraphy and facies analysis of the NE Gondwana Middle Jurassic inner-platform mixed carbonate-siliciclastic succession, Saudi Arabia. Arabian J. Geosci. 15, 820.

Al-Mojel, A., Razin, P., Dera, G., 2020. High-resolution sedimentology and sequence stratigraphy of the oxfordian- kimmeridgian, Hanifa, jubaila and arab outcrops along jabal Tuwaiq, Central Saudi Arabia. J. Afr. Earth Sci. 165, 103803.

Ambroggi, R., 1963. Étude géologique du versant méridional du Haut-Atlas Occidental et de la plaine de Souss. Notes et Mémoires du Service géologique du Maroc, Rabat 157, 1–321.

Andrieu, S., Brigaud, B., Barbarand, J., Lasseur, E., Saucède, T., 2016. 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. Sediment. Geol. 345, 54–84.

Atrops, F., Meléndez, G., 1993. Current trends in systematics of jurassic ammonoidea: the case of oxfordian-kimmeridgian perisphinctids from southern Europe. Geobios, M.S. 15, 19–31.

Aurell, M., Bádenas, B., Bosence, D., Waltham, D.A., 1998. Carbonate production and o! shore transport on a Late Jurassic carbonate ramp (Kimmeridgian, Iberian basin, NE Spain): evidence from outcrops and computer modelling. In: Wright, V.P., Burchette, T.P. (Eds.), Carbonate Ramps, vol. 149. Geological Society, London, Special Publications, pp. 137–161.

Aurell, M., Fernández López, S., Meléndez, G., 1994. The Middle-Upper Jurassic oolitic ironstone bed in the Iberian Range (Spain), eustatic implications. Geobios 17, 549–561.

Azerêdo, A.C., 1998. Geometry and facies dynamics of Middle Jurassic carbonate ramp sandbodies, west-central Portugal. In: Wright, V.P., Burchette, T.P. (Eds.), Carbonate Ramps, vol. 149. Geological Society, London, Special Publications, pp. 281–314.

Azerêdo, A.C., Wright, V.P., Ramalho, M.M., 2002. The Middle-Late Jurassic forced regression and disconformity in central Portugal: eustatic, tectonic and climatic effects on a carbonate ramp system. Sedimentology 49, 1339–1370.

Banner, F.T., Whittaker, J.E., 1991. Redmond's" new lituolid foraminifera" from the Mesozoic of Saudi Arabia. Micropaleontology 41–59.

Banner, F.T., Simmons, M.D., Whitaker, J.E., 1991. The mesozoic chrysalidinidae foraminifera, (textulariacea) of the Middle East: the Redmond aramco taxa and their relatives. Bulletin of the British Museum of Natural History (Geology) 47, 101–152.

Bardhan, S., Datta, K., Chanda, P., Mallick, S., 2012. Systematic revision and sexual dimorphism in *Choffatia* (ammonoidea: perisphinctoidea) from the callovian of kutch, India. Palaeoworld 21, 29–49.

Bassoullet, J.-P., 1997. Les grands foraminifères. In: Cariou, E., Hantzpergue, P. (Eds.), Biostratigraphie du Jurassique ouest européen et méditerranéen. Zonation parallèle et distribution des invertébrés et microfossiles. Bulletin des Centres de Recherches Exploration-Production Elf-Aquitaine, Pau, Mémoire, vol. 17, 293-30.

Bassoullet, J.-P., Colchen, M., 1987. Découverte d'*Archaeosepta platierensis* foraminifère du dogger alpin, Himalaya du Ladakh (Inde). Geobios 20, 549–555.

Bayet-Goll, A., Daraei, M., Seginsara, M.I., 2022. Palaeogeographic reconstruction and sequence architecture of the middle-upper Jurassic successions of Hawraman Basin (NW Iran): implications for tectono-depositional processes of the northeastern passive margin of the Arabian Plate. Geol. J. 57, 2058–2093.

Ben Ismail, M., Bouaziz, S., Alméras, Y., Clavel, B., Donze, P., Enay, R., Ghenmi, M., Tintant, H., 1989. Nouvelles données biostratigraphiques sur le Callovien et les

faciès "purbecko-wealdiens" (Oxfordien à Vraconien) dans la région de Tataouine (Sud-tunisien). Bulletin de la Société Géologique de France, 8ème série 5, 353–360.

Bonnot, A., 1995. Les Aspidoceratidae (Ammonitina) en Europe Occidentale, au Callovien supérieur et à l'Oxfordien inférieur. Thèse de Doctorat de l'Université de Dijon, 1–487.

Bonnot, A., Boursicot, P.-Y., Ferchaud, P., 2014. Les genres *Subgrossouvria* Spath et *Orionoides* Spath (Ammonitina, Perisphinctidae) de l'Horizon à Leckenbyi (Callovien supérieur, Zone à Athleta) de Montreuil-Bellay (Maine-et-Loire, France). Carnets Géol. 14, 351–399.

Bouaouda, M.S., 2002. Micropaléontologie de la plate-forme du Bathonien-Oxfordien des régions d'Imi'N-Tanout et du Jbilet Occidental (Maroc), Essai de biozonation. Rev. Paléobiol. 21, 223–239.

Bouaouda, M.S., 2004. Le bassin atlantique marocain d'El Jadida-Agadir : stratigraphie, paléogéographie, géodynamique et biostratigraphie de la série Lias-Kimméridgien. Unpubl. Thesis (Doct. Etat) Univ. Mohamed V Rabat, p. 208.

Bouaouda, M.S., 2006. Découverte de *Sievoides kocyigiti* Farinacci & Ekmekci dans le Jurassique supérieur du bassin d'Essaouira (Maroc). Bulletin de l'Institut Scientifique, Rabat 28, 1–7.

Bouaouda, M.S., 2007. Lithostratigraphie, Biostratigraphie et Micropaléontologie des formations du Lias au Kimméridgien du bassin atlantique marocain d'El Jadida-Agadir, travaux de l'Institut Scientifique. Rabat, série géologie et géographie physique no. 22.

Bouaouda, M.S., 2009. Distribution de *Sievoides kocyigiti* Farinacci & Ekmekci (foraminifère benthique) dans le Jurassique supérieur du bassin d'Agadir (marge atlantique marocaine). Rev. Paléobiol. 28, 33–42.

Bouaouda, M.S., Barattolo, F., Mohammed-Rida Kharrim, M.-R., El Kamar, A., 2009. Distribution de *Megaporella boulangeri* Deloffre et Beun, 1986 (algue dasycladale) dans le Jurassique du bassin atlantique marocain. Rev. Micropaleontol. 52, 107–122.

Bouaouda, M.S., Peybernès, B., Boutakiout, M., 2004. Foraminifbres benthiques complexes du Bathonien supérieur au Kimméridgien inférieur de la marge atlantique marocaine («Bassin» d'Agadir): stratigraphie et paléobiogéographie. Rev. Micropaleontol. 47, 13–27.

Bouatmani, R., Medina, F., Aït Salem, A., Hoepffner, C., 2003. Thin-skin tectonics in the Essaouira basin (western High Atlas, Morocco): evidence from seismic interpretation and modelling. J. Afr. Earth Sci. 37, 25–34.

Boudagher-Fadel, M.K., 2018. Evolution and Geological Significance of Larger Benthic Foraminifera. UCL press, p. 693.

Bromley, R.G., Ekdale, A.A., 1984. Trace fossil preservation in flint in the European chalk. J. Paleontol. 298–311.

Burchette, T.P., Wright, V.P., 1992. Carbonate ramp depositional systems. Sediment. Geol. 79, 3–57.

Burchette, T.P., Wright, V.P., Faulkner, T.J., 1990. Oolitic sandbody depositional models and geometries, Mississippian of southwest Britain: implications for petroleum exploration in carbonate ramp settings. Sediment. Geol. 68, 87–115.

Cariou, E., Hantzpergue, P., 1997. Biostratigraphie du Jurassique Ouest-Européen et Méditerranéen. Zonations parallèles et distribution des invertébrés et microfossiles. Bulletin des Centres de Recherches Exploration–Production Elf-Aquitaine, Pau, Mémoire 17, 1–422.

Cariou, E., Meléndez, G., 1990. A modified perisphinctid zonation for the middle oxfordian of southern Europe, submediterranean province. First oxfordian meeting, zaragoza. Publicaciones del Seminario de Paleontología de Zaragoza (Sepaz) 2, 129–151.

Carmeille, M., Bourillot, R., Pellenard, P., Dupias, V., Schnyder, J., Riquier, L., Mathieu, O., Brunet, M.-F., Enay, R., Grossi, V., Gaborieau, C., Razin, P., Visscher, P. T., 2020. Formation of microbial organic carbonates during the late jurassic from the northern Tethys (Amu Darya Basin, Uzbekistan): implications for jurassic anoxic events. Global Planet. Change 186, 103127.

Carr, I.D., 2003. A sequence stratigraphic synthesis of the North African Mesozoic. J. Petrol. Geol. 26, 133–152.

Cecca, F., Garin, B.M., Marchand, D., Lathuiliere, B., Bartolini, A., 2005. Paleoclimatic control of biogeographic and sedimentary events in Tethyan and peri-Tethyan areas during the Oxfordian (Late Jurassic). Palaeogeogr. Palaeoclimatol. Palaeoecol. 222, 10–32.

Chamberlain, C.K., 1975. Trace fossils in DSDP cores of the Pacific. J. Paleontol. 1074–1096.

Charton, R., Bertotti, G., Duval-Arnould, A., Storms, J.E.A., Redfern, J., 2021. Low-temperature thermochronology as a control on vertical movements for semi-quantitative source-to-sink analysis: a case study for the Permian to Neogene of Morocco and surroundings. Basin Res. 33, 1337–1383.

Chiocchini, M., Mancinelli, A., 1996. *Archaeosepta platierensis* Wernli, 1970 (foraminiferida) from the middle jurassic near Mt boragine (north-eastern latium) and taxonomic position of the genus *Archaeosepta* Wernli, 1970. Palaeopelagos 6, 237.

Chiocchini, M., Chiocchini, R.A., Didaskalou, P., Potetti, M., 2008. Upper Triassic, Jurassic and Cretaceous microbiostratigraphy of the carbonatic platform facies in the central-southern Latium and Abruzzi: final revision. Mem. Descr.lla Carta Geol. Italia 84, 5–170.

Christ, N., Immenhauser, A., Wood, R.A., Darwich, K., Niedermayr, A., 2015. Petrography and environmental controls on the formation of Phanerozoic marine carbonate hardgrounds. Earth Sci. Rev. 151, 176–226.

Cooper, G., 1989. Jurassic brachiopods of Saudi Arabia. Smithsonian Contrib. Paleobiol. 65, 1–213.

Darga, R., Schlagintweit, F., 1991. Mikrofazies, Paläontologie und Stratigraphie der Lerchkogelkalke (Tithon-Berrias) des Dietrichshorns (Salzburger Land, Nördliche Kalkalpen). Jahrbuch der Geologischen Bundesanstalt 134, 205–226.

Davies, R.J., Stephen, K.J., Underhill, J.R., 1996. A re-evaluation of middle and upper jurassic stratigraphy and the flooding history of the Moray Firth rift system, North Sea. In: Hurst, A., Johnson, H., Burley, S.D., Canham, A.C., Mackertich, D.S. (Eds.), Geology of the Humber Group: Central Graben and Moray Firth, UKCS, vol. 114. Geological Society, London, Special Publications, pp. 81–108.

Davison, I., 2005. Central atlantic margin basins of north west Africa: geology and hydrocarbon potential (Morocco to Guinea). J. Afr. Earth Sci. 43, 254–274.

De Matos, J.E., 1997. Stratigraphy, Sedimentation and Oil Potential of the Lower Jurassic to Kimmeridgian of the United Arab Emirates: Outcrop and Subsurface Compared. Doctoral dissertation, University of Aberdeen.

Dera, G., Prunier, J., Smith, P.L., Haggart, J.W., Popov, E., Guzhov, A., Rogov, M., Delsate, D., Thies, D., Cuny, G., Pucéat, E., 2015. Nd isotope constraints on ocean circulation, paleoclimate, and continental drainage during the Jurassic breakup of Pangea. Gondwana Res. 27, 1599–1615.

Dominguez, J.M.L., Wanless, H.R., 1991. Facies architecture of a falling sea-level strandplain, Doce river coast, Brazil. In: Swift, D.J.P., Oertel, G.F., Tillman, R.W., Thorne, J.A. (Eds.), Shelf Sand and Sandstone Bodies: Geometry, Facies and Sequence Stratigraphy, vol. 14. International Association of Sedimentologists, Special Publication, pp. 259–289.

Dromart, G., Garcia, J.-P., Picard, S., Atorps, F., Lecuyer, C., Sheppard, S.M.F., 2003a. Ice age at the middle-late jurassic transition. Earth Planet Sci. Lett. 213, 205–220.

Dromart, G., Garcia, J.-P., Gaumet, F., Picard, S., Rousseau, M., Atrops, F., Lecuyer, C., Sheppard, S.M.F., 2003b. Perturbation of the carbon cycle at the Middle/Late Jurassic transition: geological and geochemical evidence. Am. J. Sci. 303, 667–707.

Du Dresnay, R., 1988. Répartition des dépôts carbonatés du Lias inférieur et moyen le long de la côte Atlantique du Maroc: conséquences sur la paléogéographie de l'Atlantique naissant. J. Afr. Earth Sci. 7, 385–396.

Dunham, R.J., 1962. Classification of carbonate rocks according to depositional textures. In: Ham, W.E. (Ed.), Classification Carbonate Rocks – a Symposium, vol. 1. AAPG Memoir, pp. 108–121.

Duval-Arnould, A., Schröder, S., Charton, R., Joussiaume, R., Razin, P., Redfern, J., 2021. Early post-rift depositional systems of the central atlantic: lower and middle jurassic of the Essaouira-Agadir Basin, Morocco. J. Afr. Earth Sci. 178, 104164.

Duval-Arnould, A., Bulot, L. Pomar, L., Charton, R., Redfern, J., Schröder, S., in review. Architecture of Oxfordian coral buildups along the Atlantic margin of Morocco. J. Afr. Earth Sci..

Ekdale, A.A., Bromley, R.G., Pemberton, G.S., 1984. Ichnology: the use of trace fossils in sedimentology and stratigraphy. SEPM Short. Course 15, 317p.

Ellis, P.M., Crevello, P.D., Eliuk, L.S., 1985. Upper jurassic and lower cretaceous deep-water buildups, Abenaki formation, Nova scotian shelf. In: Crevello, P.D., Harris, P. M. (Eds.), Deep-water Carbonates, vol. 6. SEPM Core Workshop, pp. 212–248.

Elmi, S., 1990. Stages in the evolution of late triassic and jurassic platforms: the example from the western margin of the Subalpine Basin (ardèche, France). In: Tucker, M.E., Wilson, J.L., Crevello, P.D., Sarg, J.R., Read, J.F. (Eds.), Carbonate Platforms: Facies, Sequences and Evolution, vol. 9. SEPM Special Publication, pp. 109–144.

Embry, A.F., Klovan, J.E., 1971. A late Devonian reef tract on northeastern Banks Island, NWT. Bull. Can. Petrol. Geol. 19, 730–781.

Énay, R., Le Nindre, Y.-M., Mangold, C., Manivit, J., Vaslet, D., 1987. Le Jurassique d'Arabie saoudite centrale : nouvelles données sur la lithostratigraphie, les paléoenvironnements, les faunes d'Ammonites, les âges et les corrélations. Geobios, M.S. 9, 13–66.

Énay, R., Mangold, C., Alméras, Y., Hughes, G.W.A.G., 2009. The Wadi ad Dawasir "delta". central Saudi Arabia: A relative sea-level fall of Early Bathonian age. GeoArabia 14, 17–52.

Farinacci, A., Ekmekci, E., 2004. *Sievoides kocyigiti* n.gen., n.sp., *Mesoendothyra altineriana* n.sp. and *Kurnubia feleensis* n.sp. from Kimmeridgian shallow water limestones intercalated with alkaline igneous rocks, Western Taurus, Turkey. Rev. Micropaleontol. 47, 57–69.

Feldman, H.R., Owen, E.F., Hirsch, F., 1991. Brachiopods from the jurassic of gebel el-maghara, northern Sinai. Am. Mus. Novit. 3006, 1–28.

Feldman, H.R., Owen, E.F., Hirsch, F., 2001. Brachiopods from the jurassic (callovian) of hamakhtesh hagadol (Kurnub Anticline), southern Israel. Palaeontology 44, 637–658.

Fernández-Blanco, D., Gouiza, M., Charton, R., Kluge, C., Klaver, J., Bräutigam, K., Bertotti, G., 2020. Anticline growth by shortening during crustal exhumation of the Moroccan Atlantic margin. J. Struct. Geol. 140, 104125.

Flügel, E., 2010. Microfacies of Carbonate Rocks: Analysis, Interpretation and Applications. Springer-Verlag Berlin, Heidelberg, p. 976.

Forbes, G.A., Jansen, H.S.M., Schreurs, J., 2010. Lexicon of Oman Subsurface Stratigraphy, vol. 5. GeoArabia Special Publication.

Fortwengler, D., Marchand, D., Pellenard, P., Thierry, J., Chateau-Smith, C., 2013. Integrated stratigraphy of some key callovian-oxfordian boundary sections in south-east France. Contribution to the choice of the global boundary stratotype section and point (GSSP) of the oxfordian stage. Field Guide Oxfordian GSSP Workshop. Groupe Français d'Etude du Jurassique, p. 79.

Frey, R.W., Seilacher, A., 1980. Uniformity in marine invertebrate ichnology. Lethaia 13, 183–207.

Fürsich, F.T., Oschmann, W., Singh, I.B., Jaitly, A.K., 1992. Hardgrounds, reworked concretion levels and condensed horizons in the Jurassic of western India: their significance for basin analysis. Journal of the Geological Society, London 149, 313–331.

Fürsich, F.T., Wilmsen, M., Seyed-Emami, K., Schairer, G., Majidifard, M.R., 2003. Platform-basin transect of a middle to late jurassic large-scale carbonate platform system (shotori Mountains, tabas area, east-Central Iran). Facies 48, 171–196.

Fugagnoli, A., 2000. First record of *Everticyclammina* Redmond 1964 (*E. praevirguliana* n. sp.; foraminifera) from the early jurassic of the Venetian prealps (calcari grigi, trento platform, northern Italy). J. Foraminifer. Res. 30, 126–134.

Gardosh, M., Weimer, P., Flexer, A., 2011. The sequence stratigraphy of Mesozoic successions in the Levant margin, southwestern Israel: a model for the evolution of southern Tethys margins. AAPG (Am. Assoc. Pet. Geol.) Bull. 95, 1763–1793.

Gentil, L., Lemoine, P., 1906. Sur le Jurassique du Maroc Occidental. Association française pour l'avancement des sciences, Comptes Rendus de la 34ème Session. Cherbourg 331–340, 1905.

Gerard, J.R.F., Bromley, R.G., 2008. Ichnofabrics in Clastic Sediments: Applications to Sedimentological Core Studies, p. 97.

Gill, G.A., Thierry, J., Tintant, H., 1985. Ammonites calloviennes du Sud d'Israël: systématique, biostratigraphie et paléobiogéographie. Geobios 18, 705–751.

Głowniak, E., 2002. The ammonites of the family perisphinctidae from the plicatilis zone (lower middle oxfordian) of the polish Jura chain (Central Poland); their taxonomy, phylogeny and biostratigraphy. Acta Geol. Pol. 52, 307–364.

Gradstein, F.M., Ogg, J.G., Schmitz, M.D., Ogg, G.M. (Eds.), 2020. Geologic Time Scale 2020. Elsevier.

Gravestock, C., van Buchem, F., Simmons, M.D., Davies, R., 2020. Using Sequence Stratigraphy to Resolve Depositional Misconceptions on the Arabian Plate – the Example of the Gotnia Basin Creation and Infill. Abu Dhabi International Petroleum Exhibition & Conference. SPE 203114.

Grossowicz, L.P., Bassoullet, J.P., Hirsch, F., Peri, M., 2000. Jurassic large foraminifera from Israel. Geological Survey of Israel, Current Research 12, 132–144.

Gygi, R.A., Coe, A.L., Vail, P.R., 1998. Sequence stratigraphy of the oxfordian and kimmeridgian stages (late jurassic) in northern Switzerland. In: de Graciansky, P.-C., Hardenbol, J., Jacquin, T., Vail, P.R. (Eds.), Mesozoic and Cenozoic Sequence Stratigraphy of European Basins, vol. 60. SEPM Special Publication, pp. 3–29.

Haas, J., Görög, Á., Kovács, S., Ozsvárt, P., Matyók, I., Pelikán, P., 2006. Displaced Jurassic foreslope and basin deposits of Dinaridic origin in Northeast Hungary. Acta Geol. Hung. 49, 125–163.

Haas, J., Pelikán, P., Görög, Á., Józsa, S., Ozsvárt, P., 2013. Stratigraphy, facies and geodynamic settings of Jurassic formations in the Bükk Mountains, North Hungary: its relations with the other areas of the Neotethyan realm. Geol. Mag. 150, 18–49.

Hafid, M., 2000. Triassic–early Liassic extensional systems and their Tertiary inversion, Essaouira Basin (Morocco). Mar. Petrol. Geol. 17, 409–429.

Hafid, M., Zizi, M., Bally, A., Ait Salem, A., 2006. Structural styles of the western onshore and offshore termination of the High Atlas, Morocco. Compt. Rendus Geosci. 338, 50–64.

Hallam, A., 1976. Stratigraphic distribution and ecology of European Juassic bivalves. Lethaia 9, 245–259.

Hallam, A., 1977. Jurassic bivalve biogeography. Paleobiology 3, 58–73.

Hallam, A., 2001. A review of the broad pattern of Jurassic sea-level changes and their possible causes in the light of current knowledge. Palaeogeogr. Palaeoclimatol. Palaeoecol. 167, 23–37.

Haq, B.U., 2018. Jurassic sea-level variations: a reappraisal. GSA Today (Geol. Soc. Am.) 28, 4–10.

Haq, B.U., Hardenbol, J., Vail, P.R., 1988. Mesozoic and Cenozoic Chronostratigraphy and Cycles of Sea-Level Change, vol. 42. SEPM Special Publication.

Henson, F.R.S., 1948. Larger Imperforate Foraminifera of South-Western Asia. British Museum, London, p. 127.

Hirsch, F., Bassoullet, J.-P., Cariou, E., Conway, B., Feldman, H.R., Grossowicz, L., Honigstein, A., Owen, E.F., Rosenfeld, A., 1998. The Jurassic of the southern Levant. Biostratigraphy, palaeogeography and cyclic events. In: Crasquin-Soleau, S., Barrier, E. (Eds.), Peri-Tethys Memoir 4. Epicratonic Basins of Peri-Tethyan Platform, Mémoires du Muséum National d'Histoire Naturelle, vol. 179, pp. 213–235.

Hottinger, L., 1967. Foraminifères imperforés du Mésozoïque marocain. Notes et Mémoires du Service géologique du Maroc 209, 1–168.

Howarth, K.M., Morris, J.N., 1998. The jurassic and lower cretaceous of wadi hajar, southern Yemen. Bulletin Natural History Museum London (Geology) 54, 1–32.

Hudson, R.G.S., 1958. The Upper Jurassic faunas of southern Israel. Geol. Mag. 95, 415–425.

Hudson, J.D., Clements, R.G., Riding, J.B., Wakefield, M.I., Walton, W., 1995. Jurassic paleosalinities and brackish-water communities: a case study. Palaios 10, 392–407.

Hughes, G.W., 2018. A new thin-section based micropaleontological biozonation for the Saudi Arabian Jurassic carbonates. Micropaleontology 64, 331–364.

Husinec, A., Read, J.F., Prtoljan, B., 2022. Middle and late jurassic record of seasea-level, sequence development, and carbon-isotope fluctuations, tethyan adriatic carbonate platform, Croatia. Palaeogeogr. Palaeoclimatol. Palaeoecol. 599, 111030.

Insalaco, E., 1996. Upper Jurassic microsolenid biostromes of northern and central Europe: facies and depositional environment. Palaeogeogr. Palaeoclimatol. Palaeoecol. 121, 169–194.

Jacquin, T., Dardeau, G., Durlet, C., de Graciansky, P.-C., Hantzpergue, P., 1998. The North Sea Cycle: an overview of 2nd-order transgressive/regressive facies cycles in western Europe. In: de Graciansky, P.-C., Hardenbol, J., Jacquin, T., Vail, P.R. (Eds.), Mesozoic and Cenozoic Sequence Stratigraphy of European Basins, vol. 60. SEPM Special Publication, pp. 445–466.

Jeannet, A., 1951. Die Eisen- und Manganerze der Schweiz – stratigraphie und Palaeontologie des Oolitischen Eisenerzlagers von Herznach und seiner Umgebung. Beiträge zur Geologie der Schweiz, Geotechnische Serie 5, 1–240.

Kadar, A.P., De Keyser, T., Neog, N., Karam, K.A., Le Nindre, Y.M., Davies, R.B., 2015. Calcareous nanofossil zonation and sequence stratigraphy of the Jurassic System, onshore Kuwait. GeoArabia 20, 125–180.

Kendall, C.G.S.C., Schlager, W., 1981. Carbonates and relative changes in sea level. Mar. Geol. 44, 181–212.

Kuznetsova, K.I., Sorkina, R.E., 1996. Zonal Stratigraphy and Foraminifera of the Tethyan Jurassic (Eastern Mediterranean). Gordon and Beach, London.

Laurin, B., 1984. Les Rhynchonellidés des plates-formes du Jurassique moyen en Europe Occidentale. Dynamique des populations, évolution, systématique. Cahiers de paléontologie (section Invertébrés). CNRS Edit., Paris, p. 465.

Laville, E., Piqué, A., Amrhar, M., Charroud, M., 2004. A restatement of the mesozoic atlasic rifting (Morocco). J. Afr. Earth Sci. 38, 145–153.

Le Roy, P., Piqué, A., 2001. Triassic-Liassic Western Morocco synrift basins in relation to the Central Atlantic opening. Mar. Geol. 172, 359–381.

Leleu, S., Hartley, A.J., Van Oosterhout, C., Kennan, L., Ruckwied, K., Gerdes, K., 2016. Structural, stratigraphic and sedimentological characterisation of a wide rift system: the Triassic rift system of the Central Atlantic Domain. Earth Sci. Rev. 158, 89–124.

MacEachern, J.A., Pemberton, S.G., 1992. Ichnological aspects of cretaceous shoreface successions and shoreface variability in the western interior seaway of north America. In: Pemberton, S.G. (Ed.), Applications of Ichnology to Petroleum Exploration, vol. 17. SEPM Core Workshop, pp. 57–84.

MacEachern, J.A., Pemberton, S.G., Gingras, M.K., Bann, K.L., 2006. The ichnofacies paradigm: a fifty-year retrospective. In: Miller III, W. (Ed.), Trace Fossils: Concepts, Problems, Prospects. Elsevier, Amsterdam, pp. 52–77.

Mader, N., Redfern, J., 2011. A sedimentological model for the continental upper triassic Tadrart ouadou sandstone member: recording an interplay of climate and tectonics (Argana Valley; south-west Morocco). Sedimentology 58, 1247–1282.

Mahboubi, C.Y., Nadir Naimi, M., Ouali Mehadji, A., Garah, A., Langer, M.R., Limam, H., 2023. First evidence of the benthic foraminifera *Alveosepta jaccardi* from the Oxfordian of the Saharan Atlas (Algeria): palaeoenvironmental and palaeoecological implications. Neues Jahrbuch Geol. Palaontol. Abhand. 91–100.

Manceñido, M.O., 2002. Paleobiogeography of Mesozoic brachiopod faunas from Andean–Patagonian areas in a global context. Geobios M.S. 24, 176–192.

Mangold, C., 1970. Les Perisphinctidae (Ammonitina) du Jura méridional au Bathonien et au Callovien. Doc. Lab. Geol. Lyon 41, 1–246.

Mangold, C., Marchand, D., Thierry, J., Tarkowski, R., 1996. Les ammonites de l'oolithe de Balin (Pologne): nouvelles données et réinterprétation stratigraphique. Rev. Paléobiol. 15, 55–77.

Manivit, J., Le Nindre, Y.-M., Vaslet, D., 1990. Le Jurassique d'Arabie Centrale. Documents du Bureau de Recherches Géologiques et Miniéres. Orléans 194, 1–560.

Maync, W., 1966. Microbiostratigraphy of the jurassic of Israel. Bull. Geol. Surv. Isr. 40, 56.

Martin-Garin, B., Lathuilière, B., Geister, J., Chellai, E.H., Huault, V., 2007. Geology, facies model and coral associations of the late jurassic reef complex at cape Ghir (atlantic high Atlas, Morocco). Compt. Rendus Geosci. 339, 65–74.

Marzoli, A., Renne, P.R., Piccirillo, E.M., Ernesto, M., Bellieni, G., Dein, A., 1999. Extensive 200 million-year-old continental flood basalts of the central atlantic magmatic province. Science 284, 616–618.

McCubbin, D.G., 1982. Barrier-island and strand-plain facies. In: Scholle, P.A., Spearing, D. (Eds.), Sandstone Epositional Environments, vol. 31. AAPG Memoir, pp. 247–279.

Medina, F., 1995. Syn-and postrift evolution of the El Jadida–Agadir basin (Morocco): constraints for the rifting models of the central Atlantic. Can. J. Earth Sci. 32, 1273–1291.

Medina, F., Et-Touhami, M., Olsen, P.E., Bouaouda, M.-S., Hafid, M., Ettachfini, E.M., 2011. Circuit C11/tour C11, haut Atlas Occidental, western high Atlas. Notes et Mémoires du Service Géologique du Maroc 562, 9–62.

Meléndez, G., Atrops, F., Ramajo, J., Perez-Urresti, I., Delvene, G., 2006. Upper Oxfordian to lower Kimmeridgian successions in the NE Iberian Range (E Spain): some new stratigraphical and palaentological data. Neues Jahrbuch Geol. Palaontol. Abhand. 241, 203–224.

Meléndez, G., Ramajo, J., Martínez-Cotanda, S., 2005. El desarrollo de la Capa de Arroyofrío (límite Calloviense-Oxfordiense) al Sur de Zaragoza, entre Ricla y Aguilón : bioestratigrafía y facies. Geogaceta 38, 3–6.

Mette, W., 1997. Palaeoecology and palaeobiogeography of the Middle Jurassic ostracods of southern Tunisia. Palaeogeogr. Palaeoclimatol. Palaeoecol. 131, 65–111.

Morabet, A.M., Bouchta, R., Jabour, H., 1998. An overview of the petroleum systems of Morocco. In: Macgregor, D.S., Moody, R.T.J., Clark-Lowes, D.D. (Eds.), Petroleum Geology of North Africa, vol. 132. Geological Society, London, Special Publications, pp. 283–296.

Muir-Wood, H.M., 1935. Jurassic brachiopoda. In: MacFayden, W.A. (Ed.), The Mesozoic Palaeontology of British Somaliland. Geology and Palaeontology of British Somaliland, II. Government of Somaliland Protectorate, London, pp. 75–147.

Mukherjee, D., 2007. A taxonomic and phylogenetic study of *Kutchithyris* – a Jurassic terebratulid from Kachchh India. J. Asian Earth Sci. 30, 213–237.

Mukherjee, D., Bardhan, S., Datta, K., Ghosh, D.N., 2003. The terebratulid *Kutchithyris* (brachiopoda) from the jurassic sequence of kutch, western India-revisited. Palaeontological Research 7, 111–128.

Murris, R.J., 1980. Middle East: stratigraphic evolution and oil habitat. AAPG (Am. Assoc. Pet. Geol.) Bull. 64, 597–618.

Nasiri, Y., Bayet-Goll, A., Mahboubi, A., Moussavi-Harami, R., Monaco, P., 2020. Paleoenvironmental control on trace fossils across a Mississippian carbonate ramp succession, Mobarak Formation, east of Central and Eastern Alborz, Iran. J. Afr. Earth Sci. 165, 103800.

O'Dogherty, L., Aguado, R., Baumgartner, P.O., Bill, M., Gorican, S., Sandoval, J., Sequeiros, L., 2018. Carbon-isotope stratigraphy and pelagic biofacies of the middle–upper jurassic transition in the tethys–central atlantic connection. Palaeogeogr. Palaeoclimatol. Palaeoecol. 507, 129–144.

Olivier, N., Pittet, B., Werner, W., Hantzpergue, P., Gaillard, C., 2008. Facies distribution and coral-microbialite reef development on a low-energy carbonate ramp (Chay Peninsula, Kimmeridgian, western France). Sediment. Geol. 205, 14–33.

Olivier, N., Martin-Garin, B., Colombie, C., Cornee, J.J., Giraud, F., Schnyder, J., Kabbachi, B., Ezaidi, K., 2012. Ecological succession evidence in an Upper Jurassic coral reef system (Izwarn section, High Atlas, Morocco). Geobios 45, 555–572.

Olóriz, F., 2002. *Idoceras hoderniae* (Gemmellaro, 1878). In: Pavia, G., Cresta, S. (Eds.), Revision of Jurassic Ammonites of the Gemmellaro Collections, vol. 6. Quaderni del Museo Geologico G. G. Gemmellaro, pp. 292–293.

Palma, R.M., López-Gómez, J., Piethé, R.D., 2007. Oxfordian ramp system (La manga formation) in the bardas blancas area (mendoza province) neuquén basin, Argentina: facies and depositional sequences. Sediment. Geol. 195, 113–134.

Pellenard, P., Deconinck, J.-F., 2006. Mineralogical variability of callovo-oxfordian clays from the Paris Basin and the Subalpine Basin. Compt. Rendus Geosci. 338, 854+866.

Pellenard, P., Fortwengler, D., Marchand, D., Thierry, J., Bartolini, A., Boulila, S., Pierre-Yves Collin, P.-Y., Enay, R., Galbrun, B., Gardin, S., Huault, V., Huret, E., Martinez, M., Chateau Smith, C., 2014a. Integrated stratigraphy of the Oxfordian global stratotype section and point (GSSP) candidate in the Subalpine Basin (SE France). Volumina Jurassica 12 (1), 1–44.

Pellenard, P., Tramoy, R., Pucéat, E., Huret, E., Martinez, M., Bruneau, L., Thierry, J., 2014b. Carbon cycle and sea-water paleotemperature evolution at the Middle-Late Jurassic transition, eastewrn Paris Basin, France. Mar. Petrol. Geol. 53, 30–43.

Pemberton, S.G., MacEachern, J.A., 1997. The ichnological signature of storm deposits: the use of trace fossils in event stratigraphy. In: Brett, C.E., Baird, G.C. (Eds.), Paleontological Events. Stratigraphic, Ecological and Evolutionary Implications. Columbia University Press, New York, pp. 73–109.

Pemberton, S.G., MacEachern, J.A., Saunders, T., 2004. Stratigraphic applications of substrate-specific ichnofacies: delineating discontinuities in the rock record. In: McIlroy, D. (Ed.), The Application of Ichnology to Palaeoenvironmental and Stratigraphic Analysis, vol. 228. Geological Society, London, Special Publublications, pp. 29–62.

Peybernès, B., Bouaouda, M.S., Alméras, Y., Ruget, C., Cugny, P., 1987. Stratigraphie du Lias et du Dogger du bassin côtier d'Essaouira (Maroc) avant et pendant le début de l'expansion océanique dans l'Atlantique central. Comparaison avec le bassin d'Agadir. Comptes-Rendus de l'Académie des Sciences, Paris 305, 1449–1455.

Picard, L., Hirsch, F., 1987. The Jurassic Stratigraphy in Israel and the Adjacent Countries. Israel Academy of Sciences and Humanities, p. 107.

Pierre, A., Durlet, C., Razin, P., Chellai, E.H., 2010. Spatial and temporal distribution of ooids along a Jurassic carbonate ramp: Amellago outcrop transect, High-Atlas, Morocco. Geological Society, London, Special Publications 329, 65–88.

Piqué, A., Le Roy, P., Amrhar, M., 1998. Transtensive synsedimentary tectonics associated with ocean opening: the Essaouira–Agadir segment of the Moroccan Atlantic margin. Journal of the Geological Society, London 155, 913–928.

Price, G.D., Heszler, B., Tansley Charlton, L.-M., Cox, J., 2023. A climate perturbation at the Middle-Late Jurassic transition? Evaluating the isotopic evidence from south-central England. Palaeogeogr. Palaeoclimatol. Palaeoecol. 628, 111755.

Ramajo, J., Aurell, M., 2008. Long-term Callovian-Oxfordian sea-level changes and sedimentation in the Iberian carbonate platform (Jurassic, Spain): possible eustatic implications. Basin Res. 20, 163–184.

Reolid, M., Abad, I., 2018. The Middle-Upper Jurassic unconformity in the South Iberian Palaeomargin (Western Tethys): a history of carbonate platform fragmentation, emersion and subsequent drowning. J. Iber. Geol. 45, 87–110.

Rigaud, S., Blau, J., Martini, R., Rettori, R., 2013. Taxonomy and phylogeny of the trocholinidae (involutinina). J. Foraminifer. Res. 43, 317–339.

Roch, E., 1930. Étude géologique dans la région méridionale du Maroc Occidental. Notes et Mémoires du Service géologique du Maroc, Rabat 9, 1–542.

Ruban, D.A., 2012. Erosional surface at the middle-upper callovian (middle jurassic) transition in the greater caucasus basin (northern neo-tethys) and tracing its presence in western Europe, North Africa and Arabia: the influence of regional tectonics. Comunicações Geológicas 99, 69–76.

Schlagintweit, F., Velić, I., 2011. New and poorly known Middle Jurassic larger benthic foraminifera from the Karst Dinarides of Croatia. Geol. Croat. 64, 81–100.

Schlagintweit, F., Gawlick, H.J., Lein, R., 2005. Mikropaläontologie und Biostratigraphie der Plassen-Karbonatplattform der Typlokalität (Ober-Jura bis Unter-Kreide, Salzkammergut, Österreich). Journal of Alpine Geology 47, 11–102.

Septfontaine, M., 1978. In: Présence d'*Archaeosepta platierensis* Wernli, 1970 dans le Jurassique briançonnais des Préalpes. Importance stratigraphique; relation avec le microfaciès et la paléogéographie, vol. 1. Notes du Laboratoire de Paléontologie de l'Université de Genève, pp. 1–6.

Septfontaine, M., 1988. Vers une classification évolutive des lituolidés (foraminifères) Jurassiques en milieu de plate-forme carbonatée. Revue de Paléobiologie, volume special 2, 229–256.

Septfontaine, M., Arnaud-Vanneau, A., Bassoullet, J.-P., Gusic, Y., Ramalho, M., Velic, I., 1991. Les Foraminifères imperforés des plates-formes carbonatées jurassiques. Etat des connaissances et perspectives d'avenir. Bull. Soc. Vaudoise Sci. Nat. 80, 255–277.

Sharland, P.R., Archer, R., Casey, D.M., Davies, R.B., Hall, S.H., Heward, A.P., Horbury, A.D., Simmons, M.D., 2001. Arabian Plate Sequence Stratigraphy, vol. 2. GeoArabia Special Publication.

Simmons, M.D., Miller, K.G., Ray, D.C., Davies, A., van Buchem, F.S.P., Gréselle, B., 2020. Phanerozoic eustasy. In: Gradstein, F.M., Ogg, J.G., Schmitz, M.D., Ogg, G.M. (Eds.), Geologic Time Scale 2020. Elsevier, pp. 357–400.

Steinhoff, I., Strohmenger, C., 1996. Zechstein 2 carbonate platform subfacies and grain-type distribution (Upper Permian, northwest Germany). Facies 35, 105–132.

Strasser, A., 1986. Ooids in purbeck limestones (lowermost cretaceous) of the Swiss and French Jura. Sedimentology 33, 711–727.

Tari, G., Jabour, H., 2013. Salt tectonics along the Atlantic margin of Morocco. Geological Society, London, Special Publications 369, 337–353.

Tasli, K., Altiner, D., Koç, H., Eren, M., 2008. Benthic foraminiferal biostratigraphy of the jurassic platform carbonate succession in the bolkar Mountains (southern Turkey). Micropaleontology 54, 425–444.

Thompson, W.O., 1937. Original structures of beaches, bars, and dunes. Bull. Geol. Soc. Am. 48, 723–752.

Tudoran, A., 1997. Biostratigraphic and Paleoenvironmental Significance of Jurassic Microfossils from Romania. Louisiana State University and Agricultural & Mechanical College.

Velić, I., 2007. Stratigraphy and palaeobiogeography of mesozoic benthic foraminifera of the karst dinarides (SE europe)-part 1. Geol. Croat. 60, 1–60.

Vuks, V.J., 2012. Middle-Late Jurassic foraminifers of the western Caucasus: evolution of foraminiferal assemblages and some environmental aspects. Paleontol. J. 47, 1163–1169.

Walley, C.D., 1985. Depositional history of southern Tunisia and northwestern Libya in Mid and Late Jurassic time. Geol. Mag. 122, 233–247.

Weir, J., 1925. Brachiopoda, Lamellibranchiata, Gastropoda and Belemnites in the Collection of Fossils and Rocks from Somaliland Made by Mssr B.K.N. Wyllie and W. R. Smellie, vol. 1. Monographs of the Geological Department of the Hunterian Museum, pp. 1–63.

Weir, J., 1929. Jurassic Fossils from Jubaland, East Africa, Collected by G.V. Glenday, and the Jurassic Geology of Somaliland by Mssr B.K.N. Wyllie and W.R. Smellie, vol. 3. Monographs of the Geological Department of the Hunterian Museum, pp. 1–63.

Weissenberger, J.A.W., Wierzbicki, R.A., Harland, N.J., 2006. Carbonate sequence stratigraphy and petroleum geology of the Jurassic deep Panuke field, offshore Nova Scotia, Canada. In: Harris, P.H., Weber, L.J. (Eds.), Giant Hydrocarbon Reservoirs of the World: from Rocks to Reservoir Characterization and Modeling, vol. 88. AAPG Memoir, pp. 395–431.

Wentworth, C.K., 1922. A scale of grade and class terms for clastic sediments. J. Geol. 30, 377–392.

Wernli, R., 1970. *Archaeosepta platierensis* Wernli, n. gen., n. sp., un nouveau Foraminifère du Dogger du Jura méridional. Comptes Rendus des Séances de la Société de Physique et d'Histoire Naturelle de Genève 5, 87–93.

Wernli, R., Görög, Á., 2007. Protoglobigérines et Oberhauserellidae (Foraminifères) du Bajocien-Bathonien du Jura méridional, France. Rev. Micropaleontol. 50, 185–205.

Weston, J.F., MacRae, R.A., Ascoli, P., Cooper, M.K.E., Fensome, R.A., Shaw, D., Williams, G.L., 2012. A revised biostratigraphic and well-log sequence-stratigraphic framework for the Scotian Margin, offshore eastern Canada. Can. J. Earth Sci. 49, 1417–1462.

Wierzbowski, H., Dembicz, K., Praszkier, T., 2009. Oxygen and carbon isotope composition of Callovian–Lower Oxfordian (Middle–Upper Jurassic) belemnite rostra from central Poland: a record of a Late Callovian global sea-level rise? Palaeogeogr. Palaeoclimatol. Palaeoecol. 283, 182–194.

Youssef, M., El-Sorogy, A.S., 2015. Paleoecology of benthic foraminifera in coral reefs recorded in the Jurassic Tuwaiq Mountain Formation of the Khashm Al-Qaddiyah area, Central Saudi Arabia. J. Earth Sci. 26, 224–235.

Zagorchev, I., Trifonova, E., Ivanova, D., Stoykova, K., Goranov, E., Pemov, P., 2000. New data on the middle and upper jurassic stratigraphy in the konyava mountain, SW Bulgaria. Geol. Balc. 30, 59–68.