Controls on stratigraphic development of shelf margin carbonates: Jurassic Atlantic margin - Essaouira-Agadir Basin, Western Morocco

A thesis submitted to The University of Manchester
for the degree of Doctor of Philosophy
in the Faculty of Science and Engineering

2019

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List of abbreviations

BST	Boundstone
CAMP	Central Atlantic Magmatic Province
CHA	Central High Atlas
DSDP	Deep Sea Drilling Project
EAB	Essaouira-Agadir Basin
Fm.	Formation
FST	Floatstone
GST	Grainstone
MAM	Massif Ancien de Marrakech
MFS	Maximum Flooding Surface
MST	Mudstone
PST	Packstone
RST	Rudstone
SB	Sequence Boundary
Tjb	Triassic-Jurassic Boundary
TS	Transgressive Surface
WM	Western Messeta
WMA	Western Moroccan Arch
WST	Wackestone

Abstract

Controls on stratigraphic development of shelf-margin carbonates: Jurassic Atlantic margin - Essaouira-Agadir Basin, Western Morocco

The Jurassic succession along the Atlantic margin offshore Morocco and on the Nova Scotia conjugate margin is marked by development of extensive carbonate platforms, alternating with siliciclastic influx. This thesis aims to improve understanding of the development and architecture of carbonate dominated sequences along a passive margin, and the relative controls of eustasy, climate and tectonics on facies distribution.

Early-post rift deposition started with development of a carbonate ramp during the Upper Sinemurian to Lower Pliensbachian, later exhumed and tilted before deposition of a fluvial package. This is interpreted to mark an important regional inversion event of Toarcian age, with a fluvial system eroding down to Triassic deposits. This is followed by an upper Toarcian transgression that established a shallow marine carbonate platform, dominated by peritidal deposits. These conditions persisted throughout the Middle Jurassic in the Essaouira part of the basin, while around Agadir a siliciclastics influx from the south occurred. The origin of the siliciclastic intervals can be linked to uplift of the Anti-Atlas documented by recent thermochronology studies.

The Middle to Upper Jurassic interval is dominated by carbonate ramp deposits with development of discrete bioherm buildups. This is a proven reservoir offshore Morocco and Nova Scotia, however facies distribution and geometry is poorly documented. The interval has been extensively sampled for biostratigraphy, with the existing collections re-evaluated, resulting in new data that improves dating. Detailed sedimentological logging of extensive outcrops has identified the establishment of a carbonate ramp during the Callovian transgression containing two potential reservoir intervals: the lower Callovian oolities and middle Oxfordian buildups. Biostratigraphic analysis has identified a previously unrecognised upper Callovian to middle Oxfordian hiatus. The age of the buildups has been more precisely constrained to middle Oxfordian, suggesting their establishment was broadly synchronous. The buildups show a range of sizes and geometries, with a typical evolution from low relief microsolenid colonies to higher relief diversified colonies. The major bioherms are fringed by a rubble facies that forms clinoforms. Highly dolomitised, these deposits enhance the reservoir potential of the formation by creating connections between the coral buildup. This formation is followed during the upper Oxfordian by a major regression that established peritidal conditions.

This study provides the first detailed analysis of the entire lower to upper Jurassic interval undertaken using a unified systematic analysis. The influence of tectonics at certain times is shown by the repeated influx of siliciclastics and deformation of the basin. At outcrop scale, integrated mapping has allowed the facies relationships and control of facies development to be addressed. The results provide further insight into the size of bioherms in the Oxfordian and their typical distribution, that can be extrapolated to the subsurface to give a better understanding of buildup geometries Offshore Morocco and the Scotian Basin, where they are potential hydrocarbon reservoirs.

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Declaration

No portion of the work referred to in the thesis has been submitted in support of an application for another degree or qualification of this or any other university or other institute of learning.

Aude Duval-Arnould

Janvier 2019

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Copyright statement

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Acknowledgments

I would like first to thank Stefan Schröder, my main supervisor for his constant support, his great availability, including for long field seasons and hours behind the microscope. I am very grateful for his continuous encouragements and tireless efforts to discuss and review my work during these 4 years.

I would also like to thank my supervisor Jonathan Redfern for his efforts and patience to revise my work and my writing, and all the great networking opportunities he provided me through conferences and fieldwork.

My supervisor, Luc Bulot, has also contributed vastly to improve the quality of this thesis, by joining long field seasons focused on biostratigraphy, providing his palaeontological expertise, and his remarkable knowledge of the study area. I am grateful for all the hours he spent trying to sort out the old Moroccan paleontology, and his insightful comments and corrections.

This study was sponsored by the North Africa Research Group (NARG) and I would like to acknowledge the support of the sponsoring companies and the Office National des Hydrocarbures et des Mines (ONHYM). Thank you to NARG and my supervisors for trusting me and giving me the great opportunity to pursue this PhD project.

I would also like to thank Giovani Bertotti and our sponsors for many enlightening discussion and shared ideas during the steering meetings.

Many thanks to Mike Simmons for his foraminifera identifications, and further discussion and comments.

A special thank you to Moussa Masrour for his great help on the field and for the identification of the echinoderms. His knowledge of the area and the stratigraphy has been very helpful. Merci pour votre aide sur le terrain, vos conseils et votre gentillesse Moussa.

I would like to thank Alain Bonnot for his help in identifying the more rebellious ammonites.

Thank you to Rémi Charton, Angel Arantagui, and Leonardo Muniz Pichel for all the good discussion and helping me understand the big picture.

Thank you to Rémi Joussiaume, Philippe Razin and Carine Grélaud for presenting us their work on the Central High Atlas and all the good discussion comparing it to the Western High Atlas.

Thank you to Luis Pomar for all the good discussion and helping me have a different perspective on carbonate production. I will not forget to “follow the food”.

Thank you to all my field assistants:
Frederic Amour (I cannot thank you enough for everything I learned with you... and for your patience teaching me), Paul Goodrich (Thank you for coming with me during so many field seasons, for all the good moments, your tireless patience, and your contagious enthusiasm and cheerfulness!), Benedict Campbell (thank you for being there for this last, stressfull field season... I will never forget your yellow socks), Awatif Habid (Thank you for that first field season, all the discussion back in Manchester, and your true friendship), Jianpeng Wang (thank you for all the very memorable moments spent on the field, I had my best and worst days with you!), Lahcen Aabi (Thank you very much for all your help and support in the field, your constant good mood and for teaching me relentlessly the rudiments of Moroccan. Thank you my friend), Emmanuel Roquette (Thank you for your incredible willingness in the field, you constant good spirit and for saving my ankle... twice), Christian Klüge (Thank you for the joint field season, for bringing such a good knowledge of the area with you, it was very good having another perspective on the field. And thank you for you for your cheerfulness and all the good moments!)

And last but not least, I would like to thank all my NARG fellows, and especially Nawwar al Sinawi for all the good discussion we had, the very nice time spent on the field and everything she taught me about the dolomitization of *our* Jurassic carbonates. Thank you to Rémi Charton, Jianpeng Wang, Emmanuel Roquette, Leonardo Muniz Pichel, and especially to my officemates Angel Arantegui and Tim Luber for all the help they gave me starting the PhD, and for all the good moments! Thank you to my new officemates, Max Casson, James Lovell-Kennedy and Orrin Bryers for their encouragements and for being here for me during these last months.

Thesis outline

The thesis consists of 8 chapters, three of which constitute the core of the study and are presented as papers intended for publication. This section introduces the different chapters and for chapters 4, 5 and 6, indicates the intended journals and the input of the different co-authors.

Chapter 1: Introduction

This chapter introduces the rationale, aims and objectives of the project and highlights the research questions addressed. It also presents the dataset and methodology applied.

Chapter 2: Literature review

This chapter presents previous work published on the study area; highlighting certain concepts and geological features used during the thesis, and introducing the location and setting of the Moroccan and Nova Scotian conjugate margin.

Chapter 3: Geological setting

This chapter provides an overview of current understanding of the tectonostratigraphic evolution and sedimentological context of the study area.

Chapter 4: Evolution of early post-rift depositional systems along the Moroccan Atlantic Margin: the Agadir-Essaouira Basin, Lower and Middle Jurassic.

This chapter is intended to be published in Sedimentology. It describes the initial post-rift carbonates deposits, the influence of siliciclastic influx on the system and links with hinterland movements, and correlation with the Central High Atlas.

Co-Authors:

Dr. Stefan Schröder: data collection, discussion, manuscript review

Prof. Jonathan Redfern: discussion, manuscript review

Dr. Rémi Charton: discussion, manuscript review

Dr. Rémi Joussiaume: discussion, manuscript review

Chapter 5: The development and evolution of Atlantic margin carbonate ramp during the Callovian to middle Oxfordian: Agadir-Essaouira Basin, Morocco.

This chapter intended to be published in Sedimentary Geology. It presents the biostratigraphic framework and the stratigraphic evolution of the lower Callovian to middle Oxfordian open ramp carbonates.

Co-authors:

Dr. Luc Bulot: data collection, biostratigraphy (ammonites & brachiopods), manuscript review

Dr. Stefan Schröder: data collection, discussion, manuscript review

Prof. Jonathan Redfern: manuscript review

Dr. Moussa Marour: data collection, biostratigraphy (echinoderms), manuscript review

Dr. Mike Simmons: biostratigraphy (foraminifera), manuscript review

Prof. Alain Bonnot: biostratigraphy (ammonites), manuscript review

Chapter 6: Development and architecture of upper Jurassic coral buildups along the Atlantic Margin (Essaouira-Agadir Basin, Morocco).

This chapter is intended to be published in Palaeogeography, Palaeoclimatology, Palaeoecology. It describes the facies associations variability, the evolution and geometries of the middle Oxfordian buildups.

Co-authors:

Dr. Stefan Schroeder: data collection, discussion, manuscript review

Prof. Jonathan Redfern: discussion, manuscript review

Dr. Luc Bulot: discussion, manuscript review

Dr. Mike Simmons: biostratigraphy (foraminifera), manuscript review

Prof. Luis Pomar: discussion, manuscript review

Chapter 7: Upper Jurassic Stratigraphy and Discussion

This chapter describes the facies variation for all the Jurassic units across the basin, including the Kimmeridgian and Tithonian, which have not been discussed in Chapters 4 to 6, and introduces the regional palaeogeograghic maps.

Chapter 8: Conclusions

This chapter present a synthesis of the project and the conclusions related to the research objectives. It also addresses recommendations for further work.

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Chapter 1: Introduction

1.1 Introduction to the project

The North Africa Research Group (NARG) has undertaken research in North Africa since its foundation in 2000. For several years, a focus has been on the Mesozoic evolution of the Moroccan Atlantic Margin, aiming to better understand the geology of the margin with studies on the Trias rift evolution, offshore salt tectonics, the sedimentologic and biostratigraphic evolution of the Cretaceous, the thermochronology and source-to-sink evolution of the passive margin. This study of the Jurassic interval sits within this framework and aims to develop a better understanding of facies and the controls on the evolution of the depositional systems that can be integrated to this ambitious multi-disciplinary project. The results have important significance to understand the potential for hydrocarbon systems offshore Morocco and within the wider Atlantic margin, to constrain the early post-rift geological evolution of this passive margin.

1.2 Rationale of the project

Upper Jurassic carbonates are proven reservoirs along the Atlantic Margin, yet they are still poorly constrained. In a number of wells in Morocco, hydrocarbons have been found in these carbonates (Sidi Rhalem, Cap Juby; Morabet et al., 1998) and gas has been produced in Nova Scotia (Deep Panuke; Weissenberger et al., 2006). But in Morocco, no oil or gas is produced as reservoir quality and distribution has proved elusive. Recent wells drilled for carbonate objectives proved to be dry or of poor reservoir quality and this suggests our understanding of depositional models that allow prediction of reservoirs is still at an early stage. The main reservoir target along this portion of the Atlantic margin consists of Middle Jurassic Oxfordian coral buildups. Any effective exploration must fundamentally understand the nature, extent and geometry of the buildups and associated facies. In the subsurface this is difficult, with the limited data. However, time equivalent units crop out extensively onshore Morocco, and this study aims to examine these superbly exposed sections to improve understanding of the systems and thus reduce the risk of petroleum exploration offshore on both margins. In the course of this study, other potential Jurassic plays have been identified, including middle Jurassic siliciclastics and oolites.

The main goal of this study is to refine the understanding of the sedimentology of the lower to upper Jurassic stratigraphy by analysing the extensive outcrops in the Agadir Basin (Western High Atlas), as well as onshore and offshore wells from the Essaouira Agadir Basin (EAB).

1.3 Aims and objectives of the project

This project examines the extensive Jurassic outcrops along the Atlantic margin of western Morocco. The thesis reports the final results, providing a high-resolution facies analysis based on extensive fieldwork, revised biostratigraphy that modifies the existing stratigraphic framework and finally building new paleogeographic reconstructions. The first aim of this project was to construct a coherent stratigraphic framework to constrain the spatial and temporal distribution of the facies suites. Of particular interest was the tectonostratigraphic post-rift evolution of the basal Jurassic strata; and an improved characteristation of the main carbonate reservoir intervals in the Middle to Upper Jurassic, their evolution, distribution, facies variation and geometry of contained buildups.

The results aim to improve understanding of the relative control of eustacy, climate, tectonics, and siliciclastic input on the development of a carbonate-dominated sequence developing along a passive margin.

The objectives of this project were defined as:

- Reinterpret and refine the basin stratigraphy in order to gain a better understanding of the depositional environments, their temporal succession and lateral distribution.
- Collect new biostratigraphic material and evaluate dating of the succession to improve chronostratigraphy and regional correlations (using ammonites, foraminifera, brachiopods and echinoderms).

- Identify any potential reservoir units, their extent and lateral facies variation.
- Establish the timing of siliciclastic influx into the basin and constrain the controls on their origin and their influence on the carbonate succession.
- Compare the Jurassic stratigraphy in the EAB with Nova Scotia and the Central High Atlas.
- Resolve the relative controls of climate, tectonic and eustacy on the development of a carbonate dominated system that develops along a passive margin.

1.4 Methodology and dataset

The principal aim of this study is to present a detailed sedimentological and stratigraphical analysis of the Jurassic deposits in the EAB and to understand the facies evolution and geometries of the different units. To this end, extensive field work was undertaken for a total of 8 months over a three-year period in the EAB, together with sampling for petrographic and biostratigraphic studies.

1.4.1 Dataset collection

Detailed sedimentary logging of 32 stratigraphic sections (Figure 1.2) from 19 locations (Figure 1.1) provided most of the data used in this study. This was complimented by a comprehensive review of the published literature for the study area, including available peer reviewed papers, PhD studies, field guides, well data and proprietary reports. Over 400 fauna specimens (400+ brachiopods, 14 echinoderms and 25 ammonites) have been collected for palaeontological studies. This faunal collection augmented data from the comprehensive review of brachiopods and ammonites collected by Gentil and Lemoine (1906) and by Ambroggi (1963). All sections studied were described bed by bed at a 5 cm scale and every bed was photographed, with GPS reference. Some 1770 hand samples were collected from the sections logged and available for more detailed description under a binocular magnifier. A total of 193 representative thin sections were made of the major macro-facies observed and studied for petrography. Multiple photographic panels were used to trace horizons, interpret geometries and to support the interpretations. Study of onshore and offshore wells (both exploration well data supplied by ONHYM and publically available DSDP wells) and their associated reports have been used to support and expand interpretations where outcrops are not available.

Sections from 4 locations: Amsittène (AOS), Tizgui barrage (C1 and C2), Assif el Hade (AHD, AS) and Paradise Valley (Pk, B1 and B2) have been used as reference composite section for the Jurassic (Figure 1.2).

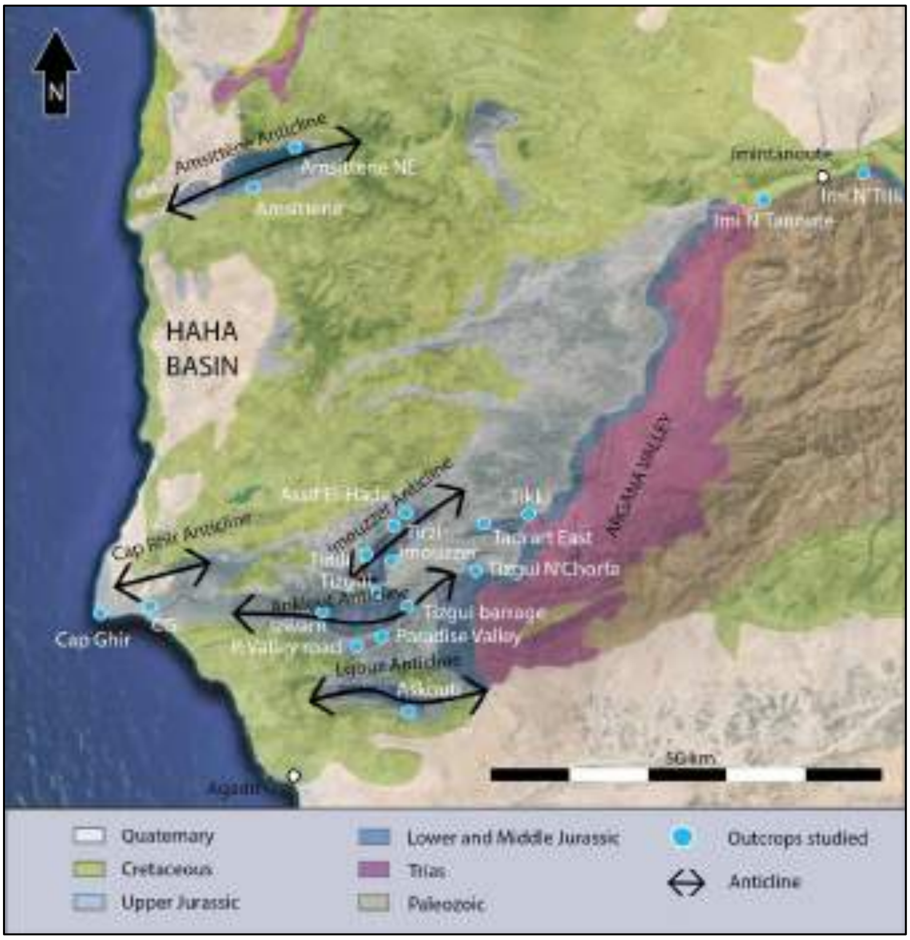

Figure 1.1: Map of the Essaouira-Agadir Basin and outcrops studied. Map after Adams, 1979 and Zühlke et al., 2004.

1.4.2 Biostratigraphy

Micro- and macro-fauna were used to identify the environment of deposition, and where possible, for biostratigraphic dating. The foraminifera were identified by Dr. Mike Simmons (Halliburton) and provided broad range dating on the Callovian to Kimmeridgian interval and to constrain the palaeoenvironments. Echinoderms were identified by Prof.

Moussa Massrour (University of Agadir) and used for their environmental indication and to the complement the biostratigraphic framework. The brachiopod and ammonite fauna were identified by Dr. Luc Bulot (CEREGE-University of Manchester), with some particular ammonites were also revised / verified by Dr. Alain Bonnot (University of Bourgogne).

Brachiopod fauna in the EAB are very diverse and allow identification of the different units across the basin and help environmental interpretation. The ammonites were used for high-precision dating of the different horizons and allowed a coherent biostratigraphic framework for the Essaouira-Agadir Basin to bo built.

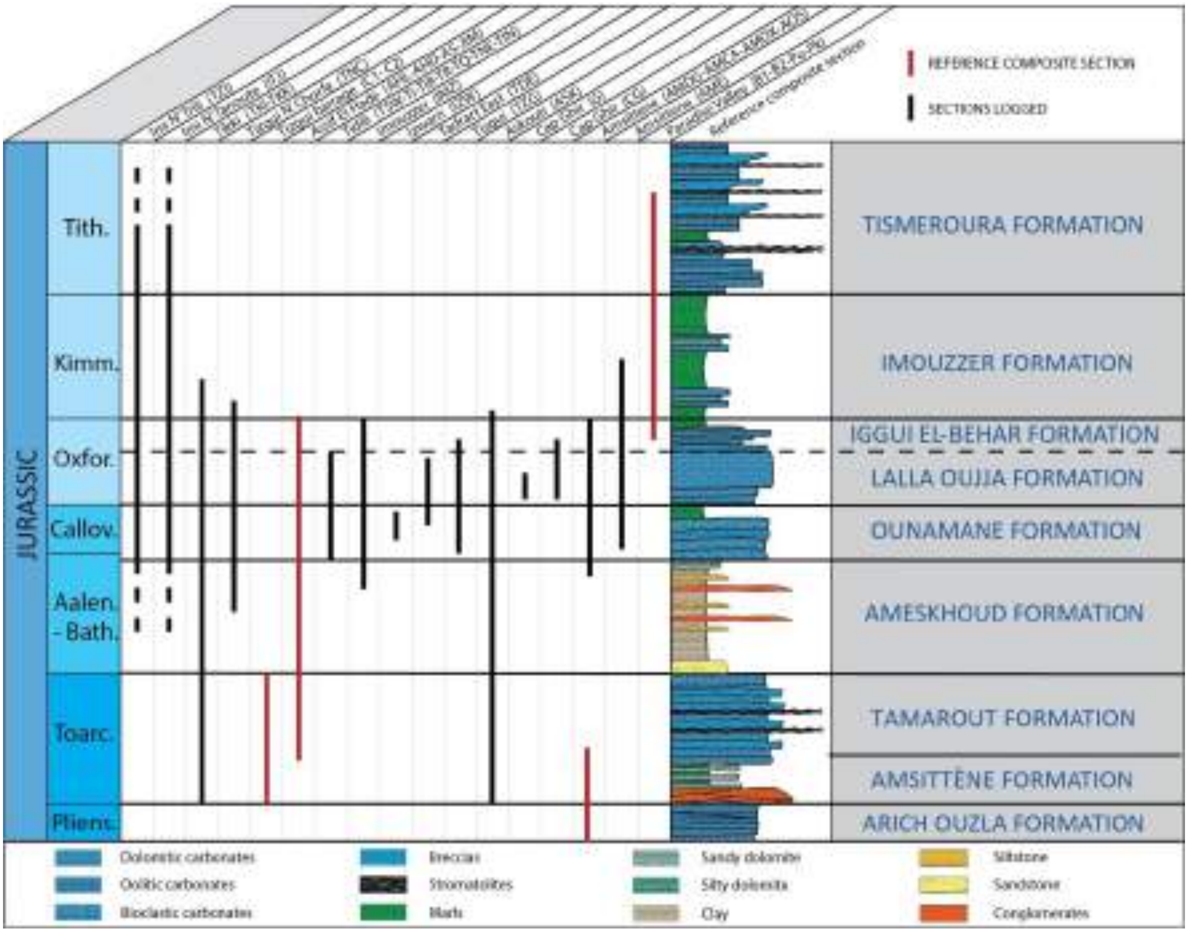

Figure 1.2: Reference section and outcrops studied with their stratigraphic extension.

1.4.3 Petrography

Most of the thin sections have been blue dye impregnated for better porosity observation and stained with Alizarin-red S to differentiate calcite from dolomite. The thin sections have been described regarding skeletal and non-skeletal elements, grain size, roundness, sorting, porosity and the matrix or associated cements. They were categorized following Dunham's (1962) classification of limestones according to their depositional texture with the modifications by Embry and Klovan (1971). The classified thin sections have been organised into microfacies, presented in the relevant chapters.

Chapter 2: Literature review

2.1 Essaouira-Agadir Basin stratigraphy

The continuous Jurassic succession outcropping within the EAB (Figure 2.1) has been studied since the early 1900's (Lemoine, 1904), but it was only in 1963 that Robert Ambroggi identified the different formations and determined the associated ages (Figure 2.2). Since then the Jurassic deposits have been studied by multiple authors (Duffaud, 1960; Duffaud et al., 1966, 1981; Du Dresnay, 1988, Bouaouda, 1987, 2004;), with a strong emphasis on the late Middle to Upper Jurassic (Ager, 1974; Adams, 1979; Martin-Garin et al., 2007; Olivier et al., 2012), which is a potential reservoir offshore Morocco (Morabet, 1998; Davison, 2005) and has produced gas on the conjugate margin of Nova Scotia (Weissenberger et al., 2006).

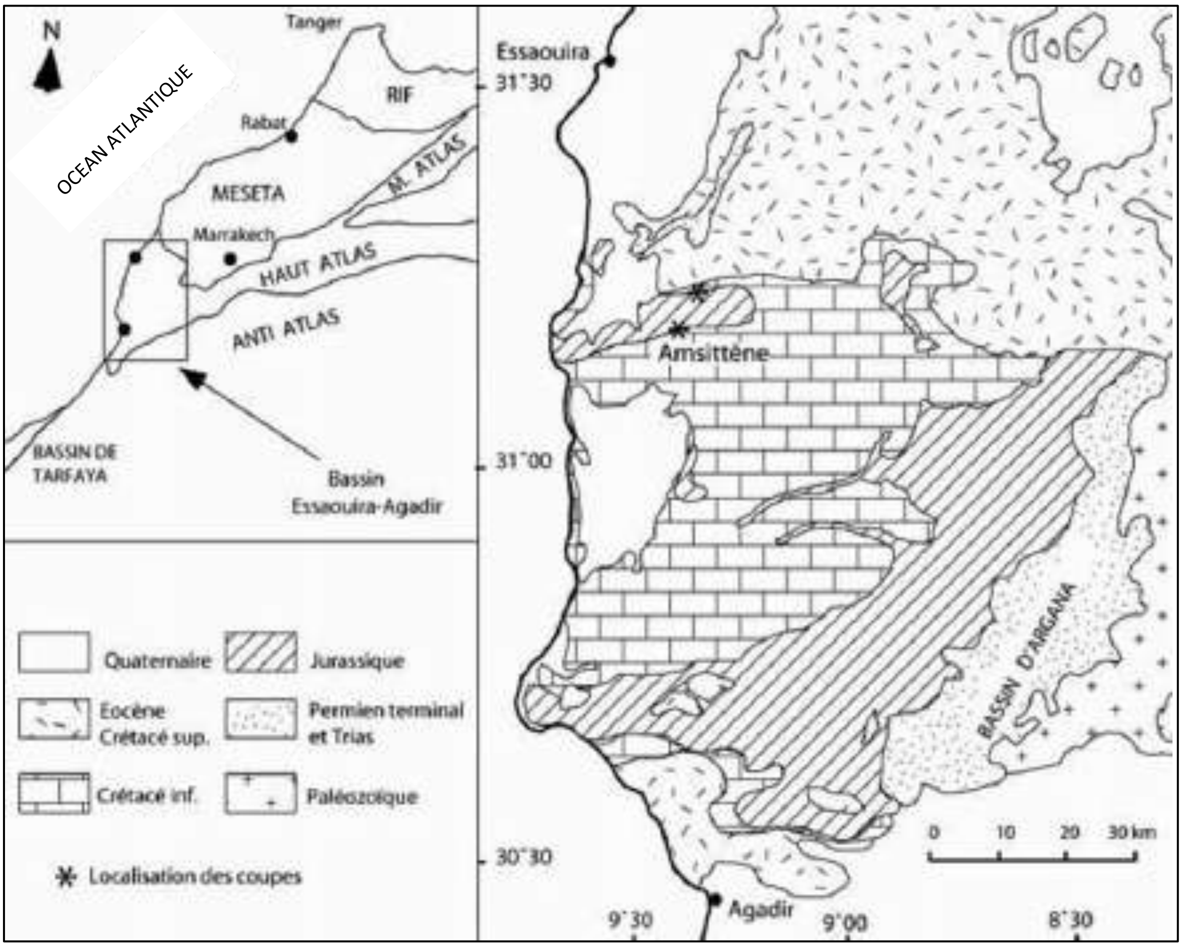

Figure 2.1: Geological map of the EAB. From Ouajhain et al., 2009.

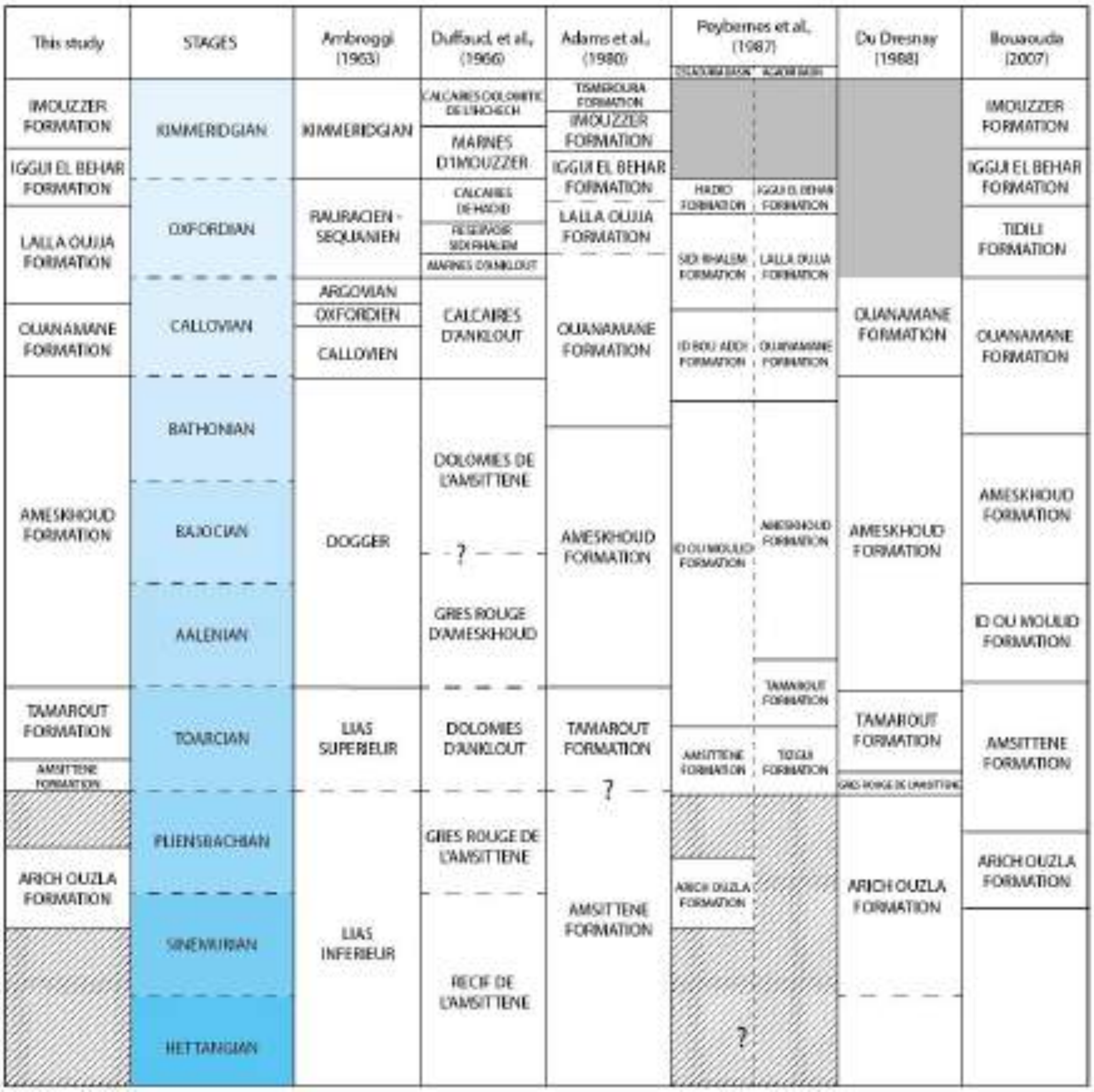

Figure 2.2: Stratigraphic overview, formations names and ages interpreted by the different authors.

2.1.1 Lower Jurassic

Arich Ouzla Formation – Sinemurian to Pliensbachian

The Arich Ouzla Formation is the oldest carbonate formation identified in the basin for the Mesozoic era. This formation is observed onshore in the core of the Jbel Amsittène (Arich Ouzla and Ida Ou Azza salt mines) in the North of the EAB and a similar unit is observed in the Tidzi diapir and in the well Essaouira-1 (ESS-1). Offshore, the wells Essaouira-1X (ESS-1X), Essaouira West-1 (ESW-1) and DSDP site 547 reached the Trias and show levels equivalent to the Arich Ouzla Formation. The formation was first identified by Roch (1930) as part of the “Jurassique lagunaire”, where he found bivalves, gastropods and belemnites. The same outcrop was later re-interpreted as a “reef bioherm” and dated from the Upper Sinemurian (Lotharingien) using brachiopods, bivalves and crinoids fauna identification by Golo and Dubar (Duffaud, 1960). The name Arich Ouzla Formation comes from a revision of the outcrops which stresses that the facies and the macrofauna were not diagnostic of a reefal environment and that the name “récif de l’Amsittène” should be abandoned (Du Dresnay, 1985; Du Dresnay, 1988). The thickness of this formation is between 0 and 70 m in outcrop, and up to 630 m in well ESS-1 (Du Dresnay, 1988).

This succession represents the first carbonate sedimentation in the region since the Devonian. Three facies are present (Bouaouda, 2007). The base is composed of massive dark dolomite, withg moldic porosity (Du Dresnay, 1988), where ooid phantoms can be recognised towards the top. A second facies consists of oncoidal wackestone with abundant brachiopods, foraminifera, gastropods, bivalves, and few solitary corals and bryozoans. These wackestones have been interpreted by Bouaouda (2007) as distal ramp carbonates. The upper part of the Arich Ouzla Formation is composed of oolithic dolomitic limestones overlain by fossiliferous pink dolomites with brachiopods and bivalves. The top of the formation has been eroded by the red conglomerates of the Amsittène Formation.

Amsittène Formation – Pliensbachian?-Toarcian

This formation has been revised by Hofmann (2000), who discriminates two members, Imerhrane (present only in the central part of the Argana Valley) and Aït Mbarek. The Imerhrane Member consists of sedimentary cycles of carbonate, mudstone intercalated with clayey mudstone and muddy siltstone, changing in each cycle to shaly and clayey mudstones. At their top, cycles consist of muddy sandstones, horizontally laminated and cross-bedded sandstones. Halite pseudomorphs are common throughout the succession. Up to 10 cycles occur in the succession. The clayey mudstones and muddy siltstones intercalations were interpreted as flooding over playa flats. Arid periods would have dried out the lakes and allow the aeolian sheet sands and sand dunes to cover the saline mudflats. The Aït Mbarek Member fluvial conglomerate lays uncomformably on top of the Imerhrane Member. A thin sabkha unit overlies this member with an angular unconformity before deposition of marine limestones.

More commonly, clay and red conglomerates of the Aït Mbarek Member rest directly with an erosive unconformity on doleritic basalts. The conglomerates present Paleozoic brecciated elements (quartzites, green schists, and few eruptive rocks) and are difficult to distinguish from the Triassic deposits where the basalts are missing. In most of the outcrops observed, the red conglomerates characterise the base of this succession, and are followed by red sandstones alternating with gypsiferous clay or marl levels. This succession has been interpreted to reflect evolution of the depositional environment from braided rivers to lakes and lagoons (Ambroggi, 1963; Bouaouda, 2007) or to a flood plain (Ouajhain et al., 2011).

Tamarout Formation – Toarcian

The Amsittène Formation is overlain by transgressive limestones which mark the end of a small deformation phase during the Lower Lias (Ambroggi, 1963). Around the SE of the Agadir-Essaouira Basin (Amskroud Anticline, Imouzzer anticline and Lgouz anticline), the Upper Lias is about 150 to 250 m thick and has been divided into three units. A first unit made of massive dolomitic limestones is followed by a middle unit composed of an alternation of marls, sandy marls and limestones. The upper unit is composed of limestone beds, with some sandy, marly or gypsiferous interbeds. These are associated with birds eye structures and stromatolites (Bouaouda, 2007). The basal beds of this formation indicate a marine influence characterised by carbonate sedimentation with brachiopods and bivalve debris. Towards the top of the succession, the facies become more marly and sandy, especially in the eastern part of the basin. Some anhydrite levels are developed on the western part. These changes in sedimentation show the evolution from open marine to lagoonal sedimentation which mark a regression (Ambroggi, 1963).

On the Jbel Amsittène, in the NW of the basinthe relation between this formation and the formation below is comformable. The Tamarout formation is mainly composed of laminated evaporitic dolomites, with stromatolite levels, and some few marine facies such as oolithic and peloidal limestones. In the first member, vuggy pink to yellow dolomites alternate with marls, marly gypsiferous limestones, cyclic stromatolites and breccia dolomites. The upper part of this formation contains ooidal and peloidal carbonates. In addition, gypsiferous marls and levels of collapse-breccias have been identified. These facies have been interpreted as sediments of an evaporitic carbonate ramp, with supratidal, lagoonal and sabkha environments (Bouaouda, 2007; Michard et al., 2011; Ouajhain et al., 2011, 2009).

2.1.2 Middle Jurassic

Ameskhoud Formation – Aalenian to Bathonian

Above the carbonate series of the Upper Lias, the Ameskhoud Formation is a siliciclastic red-brown unit. This formation is missing along the Amsittène Anticline, in the NW of the basin. In the SW part of the basin, the contact with the underlying unit is defined by a hardground level or an abrupt transition from carbonates to terrigenous facies. In the eastern part of the Agadir-Essaouira Basin, towards Imi n'Tanout, this formation locally lies directly on the Triassic red clays of the Argana Group, or uncomformably above Paleozoic basement (Ambroggi, 1963; Bouaouda, 2007).

The Ameskhoud Formation is up to 350 m thick with conglomerates, coarse or fine sandstones, sandy clays and marls. In the Imouzzer Anticline, the marls are dominant in alternation with sandy clays, fine and coarse poorly consolidated sandstones, and few conglomerate levels. Close to the top of the formation, few dolomitic sandy carbonates were deposited below a least 20 m of red marls and sandstones (Ambroggi, 1963). In the Anklout Anticline, the base of the succession is composed of a succession of erosive conglomerates with Paleozoic pebbles, red clays and up to 1 m thick layers of fine to coarse sandstones which present cross-bedding, parallel laminations and ripple marks. The sediments fine upsection, and the middle of the succession is mainly composed of clays and fine to medium sandstones. The upper part of the succession is an alternation between carbonates and terrigenous levels. The first carbonate level is made of oolithic limestone followed by stromatolites and marly dolomites. This series is still dominated by cross-bedded thin sandstones levels and yellowish to greenish clays, probably lagoonal. The carbonate levels are partly dolomitic, silty, often micritic and contain ooliths, fragments of gastropods and bivalves, echinoderms, foraminifers (*Pseudocyclamina*) and brachiopods (*Rhynchonella*). As in the Imouzzer Anticline, the top of the formation shows red clay levels alternating with sandstones (30 m thick; (Bouaouda, 2007).

Ambroggi (1963) interpreted the conglomerates as marine deposits, probably deltaic, which implies the uplift of a continent at the end of the Lias, at the east of the basin. The sandstones and conglomerates have been interpreted by Bouaouda (2003) as fluvial deposits. The Upper part of the unit, an alternation between thin sandstones and clays, was interpreted as lagoonal sediments (Bouaouda, 2007; Michard et al., 2011; Ouajhain et al., 2011).

Ambroggi (1963) interpreted the conglomerates as marine deposits, probably deltaic, and implies the uplift of a continent at the end of the Lias, at the east of the basin. The sandstones and conglomerates have been interpreted by Bouaouda (2003) as fluvial deposits. The Upper part of the unit, alternation between thin sandstones and clays was interpreted as lagoonal sediments (Bouaouda, 2007; Michard et al., 2011; Ouajhain et al., 2011).

Ounamane Formation – Callovian

The Callovian deposits were first described as grey and green dolomites, with very abundant brachiopods (Terebratula and Rhynchonella), gastropods (Nerinea) and oysters. Those dolomitic levels are alternating with marly carbonates also rich in brachiopods (Ambroggi, 1963). The Ouanamane Formation was later recognised by Adams (1979) as an association of four lithofacies at the base of the formation which are oolite grainstone, mudstone, dolomite and mixed bioclast and intraclast wackestones and packstones. The first facies consist of well-sorted fine oolithic grainstone with few shell fragments and quartz grains. The mudstone facies is more common at the base of the formation and present up to 10 percent of complete bivalves and foraminifera. The dolomites make the main part of the succession in the eastern part of the basin while in the middle and western part the wackestones / packstones dominate. The dolomitic base of the Ouanamane Formation, essentially developed towards the East and North-East was differentiated by Bouaouda (2007) and named Formation Oumsissène. The dolomitic levels are often silty, and some layers present structures that have been identified as potential algal mats (Adams, 1979; Bouaouda, 2007). The wackestones and packstones facies contain few ooids and more bioclasts (bivalves, brachiopods, gastropods, echinoderms, foraminifers, corals, calcareous algae and bryozoa) and intraclasts. The lower part of the formation presents a small proportion of siliciclastics and is characterised by hard-grounds with a high bioturbation rate and oysters encrustments. The middle and upper part of the formation is characterised by fossiliferous limestones, well-bedded and rich in brachiopods, including different species of rhynconella and terabratula, alternating with marls and marly limestones. The fossiliferous limestones can be wackestones to oolithic grainstone and present a remarkable regularity in the fauna composed of brachiopods, bivalves including trichites, oysters, fragments of echinoderms and bryozoa. Encrusted and burrowed levels also occur regularly, the uppermost part of the formation gets more marly and the fauna diversity decreases with few ammonites, brachiopods, corals, ostracods and foraminifers (Adams, 1979; Bouaouda, 2007; Bouaouda et al., 2004; Michard et al., 2011; Ouajhain et al., 2011, 2009).

The Ouanamane Formation records show the development of open marine conditions, with a progressive growth of a carbonate platform from the West to the East (Adams, 1979; Ambroggi, 1963; Bouaouda, 2007; Ouajhain et al., 2011). The presence of microbial laminites or stromatolites at the base of the succession, have been interpreted as the development of a tidal flat during the beginning of the transition from continental to marine deposits (Adams, 1979). In the lower part of the formation, some levels contain gypsum pseudomorphs that characterise the development of rare sabkha environment (Michard et al., 2011). The mudstones, bioclastic wackestones and packstones, and the oolithic grainstones are interpreted as lagoonal to shoal deposits. The fossiliferous limestones are shallow marine, low energy subtidal deposits, more distal for the marls (Adams, 1979; Bouaouda, 2007; Ouajhain et al., 2011, 2009).

Biostratigraphic studies on foraminifera (Bouaouda, 2007; Bouaouda et al., 2004) show that the lower limit of the formation is Upper Bathonian to Lower Callovian age. The upper part of the formation is diachronic along the basin, from Middle Callovian towards the NW up to Upper Oxfordian toward the eastern part of the basin.

2.1.3 Lower Jurassic

Lalla Oujja Formation – Oxfordian

The Lalla Oujja Formation overlies the marls of the Ouanamane Formation throughout the Agadir-Essaouira basin. This formation is mainly made of dolomitic reefal limestones, and always capped by the Iggui-El-Behar Formation.

The Lalla Oujja Formation has been dated from the Upper Jurassic based on gastropods (Nerinea) and corals identification, the Cap Ghir outcrops have been particularly studied because the corals are very well preserved there, and the reef present a high diversity is skeletal assemblage. The Lalla Oujja Formation was first described by Ambroggi (1963) as a massive reefal dolomite unit that composes the crest of most of the anticlines studied. In the western part of the area, this unit shows lens shaped carbonated buildups that are surrounded by bedded bioclastic packstones, comprising oysters, echinoderms, calcareous algae, sponges and corals. Those beds thinabove the builds-up, which indicates a positive topography made by the reef units (Adams, 1979). In this part of the basin, the Igui-El-Behar Formation is reduced and hard to distinguish from the Lalla Oujja Formation, it is about 15 m thick and composed of bedded light limestones. Towards the East, the Lalla Oujja Formation and the Iggui-El-Behar Formation form two distinct units separated by a marly interval up to 10 m thick (Ambroggi, 1963; Adams, 1979; Bouaouda, 2007). Further east, in the Ameskhoud Anticline, this formation is composed of bioclastic wackestones, locally boundstones at the base, comprising oysters, bryozoans, echinoids, calcareous algae, a few corals and serpulites. These levels are quickly overlaid by three levels of massive dolomite with few phantoms of ooids, and large vugs, the massive aspect of those units is interpreted as showing a reef structure. Those massive levels are separated by bioclastics layers, and overlaid by non-fossiliferous marly limestones followed by a bioturbated bioclastic unit with bivalves, echinoderms, brachiopods, bryozoans, annelids and foraminifers (Bouaouda, 2007; Michard et al., 2011).

Studies focused on the Upper Jurassic coral reef system have been realised along Cap Ghir and the Anklout Anticline:

- The Tiguert-Cap Ghir Anticline is a W-E structure identified by Ambroggi (1963) and located about 40 km north of Agadir along the coast of the Atlantic Ocean. The age of the reef complex of Cap Ghir is still not well constrained, several dates have been given from Upper Oxfordian to Lower Kimmeridgian (Ambroggi, 1963; Ourribane et al., 2000). Two interpretations of the reef structures have been proposed. A first study on Cap Ghir discriminated two reefs of Oxfordian and Kimmeridgian age, on the base of benthic foraminifera (Ourribane et al., 2000). In a second study, the Cap Ghir reef outcrop is interpreted as a single reef, with three different coral associations, that presents four different erosive phases due to Plio-Pleistocene abrasive terraces. The interpretation for such a large reef structure located in this particular area and disappearing further north and further south would be a palaeorelief formed by a tilted bloc (Martin-Garin et al., 2007). The Cap Ghir outcrops show three different coral associations in the reef complex, and an important microbialites and microencrusters development.

    - The first coral association comprise 24 genera of corals, mostly massive and branching, the dominant genera identified are *Stylina, Psammogyra*, *Thamnasteria*, *Aplosmilia* and *Rhipidogyra*; some giant colonies of Complexastrea have also been found. The coral association is typical of a photoheterotrophic back-reef environment (Martin-Garin et al., 2007). Secondary builders are stromatoporoids, bryozoan, mollusks, microbialites (laminated stromatolites and peloidal thrombolites). The microencrusters are mainly *Bacinella*, some encrusting foraminifers, *Lithocodium*, and *Tubiphytes* (Ourribane et al., 2000). The *Tubiphytes* have been recently renamed *Crescentiella* and identified as possible symbiosis or encrustation between cyanobacteria and foraminifer (Senowbari-Daryan et al., 2008)

> The second coral association is dominated by colonies of *Dimorpharaea*, laminar microsolenoid corals that develop very thin platy-shapped colonies. This assemblage is typical of lower fore-reef slope (Martin-Garin et al., 2007). These colonies are associated with corals and sponges, and have a ground mass composed of reef debris and echinoids clasts with rare molluscs and bryozoans. Laminated stromatolites are also very common in this reef and are associated with *Tubiphytes*, *Bacinella* and *Lithocodium*. The role of the microbial crusts appears to be very important because they grew on the corals and stabilised the debris (Ourribane et al., 2000). We can notice that the same platy-shaped elements, interpreted by Martin-Garin et al. (2007) as *Dimorpharaea* colonies have been interpreted by Ourribane et al. (2000) as stromatoporoids.

> The third coral association is composed of 12 genera of corals, and dominated by Microsolena, Isastrea, Stylina, Etallonasteria and Thamnasteria. The size of the coral colonies are reduced compared to the colonies of the first coral association, which has been attributed to the wave action or a lower light input. This observation associated with the disparity and the reduced diversity of corals characterise an upper fore-reef slope (Martin-Garin et al., 2007). Like in the second association, the colonies are associated with sponges, and have a ground mass composed of reef debris and echinoids clasts with rare molluscs and bryozoans. Laminated stromatolites are also very common and associated with Tubiphytes, Bacinella and Lithocodium(Ourribane et al., 2000).

The work on the Izwarn section (Anklout Anticline) is particularly relevant as the authors identified the ecological succession stages for the settling of a reef initiated by *Dimorpharaea* colonies (Figure 2.3), which is the main type of reef builder we encountered in the area of study. The pioneer stage is the stabilisation stage preceding the coral reef settlement. It is composed of a muddy substrate with bioclastc deposits made of echinoderms (echinoid spines and crinoid ossicles) and gastropods, and a small portion of brachiopods, bivalves and coral fragments. A few colonies of *Dimorpharaea* are also present; those corals developed platy-shapes in a low light environment. Associated with the absence of sedimentary structures, these observations leaded to conclude that the pioneer stage of the reef development was below the storm wave base. The colonisation stage is characterised by the development in situ of small platy colonies of *Dimorpharaea* that are interpreted to develop in a deep fore-reef environment. The last stage of the reef settlement was the diversification stage that is characterised by the association of *Dimorpharaea*, which is still the most abundant genus, with other corals genus such as *Enallhelia*, *Stylosmilia*, and *Calamophylliopsis*. The reef framework has been highly supported by thick stromatolitic crusts that developed at the surface of the bioconstruction. While the conditions for coral growth were not optimal, the impact of the microbial carbonates on the reef development is prominent (Olivier et al., 2012).

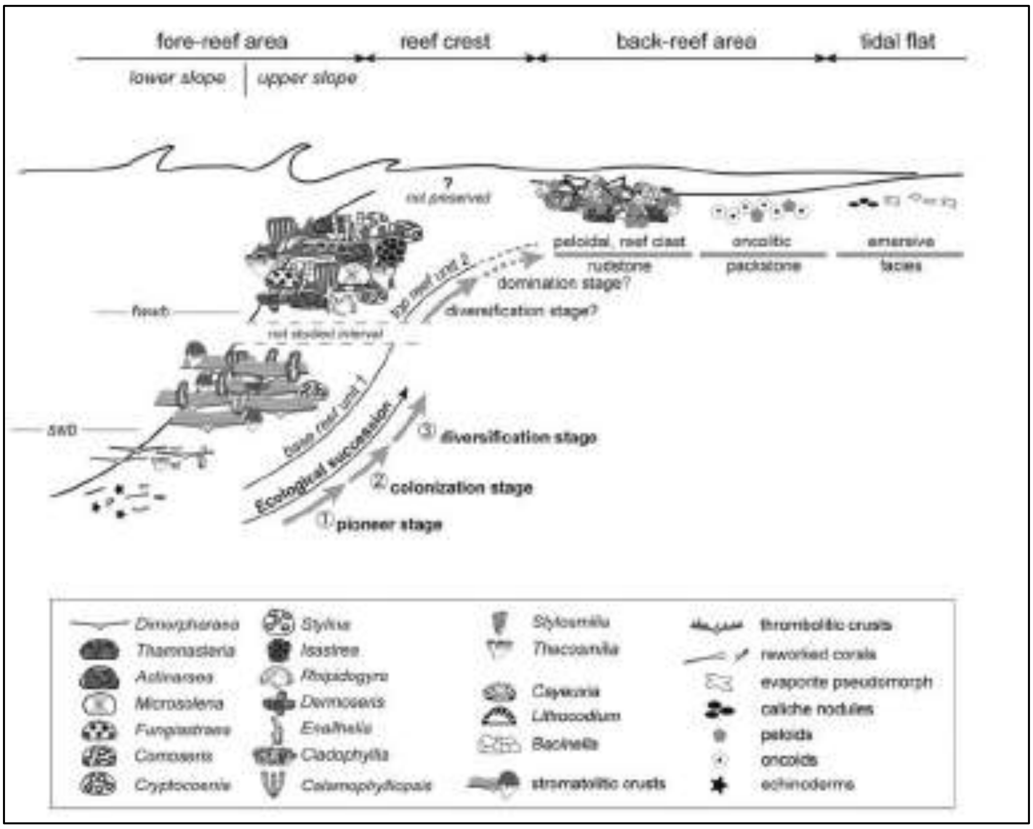

Figure 2.3: Schematic illustration showing inferred palaeoenvironmental settings
and ecological succession during the deposition of the studied reefs along the Izwarn section. From Olivier et al., 2012.

In the northern part of the study area, along the Jbel Amsittène, the base of the Lalla Oujja Formation is composed of red siliciclastic grainstones, with cross-beds and channels which is the stabilisation phase of the reef edification. A bioclastic unit has developped on top of this massive unit, it is rich in bivalves, echinoderms and gastropods, with still a sandy fraction. This unit is regularly bedded in an alternation of more or less sandy or silty beds, sometimes oolithic or rich in benthic foraminifera and skeletal grains. After this transition phase, a first reef complex develops; this bindstone unit is mainly composed of laminar microsolenoid corals (*Dimorpharaea?*) with bivalves and echinoids. This facies should be the colonisation stage of the reef development. Above this level, the diversification stage of the reef complex is characterized by a boundstone, sometimes floatstone texture with various corals associated with echinoderms and gastropods. Some serpulids and encrusting organismes are also common such as red algae and tubiphytes. This formation has been interpreted as a reef barrier environment (Bouaouda, 2007; Ouajhain et al., 2011, 2009).

In all the Agadir-Essaouira Basin, the Lalla Oujja Formation is dominated by reef deposits, with laterally bioclastic and marly deposits. The deposits environments associated with those facies vary from fore-reef to back-reef. Finally different deposits times can be notice which results from eustatic or tectonic movements (Michard et al., 2011).

Iggui-El-Behar Formation

The Iggui-El-Behar Formation can reach 140 m thick in the NW of the area of study (Jbel Amsittène), but is just a few meters thick further south, on top of the Cap Ghir reef complex and up to 20 m thick in the eastern part of the Agadir-Essaouira Basin, around the Imouzzer Anticline (Adams, 1979; Ambroggi, 1963; Bouaouda, 2007). A publication about the Jurassic foraminifers *Sievoides kocyigtti* has been used to date the Iggui-El-Behar Formation (Bouaouda, 2009). According to this study, the age of this formation would be Upper Jurassic to Lower Kimmeridgian, which is consistent with the previous datations (Adams, 1979; Ambroggi, 1963).

In the NW of the study area, along the Jbel Amsittène, the Iggui-El-Behar Formation is composed an alternation of facies that shows a high energy variability in the regime of deposition (Ouajhain et al., 2011). Beds with algally-coated grains and nereids gastropods constitute the base of the formation, which constitute back-reef lagoon deposits (Adams, 1979). Above this layer, the formation, is well stratfied andcomposed of peloidal limestones associated with some algal stromatolites, fossiliferous oolithic packstones and grainstones, and micritic limestones with dasyclad algae. This facies association is characteristic of a lagoonal environment, and some pink dolomite beds might be linked to emersive episodes (Adams, 1979; Bouaouda, 2007; Ouajhain et al., 2011, 2009). Toward the center and the east of the onshore basin, the Iggui-El-Behar Formation comprises peloidal packstones to mudstones with a sparse fauna, usually ostracods, nereids gastropods, foraminifers, dasyclad algae and thin-shelled bivalves. Those beds with stromatolites, bird-eyes structures, evidences of evaporites, and few brecciated beds, interpreted as dissolution breccias, are characteristic of lagoon to sabkha environments (Adams, 1979; Ambroggi, 1963; Bouaouda, 2007). The base of some beds of pale micrites, show the presence of darker angular mud fragments, with a fining-upward pattern passing into shales. Those deposits, characteristics of storm deposits reworking the layer below are quite common in this formation (Ager, 1974). The formation records a shallowing-upward sediments deposition, which continues as the formation is overlain by red marly deposits (Adams, 1979).

Imouzzer Formation – Kimmeridgian

The Imouzzer Formation conformably overlays the Iggui-El-Behar Formation. It is defined by red clay and marls alternating with dolomitic levels. This formation is well displayed across the basin, and has been studied along the Amsittène Anticline, the Cap Ghir Anticline and the Ameskhoud Anticline (Ambroggi, 1963; Adams, 1979; Adams et al., 1980; Bouaouda, 2007; Ouajhain et al., 2009). In the north of the area, along the Jbel Amsittène, the base of the formation is more silty than along Cap Ghir. The base of the formation is described as an alternation between thick beds of "chocolate" marls and bioturbated sandy dolomitic wackestones to packstones with oysters, bivalves, foraminifera, and some echinoids, gastropods and bryozoans. In the chocolate marls, few siltstones and yellow marls have been interpreted as lacustrine deposits. The middle part of the formation is an alternation between "chocolate" marls with thin siltstone beds, and bioturbated dolomitic wackestones to packstones, with peloids, echinoids, gastropods, bivalves and foraminifers. The upper part of the formation is dominated by "chocolate" marls, with few dolomite levels including anhydrite and foraminifera (Bouaouda, 2007). Further south along the Cap Ghir, and further east along the Imouzzer Anticline, the Imouzzer Formation is similar but not silty, or sandy. It is composed of an alternation of thick beds of choclate marls and grey to yellow carbonates, often dolomitic, that present a fauna essentially composed of oyesters (and also some gastropods in Cap Ghir only). Around the Ameskhoud and Lgouz anticlines, the marls are remarkably thick at the base, and the limestone levels are very dolomitic (Ambroggi, 1963).

The "chocolate" marls of the Imouzzer Anticline have been interpreted as emerged and proximal deposits (Ambroggi, 1963) and as shallow marine deposits with a strong silic-clastic influx (Bouaouda, 2007). The limestones deposits would be open marine, up to lagoon deposits toward the top (Ambroggi, 1963).

Tismerroura Formation – Tithonian

The Imouzzer Formation evolves slowly to the Tismerroura Formation, with an increase of the limestone beds deposits and the disappearance of the redish marls deposits. This formation shows strong variations into the Basin. In the very north of the area, the Kimmeridgian deposits are essentially composed of a thick gypsum level, around 100 m thick. To the south, along the Imouzzer Anticline, the base of the formation show some gypsum beds in the marls, and above is composed of dolomitic marly limestones with some brachiopods, bivalves and oysters. This formation has been here interpreted as lagoonal deposits. This is similar to the deposits in the Cap Ghir, but in the Cap Ghir, the carbonates are not marly nor dolomitised, and present a higher diversity of fauna, including calpionella, gastropods, brachiopods, bivalves and oysters. The upper part of the formation is also different because it is composed of a thick level of grey and red marls without fauna. Toward the SE of the basin, along the Anklout Anticline, the gypsum at the base of the formation is not present, and the formation is composed of dolomtic marly limestones at the base, and dolomitic limestones in a regular alternation with greenish marls levels in the middle and upper part of the formation (Ambroggi, 1963; Adams, 1979; Adams et al., 1980 Zühlke et al., 2004).

In the eastern part of the basin, inner lagoonal sandstones, anhydrites and dolomites developed. The formation is considerably thinner than in the western part of the basin (Zühlke et al., 2004). This entire formation has been interpreted as a succession of cyclic lagoonal sediments, deposited in a wide lagoon caracterise by a lack of energy (Ambroggi, 1963).

2.2 Moroccan Atlantic salt basin

During the Late Trias to Early Jurassic, thick salt deposits formed along the Moroccan Atlantic Margin (Hafid et al., 2000, 2006; Davison, 2005) (Figure 2.4). The long-lasting salt deformation has been controlling the distribution of the sedimentary depocenters across the basin (Hafid et al., 2000; Davison, 2005; Muniz-Pichel, 2018).

In the offshore EAB, two different salt domains were identified (Tari et al, 2003, 2012; Muniz-Pichel, 2018). The Essaouira segment, in the north is dominated by allochtonous salt bodies, salt tongues and seaward-verging diapirs (Tari et al., 2000, 2017). The allochtonous salt sheets are associated to multiple feeders, which indicates the formation of mini-basins during the middle Jurassic to Aptian (Tari and Jabour, 2013; Muniz-Pichel., 2018). The Agadir segment is characterised by dominantly symmetric salt diapirs oriented NE-SW which delimitate multiple mini-basins. The geometries and orientation of the diapirs indicate their formation by sedimentary loading and salt expulsion (Muniz-Pichel., 2018).

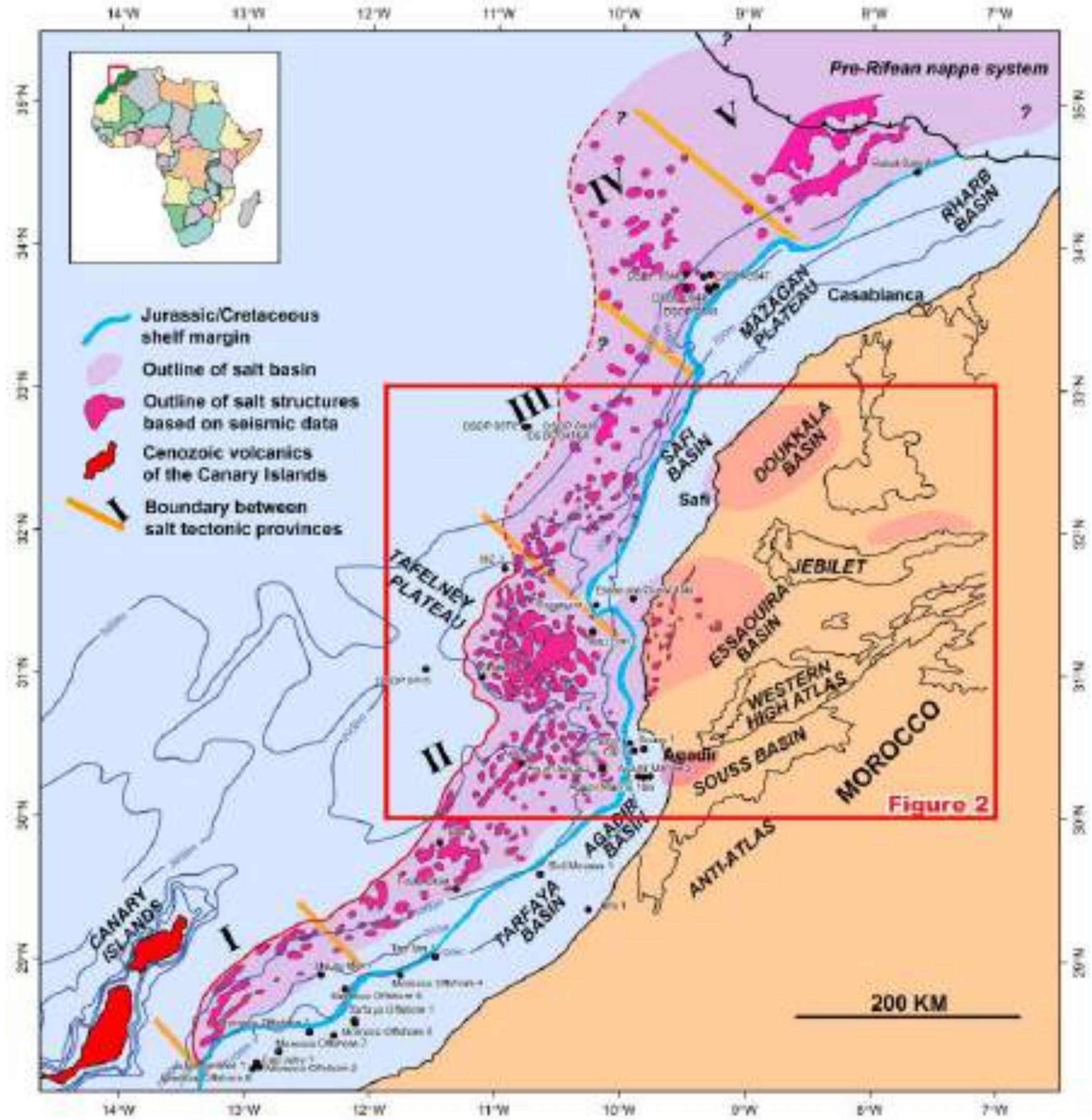

Figure 2.4: Salt basins along the Moroccan Atlantic Margin.
From Tari et al., 2017.

In the Central High Atlas, folding during the opening of the Jurassic basins has been attributed to salt diapirism (Michard et al., 2011; Saura et al.,2014; Verges et al., 2017; Teixell et al., 2017; Malaval, 2017; Joussiaume, 2017). Onshore, the EAB has multiple NE-SW and E-W anticlines, but only the Tidzi and Jbel Hadid salt diapirs in the North of the Essaouira segment reached the surface.

2.3 Dissolution-collapse breccias

Dissolution-collapse breccias, also called solution-collapse breccias result from the withdrawal of anhydrite or gypsum. The karstic features formed by the dissolution of the evaporites induce the collapse of the overlying carbonate layer. The breccias formed are composed of angular carbonate clasts and present a sharp flat base and an irregular top (Friedman, 1997).

In the Chiantla Quadrangle of north-west Guatemala, dissolution breccias have been first identified as non-depositional features, dominated by dolomite, entirely clast supported and presenting an extremely poorly sorting (Blount and Moore, 1969). These authors also identified the matrix surrounding the dolomite clasts as sparry calcite and dolomite. They also indicate that lithologies of clasts compatible with evaporite deposits and lateral presence of evaporites in the neighboring outcrops remain the most definitive argument towards this mode of breccias formation.

Finally, main types of dissolution-collapse breccias were distinguished in the gypsum palaeokarst system of the Central Spitsbergen in Svalbard. Collapses of dolines or cavities form thick breccias (up to 200 m) associated with V-structures or pipes and are common towards the centre of the basin, where the salt deposited was thicker. The margins of the basins are characterised by more continuous breccias (30 cm to 10 m) interbedded with non-deformed carbonate beds. These breccias are dominantly coarsening-uppard and present a sharp and flat lower boundary (Eliassen and Talbot, 2005).

2.4 Scotian Basin

The opening of the Central Atlantic Ocean separated North America and North-west Africa, creating syn-rift deposition during the Trias, followed by the continental break-up in the Sinnemurian (Labails et al., 2010). These authors indicate a very slow oceanic accretion until the Bajocian and a severe change to faster spreading rates until the Tithonian, when the accretion rate slowed again. The Scotian Basin and Moroccan Atlantic basins therefore recorded very similar deposits during the Jurassic (Figure 2.5).

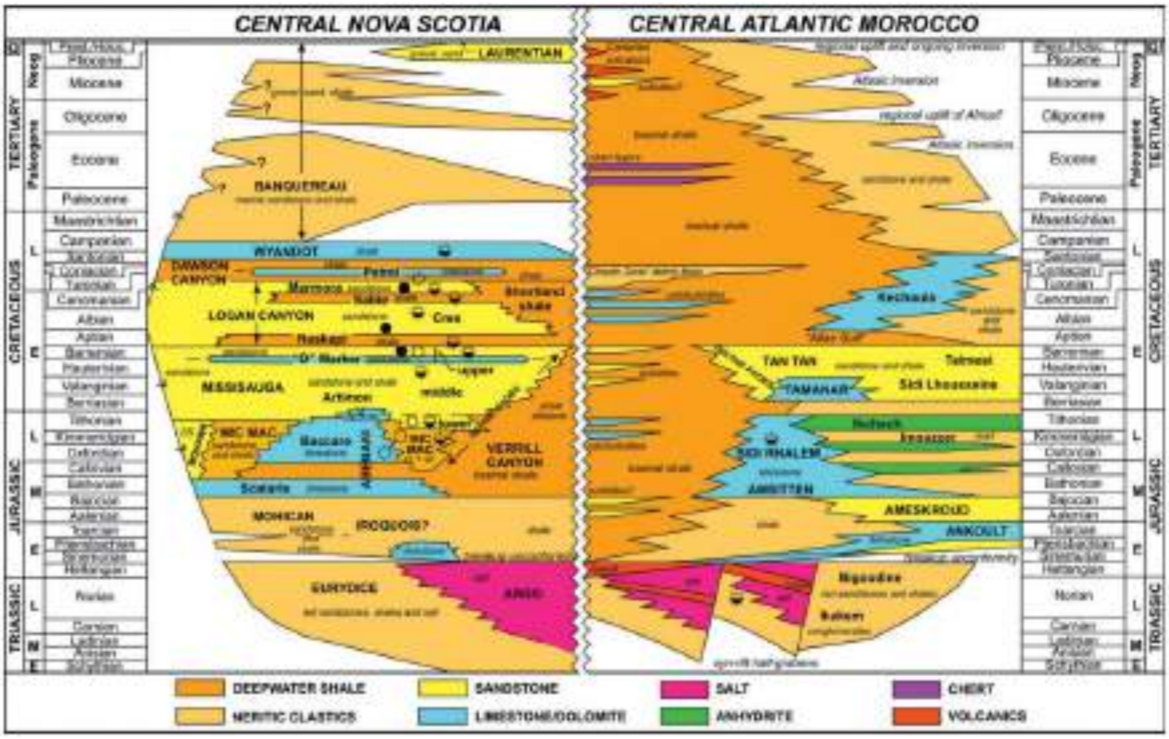

Figure 2.5: Simplified stratigraphy of the Scotian Basin and Moroccan Atlantic Margin. From Tari et al., 2012.

Both basins present continental syn-rift sequences, dominated by siliciclastic and salt deposits. The salt deposits have been mapped along both basins, and similar trends have been identified, with distinction of 5 salt domains distributed identically along the margins (Tari et al., 2012). Both basins record the CAMP basalts (Marzoli et al., 1999; Cirilli et al., 2009), which are a good chronological marker. The Jurassic sedimentation along the two margins was dominated by carbonate deposition widely influenced by siliciclastic influx.

The early Jurassic post-rift sequence of the Scotian basin is dominated by evaporites and tidally-influenced dolomites of the Iroquoi Formation (Jansa and Wade, 1975). It is followed by the Mohican Formation siliciclastic and shale deposits. During the Middle Jurassic, a transgression affected the Atlantic Ocean, and thick carbonate deposition persisted, until the Early Cretaceous (Wade and MacLean, 1990).

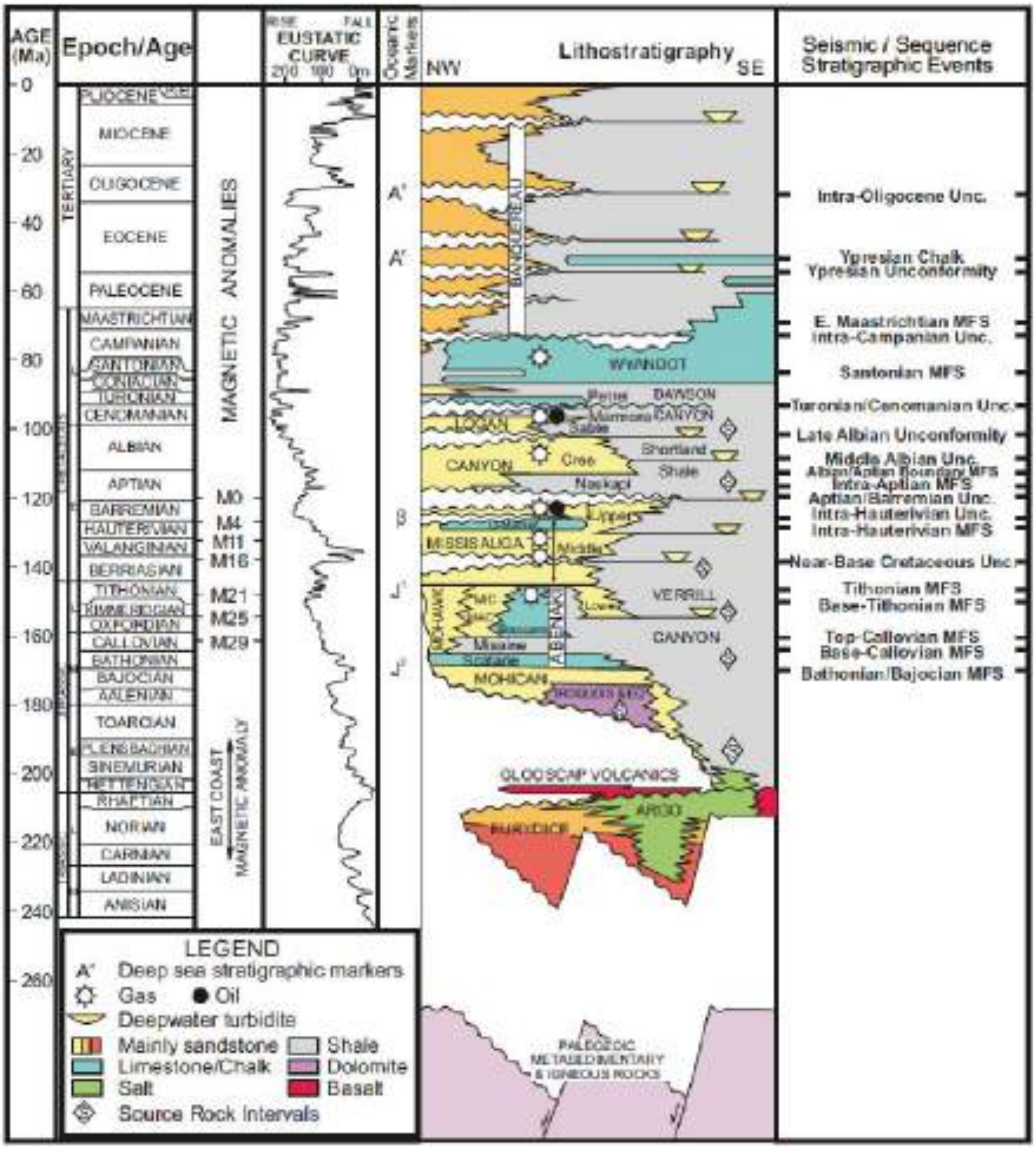

Figure 2.6: Stratigraphic chart of the Scotian Basin. Modified from Weston et al., 2012.

The Middle Jurassic to Early Cretaceous records ths deposition of the Abenaki Formation. This formation is divided in 4 members characterised by different lithologies (Jansa and Wade, 1975; Eliuk, 1978). The base of the Scatarie Member, directly following the Mohican Formation is diachronous, dated Bathonian to early Callovian (Weston et al., 2012). This member is dominated by oolitic packstones and grainstones, alternating in cycles with oncolitic horizons, fossiliferous wackestones and silty or sandy interbeds. This member has been interpreted recording lagoonal to shallow shelf deposits (Eliuk, 1978).

The second member of the Abenaki Formation is the Misaine Shale Member, which recorded the late Callovian maximum flooding surface (Weston et al., 2012). It is composed of dark calcareous shales with thin calcareous siltstones and sandstones developing towards the North, where the Sable delta developed (Eliuk, 1978; Eliuk, 2016).

The following Baccaro Member was deposited form Oxfordian to Early Tithonian (Weston et al., 2012). It is composed of a thick continuous platform in the SW, and the laterally equivalent Mic-Mac Formation in the NE is dominted by siliciclastic deposits formed by the activity of the Sable delta (Eliuk, 2016). The Baccaro Limestone Member present various facies, with minor oolitic, stromatoporoidal packstones and floatstones, coral-stromatoporoids reefs and sponge bioherms (Eliuk, 1978, 2016; Ellis et al., 1985; Pratt and Jansa, 1989) which compose reef, bank and platform environments. This interval in particularly important as is composed the reservoir unit of the Deep Panuke field (Weissenberger et al., 2006).

The last Member is the Artimon Member, dated Tithonian to Beriasian (Weston et al., 2012) and dominated by deep water sponge mounds (Eliuk, 1978, 2016; Ellis et al., 1985; Pratt and Jansa, 1989).

Chapter 3: Geological Setting

The Moroccan Atlas massif is composed of three mountain ranges: The High Atlas, Middle Atlas and Anti-Atlas. The High Atlas has a SW-NE trend resulting from the inversion of the intra-continental Atlas basin (Figure 3.1). The Atlasic continental rifting began during Middle-Late Triassic (Michard et al, 2008) and is part of the opening of the Alpine Tethys which is a consequence of a major extensive phase linked with the opening of the Central Atlantic Ocean. The first Triassic deposits are composed of red siliciclastics and evaporites. These layers, are known to act as decollement levels.

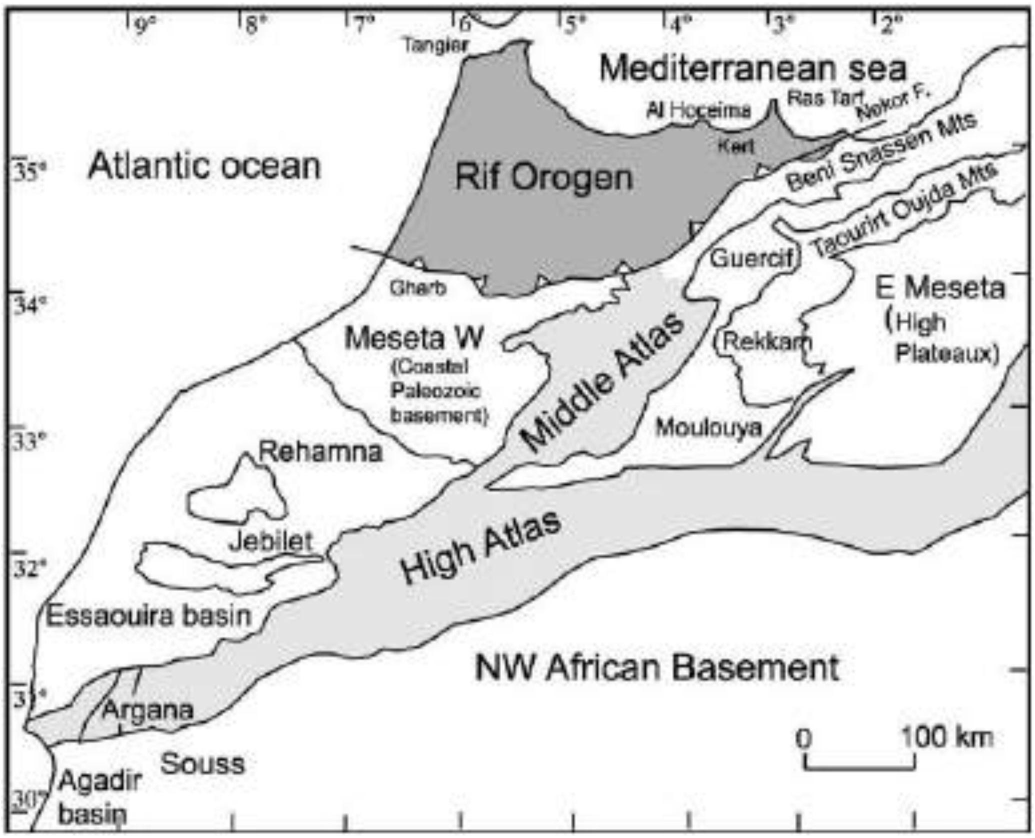

Figure 3.1: Tectonic sketch map of NW Morocco, with location of the Agadir Basin.
From Ait Brahim et al., 2002.

Normal faulting during rifting created a horst and graben system, thinning the continental crust and deepening the Atlas Basin. This tectonic subsidence created significant accommodation space, which was filled by the deposition of thick sediments from the

Triassic until the Cretaceous. During the Middle Jurassic in the Tethys Gulf, the thinning of the continental crust was significant enough to allow the development of magmatism along normal faults. The extensional phase of the Atlasic Basin ceased immediately after this event and was followed by thermal subsidence. The accommodation space began to decrease as the sediments filled the basin.

The Moroccan Atlantic Basin, outcroping in Western High Atlas resulted from the opening of the Atlas Basin and the Atlantic rifting and drifting phases. This basin was partially separated from the Tethys Gulf by the Western Moroccan Arch (Frizon de Lamotte et al, 2009; Saddiqi et al., 2009) which caused this basin to evolve differently from the Atlas Basin. The Moroccan Atlantic margin formed during the Late Permian to Early Jurassic rifting phase and constituted NE trending half-grabens. From Permian to Sinemurian, those half-grabbens were filled by over 2000 m of red continental deposits including conglomerates, sandstones and clays (Figure 3.5), overlain by basalt flows (Hafid et al., 2008).

3.1 Regional structures

The Atlas system trends E-W and consists of intra-continental fold and thrust belts with plateau areas such as the Moroccan Meseta. The geometry of the Atlas system is inherited from the Triassic rifting of the Central Atlantic as well as the Western part of the Tethys rift (Favre and Stampfli, 1992). The Atlas Mountain range presents a strong longitudinal asymmetry with a mean elevation of 1500 m the Moroccan High Atlas, 1050 in the Saharan Atlas, and only 600 m in the Tunisian Atlas. South of the South Atlas Front (Figure 3.2), the only uplifted area, separated from the Atlas by foreland basins is the Anti Atlas domain, SW of Morocco (Frizon de Lamotte et al., 2009).

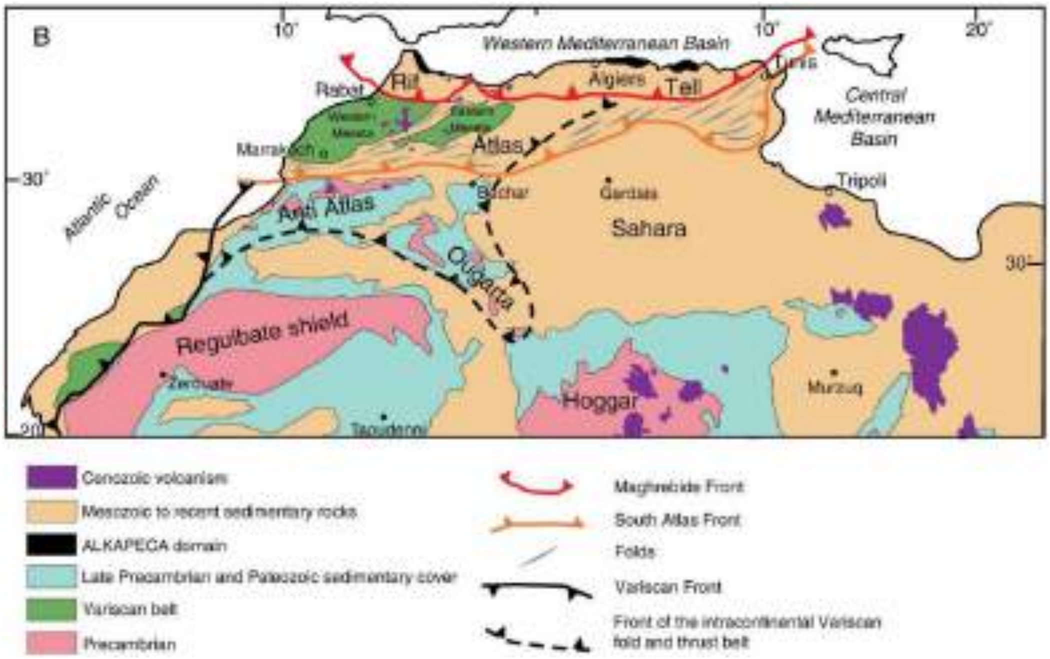

Figure 3.2: Main structural domains of NW Africa. From Frizon de Lamotte et al., 2009

The Essaouira Basin is located along the West coast of Morocco and belongs to the Western High Atlas domain (Figure 3.4). It is divided in four main structural domains, separated by major faults systems (Figure 3.3).

3.2 Permian pre-rift

In Morocco the Palaeozoic basement, mostly composed of igneous and continental sedimentary Permian rocks, crops out only in a few areas in the High Atlas and Middle Atlas (Figure 3.4). During the Variscan Orogeny, these rocks experienced an inhomogeneous deformation along shear zones showing two main orientations N20–45°E and N70–90°E and acting as weakness zones during the end of the Paleozoic (Laville et al., 2004; Le Roy and Piqué, 2001; Pique et al., 1998).

One of the most extensive outcrops of Paleozoic rocks is located along the Argana Valley in the Western High Atlas (Figure 3.8). The Argana Basin is situated 30km NW of Agadir and corresponds to a 80 km long SW-NE valley between Ameskhoud and Imi n'Tanout with Permian and Triassic strata,composed of three formations: Ikakem Formation, Timezgadiwine Formation and Bigoudine Formation, which rest with an angular unconformity on the Paleozoic basement (Figure 3.5). Upper Permian dykes pass through the Cambrian basement schists and the Ikakem Formation (Aït Chayeb et al., 1998).

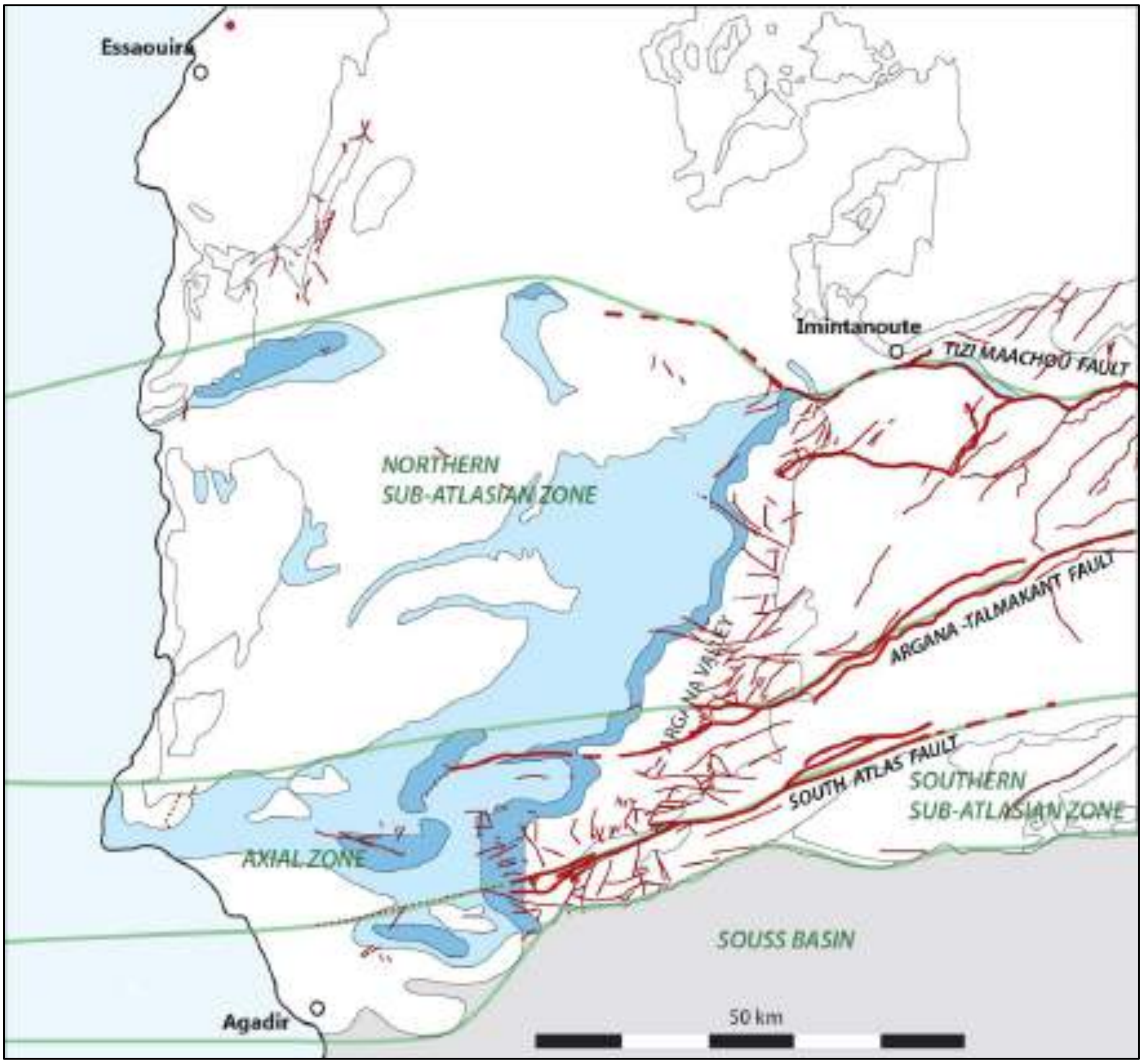

Figure 3.3: Major faults and structural domains in the Essaouira-Agadir Basin. Faults location from the geological maps 1:10000 of Tamanar, Imi'n Tanoute, Argana, Taghazout and Khemis Meskala. Structural domains Zühlke et al. (2004).

The Ait Driss Member of the Ikakem Formation is Pensylvanian in age (Stephanian) (Ambroggi, 1963; Doubinger and Roy-Dias, 1985). The Tourbihine member of the Ikakem formation has been dated Lower Permian (Autunian) by the same authors, with the top of this member dated as Upper Permian by vertebrate fossils and ichnofauna studies (Hminna et al., 2012; Jalil, 1999).

The basal conglomerates have been interpreted as debris-flow with large blocks, which indicate a tectonic activity as well as the existence of high relief during this period. The sediment was sourced from the Paleozoic formations surrounding this area, which were outcropping at this time. The conglomerates evolve to sandstones and silts, with less evidence of synsedimentary tectonics events as the basin deepened during the Upper Permian. This facies evolution is observaed in different small-scale basins along the Argana Valley which were probably part of a single large basin called the Souss Basin and faulted between Lower Permian and Upper Permian by the NNE-SSW Agadir ou Anzizen fault. (Saber and Broutin, 2002; Saber, 1995). One interpretation of the opening of such intracontinental basins by Saber at al. (1995, 2002, 2007) is a Hercynian late-orogenic transpressional system replaced in the Upper Permian by an extensive phase associated with the relaxation or collapse of the chain. The geochemistry of the Upper Permian and Lower Trias volcanism identified by Aït Chayeb (1998) is dominated by alkaline to transitional basalts, influenced by their establishment in an extensive geodynamic context. These deposits and magmatic rocks mark the pre-rifting phase of the Upper Permian which predates the Triassic extension.

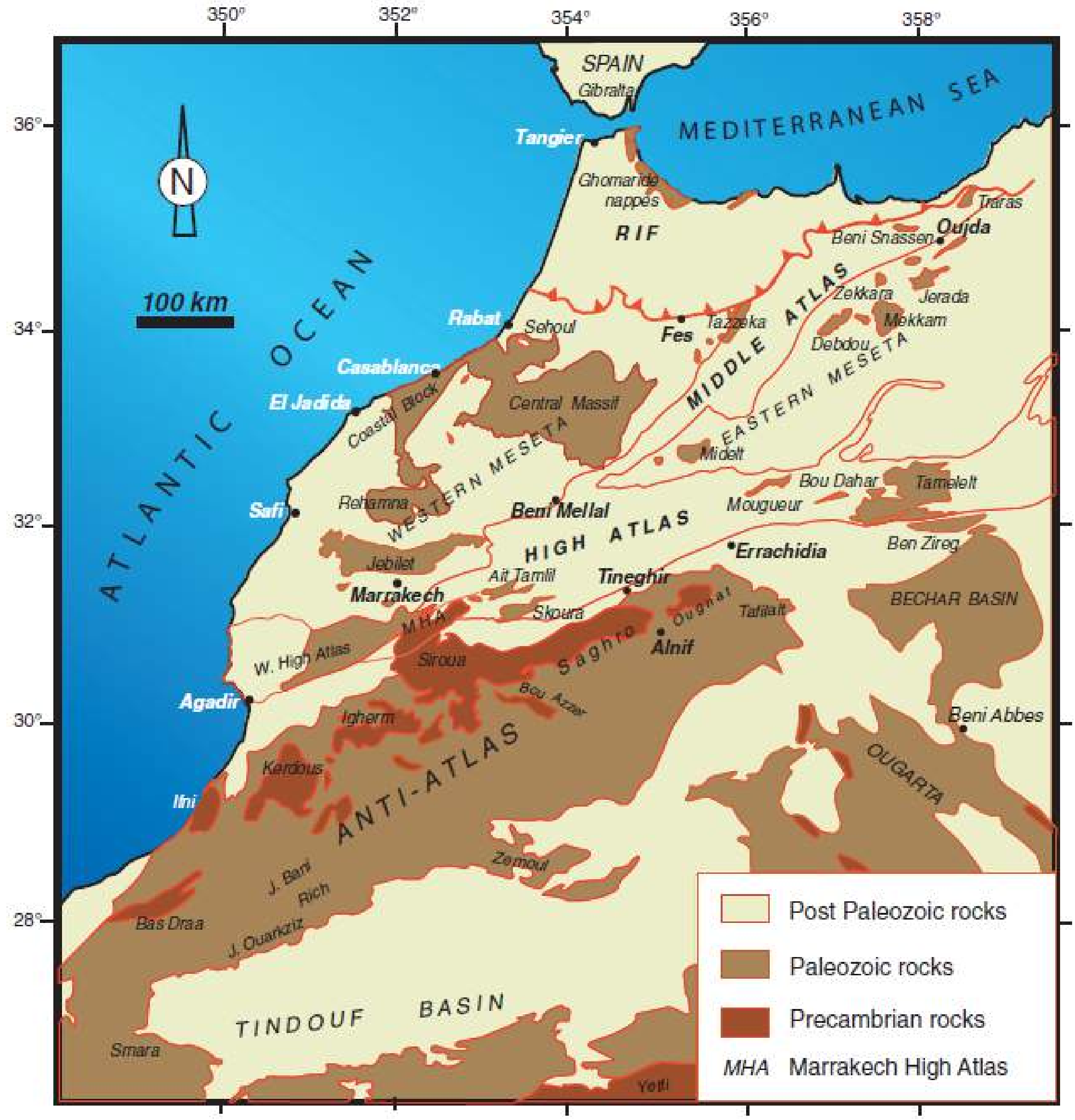

Figure 3.4: Paleozoic and Precambrian oucrops in Morocco and westernmost Algeria, from Michard et al.(2008).

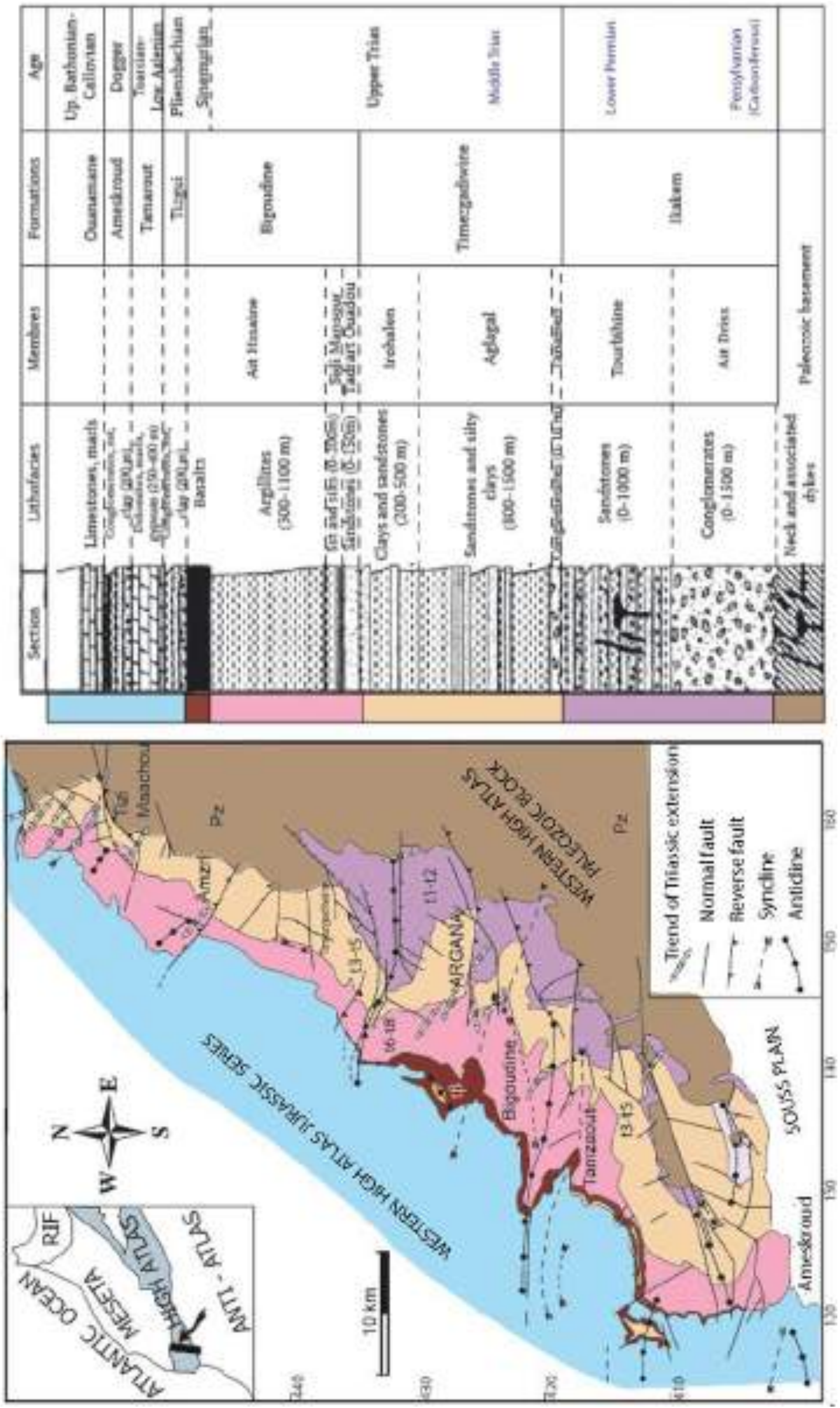

Figure 3.5: Geological map of the Argana Valley from Tixeront (1976) in Frizon de Lamotte et al. (2008).
Lithostratigraphic synthetic section of the Argana Basin formations (Brown, 1980; Medina, 1991) modified from Aït Chayeb at al., 1998 (after Doubinger and Roy-Dias, 1986, 1985; El Arabi et al., 2006; Tourani et al., 2010).

3.3 **Rifting**

Triassic extension

In the High Atlas, the Triassic sequence presents evidence of extensional movements due to the intracontinental rift: synsedimentary normal faults, rapid facies evolutions (conglomerates, sandstones, claystones) and extensive evaporite deposits. Initiated during the Trias, normal faulting led to the formation of numerous half-grabens, filled by fluviatile-deltaic sandstones.

Before the Central Atlantic opening, a westward-dipping half-graben system comprising NNE-SSW to NE-SW trending faults developed and was linked by E-W transfer faults (Bouatmani et al., 2003; Medina, 1988). These transfer faults have been interpreted as reactivated Variscan thrust faults (Laville and Piqué, 1992). This extensional system corresponds to the beginning of continental rifting which led to the formation of the Moroccan Atlantic Margin. A recent study on magnetic anomalies refined the work of Klitgord and Schouten (1986) by comparing the magnetic anomalies in North America to the magnetic anomalies in Western Africa and taking into account the Atlas Domain intra-continental rifting in the reconstruction of the paleogeography. The Moroccan Meseta was then disconnected from the West-African Craton (Figure 3.7) which improved drastically the fit of the ECMA (East Coast Magnetic Anomaly) and the WACMA (West African Coast Magnetic Anomaly) (Sahabi et al., 2004) (Figure 3.6) The formation during the Triassic rifting of several microplates between Laurasia and Gondwana, including a Morocco-Algeria Plate, have been also considered at a larger scale (Schettino and Turco, 2011, 2009). These tectonic studies led the authors to consider a Ladinian (Middle Triassic) age for the onset of the Central Atlantic rifting. Rifting terminated with the formation of the first oceanic crust in the proto-Atlantic, during the Sinemurian (Hafid, 2000; Pique et al., 1998).

In the EAB, Paleozoic basement is exposed on its eastern rim, in the Massif Ancien de Marrakesh. The hypothesis of a half-graben basin with NW tilted blocks forming the Argana Basin, which is the only outcropping domain for the Trias in the EAB (Brown, 1980; Laville and Petit, 1984; Medina, 1988, 1991, 1995), has been challenged based on detailed structural mapping, with a new hypothesis suggesting a wide rift to sag system (Baudon et al., 2012).

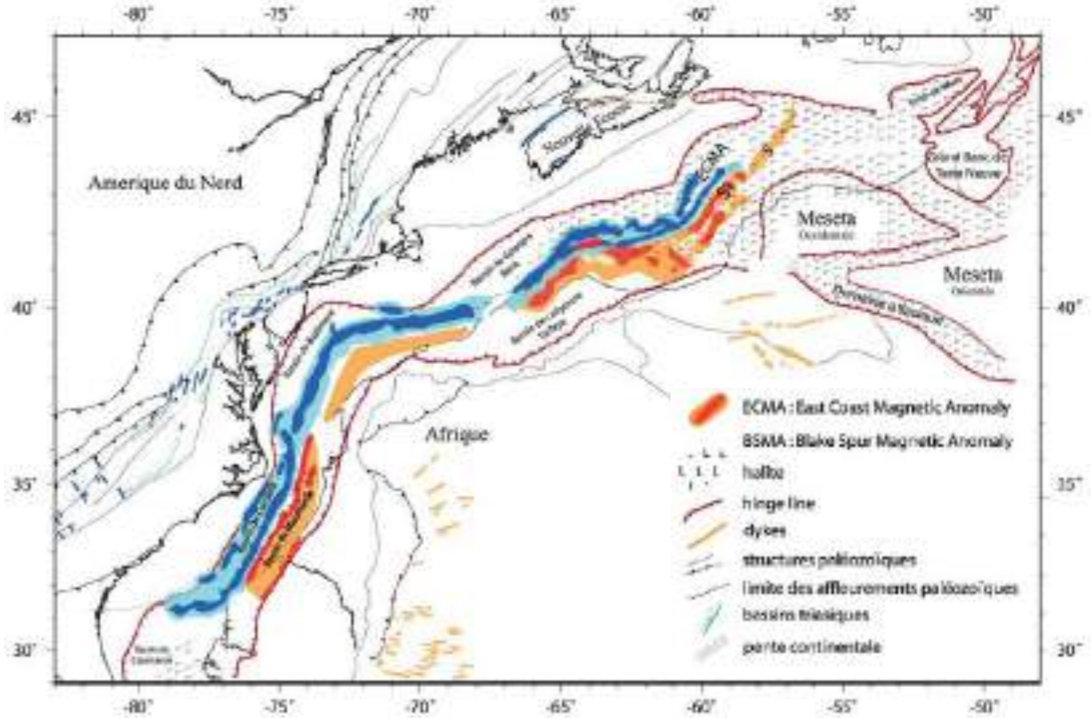

Figure 3.6: Magnetic anomalies fit at 195 Ma (Late Sinemurian) of the African–Mesetan–North-American. In blue is the West Atlantic Coast Magnetic Anomaly (WACMA) and in orange is the East Coast Magnetic Anomaly (ECMA). The ECMA and WACMA present a strong symmetry. The authors disconnected the Moroccan from the West-African Craton in order to take into account the younger Atlasic compression and to improve the fit. This fit represents the relative positions of the continents after the rifting. . (From Sahabi et al., 2004).

The Triassic series outcrops mainly along the Argana Valley and rests on Permian rocks with an erosive unconformity. The depression formed during the Lias were filled by at least 2000 m of continental red beds, lacustrine shales, evaporites and basalts. The first Triassic deposits were composed of alluvial-fan red conglomerates which have eroded the units below. The facies assemblage is composed of coarse to fine siliciclastic deposits resulting from a lacustrine and deltaic complex, and an upper mud-plain sequence that passes westward to evaporite deposits (Figure 3.5). Triassic and Liassic tholeiitic basalts with a wide regional extension are present on top of these sediments (Ambroggi, 1963;

Brown, 1980). An intrusive basalt sill cuts through the Bigoudine Formation and overlies it conformably in the south of the valley. Those basalts are part of the Central Atlantic magmatic province which extends over four continents and marks the boundary between Triassic and Jurassic (Marzoli et al., 2004).

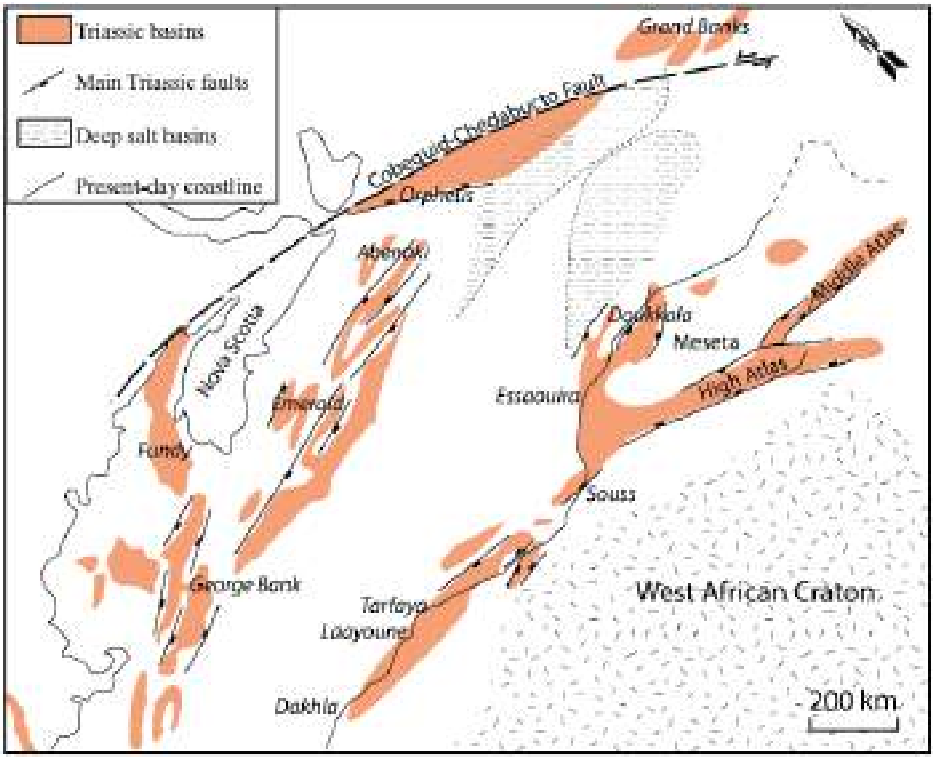

Figure 3.7: Map of the Central Atlantic domain
Showing the correlation between the main Triassic basins just prior to continental break-up. From Gouiza et al. (2010); after Le Roy and Piqué, (2001) and Davison, (2005).

The Carnian is usually cited as the onset of Atlas Rifting (Beauchamp et al., 1999; Laville et al., 2004; Le Roy and Piqué, 2001), but it has been revised to an earlier date in the Western High Atlas. A palynological study of the Argana Valley proved that Anisian siltstones with evidence of syntectonic deposition have been recorded (El Arabi et al., 2006). More recently, chirotherioid traces have been studied in the Tanamert member of the Timezgadiwine Formation (Figure 3.5).The trace fossils have been identified as being created by Archosaurs living in a floodplain close to alluvial-fans. The traces have been dated as Upper Olenekian (Klein et al., 2010; Tourani et al., 2010). The Permian-Triasic hiatus, interpreted to be Permian to Carnian is significantly shortened, as outcrops show a footprints sequence more or less complete from Permian to Lower Jurassic. The first Atlas rifting phase is then occurring in the Early Triassic. The extensional phase in a NW direction occurring during Late Triassic has been interpreted as a response to the early stage of the Atlantic rifting (Laville et al., 2004).

Syn-rift salt deposits form an extensive diapiric province along the Moroccan Atlantic margin (Figure 3.8). This province extends over 900 km along the margin. The salt was deposited in the deepest half-graben during the rifting i. e. in the western part of the Agadir-Essaouira Basin, but make an incursion onshore and outcrop in the core of the Amsittène Anticline. The size and frequency of the salt diapirs leads to the conclusion that the original thickness of the salt offshore Morocco probably exceeded 1.5 km. This level of salt mobility strongly influences the post-rift sedimentary and structural evolution of the western part of the basin (Davison, 2005; Hafid et al., 2008).

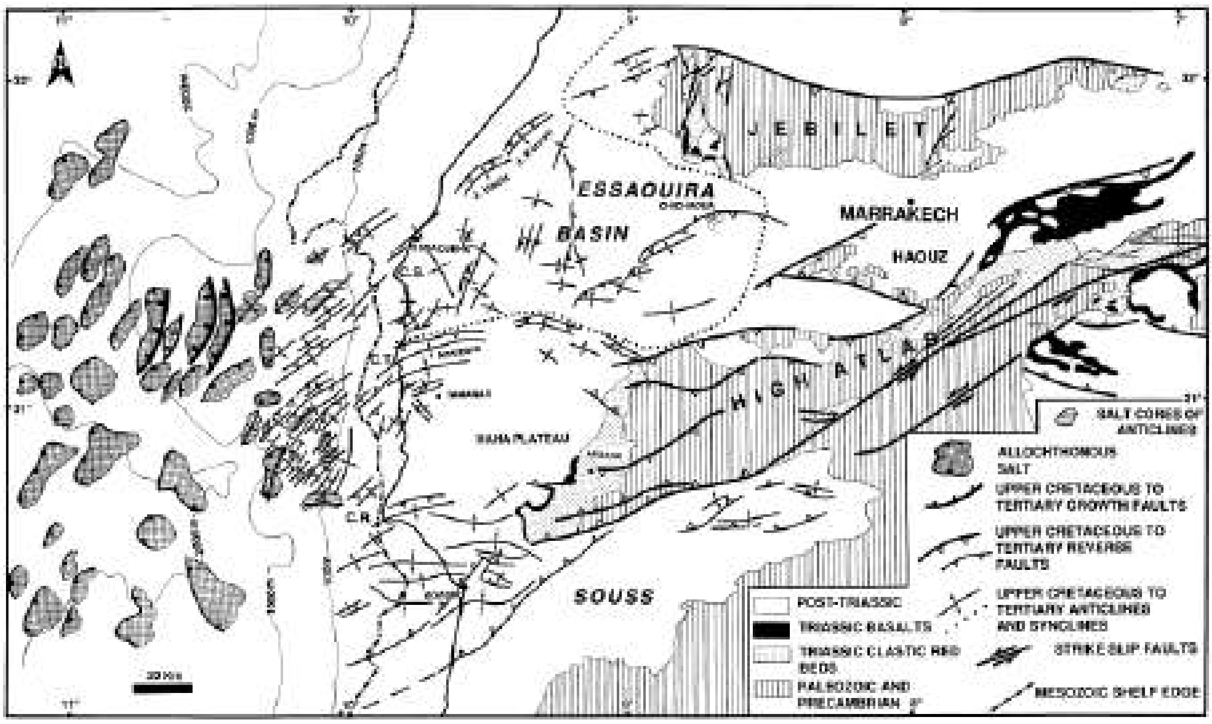

Figure 3.8: Structural map of the western termination of the High Atlas system.
From Hafid, (2000).

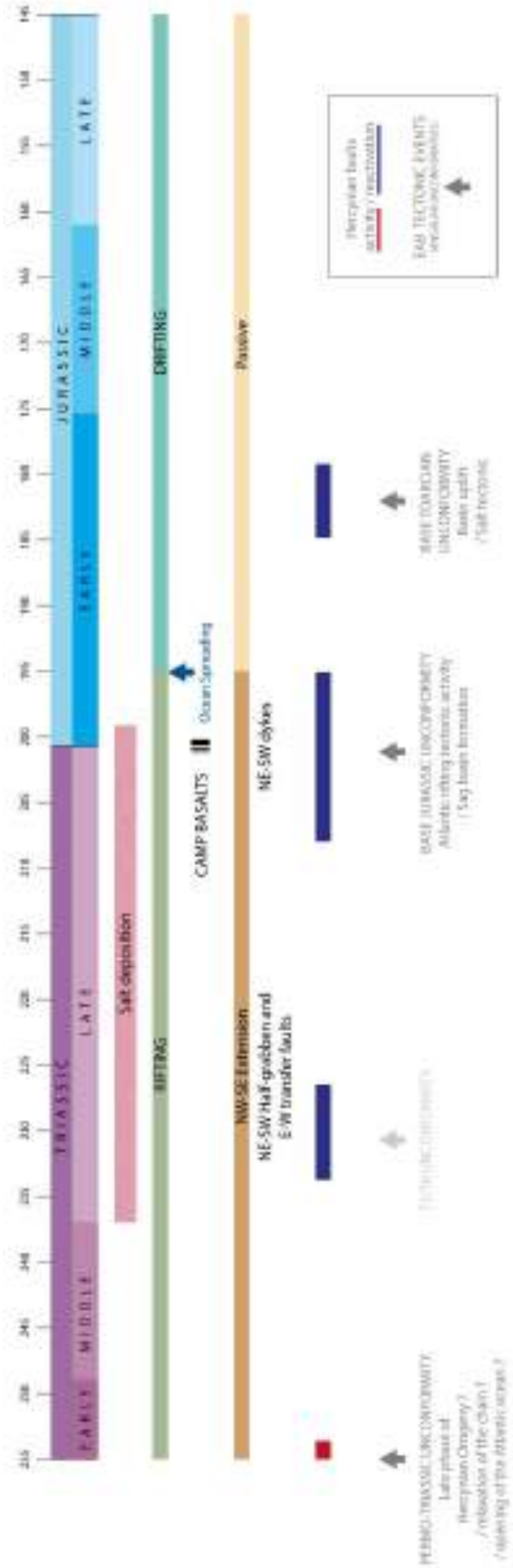

Figure 3.9: Summary of the principle tectonostratigraphic events affecting the Essaouira-Agadir Basin during the Triassic and Jurassic. Data from Tari et al., 2003; Hafid, et al., 2008; Leleu et al., 2016.

Abortion of the Atlasic rift

Surface and subsurface geometries suggest an unconformable contact between the Triassic deposits and the overlying Lower Lias sag basin, which pinch out eastward and northeastward on the western flank of the western Moroccan Arch (Medina, 1994; Hafid et al., 2006). Seismic profiles along the Middle and High Atlas shows the presence of widespread Triassic-Early Liassic sag basins (formed just after rifting) (Figure 3.9) containing evaporites and important basalt flows (Hafid et al., 2000, 2006). The sediments recorded active faulting during the Upper Triassic, and further show the diminution of fault activity from Early to Middle Liassic. The Liassic carbonate platform which overlaps the Triassic sediments is considered as a post-rift sequence of the Atlasic aborted rift (Medina, 1991; Piqué and Laville, 1996). During the Early and Middle Liassic thermal subsidence, relative sea level rise created accommodation and allowed the development of a carbonate platform. This shelf margin was developed in the whole rifted basin (Laville et al., 2004).

CAMP Basalts

The Central Atlantic Magmatic Province (CAMP) basaltic event coincides with the Triassic-Jurassic boundary (TJb) (Marzoli et al., 1999). The basalts resulting from this event are present across four continents but best preserved in rift basins of NW Africa and NE America. CAMP magmatism spread over 10.10^6 km^2 and is principally composed of tholeiitic basalts (Mc Hone, 2003). In Morocco, CAMP-related dykes outcrop in the Anti-Atlas and CAMP lava-flows, sometimes associated with dolerite dykes outcrop in the Eastern Messeta, in the Central High Atlas and in the Western High Atlas (Youbi et al., 2003; Hachimi et al., 2011). During the last thirty years, several authors have tried to refine their dating by using Argon isotopes analysis ($^{40}\mathrm{Ar}/^{39}\mathrm{Ar}$) (Fiechtner et al., 1992; Marzoli et al., 1999; Knight et al., 2004; Nomade et al., 2007a; Verati et al., 2007; Jourdan et al., 2009; Marzoli et al., 2011). They found that CAMP magmatism did not exceed 10 Ma with a 1 Ma dominant period of magmatism. The more recent global dating for the CAMP at this point gives the magmatism ranging from 202 to 190 Ma and the peak of activity around 200 MA (Nomade et al., 2007b). These authors also provide a more accurate dating for Moroccan basalts with an age range from 199.0 ± 2.0 to 195.6 ± 8.9 Ma for the lower to uppermost lava flows.

The $^{40}\mathrm{Ar}/^{39}\mathrm{Ar}$ dating associated with zircon uranium-lead (U-Pb) analysis and supported by palaeontological, geochemical and magnetostratigraphic data (Knight et al., 2004; Marzoli et al., 2004) linked the CAMP activity with the Triassic-Jurassic Boundary (Palfy et al., 2000; Jourdan et al., 2009; Marzoli et al., 2011; Blackburn et al., 2013). This hypothesis has been challenged by the analysis of palynological assemblage studies and vertebrate extinction in North America (Olsen et al., 2003; Whiteside et al., 2007). These authors correlated the last occurrence of *Patinasporites densus* and the magnetostratigraphic marker chron E23r (brief period of reversed magnetic polarity) to the TJb using carbon isotope. The occurrence of this event 40 cm below the first basalt flow concluded the CAMP event postdated it. This interpretation was refuted according to the acquisition of new Rhaetian palaeontological evidences above the base of the basalts in the Fundy basin of Nova Scotia (Cirilli et al., 2009; Kozur and Weems, 2010). Whereas the link between the CAMP lava flows and the end-Triassic extinction is still debated, a new high-precision U-Pb dating links the TJb with intrusive CAMP activity (Davies et al, 2017). It correlates maffic intrusive units, occurring ≈100 Kyrs before the earliest eruptions, to the onset of changes in the biologic and climatic records.

3.4 **Drift Phase**

From seismic sections and well data, it has been observed that the faults affecting the Triassic-Early Liassic deposits of the Atlantic Moroccan basins are capped by a Late Liassic-Middle Jurassic post-rift uncorformity (Hafid, 2000; Pique et al., 1998) (Figure 3.9). During the Liassic, fault activity declined and the sea level rise accompanying thedeposition of the marine carbonate, overlying the synrift sequence, was mainly controlled by a thermal subsidence (Hafid et al., 2000; Laville and Piqué, 1992; Mridekh et al., 2000).

Offshore Morocco, the Dogger and Upper Lias are dominated by shales, dolomites and anhydrite deposits, while the Upper Jurassic is composed of limestones. Onshore, the shales and dolomites of the Lower Jurassic are consistently thinner than the offshore deposits. The seismic profiles show that the sediments were deposited on a ramp type platform, with occasionally development of a flat-topped platform. During the Valanginian, the carbonate platform were covered by siliciclastic deltaic sediments (Davison, 2005; Hafid, 2000).

3.5 Moroccan basins

During the Early Mezosoic, rifting controled the evolution of the Atlasic basins. The Western Moroccan Arch was a shallow sea or emerged landmass separating the Tethyan and Atlantic realm (Frizan de Lamotte et al, 2008) (Figure 3.10). The area of study is located in the North-West of Morocco, in the Agadir-Essaouira Basin.

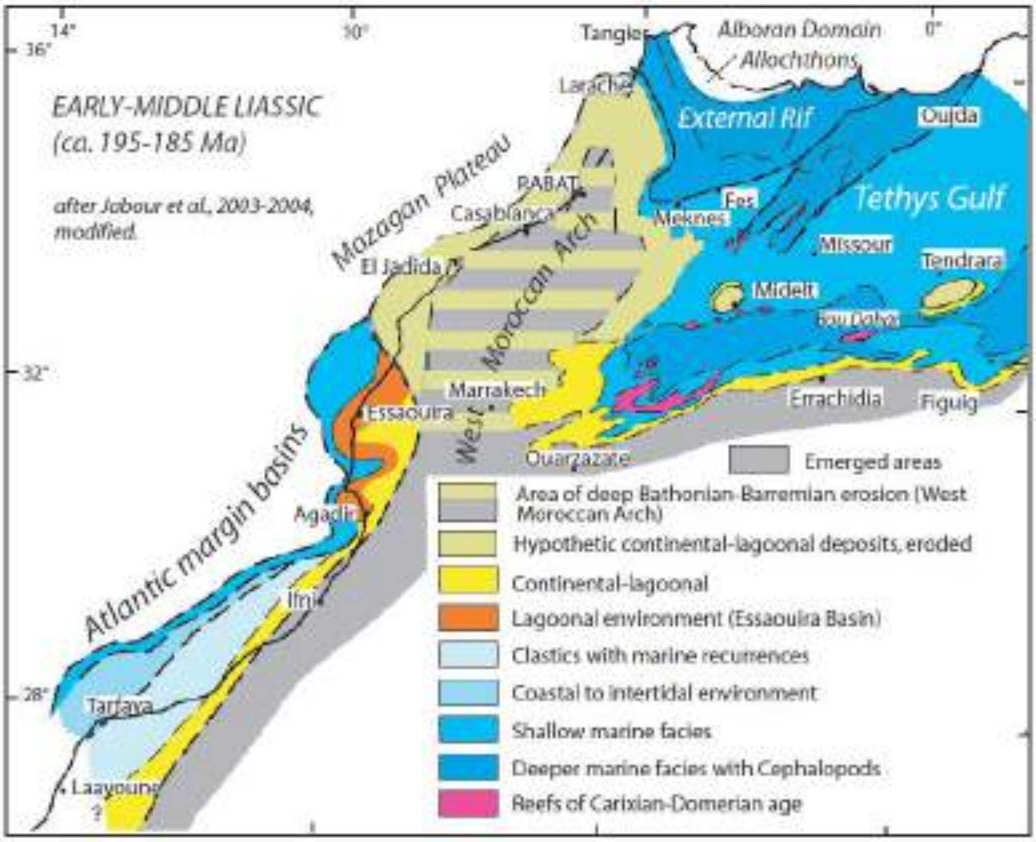

Figure 3.10: Paleogeographic map of the Atlas system during the lower Jurassic. From Frizon de Lamotte et al., 2008.

During Late Trias to Middle Lias times, two large provinces can be differentiated along the rift. The eastern province connected to the Tethys Gulf was separated from the western province that was open to the Central Atlantic. These two provinces are supposed to be delimited at this epoch by the Western Moroccan Arch which has been interpreted as emergent land from the Lias until the Early Cretaceous (Faure-Muret, 1974) and is now interpretad as a shallow platform, eroded during the Middle Jurassic - Early Cretaceous (Ghorbal et al, 2007; Saddiqi et al, 2009) (Figure 3.10 andFigure 3.11).

Two extensional members have been distinguished from seismic and field data in the eastern and western Atlasic domains (Frizon de Lamotte et al., 2008). During the Late Triassic to lower Jurassic, in the western domain, a sag basin dominated by salt deposits and basalts flows (Figure 3.9) formed the Argana Valley and Agadir-Essaouira Basin (Figure 3.11). The post-rift sedimentation started during Middle Jurassic and recorded a tectonic relaxation phase during the Oxfordian (Figure 3.12) followed by a transgressive trend during the Upper Jurassic (Ellouz et al., 2003; Frizon de Lamotte et al., 2008).

In the north-east domain, the Middle Atlas, the Triassic extension was limited to the Lower and Middle Jurassic were characterised by a new extensional phase (Figure 3.12) characterised by shallow marine deposits (Ellouz et al., 2003; Frizon de Lamotte et al., 2008). The High Atlas domain, which corresponds to the eastern part of the Atlas system, presents an intermediate situation with the development of a rift basin filled with fluvial deposits during the Late Triassic (Figure 3.11). The extension in the High Atlas basin stopped during the Jurassic, and a carbonate platform grew during the Lower Jurassic with deltaic sediments deposited during the Middle Jurassic (Ellouz et al., 2003; Frizon de Lamotte et al., 2008).

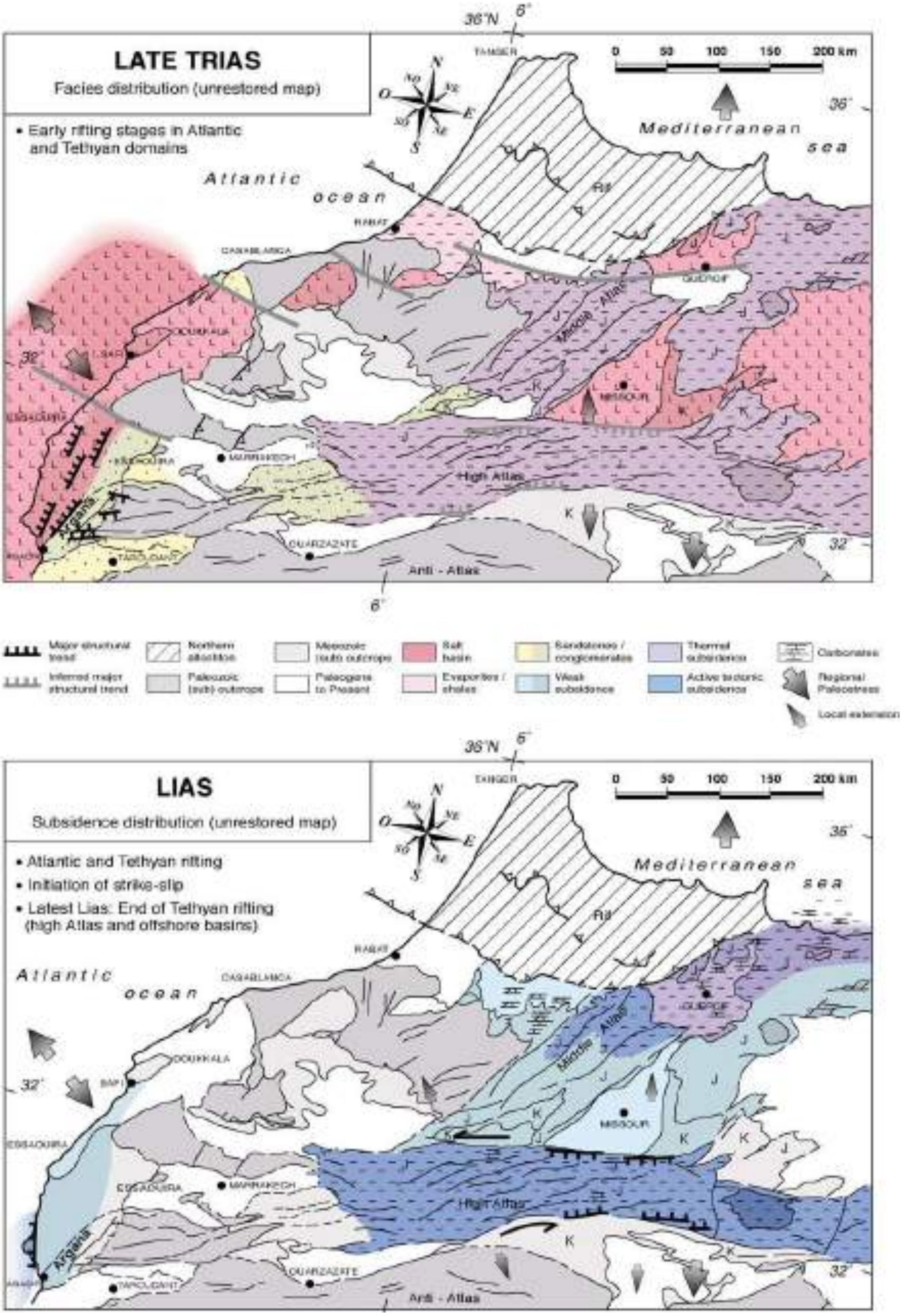

Figure 3.11: Late Trias: Facies distribution map for the Late Trias.
Lias: Subsidence distribution in Morocco during the lower Jurassic. From Ellouz et al., 2003

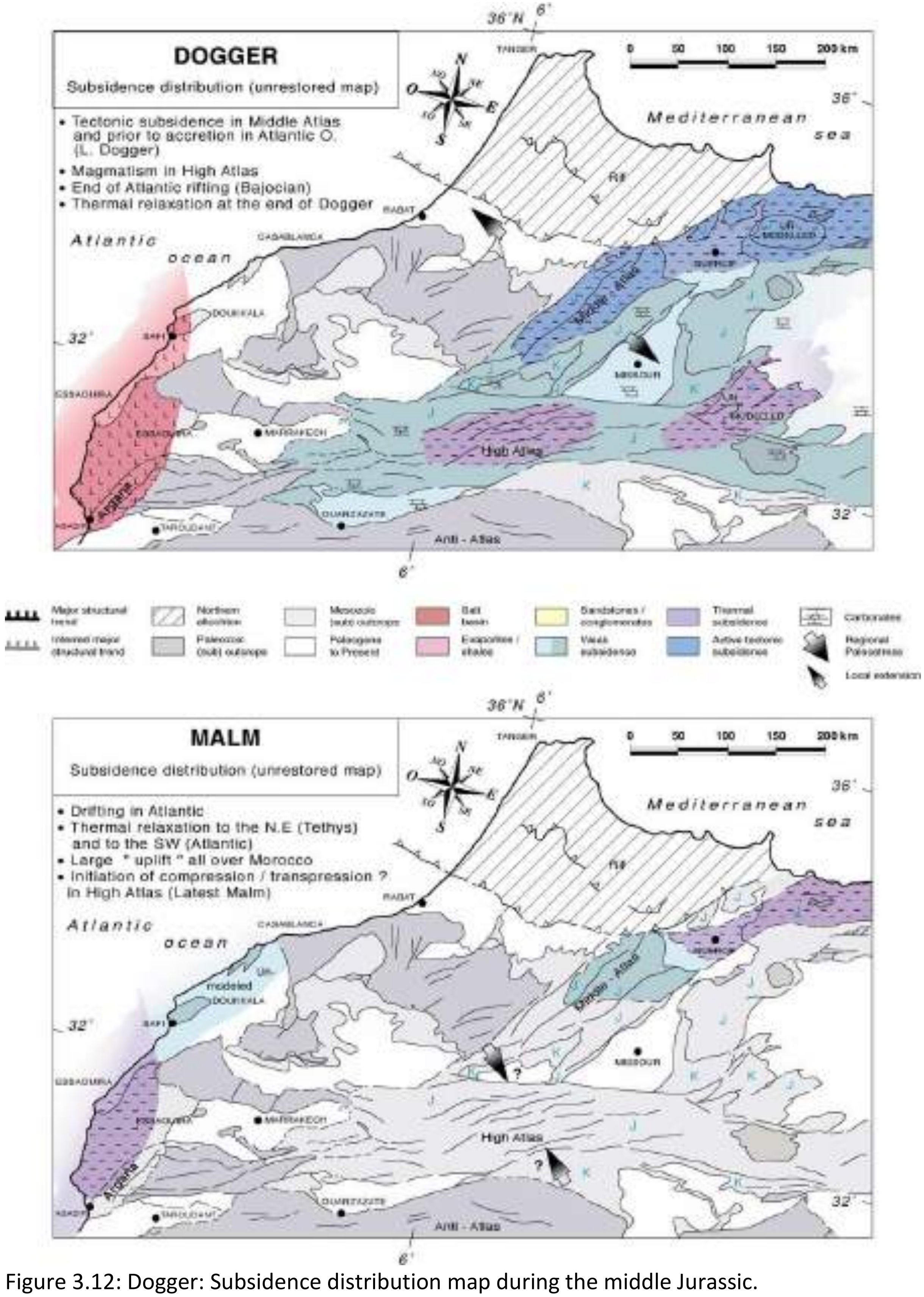

Figure 3.12: Dogger: Subsidence distribution map during the middle Jurassic.
Malm: Thermal subsidence distribution during the upper Jurassic. From Ellouz et al., 2003

3.6 Environmental and climatic variations during the Jurassic

The Mesozoic era coincides with a long greenhouse period of generally elevated temperatures and high sea level. The Jurassic period lasted around 55.6 m.y. (Ogg et al., 2016) and is divided into 11 stages. The base of the Jurassic is relatively well constrained corresponding to a biotic mass extinction and the Central Atlantic Magmatic Province (CAMP) basalts event. The top of the Jurassic is not defined by such a widespread extinction event and the age of the Tithonian varies regionally, according to the palaeontological or secondary method used (Ogg et al; 2012).

The end of the Triassic is characterised by a warm and semi-arid to arid climatic conditions (Preto et al., 2010; Whiteside et al., 2011; Iqbal et al., 2019). The end-Triassic period is linked to climate change, with increased precipitations, anoxia and a warming trend towards a hothouse period (Ahlberg et al., 2003; Jaraula et al., 2013; Ge et al., 2018; Iqbal et al., 2019). A direct link between the eruption of the CAMP basalts and the P_{CO_2} variations due to volcanogenic CO_2 and thermogenic methane release in the atmosphere has been established (Pálfy et al., 2001; 2007; Hesselbo et al., 2002; Ruhl et al., 2011; Fujisaki et al., 2018).

Lower Jurassic

The Hettangian (earliest Jurassic age) directly follows the T/J mass extinction event and oceanic anoxia and changes in the global geochemical cycles are inferred to have been a limiting factor for the biotic recovery (Ruhl et al., 2011; Jaraula et al., 2013). H_2S-enriched anoxia (euxinia) persisted throughout the Hettangian (Jaraula et al., 2013; Luo et al., 2018). These authors also state that the expansion of the anoxia during the Hettangian resulted from high oceanic temperatures reducing the potential of oxygen solubility in seawater and creating vertical stratification.

During the Jurassic, when the atmospheric CO_2 was notably higher than current levels, and sea level higher; tropical to sub-tropical conditions were present over most of the Pangea and no evidence of polar ice cap have been found (Hallam, 1982; Selwood and Valdes., 2008). Equitable climate throughout the Jurassic has been long inferred, but cooler periods are now suggested by oxygen isotopic and paleobiogeographic evidence (Veizer et al., 1999; Dromart et al., 2003; Cecca et al., 2005).

The Lower Jurassic records an extensive spread of evaporites facies, indicating dry conditions, and an average summer temperature above 35 °C calculated for the regions of western Pangea by general circulation modelling (Chandler et al., 1992). But the relative warm climate of the Lower Jurassic seems to have been interrupted by a cooler event during the late Pliensbachian to early Toarcian (Dera et al 2009; Rogov and Zakharov, 2010; Korte and Hesselbo, 2011; Korte et al., 2015) marked by a negative $\delta^{13}C$ and a decrease of mean $\delta^{18}O$ values (Grossman, 2012; Saltzman and Thomas, 2012). A particularly warm period during the Toarcian Oceanic Anoxic Event (T-OAE) followed (Jenkyns, 1988; Bailey et al., 2003).

Middle Jurassic

However, the Middle Jurassic continued to record high CO_2 concentrations and therefore relatively high temperatures (Fletcher et al 2007; Sellwood and Valdes, 2008; Wallman, 2008). Nonetheless, the presence of abundant glendonites pseudomorphes during the Bajocian and the early Callovian has been presented as a marker of temperature fall (Rogov and Zakharov, 2010).

The Middle to Late Jurassic transition (Late Callovian to early Oxfordian) records an abrupt variation of $\delta^{13}C$ (**Error! Reference source not found.**) (Dromart et al., 2003; altzman and Thomas, 2012) and an abrupt decrease of species variability across the Tethyan and peri-Tethyan regions (Cecca et al, 2005). The Tethys sea-water temperatures for this time interval have been calculated using oxygen isotope and lower temperatures have been found for the late Callovian and early Oxfordian, followed by an increase in temperatures during the middle Oxfordian (Dromart et al., 2003). The warming of the middle Oxfordian appears to be supported by biogeographic and geochemical data (Cecca et al., 2005).

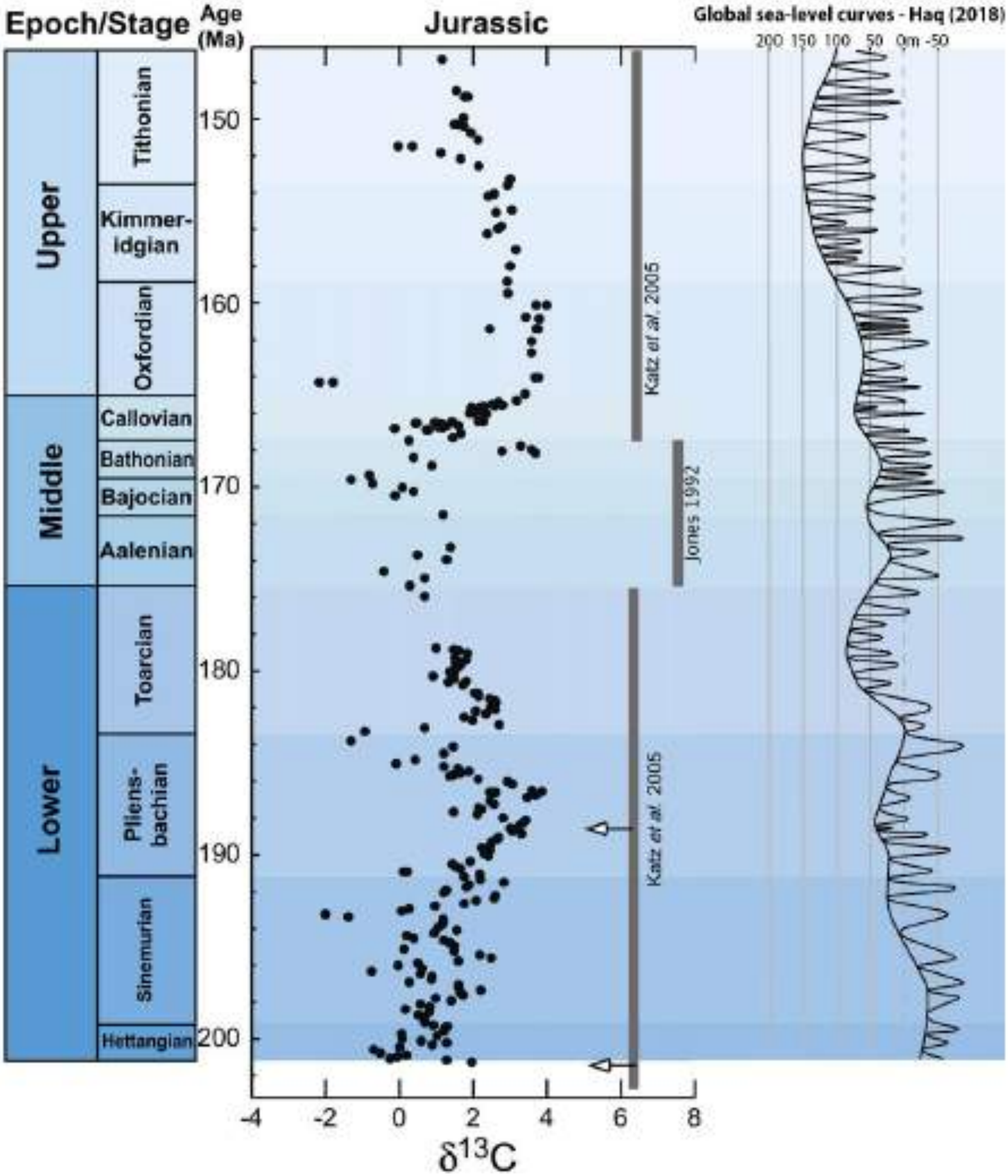

Figure 3.13: Variation of $\delta^{13}C$ and sea-level curves through the Jurassic. Data sources are indicated by vertical bars. From Saltzman and Thomas (2012) and Haq (2017).

Upper Jurassic

The Upper Jurassic is interpreted to have been slightly warmer than the rest of the Jurassic with monsoonal rainfalls (Weissert and Mohr, 1996; Brigaud et al., 2008) and maintained a higher sea-level (Takashima et al; 2006; Haq, 2017).

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Chapter 4: Evolution of early post-rift depositional systems along the Moroccan Atlantic Margin: the Essaouira-Agadir Basin, Lower and Middle Jurassic.

Abstract

The early post-rift evolution of the Central Atlantic is recorded in the Lower and Middle Jurassic outcrops of the Essaouira-Agadir Basin (EAB). This important succession, characterised by alternating deposition of marine carbonates and fluvial clastics, provides evidence for significant tectonic activity along the passive margin, rejuvenating clastic input to the basin. Field observations, well data and petrographic analysis are integrated into a coherent sedimentological model, correlated across the basin within a sequence stratigraphic framework. Comparison is drawn with equivalent aged units in the Western High Atlas, which allows constraint on the regional versus local tectonostratigraphic evolution in the Lower-Middle Jurassic. In the EAB, the Upper Sinemurian to Lower Pliensbachian records an initial transgression, which led to development of open marine ramp carbonates of the Arich Ouzla Formation. This formation is preserved locally in the North of the basin but in many locations it has been tilted and exhumed before the deposition of the overlying fluvial Amsittène Formation. The erosive base of this unit is traceable across the basin and cuts down into the Pliensbachian, CAMP basalts or the Triassic. In the Central High Atlas, the correlative fluvial erosive event has been dated as Toarcian in age. This influx of siliciclastic sediments is interpreted to be sourced from the Meseta and/or the Anti-Atlas, supported by recent apatite-fission track thermochronology that indicates uplift at this time. Repeated carbonate and clastic deposition continued throughout the Lower and Middle Jurassic. During the Upper Toarcian (Tamarout Formation), a regional carbonate platform, dominated by peritidal deposits, developed across the EAB in response to renewed marine transgression. Facies associations include oolitic and bioclastic grainstones, crystalline dolomite, stromatolites

and dissolution breccias or evaporites. The overlying Middle Jurassic Ameskhoud Formation records a regression and return to fluvial dominated conditions, followed by evolution back to shallow-marine siliciclastics. Unlike the Amsittène Formation, terrestrial conditions were established in the south of the EAB, which indicates a potential source area in the Anti-Atlas, while the North of the basin was dominated by shallow marine carbonates. These observations lead to the conclusion that movements of the hinterland can compensate the effect of the eustasy in passive margins and trigger forced regressions.

4.1 Introduction

The Essaouira-Agadir Basin (EAB) is located at the junction between the Central Atlantic and the Atlas Mountain belt. As such, this basin records the Central Atlantic syn- and post-rift evolution and was affected by the Atlasic (Tethysian) rift and later Alpine inversion. The latter resulted in uplift, providing exceptional exposures of Mesozoic sequences.

The early post-rift stage of the Eastern Central Atlantic passive margin was previously interpreted to record fairly monotonous thermal subsidence, with extensive carbonate platforms deposited over rifted basement (Lehner and De Ruiter, (1977); Le Roy, (1997); Frizon de Lamotte, (2000); Le Roy and Piqué, (2001) and Guiraud et al. (2005)). More recent studies suggest that the adjacent non-rifted continental crust was more tectonically active, with high rates of exhumation in the Western Messeta (e.g Ghorbal et al., 2008) and the Anti-Atlas during the Jurassic (Figure 4.1), which influenced sedimentation (Sehrt, 2014; Gouiza et al., 2017; Charton, 2018). Halokinesis is also recognised as an important control on subsidence and deformation (Hafid et al., 2008). Salt bodies identified by seismic imaging are diapirs and tongues offshore Agadir and allochthonous salt tongues, sheets and canopies offshore Essaouira (Davison, 2005; Jabour & Tari 2007; Tari et al., 2013; Pichel, 2018).

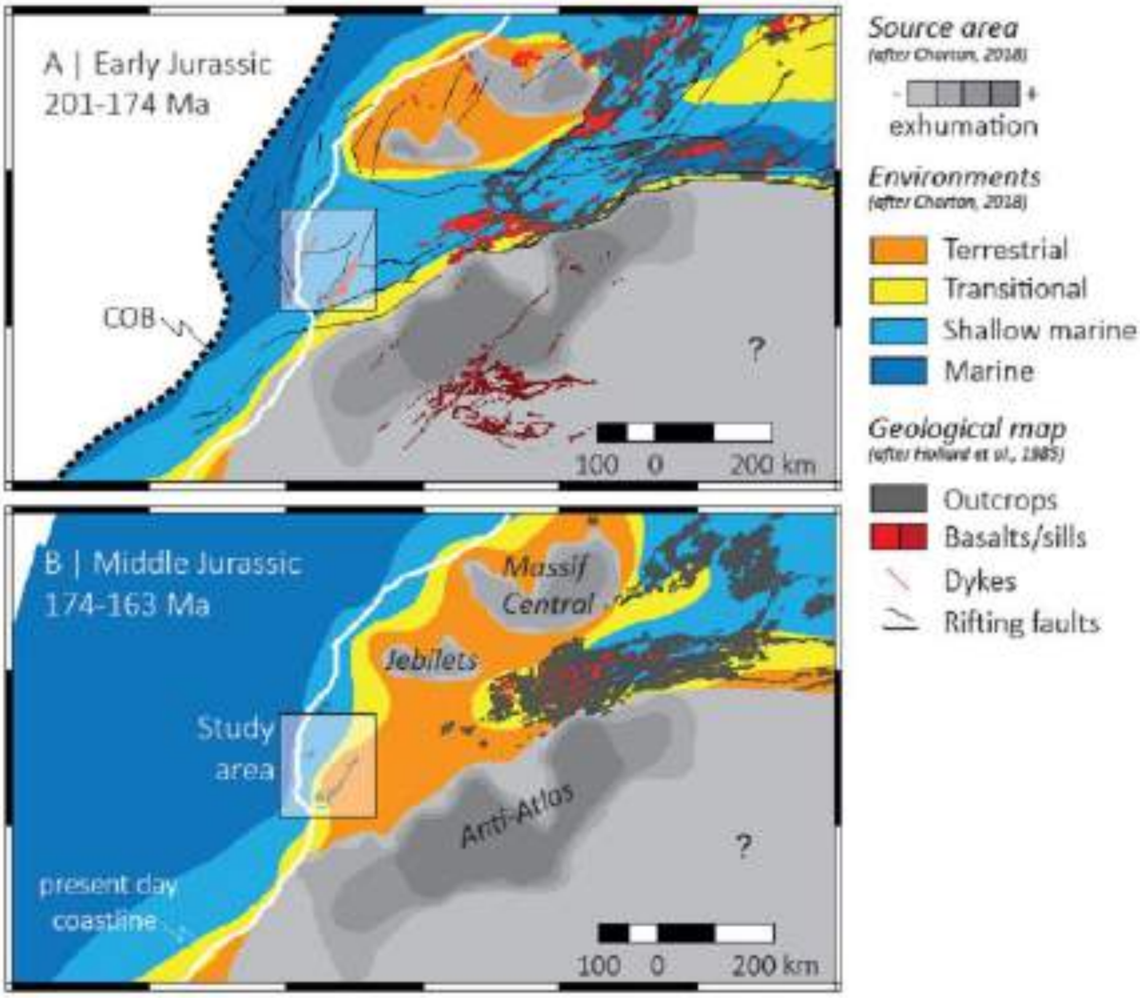

Figure 4.1: Paleogeographic maps of the Lower and Middle Jurassic. After Charton, 2018.

The oldest carbonates deposits, dated as Lower Jurassic, only crop out in the Amsittène Anticline and record open marine conditions (Duffaud, 1960; Du Dresnay, 1988). The Upper Lias deposits are more extensively exposed in the EAB and comprise conglomerates and sandstones followed by gypsiferous mudstones. They are overlain by an extensive carbonate platform resulting from Lower Jurassic transgression, followed during the Middle Jurassic by a continental unit (Ambroggi, 1963; Bouaouda, 1987; Peybernès et al., 1987). The variations of sedimentation and the periodic siliciclastic influx in the basin can result from potential tectonic activity causing increased denudation of sedimentary source(s) in the massifs surroundings of the EAB. This basin therefore offers the potential to study the relationship between tectonics and sedimentation in a syn-rift to early post-rift mixed carbonate-silicicastic sequence.

The EAB offers a rare outcrop window that exposes the most complete Mesozoic succession along the Central Atlantic Margin and is thus very valuable for understanding its early post-rift evolution. The aims of this study are to (1) refine the sedimentological understanding, establish local and regional trends in carbonate deposition, recognize the different scale of transgressive-regressive cycles affecting the sedimentation and identify the extension of the carbonate formations, (2) integrate the sections logged at basin scale within a sequence stratigraphic framework to identify the key surfaces and facies shift and determine the sedimentological controls on the facies variations, (3) place the result into regional context and extract implications on the controls of siliciclastic versus carbonate sedimentation along the western Moroccan margin. Finally, by comparing the EAB with the Central High Atlas (CHA), this study will (4) assess new evidence for tectonic uplift during the Lower and Middle Jurassic across and between both basins.

4.2 Geological setting

4.2.1 Structural evolution and syn-rift sedimentary architecture

The western High Atlas, which structurally comprises the EAB and the Massif Ancien, inherited its geometry from the underlying Triassic Atlas rift, associated with opening of the Central Atlantic (Favre and Stampfli, 1992). The rift zone developed in Precambrian-Palaeozoic basement, influenced by older structures inherited from the Variscan orogeny and potentially older lineaments. During the Variscan orogeny, the Central High Atlas experienced deformation along shear zones showing two main orientations, N20–45°E and N70–90°E, which later acted as weakness zones during Triassic rifting (Pique et al., 1998; Le Roy and Piqué, 2001; Laville et al., 2004). N-S to NNE-SSW westward-dipping half-grabens have been interpreted (Medina, 1988; Bouatmani et al., 2003) linked by E-W transfer faults, which are believed to be reactivated Variscan thrust faults (Laville and Piqué, 1992).

Rifting of the Central Atlantic began in the Ladinian (Middle Triassic) (Schettino and Turco, 2009, 2011), and terminated with the formation of the first oceanic crust in the proto-Atlantic, during the Sinemurian (Pique et al., 1998; Hafid, 2000) . In the Argana Valley, the sag basin formed during rifting has been filled by over 2000 m of continental red beds, lacustrine shales, evaporites and basalts (Olsen et al., 2003, Mader et al 2011). Although the structural control on Triassic deposition is debated (Hofmann et al., 2000; Baudon et al., 2012), the previous studies all agree on minimal tectonic influence on Lower Jurassic sedimentation.

The main structural interval to affect the region was the later Alpine/Atlas Inversion, that lasted from the Upper Cretaceous to the Neogene (Hafid, 2000; Hafid et al., 2006). This reactivated faults, uplifting and folding the sections, and ultimately exposed the Mesozoic to the surface.

4.2.2 CAMP Basalts

The Central Atlantic Magmatic Province (CAMP) basaltic event, sits at the base of the Jurassic section. The basalt flows were emplaced during the final phase of rifting that initiated the break-up of the Atlantic. The CAMP magmatism spans about 10 Ma, with a 600 000 years to 1 Ma peak of activity dated by radiometric methods to around 199 Ma (Fiechtner et al., 1992; Marzoli et al., 1999; Palfy et al., 2000; Knight at al., 2004; Nomade et al., 2007; Verati et al., 2007; Davies et al., 2017). The link between the Triassic-Jurassic mass extinction and the CAMP basalts has been extensively studied and refined (Whiteside et al., 2007; Blackburn et al., 2013; Davies et al., 2017) therefore, despite no consensus on the subject (Nomade et al., 2007; Whiteside et al., 2007), the CAMP magmatism will here be estimated to be synchronous with the Triassic-Jurassic boundary. Deposits following the CAMP basalts euption will be considered as lowermost Jurassic in this study, although there is an ongoing debate as to whether the CAMP volcanism predates (Marzoli et al., 2004) or postdates (Olsen et al., 2003; Whiteside et al., 2007) the Triassic-Jurassic boundary.

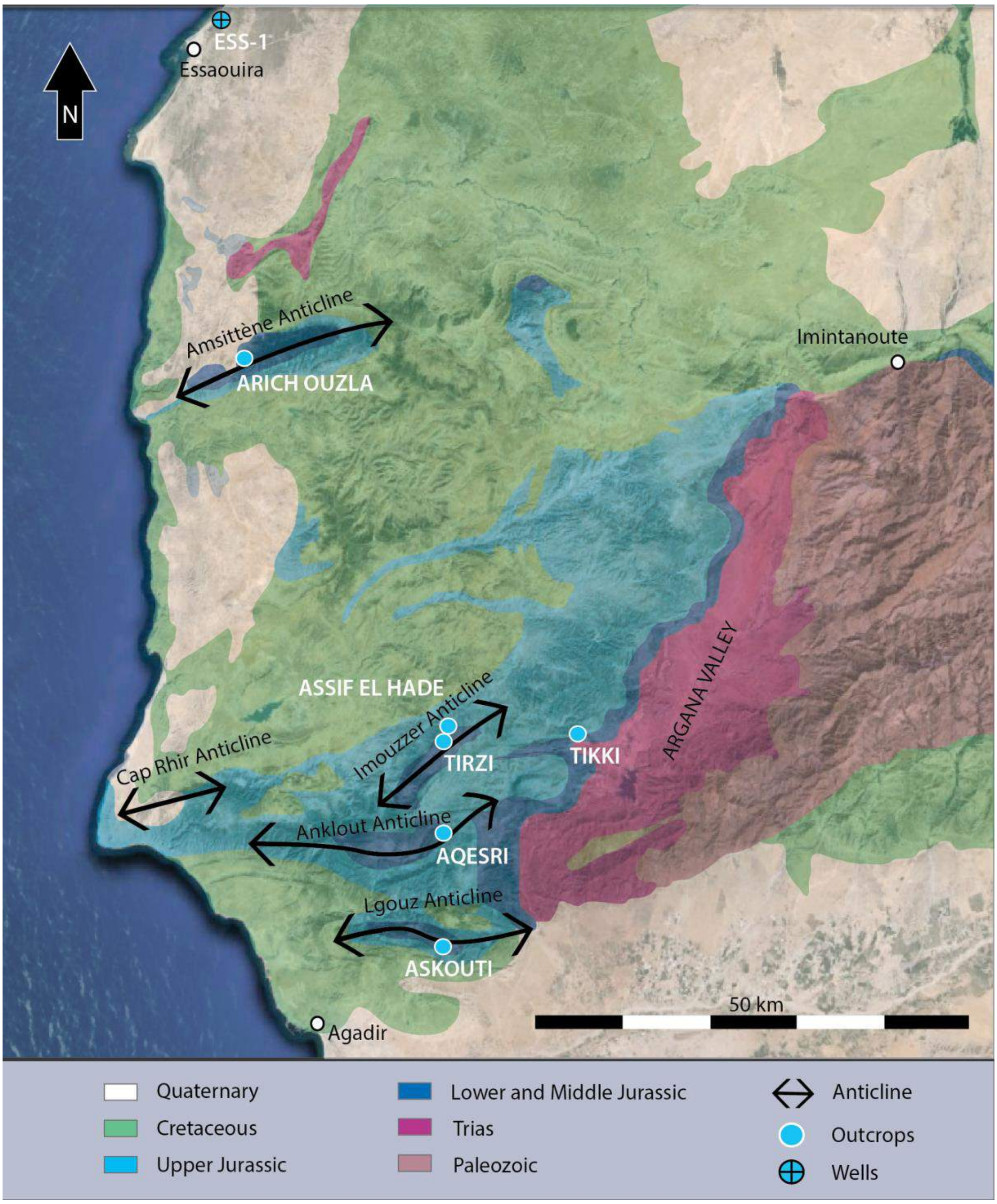

Figure 4.2: Geologial map of the Essaouira-Agadir Basin.
(Modified after Choubert, 1957; Zuhlke et al., 2004) and location of the wells and outrops superimposed on the satellite imagery of the study area (Google Earth).

4.2.3 Lithostratigraphic units

The Lower Jurassic stratigraphy of the Agadir-Essaouira Basin has historically been described based on lithostratigraphy (Figure 4.4) (e.g. Roch, 1930; Ambroggi, 1963; Duffaud, 1960; Adams, 1979; Adams et al., 1980; Peybernes et al., 1987 and Du Dresnay; 1988; Bouaouda, 2007). This approach has been retained in this paper due to a lack of precise dating for most of the syn-rift and early post-rift formations. Adams and co-authors (1980) defined three main lithostratigraphic units and distinguished them into formations. The older Arich Ouzla Formation was subsequently named by Peybernes et al. (1987).

To the west, (e.g. Jebel Amsittѐne and in offshore wells) Lower Lias carbonates have been ascribed to the Arich Ouzla Formation, immediately below the Amsittѐne Formation (Figure 4.4). Onshore, these are only locally preserved in the core of the Amsittѐne Anticline (Figure 4.2), and consist of dolomitic carbonates. This succession rests on Triassic red mudstones and evaporites (Du Dresnay; 1988). The Arich Ouzla Formation has been dated as Sinemurian to Upper Pliensbachian on the basis of brachiopod fauna (Duffaud, 1960; Peybernes et al., 1987).

In more proximal locations, i.e. along the Argana Valley (e.g. Tikki locality), red to purple coloured siliciclastic deposits of the Amsittѐne Formation directly overlie CAMP basalts (Tixeront, 1974). This unit thickens to the NE. Its base is generally erosive resting upon basalts or Triassic continental deposits. A Toarcian age was originally determined by superposition, between the underlying Arich Ouzla Formation (Upper Sinnemurian - Lower Pliensbachian) and the overlying Tamarout Farmation (Toarcian; Peybernes et al., 1987; Du Dresnay; 1988).

The Amsittѐne Formation passes gradationally into the Tamarout Formation, which contains up to 400 m of dolomites and dolomitic carbonates, with associated breccia horizons or evaporites (Ambroggi, 1963; Bouaouda, 1987, 2007). This unit is laterally heterogeneous and the proportion of evaporites, stromatolites or siliciclastic material varies around the basin. Brachiopods (*Zeilleria lycetti* (Adams et al., 1980*)* and *Terebretula withakeri ?* (Determination G. Dubar in Ambroggi, 1963) place the upper part of this formation into the Toarcian (Ambroggi, 1963).

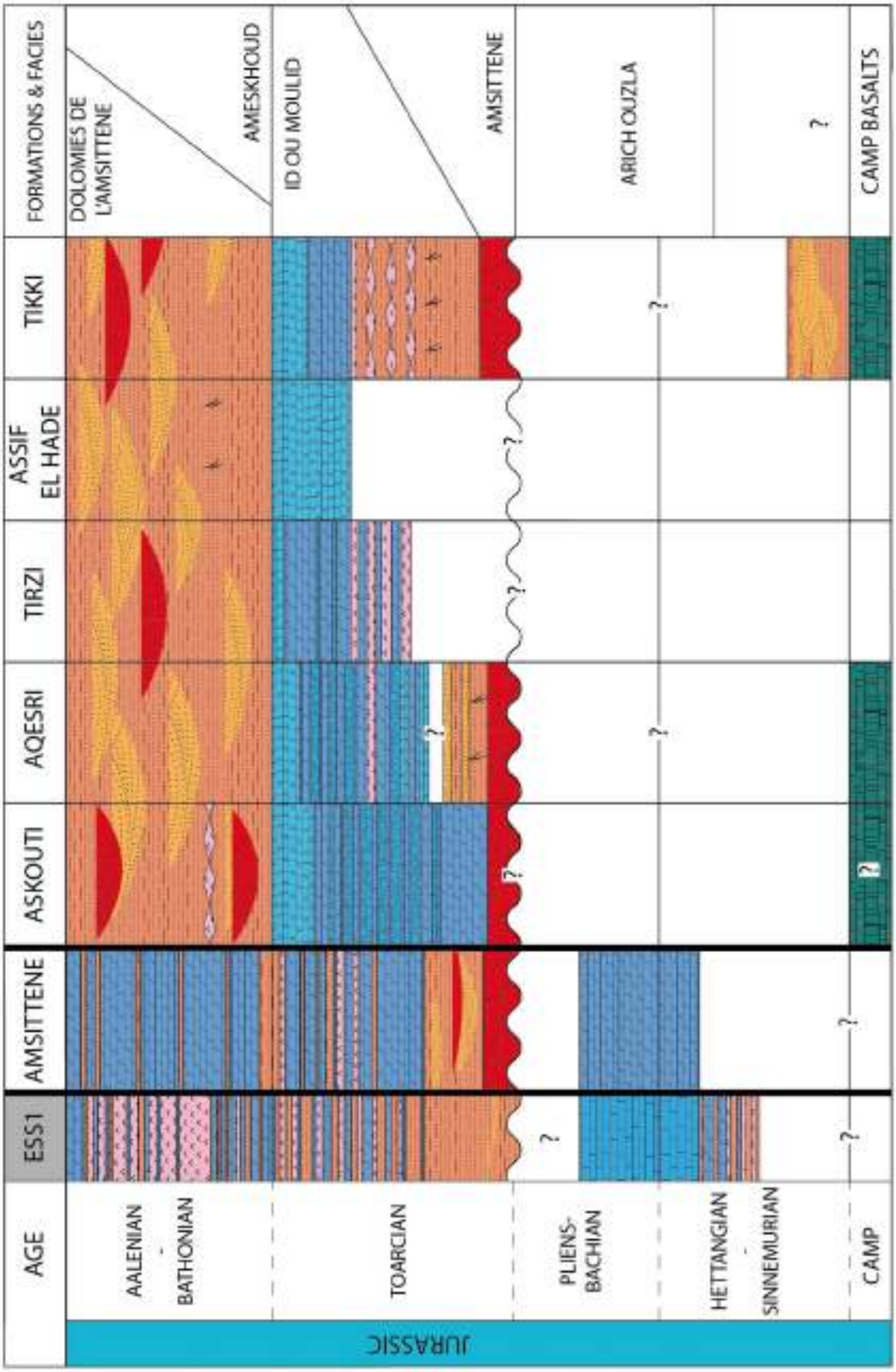

Figure 4.3: Chronostratigraphic chart of the Lower to Middle Jurassic in the Western High Atlas

The Ameskhoud Formation is the youngest Lower Jurassic lithostratigraphic unit described in the Agadir Basin and is composed of red mudstones and siltstones alternating with sandstones and conglomerates (Ambroggi, 1963). In the Essaouira Basin and offshore EAB, this formation is laterally equivalent to a thick dolomitic interval (Figure 4.3) tentatively associated with the Tamarout Formation, later renamed the Id Ou Moulid Formation (Peybernes et al. 1987). The Ameskhoud Formation is dated as Aalenian/Bajocian to Bathonian/Callovian in age (Adams et al., 1980; Du Dresnay, 1988; Bouaouda et al., 2007), based on its relative position, bracketed between the Toarcian Tamarout Formation and the overlying Ouanamane Formation, the basal age of which has been interpreted as Bathonian based on foraminifera identification (Bouaouda et al., 2007 and references therein). New biostratigraphic evidence indicates that a lower Callovian age cannot be excluded for the Ouanamane Formation (Chaper 5).

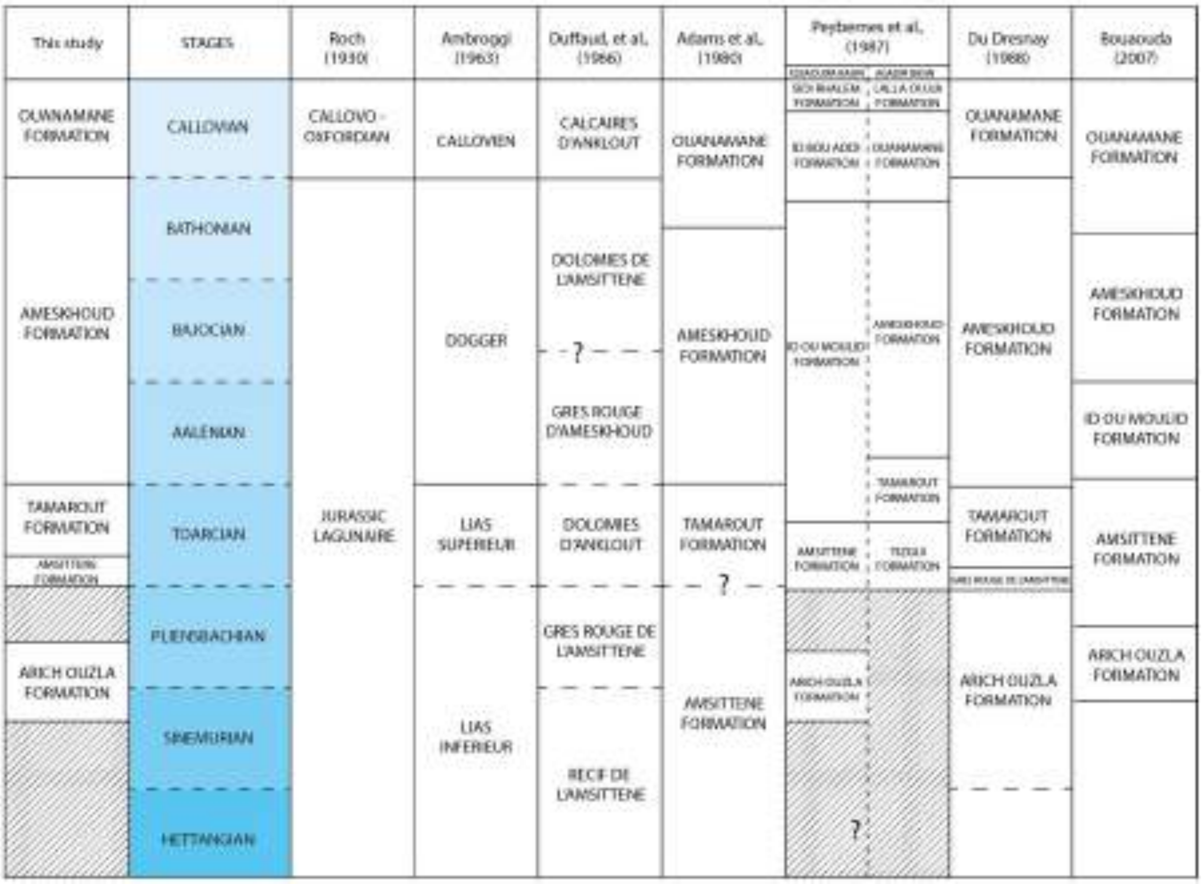

Figure 4.4: Time definition of Lower and Middle Jurassic formations of the EAB used in the present work and compared with other studies (see references therein).

4.3 Methods

The data presented in this study derive from 7 georeferenced sections that have been sedimentologically logged at high resolution across the EAB (Figure 4.2). Continuous sections of Middle and Lower Jurassic have been logged in the East of the basin. Samples were collected every 2 m on average for the Lower Jurassic part of the sections; with fewer samples collected for the Middle Jurassic, which was not the original target of this study. Petrographic analyses were conducted on the two carbonate formations: 18 thin sections for the Arich Ouzla Formation and 22 thin sections of the Tamarout Formation and completed with petrographic identification. Microfacies analyses were based on texture, diagnostic grains, grain type quantification, sedimentary structures and bioturbation. The lateral variability of the beds was studied at outcrop scale, with thicknesses, lateral extent and cyclicity of the different beds recorded to get a better understanding of the depositional environment. The onshore well Essaouira-1 (ESS-1) and the offshore wells Essaouira-1X (ESS-1X), Essaouira West-1 (ESW-1) and DSDP site 547 reached the base of the Lias and show stratigraphy equivalent to the Arich Ouzla Formation.

Depositional sequences were identified using sequence stratigraphic concepts. The observation and interpretation including lateral and vertical facies organisation and variability, unconformities and correlative conformities and geometries of the units, enable the identification of parasequences. General transgressive and regressive trends were identified and correlated across the basin in order to build a coherent framework for the Lias and Dogger deposits.

4.4 Arich Ouzla Formation

4.4.1 Facies association 7 (FA7 – Table 1 and Figure 4.5): Open platform

Description

The Arich Ouzla Formation is the oldest Mesozoic carbonate unit identified in the basin. This formation is observed onshore in the core of the Jbel Amsittène (Arich Ouzla and Ida

Ou Azza salt mines) in the North of the EAB; a similar unit occurs in the Tidzi diapir and in well Essaouira-1 (ESS-1). This formation is recognised in the Essaouira Basin and thickens towards the North, which may be an indication of the South-North basin orientation (Du Dresnay, 1985; Du Dresnay, 1988).

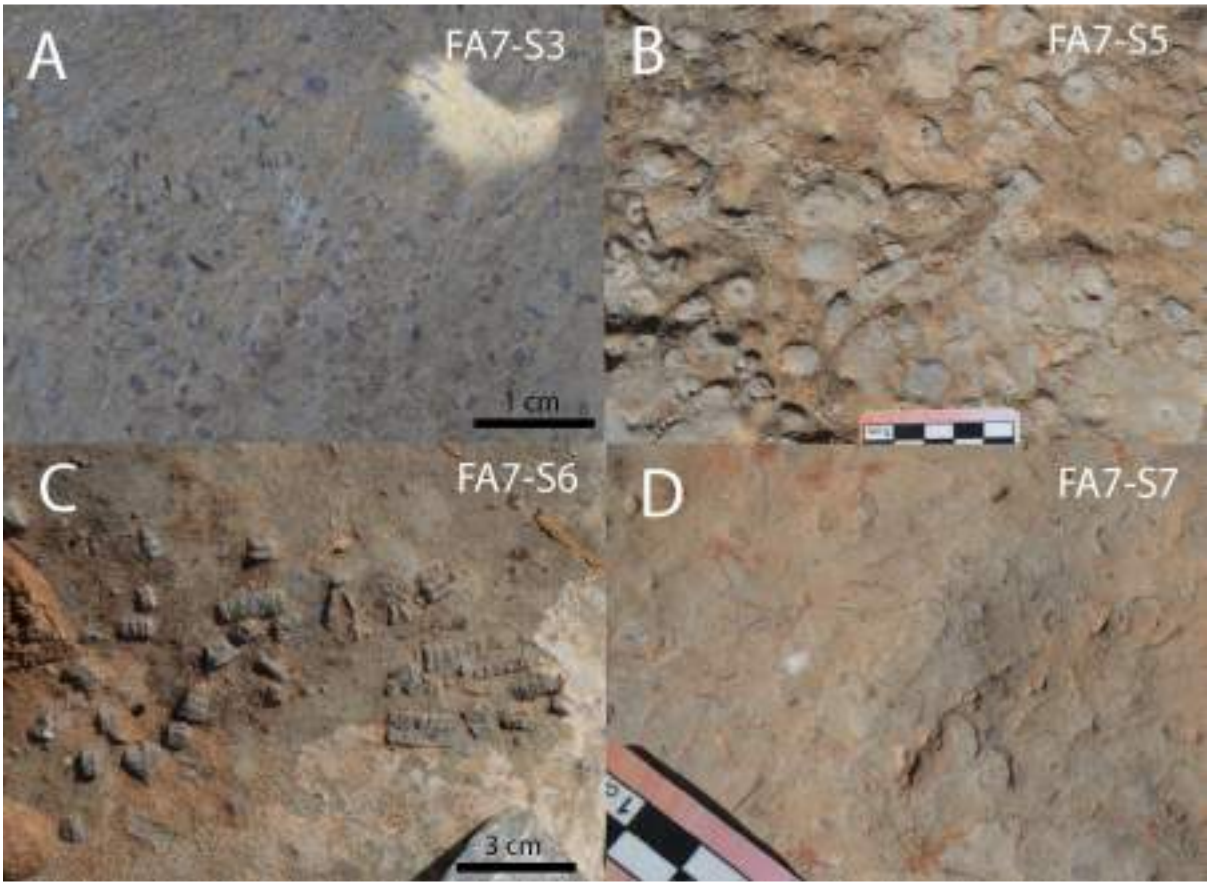

Figure 4.5: Facies of the Arich Ouzla Formation.
Facies FA7-S3, oncoidal PST and shell fragments (A); Facies FA7-S5, oncoidal RST with shell fragments and crinoids (B); Facies FA7-S6, crinoids FST (C); Facies FA7-S7, bioclastic FST with thin shell fragments (D).

In total, 84 m of the Arich Ouzla Formation was logged in the NE part of the outcrops near the Arich Ouzla salt mine. The lower boundary of the formation is not visible, but the location of the salt mine suggests only 1 to 6 m of cover between the carbonates and underlying Triassic red mudstones and evaporites. The formation is composed of three lithologic units in this locality.

Unit I is dolomitic and principally composed of ooids, peloids and small shell fragments with a PST to GST fabric (FA7-S1, FA7-S2, FA7-S3). It includes rare oncoids, crinoids and coral fragments. The top 7 m of this unit records horizontal laminations and horizontal stylolites and is characterised by a darker colour and an associated strong kerogenic smell.

Unit II is made of limestone, locally partially dolomitised. The base of this unit is very dolomitic and made of oncoidal FST with abundant shell fragments and brachiopods (FA7-S4). The middle part consists of oncoidal and peloidal FST (FA7-S3, FA7-S4) and PST with very abundant crinoids, shell fragments and some belemnites, *Trichites* bivalve shells, and gastropods (FA7-S5 and FA7-S6). Ammonites have been observed, but their poor preservation prevented further identification.

Unit III is composed of 22 m of crystalline dolomite (FA7-S7). The lower part of the unit is thinly bedded and contains abundant horizontal stylolites. The bulk of the unit is made of patches of yellow, pink, purple and white dolomite. Most of the fabric has been replaced by crystalline dolomite, although some large bioclasts can still be observed, but are not identifiable. The upper part of this unit is strongly bioturbated, with the matrix and the bioturbations composed of two different dolomite textures.

Interpretation

The high degree of dolomitization in most of Unit I prevents a continuous detailed facies identification. Where dolomitization is only partial, facies are dominated by coated grains identified as potential oncoids and shell fragments. The 2 to 4 mm long ellipsoidal non laminated oncoids associated to thin shell fragments and rare gastropods can be interpreted as upper slope deposits (Flügel, 2010). During the Jurassic, oncoids were frequently deposited over the carbonates shelves down to the basin. But the position of this unit in a transgressive sequence, between continental Triassic deposits and mid-ramp deposits of the unit II indicates the a shallow lagoon environment of deposition is more likely.

In Unit II, oolitic FST present large elliptical and concentric oncoids with abundant crinoids and associated to ammonites and belemnites. The larger size of the oncoids is due to lower water energy level. The presence of organisms related to a deeper environment suggests that this unit was deposited in more down ramp settings. The upper part of this unit displays some solitary corals and coral fragments. These associated to crinoids fragments and smaller oncoids indicate an environment with higher energy. The upper part of this unit can be considered as midramp deposits, probably not far from a potential organic buildup margin.

Unit III is composed of yellow and pink crystalline dolomite with abundant small fractures and vuggs, and has been strongly weathered.

4.4.2 Regional variations of Arich Ouzla

Offshore the well drilled by DSDP leg 79 reached the Lower Lias, which give access to deeper-water facies further North of Essaouira. The occurrence of nanofossils *Involutina ticinensis* (Schweighauser) together with *Schizosphaerella punctulata* and *S. astrea* and the well-preserved foraminifera assemblages dominated by Nodosariids led the authors to date cores 24 to 14 from the Well 547B as Late Sinemurian to Early Pliensbachian (Bernoulli and Kälin, 1984; Riegraf et al., 1984). These deposits are made of 77m black shales and pelagic limestones, directly overlying poorly dated stromatolitic boundstones (Steiger and Jansa, 1984). These authors interpret this interval as a restricted basinal stage with a deeper pelagic environment. The limestones breccias and redeposited nodular limestones horizons could derive from a carbonate ramp equivalent to the Arich Ouzla Formation.

4.5 Amsittène Formation

The Amsittène Formation outcrops in several localities around the EAB. The thickness of this formation across the basin varies between 80 and 140 m, pinching out to the NE along the Argana Valley close to Zaouiat Ouidmane (Tixeront, 1974 Carte Argana). The channelized base is erosive, cutting into Paleozoic, Triassic and older Jurassic deposits.

4.5.1 Facies Association 1 (FA1 – Table 1 and Figure 4.7): Alluvial fan

Description

In the outcrops of Tikki, located along the northern branch of the Tizi N'Test Fault, part of the Amsittène Formation is composed of massive quartzite conglomerates and interbedded sandstones. The base locally consists of massive conglomerates composed of rounded pebbles and cobbles. The pebbles are often pitted, the result of pebble impact, that gives them a characteristic whitish colour. Based on lithofacies characteristics, the facies association FA1 is composed of three main facies (FA1-S1, FA1-S2 and FA1-S3). Facies FA1-S1 is made by poorly sorted conglomerates, usually massive with cobbles up to 15 cm in diameter, floating in a pebbly groundmass. The matrix is composed of sub-rounded to sub-angular gravel and sand. Pebble petrography is dominated by quartzitic pebbles with occasional metabasalt pebbles. Some sandstones lenses separate the conglomerates foresets and pick out the local cross-bedding. Palaeocurrents based on cross beddings are variable, pointing dominantly westward, with subordinate palaeocurrents to the east and south.

FA1-S2 consists of medium to coarse grained poorly-sorted sandstones with pebbly horizons. This facies shows common planar cross-beds and small-scale (15-25 cm) trough cross-beddings with pebbles and cobbles concentrated at the base of some sets. Medium-grained sandstones are also interbedded with the massive conglomerates FA1-S1 and constitute well-sorted lenses and wedges. Thin cross-bedded sandy horizons with granules are particularly common in the upper part of the main conglomerate units. Facies FA1-S3 is composed of poorly sorted fine-grained sandstones with occasional clay horizons. This facies is mainly unstratified and presents occasional nodular beds and root traces.

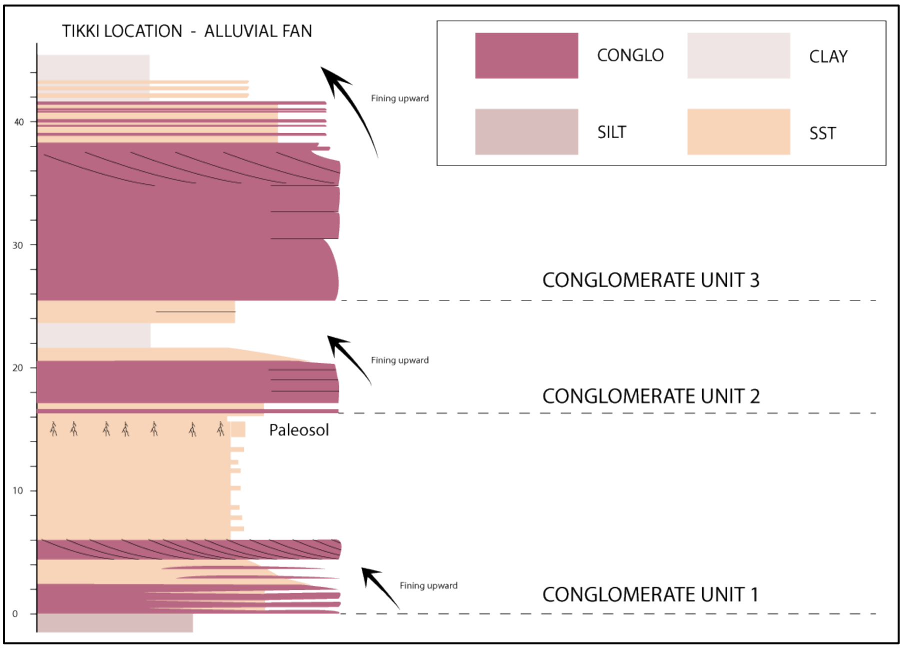

Figure 4.6: Simplified log of the Alluvial fans in Tikki section (TKI).

Three conglomerate units can be distinguished (Figure 4.6), separated by sandstone units (FA1-S2 and FA1-S3). The thickness of the conglomerate unit varies slightly laterally but the general organisation remains consistent between outcrops. Horizons of pebbly sandstones (FA1-S2) are present at the base and at the top of the two first conglomerates. The first conglomerate unit is 6m thick and is separated from the second conglomerate unit by 10 m of fine to medium grained sandstones. These sheet-like sandstones are formed by a succession of thin individual beds (10-30 cm) with rare roots traces. The second conglomerates unit is 5 m thick and displays fining upward. The bulk of these two conglomerate units (FA1-S1) are poorly sorted generally massive with rare cross-bedding indicating westward paleocurents. The upper part of these conglomerates locally contains cross-bedded sandstones lenses made of sub-angular material. The second conglomerate unit is separated from the third by 5 m of interbedded, horizontally stratified, fine grained sandstone and mudstone, with no visible current features. The third conglomerate is coarse-grained, poorly sorted and has a non-erosive base. This unit fines upward, from massive conglomerates at the base to a middle part of the unit divided in sets horizontally stacked with apparent cross-bedding and to the upper part of the unit containing cross-bedded sandstones lenses. It is directly followed by medium and fine sandstones, rapidly grading into mudstones.

Interpretation

Overall the units have a homogeneously pale purple-pink outcrop colour suggesting subaerial oxidising condition. The association of extensive units of stratified and unstratified conglomerates associated to sandstones and pebbly sandstones is interpreted as alluvial fan deposits.

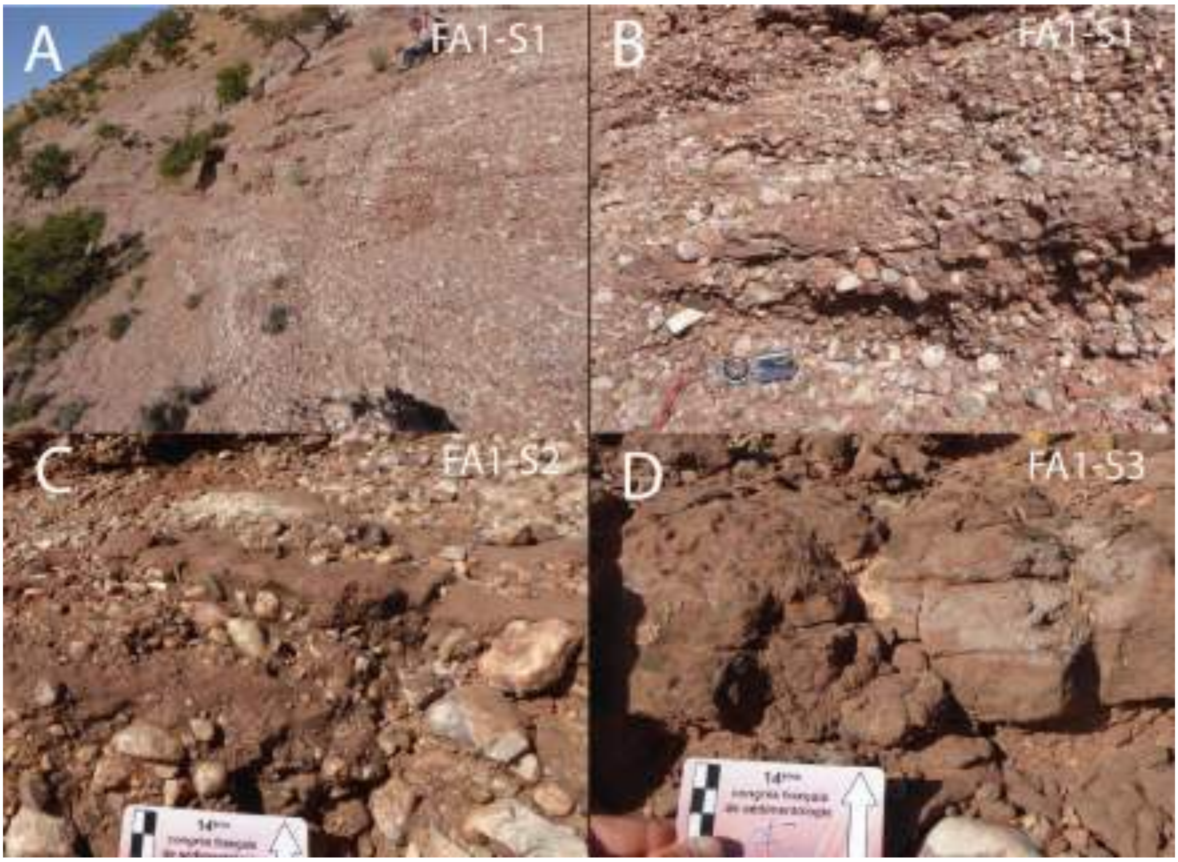

Figure 4.7: Facies of the Amsittène Formation.
Facies FA1-S1, Quartzite conglomerates (A and B); Facies FA1-S2, Medium to coarse sandstones interbeded with conglomerates (C); Facies FA1-S3, Fine sandstones (D).

The two first conglomerates units, with pebbly sandstones at the bottom and top of the units, are dominated by the facies FA1-S1 and display cross stratification. This suggests deposition in high energy environment and can be interpreted as streamflow deposits. The cross-bedded sand lenses in these conglomerates were deposited from waning traction currents (Miall, 1977, 1996; Blair, 1999). The pebbly sandstones and cross-bedded sandstones with planar and trough cross-beds are also interpreted to be deposited in a braided system, but with less energy (Miall, 1977; Heward, 1978). The non-erosive base of the third conglomerate, followed by an unstratified conglomerate unit are a common association of features characteristic of non-cohesive debris flow deposits and its large extent is indicative of a lobe deposit (Harvey et al., 2005). Large lobe deposits are more common in the distal part of an alluvial fan deposits (Miall, 1977; Blair and McPherson, 1994). The top of this third conglomerate contains cross-bedded sandstones lenses which can be interpreted as deposited by waning flow as floodwaters declined (Bluck, 1967; Heward, 1978).

The finer grained sheet-like sandstones with rare paleosols (FA1-S3) separating the different conglomerates and sandstones units have been interpreted as crevasse splay deposits.

4.5.2 Facies association 2 (FA2 – Table 1): Braided river

Description

In Askouti and Akesri, the base of the Amsittène Formation is composed of channelized conglomerates with sandstones. The red conglomerates in these two localities have an erosive base and are composed of quartzite and basaltic pebbles and cobbles. In Akesri, this formation lies uncomformably on top of the CAMP basalts.

The conglomerate horizons are thick (up to 2 m) with an erosive base or contain horizontal bedding and low angle cross-beds. The units are lenticular shaped and repeat through the stratigraphy. They comprise a pebble-supported base, with sub-rounded to sub-angular pebbles and cobbles. Their sorting is poor, but they locally contain matrix-supported lenses with aligned pebbles. The conglomerates then fine-upward into poorly sorted coarse grained sandstones with granules and medium to fine grained sandstones with moderate sorting. The fine to medium grained sandstones exhibit migrating ripples with an average height of 2 cm. The coarse sandstones are trough cross-bedded or have tabular and planar cross-bedding and are locally followed by medium-grained sandstones with well-developed current ripples and horizontal bedding. A single dominant channel can be distinguished in most locations, but its width is difficult to constrain due to the narrow outcrops.

Interpretation

The fining upward conglomerates and sandstones are interpreted to represent cyclic channel deposits in a braided river environment (Williams and Rust, 1969; Miall, 1977, 1996; Bridge, 2003).

The erosive clast-supported conglomerates with horizontal stratification are interpreted to be deposited by bar migration (Rust, 1972). The channels eroded into the floodplain deposits (FA3) and the CAMP basalts. With widths typically greater than 10 m and reaching a thickness up to 2 m, and presence of interbedded sandstones lenses, indicates a compound braid bar system. The planar and low angle cross-stratified conglomerates are typical of the initial deposition of mid-channel bars dominated by bedload transport (Bridge, 1993; Lunt and Bridge, 2004). These bars are formed by migration of gravel sheets downstream and can later be cut by second order cross-bar channels. The interbeded sandstones and presence of aligned pebbles appearing in the conglomerate units, are interpreted to be the result of ephemeral conditions (waning flow) or secondary transverse flows (Rust, 1972, 1978; Bridge, 1993). Sandy deposits were concentrated in topographically high part of the channel belt and in channel fill during the falling stage of floods. They can be interpreted in two different ways. The coarse sandstones at the top of fining-upward conglomerates which mainly exhibiting tabular cross-bedding with planar cross-bedding can be interpreted as bed-load sheets deposited on bar crests. The coarse sand-wedges and lenses with planar cross-bedding and trough cross bedding could also be interpreted as isolated active cross-bar channel (Rust, 1972; Lunt and Bridge, 2004).

The migrating ripples in the sandstones are organised as sets, where the stoss side are not preserved for most of the sets. The finer sediments and the formation and preservation of the ripples indicate a variation of the current strength. This facies is associated to a more protected environment within the river or due to waning of the flow and deposition on bar tops (Miall, 1996; Best et al, 2003).

4.5.3 Facies association 3 (FA3 – Table 1): Flood plain

Description

The deposits of FA3 are characterised dominantly by red mudstones, siltstones and very fine grained sandstones organised into thin horizontal beds. The mudstones are finely laminated and developed as massive bedded red horizons up to 50 cm thick. They alternate with thin beds of siltstones (FA3-S1) and very fine grained sandstones (FA3-S2), which record ripple lamination on the top surface of the beds. Thin horizons of very-fine grained sandstones grade upward to siltstone and display locally current ripples and flaser bedding. Fine grained sandstones sheets with sharp bases, low angle tabular cross-bedding and climbing current ripples (FA3-S3) are common. Occasional interbedded matrix supported conglomerates (FA3-S4) form 40 cm thick cross-bedded horizons, of restricted lateral extent (up to 10 m wide).

Interpretation

The mudstone and siltstones interbedded with minor sandstones are interpreted to have been deposited on floodplains in inter-fluve areas. The association of better sorted fine sand and silt with climbing ripples and cross-lamination represent multiple fining-upward episodic depositional events. These have been interpreted as overbank splay deposits, where variations in grain sorting can be related to the differences in sediment load depending on the water discharge (Lunt et al., 2004). The thin matrix-supported cross-bedded conglomerates horizons are discontinuous and laterally pass into fine grained sandstones and siltstones. The coarse sediments and lenticular shape of these units suggest that this facies is associated with small channels. The mudstone and siltstone dominated units are always closely related to the braided river deposits but are characteristic of a different water energy. This facies association have then been interpreted as flood plain deposits by analogy to the environment described by Smith (1980), being deposited during flooding of adjacent rivers.

4.5.4 Facies Association 4 (FA4 – Table 1): Coastal plain

Description

Above the braided river deposits of Aqesri, a mixed carbonate and siliciclastic unit is developed. It is mainly composed of mudstone and marls (FA4-S1) alternating with siltstones (FA4-S2), dolomitic sandstones (FA4-S3) and sandy dolomite (FA4-S4). The lower part of this unit is dominated by red mudstones and interbedded dolomitic sandstones and siltstones. These units present some mud rip-up clasts at the base, current ripples and roots traces at the top. Further up the unit is dominated by sandy and silty dolomite, alternating with marls and mudstone horizons. The sandy dolomites present carbonate-coated quartz grains and some ooids, and feature flaser bedding with bi-diectional flow indication and tabular cross-bedding.

Interpretation

Red and green mudstones and marls are the most common facies in FA4 and indicate deposition in a very quiet low energy environment. The dolomitic sandstones with marly rip up clasts at the base and tabular cross-bedding reflect an increase of energy in the environment and currents eroding the underlying deposits. The presence of root traces at the top of these beds indicates repeated exposure. The increase in sandy dolomites towards the top of the unit suggests a change in the environment of deposition as the formation of ooids and coated grains requires marine or lacustrine conditions. These sandy and silty dolomites can be interpreted either as marine intercalations or lacustrine deposits in a generally more continental environment. Abundant flaser bedding with ripples indicating opposite directions of flow in facies FA4-S4 suggests tidal influence in these deposits. Facies FA4 is dominated by continental deposits, with occasional development of shallow lakes and marine incursions. This records the transition to the overlying shallow-marine Tamarout Formation, it is interpreted as a coastal plain facies association.

4.6 Tamarout Formation

4.6.1 Facies association 5 (FA5 – Table 1): Sabkha

Description

Twelve meters of red mudstones and siltstones with horizons of nodular gypsum are developed around Tikki. At the top of this unit, sandy dolomite levels (FA5-S3), sandstones and siltstone horizons (FA5-S4) appear, still alternating with mudstone (FA5-S1) and evaporites (FA5-S2). The amount of evaporites decreases towards the top of the unit, which is capped by one meter of wavy-bedded peloidal packstone. This facies is transitional between the continental Amsittène Formation and the shallow marine deposits that dominate higher-up in the Tamarout Formation.

Interpretation

The discontinuous appearance of the evaporite levels in a silty-clay matrix can be explained by gypsum crystals growing in the capillary zone and displacing silts and clay as they grow (Warren, 1991). The sandstones and siltstones indicate a subtle siliciclastic influx, while the sandy dolomite level could indicate the proximity of the coastline. The facies association and the transitional position in the succession of this small unit tends to suggest a coastal character for this evaporitic environment. It has been interpreted a coastal sabkha, potentially linked with shallow groundwater resurgence or supratidal water flooding.

4.6.2 Facies association 6 (FA6 – Table 1 and Figure 4.10): Peritidal carbonates

Description

The upper part of the Tamarout Formation in Tikki and extending to the western part of the basin is an association of dolomites, marls and limestones (Adams et al., 1980). Facies association FA6 is dominated by dolomitic facies (FA6-S1a and FA6-S1b). The amount of siliciclastics in the system varies between 5-40%. The association of peloidal and oolitic horizons (FA6-S3, FA6-S4 and FA6-S5) with microbial mats (FA6-S8) and dolomite levels

(FA6-S1a and FA6-S1b) alternating with some marls (FA6-S7) and MST (FA6-S6) are found across the entire basin. However, in the East and North of the basin these facies alternate with extensive evaporites (FA6-S9a), while in the South and West of the basin the evaporites disappear and are replaced by thickening upward breccia units (FA6-S9b).

Different trends can be identified in this facies association, with units distinguished depending on the proportion of each facies. Unit T1 forms the base of the formation and is dominated by laminated dolomite composed of euhedral and anhydral crystals (FA6-S1a and FA6-S1b), cross-laminated peloidal and oolitic GST (FA6-S4 and FA6-S5) and breccias (FA6-S9b). In the North-East of the basin, Unit T2 is dominated by green and red marls (FA6-S7), thinly bedded dolomite horizons (FA6-S1b), nodular and continuous evaporite horizons with thin dolomite and stromatolites horizons (FA6-S9a) and thinly bedded peloidal and oolitic PST (FA6-S3). Towards the West and South, thin dark stromatolite horizons (FA6-S8), meter-thick breccias (FA6-S9b), cross-laminated oolitic and peloidal PST and GST (FA6-S3, FA6-S4 and FA6-S5) and thinly bedded dolomite (FA6-S1a) are the more common facies. Microbial horizons of 5 to 10 cm thick (FA6-S8) are noticeable by their dark grey colour, kerogenic smell and irregular bedding. This facies is followed by centimetre to meter-thick coarsening-upward breccias (FA6-S9b), including stromatolites elements at the base of the unit and evolving to thinly laminated dolomite beds and oolitic GST with cross-bedding and wave ripples (FA6-S1a, FA6-S1b and FA6-S5). The unit T3 consists of interbedded yellow crystalline and sandy dolomite (FA6-S1a and FA6-S1b) with dark grey oolitic PST and GST (FA6-S4 and FA6-S5), breccias with centimetre to decimetre-large elements (FA6-S9b) and grey marls (FA6-S7).

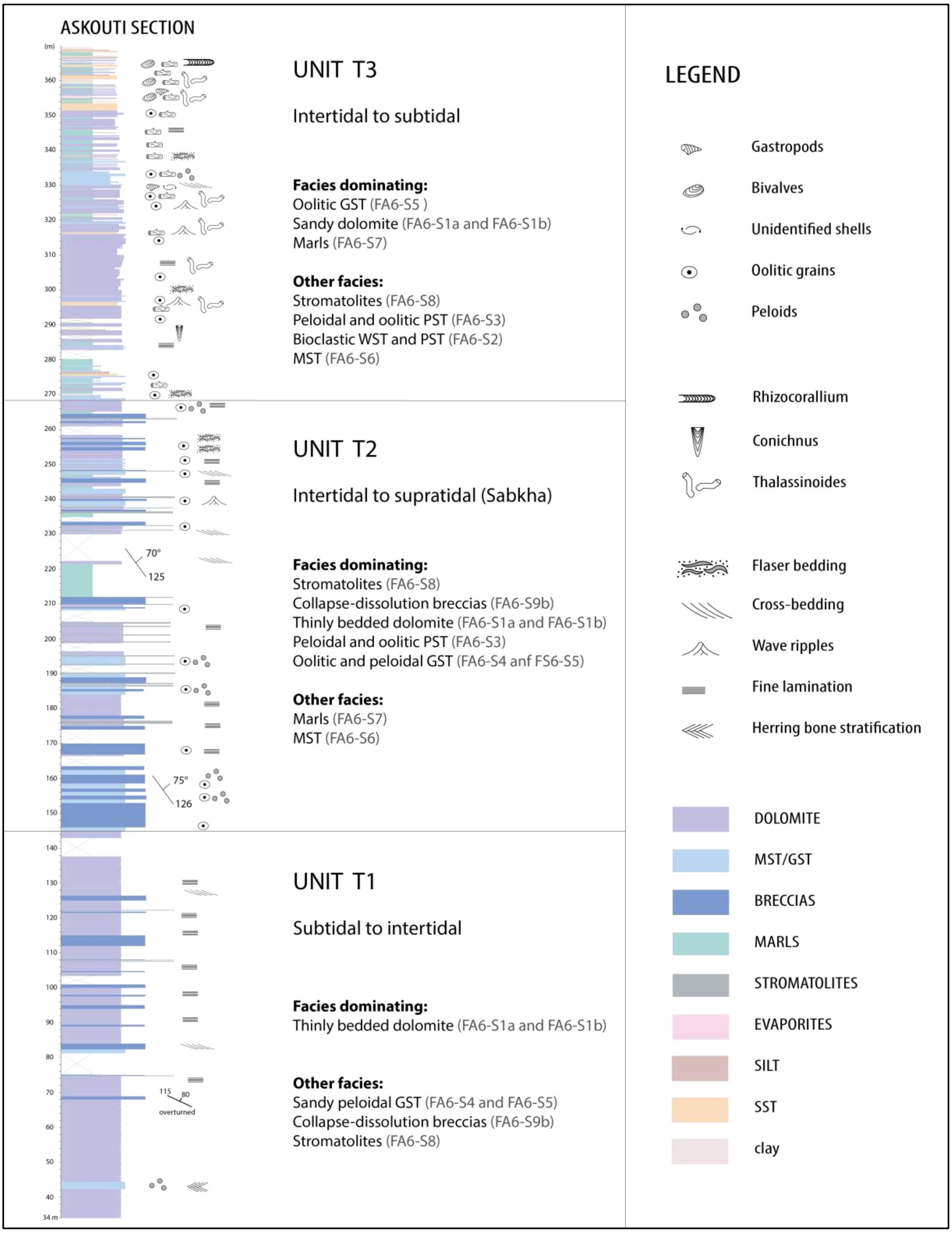

Figure 4.8: Askouti section.
Example of the Units T1, T2 and T3 facies organisation and stratigraphy.

In the breccia horizons (FA6-S9b), angular clasts are inversely graded and brecciation intensity decreases to the top of each bed, with tops commonly only consisting of fractured dolomites. The lateral continuity of the beds is variable; some of the smaller beds present a lenticular shape while the larger beds are continuous over 50 m. The clasts lithology consistent with the surrounding host sediments (laminated dolomite, oolitic GST and stromatolites).

Table 1: Facies descriptions and facies associations of the Lower Jurassic depostis.

Facies Association	Facies Name	Main elements	Sedimentary features and bioturbation	Large scale observation and bedding Thickness	Depositional environment
FA1	FA1-S1 Quartzites conglomerates	Quartzite pebbles and cobbles, well rounded, poorly sorted conglomerates, clast-supported, polymodal, medium sandstones and granules matrix Pebbles, cobbles : Quartzite 99%, Metabasalt and basalt 1%	Cross-bedding and parallel laminations in conglomerate Erosion surfaces	Very continuous lateraly m-Dm	Alluvial fan
	FA1-S2 Medium to coarse sandstones	Poorly sorted medium sandstones, sub-angular to sub-rounded, lenses and horizontal beds interbeded conglomerates	Crossbeds alternating between medium sandstones and conglomerates Planar cross bedding Trough cross-bedding	cm	
	FA1-S3 Fine sandstone	Fine sandstones, poorly sorted, subangular grains	Occasional nodular horizons and roots traces, paleosols	cm-m	
FA2	FA2-S1 Quartzites conglomerates	Quartzite pebbles and cobbles, subrounded to subangular, poorly sorted conglomerates, clast-supported, coarse sandstones matrix Pebbles, cobles : Quartzite 90%, green basalt 10% Pebble size un to 10 cm, average size 2 cm	lenticular beds planar cross-bedding low angle cross-bedding massive erosive base	Disappear laterally 30 cm -2 m	Braided river
	FA2-S2 Coarse sandstones	Poorly sorted coarse sandstones with granules, sub-rounded to subangular	Planar cross bedding Trough cross-bedding	5-30 cm	
	FA2-S3 Medium to fine sandstone	Moderately sorted medium sandstones fining upward to fine sandstones	assymetric current ripples planar cross-bedding	5-15 cm	
FA3	FA3-S1 Red clay and siltstones	Red clay and siltstones alternating in thin horizons. Thicker horizons of clay with thin cm-thick horizons of red silt Horizons of clay coarsening	Horizontal laminations Current ripples in silt lenses	Locally laterally passing to conglomerates 50 cm -5 m	Flood plain

		upward to silt			
	FA3-S2 Very fine sandstones	Very fine sandstones or fine sandstones Very thin horizons alternating with silt horizons and thicker beds thinning upward to silt locally small carbonate cement content	Current ripples Flaser bedding	cm-m	
	FA3-S3 Fine sandstones	Fine sandstones horizons alternating with thin (20 cm) silt horizons	Low angle tabular cross-laminations Climbing ripples with sinuous crests	30 - 40 cm	
	FA3-S4 Conglomerates	Matrix-supported conglomerates, thinning upward, quartzite and silty pebbles+-	Tabular cross bedding Cross-laminations	50 cm - 1.5 m	
FA4	FA4-S1	Red and green clays and marls	No bedding	cm-m	Coastal plain
	FA4-S2	Red siltstones	rip-up clasts current ripples root traces paleosoils	20 cm - 1.5 m	
	FA4-S3	Dolomitic sandstones	Planar laminations Flaser bedding Wavy bedding	20 cm - 1 m	
	FA4-S4	Sandy dolomites with carbonate-coated sand grains and thin ooids	Planar laminations Flaser bedding Tabular corss-bedding	10 cm - 80 cm	
FA5	FA5-S1 Red clay	Red clay	No bedding	cm-m	Sabkha
	FA5-S2 Evaporites nodules	Gypsum nodules horizons	Nodular	cm	
	FA5-S3 Sandy carbonates	Sandy and silty carbonate beds	Horizantal parallel bedding	cm	
	FA5-S4 Sandstones and siltstones	Thin horizons of very fine sandstones and siltstones	No bedding	mm-cm	
FA6	FA6-S1a Dolomite	Dark grey, light grey and yellow crystaline dolomite 0-10% Bivalve shell fragments Quartz rich horizons 20% very fine to fine sand Horizons with wood fragments up to 5%	Thinly bedded Horizontal laminations Heavily bioturbated horizons Rare Conichnus	Vuggy horizons cm-m	Peritidal
	FA6-S1b Dolomite	Light grey and dark grey micritic dolomite Occasional quartz rich levels, 20% very fine to fine sand Wood fragments	Horizontal laminations Faint cross bedding Cross-laminations of fine sand	Dark beds kerogenic cm-m	

	FA6-S2 Bioclastic WST/PST	Brachiopods, bivalves or gastropods rich horizons, elements up to 1cm, mostly unbroken, well sorted in a micrite matrix	Horizontal orientation of the grains	cm	
	FA6-S3 Peloidal and oolitic WST/PST	Poorly to moderately sorted pelloids and ooids fine quartz grains 0-40% Bivalves and brachiopods 0-10% Rare echinoderm fragments 0-2%	Peloids and ooids wavy bedding and lenticular bedding in muddy matrix Horizontal laminations with fine sand horizons	cm-m	
	FA6-S4 Peloidal GST	Well sorted peloids 0-40% fine quartz, well sorted	Massive beds and herringbone cross-stratification with peloids Low angle cross stratification of fine sand and peloids Wave-formed cross-lamination with discordant internal laminae	cm-m	
	FA6-S5 Oolitic GST	Medium to well sorted ooids, peloids and aggregates, beds with ooids quartz nuclei Beds with distorted ooids 0-20% shell fragments oriented parallel to the bedding 0-15% fine quartz grains, angular to subrounded Presence of authigenic quartz 0-5%	Swaley cross-stratifications Cross-bedding 5-40 cm Cross-laminations Horizontal lamination Wave ripples Rare Rhizocorallium	cm-m	
	FA6-S6 MST	Micrite with 0-15% clay content MST clasts	Occosional horizontal bioturbation	cm	
	FA6-S7 Marls	Green and grey marls	No bedding	cm	
	FA6-S8 Stromatolites	Dark micritic horizons	Stratiform irregular laminations	Kerogenic, laterally extensive cm	
	FA6-S9a Evaporites	Gypsum beds Thin mm thick horizons of grey MST and inclusions of MST granules	Waves ripples in the MST horizons Gypsum horizons massive or thinly laminated	Lateral variations of thickness Often associated with post-sedimentary deformations cm-m	

	FA6-S9b Breccias	Dolomitic angular pebbles and cobbles of micrite, oolitic GST and stromatolites	Breccias coarsening upward, top bed of the breccia horizon usually less broken	Lateraly discontinuous beds, breccia lenses Layers above not deformed cm-m	
FA7	FA7-S1 ooiltic PST/GST	dolomitic recrystallized oolitic and peloidal GST facies	Thick beds, sedimentary features averprinded by dolomitisation	dm-m	Open marine
	FA7-S2 Cristalline dolomite	Light grey to light grey homogeneous crystalline dolomite, porous 10-20% with recristallised shell fragments and phantom of rounded grains	Thick beds, sedimentary features averprinded by dolomitisation	dm-m	
	FA7-S3 coated grainsl PST	coated grains (oncoids?) 10-40%, shell fragments 10-30%, Peloids 5-20%, micrite matrix, partially dolomitised	Horizontal bedding	dm-m	
	FA7-S4 coated grains FST/RST	coated grains (oncoids?) 40-70%, shell fragments 10-30%, Peloids 5-20%, gastropods 0-10%, micrite matrix, partially dolomitised	Horizontal bedding	cm-dm	
	FA7-S5 Oncoidal RST	Oncoids 50-70%, crinoids 5-10%, shell fragments 5-10%	Horizontal bedding	cm	
	FA7-S6 Crinoid FST	Crinoids 10-30%, oncoids 10%, coated grains 20-40%, shell fragments 5-10%, gastropods 0-5%, presence of belemnites and ammonites	Horizontal bedding	cm	
	FA7-S7 bioclastic WST/PST	Shell fragments 10-50%, coated grains 5-15%, crinoids fragments 2-10%, partially dolomitised	no bedding	cm-dm	
	FA7-S8 bioclasticPST/GST	Coated grains 20-50%, gastropods 0-5%, coral fragments 0-5%, shell fragments 5-15%, crinoids fragments 5-10%, belemnites fragments 0-2%, dolomitised	Horizontal bedding	dm-m	
	FA7-S9 Dolomite	Yellow and pink dolomite, porosity made by dissolved shells	Thalassinoides bioturbation preferentially recrystallised	heavily fractured	

Facies Association	**Facies Name**	**Main elements**	**Sedimentary features and bioturbation**	**Large scale observation and bedding**	**Depositional environment**
FA4	FA4-S1a Dolomite	Dark grey, light grey and yellow crystaline dolomite 0-10% Bivalve shell fragments Quartz rich horizons 20% very fine to fine sand Horizons with wood fragments up to 5%	Thinly bedded Horizontal laminations Heavily bioturbated horizons	Vuggy horizons cm-m	
	FA4-S1b Dolomite	Light grey and dark grey micritic dolomite Occasional quartz rich levels, 20% very fine to fine sand Wood fragments	Horizontal laminations Faint cross bedding Cross-laminations of fine sand	Dark beds kerogenic cm-m	
	FA4-S2 Bioclastic WST/PST	Brachiopods, bivalves or gastropods rich horizons, elements up to 1cm, mosttly unbroken, well sorted in a micrite matrix	Horizontal elongation of the grains	cm	
	FA4-S3 Peloidal and oolitic WST/PST	Poorly to moderately sorted pelloids and ooids fine quartz grains 0-40% Bivalves and brachiopods 0-10% Rare echinoderm fragments 0-2%	Peloids and ooids wavy bedding and lenticular bedding in muddy matrix Horizontal laminations with fine sand horizons	cm-m	
	FA4-S4 Pelloidal GST	Well sorted pelloids 0-40% fine quartz, well sorted	Massive beds and herringbones cross-stratification with peloids Low angle cross stratification of fine sand and peloids Wave-formed cross-laminations with discordant internal laminae	cm-m	
	FA4-S5 Oolitic GST	Medium to well sorted ooids, peloids and aggregates, beds with ooids quartz nuclei Beds with distorted ooids 0-20% shell fragments oriented parallel to the bedding	Swalley cross-stratifications Cross-bedding 5-40 cm Cross-laminations Horizontal laminations Wave ripples	cm-m	

		0-15% fine quartz grains, angular to subrounded Presence of authigenic quartz 0-5%			
	FA4-S6 MST	Micrite with 0-15% clay content MST clasts	Occosional horizontal bioturbation	cm	
	FA4-S7 Marls	Green and grey marls	No bedding	cm	
	FA4-S8 Stromatolites	Dark micritic horizons	Stratiform irregular laminations	Kerogenic, laterally extensive cm	Protected environment

Interpretation

FEATURES COMMON TO THE THREE UNITS

The stratabound dolomites (FA6-S1a and FA6-S1b) represent dolomitisation of MST and oolitic carbonates. The dolomite horizons consist of replacive euhedral-anhedral dolomite crystals formed by reflux of low temperature dolomitizing fluids (Al-Sinawi et al., 2017). This mechanism whereby hypersaline fluids descend into the subsurface is typical of peritidal carbonates and can be linked with the presence of evaporites (Lu and Meyers, 1998; Flügel, 2010).

The heterolithic stratification (FA6-S3) of oolitic and peloidal layers with mudstones is common. The ooids are formed in a environment with some energy while the peloids need little energy to be preserved and mud is deposited in a very low energy environment. The sharp contact at the base of the oolitic-streaked and peloidal-streaked muds and more gradational top can be explained by an environment with variable current flow velocity. Stronger tidal current bring in the oolitic or peloidal grains, whereas the subordinate tidal current will be depositing the mud in the system. These heterotithic stratifications are characteristic of an intertidal environment (Flügel, 2010).

The inverse grading, variable lateral extension, absence of preferential orientation of the clasts in the breccias beds and the absence of breccias where the evaporites are intact are evidence for the collapse-dissolution origin of the breccias (Friedman, 1997). Collapse occurred after deposition when evaporite beds came in contact with under saturated fluids, such as meteoric water or seawater of lower salinity (Warren, 2016). Associated with the breccia beds, some beds present characteristics of syn-sedimentary ductile folding.

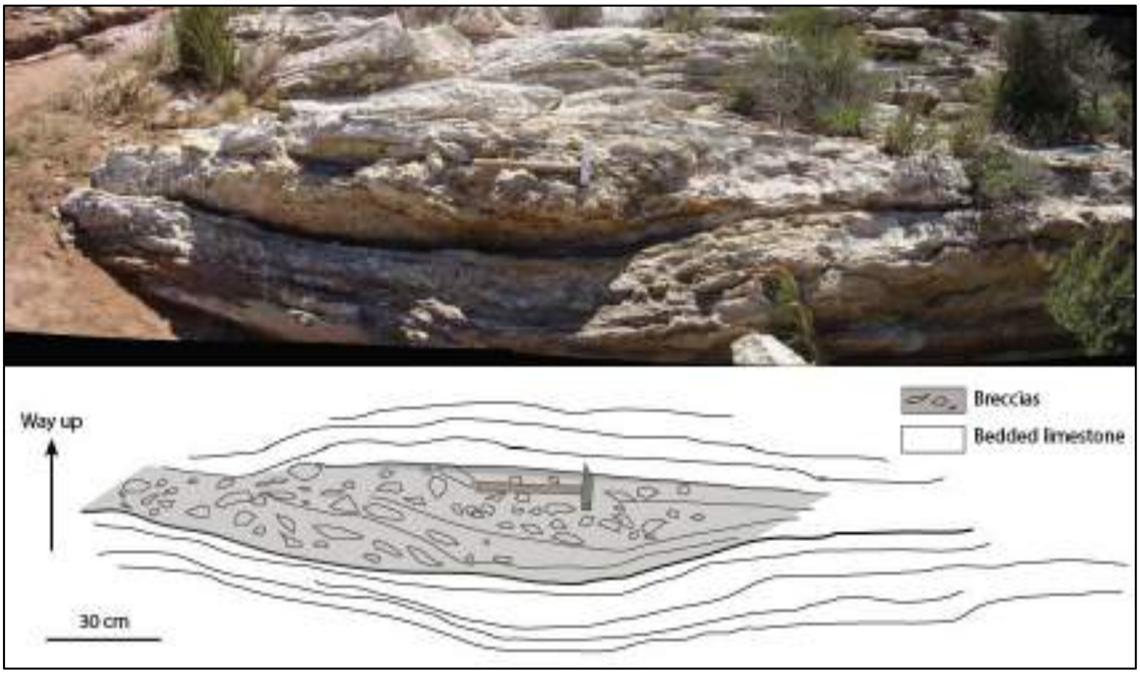

Figure 4.9: Example of lenticular breccia bed and associated syn-sedimentary ductile folding in the Tamarout Formation.

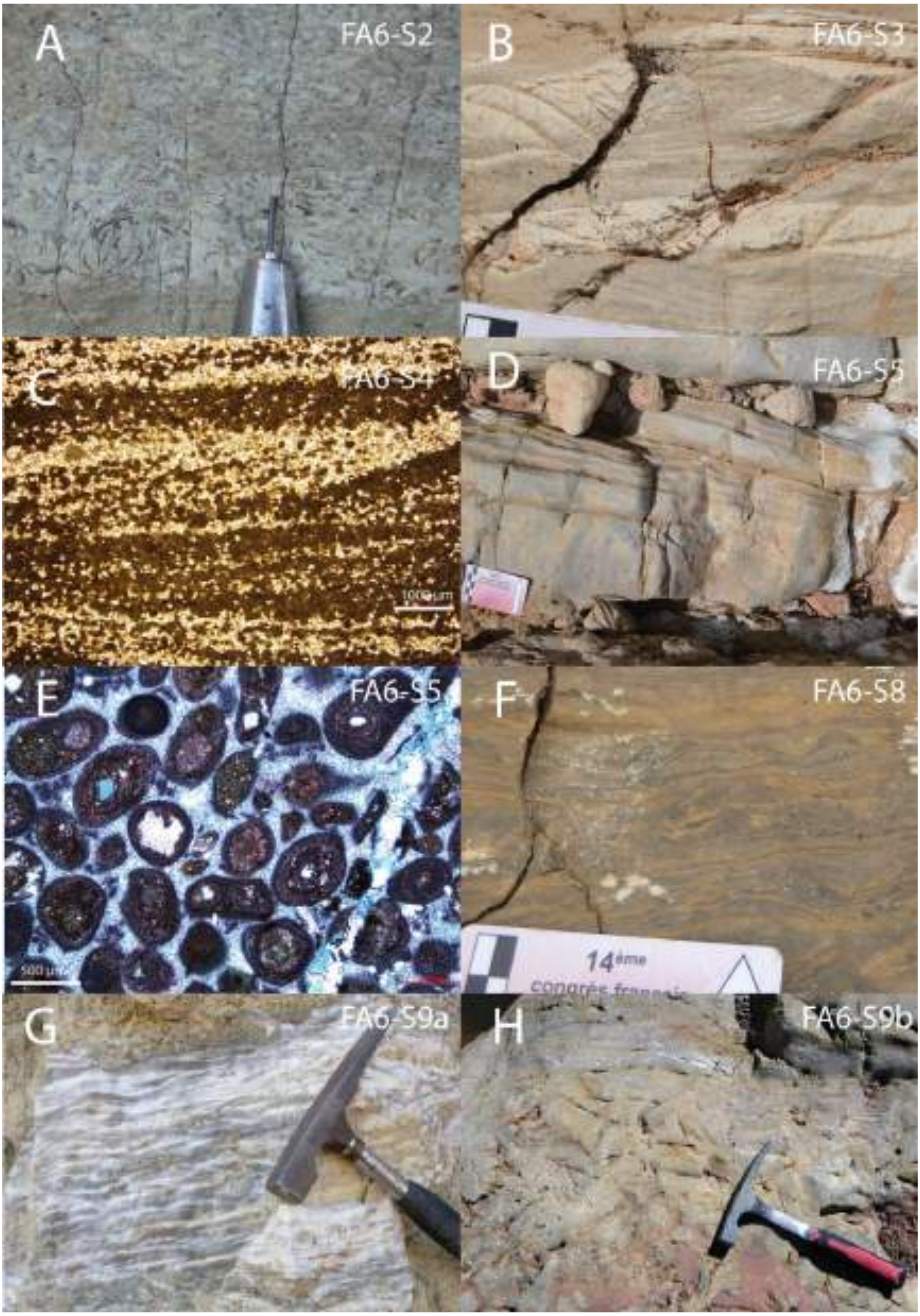

Figure 4.10: Facies of the Tamarout Formation.
Facies FA4-S2, Bioclastic PST with mm-scale bivalves shell fragments (A). Facies FA4-S3, wavy-bedded oolitic WST/PST (B). Facies FA4-S4, wave ripples in peloidal and fine sand GST (C). Facies FA4-S5, cross-beds in oolitic GST (D and E). Facies FA4-S8, Stromatolites (F). Facies FA4-S9a, Evaporites (G). Facies FA4-S9b, dissolution-collapse breccias (H).

UNIT T1 (Figure 4.8)

Unit T1 is mainly composed of thinly bedded dolomite horizons (FA6-S1a and FA6-S1b) associated with peloidal sandy GST (FA6-S4 and FA6-S5) and minor collapse-dissolution breccias (FA6-S9b).

In the peloidal GST (FA6-S4), a large proportion (up to 40%) of very well sorted fine grained sandstone and the presence of herringbone cross-stratifications are evidence for reworking by bidirectional currents in an intertidal environment. The identification of swaley cross-stratification in the oolitic GST (FA6-S5) demonstrates higher energy storm-wave currents, which are often developed in slightly deeper peritidal settings.

Bioclastic PST (FA4-S2) are rarer, but common to most outcrops the in Unit T1. The bioclasts are small bivalve shells or gastropods, always smaller than 1 cm, and either well rounded or unbroken. They are deposited in a muddy matrix with intraclasts and some fine quartz grains. The matrix indicates a relatively low energy while the good sorting and the roundness of the shell fragments indicate a higher energy environment. This facies can be attributed to subtidal environment where reworked fine grained bioclasts can be brought by tidal currents and form lag horizons (Flügel 2010).

The association of facies in the unit T1 range from rare intertidal to supratidal breccias to more common intertidal peloidal GST and dolomite down to subtidal oolitic GST. The overall environment of deposition is interpreted as dominantly subtidal to intertidal.

UNIT T2 (Figure 4.8)

The unit T2 is dominated by bedded dolomite, oolitic and peloidal sandy PST and GST, stromatolites and evaporite horizons or dissolution breccias.

The oolitic GST (FA6-S5) of unit T2 are tangential ooids partly overprinted by micritization. Tangential ooids are associated with very shallow water in high energy setting and hypersaline environment (Davies et al., 1978). In this unit, the oolitic and peloidal GST precede stromatolitic levels (FA6-S8). The stromatolites observed are biolaminites, stratiform microbially controlled carbonates, which develop in very low energy upper intertidal to supratidal environments (Hoffman, 1976; Beukes et al., 1989; Tucker, 2009).

Heterolithic stratification is also common in this unit, and displays wavy bedding and flaser bedding which are evidences of variations in the flow intensity typical for intertidal environments. Sandy peloidal GST is a common facies associated with wave ripples. The sorting of the ooids and the sand is very good which indicates a consistent flow, and the disappearance of the mud is diagnostic for higher energy tidal currents.

The evaporite levels (FA6-S9a) are composed of continuous and lenticular gypsum beds associated with dolomite levels. The thickness of the beds varies from a few centimetres to several meters. The lateral continuity of some evaporites horizons and their association to centimetre thick microbial laminated carbonates and dolomite levels indicates temporarily subaqueous conditions (Rouchy and Caruso, 2006; Warren, 2016). To develop extensive evaporitic levels, the environment must be very restricted, with little fresh-water or open-marine water coming into the system.

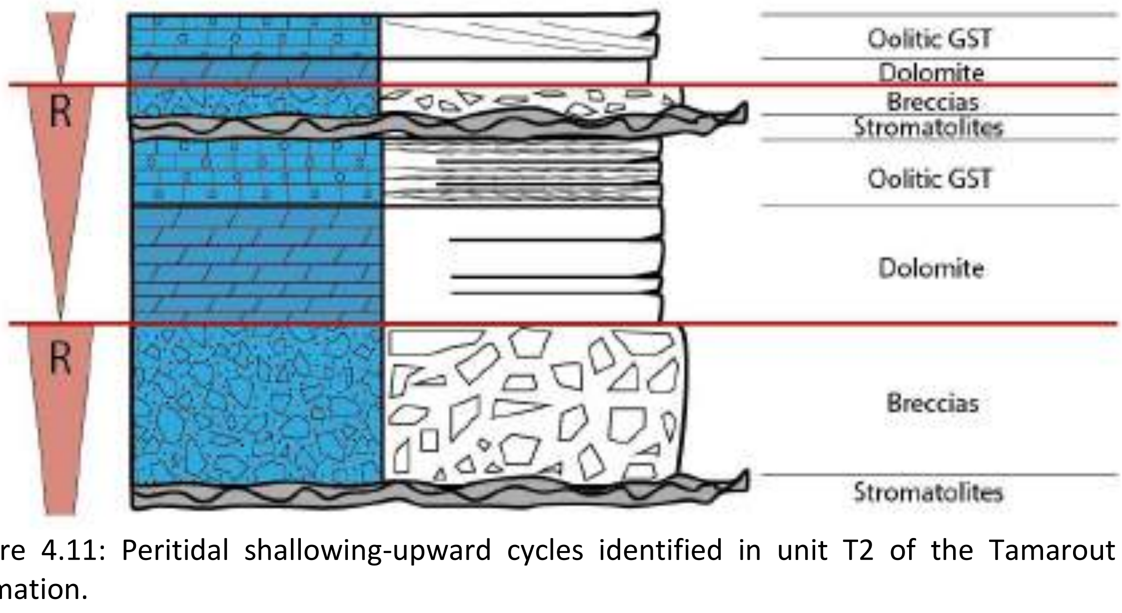

Figure 4.11: Peritidal shallowing-upward cycles identified in unit T2 of the Tamarout Formation.

Shallowing-upward cycles have been identified in T2 (Figure 4.11). The lower facies consist of intertidal dolostone (FA6-S1a and FA6-S1b) and peloidal and oolitic PST and GST (FA6-S3, FA6-S4 and FA6-S5), followed by inter- to supratidal stromatolites (FA6-S8) evaporites (FA6-S9a), often replaced by dissolution breccias (FA6-S9b).

Towards the South and West of the basin, the evaporites disappear and are replaced by breccias appearing over hundreds of meters, forming breccia beds from 20 cm up to 10m thick (FA4-S9b).

The association of facies FA6-S1b, FA6-S5, FA6-S8 and FA6-S9a/FA6-S9b is characteristic of sabkhas or small coastal salinas in a broad saline tidal flat environment (Friedman, 1997; Warren, 2016).

UNIT T3 (Figure 4.8)

Unit T3 is dominated by dark grey oolitic GST (FA6-S5) and sandy dolomite (FA6-S1a and FA6-S1b) associated to grey marls (FA6-S7) and rarer breccia horizons (FA6-S9b).

Sandy peloidal PST and GST (FA6-S3 and FA6-S4) and sandy dolomite are also very common facies in units T2 and T3. The siliciclastic elements are fine to very fine, angular to subangular quartz grains associated with low-angle cross-bedding, wave ripples and flaser bedding. They are interpreted to record inter-tidal to sub-tidal environments where oscillatory currents are common, characteristic of open-marine settings. Some horizons with up to 20% rounded shell fragments oriented parallel to the bedding indicate a high energy environment.

In Unit T3, oolitic GST (FA6-S5) are composed of compound ooids and single ooids with radial concentric fabric surrounded by tangential concentric microfabric. The radial fabric can form in marine setting as well as in saline and fresh-water lakes and the transition to tangential concentric fabric reflects a change of water energy to a more agitated environment (Tucker, 2009; Flügel, 2010). The ooids are well sorted and their nuclei are made of different type of grains, commonly fine quartz grains, they are also made of bivalves rounded shell fragments and ostracods. These microfacies features associated to planar cross-bedding are reflecting high energy marine environments, most likely oolitic shoals or tidal bars.

The association of minor marls horizons (FA6-S7) with dm-thick oolitic PST (FA6-S4), some oolitic grainstones (FA6-S5) presenting worn skeletal grain and abundant dolomite beds (FA6-S1a and FA6-S1b) is a sign of reworking of the fine-grained material in low energy settings. This is interpreted as an intertidal to subtidal inner ramp environment. The dissolution breccias only appear at the base of this unit, and mark the transition from the intertidal to supratidal environments of unit T2 to the subtidal environment of unit T3.

Tamarout Formation Interpretation

The reported presence of brachiopods in the literature (Ambroggi, 1963) and fragments of echinoderm observed in thin sections are evidence for an overall open marine environment. The association of facies FA6 is a combination of normal seawater and hypersaline environments, with tidal influence and more restricted periods. It has been interpreted as a very shallow inner ramp environment with periods of marine restrictions and salinity increase.

4.7 Ameskhoud Formation

The Ameskhoud Formation present variable facies across the basin and is here described from the location of Assif El Hade where it is 240 m thick. This location is situated in the middle of the basin and presents characteristics transitional between the two other type of facies observed (Fluvial dominated in Tikki and Askouti; shallow-marine carbonates in the Amsittène Anticline).

4.7.1 Facies interpretations

Description

Lower part of the Ameskhoud Formation (110 m):

The Ameskhoud Formation overlies the oolitic Unit T3, where the proportion of quartz grains was already up to 20-40%. The lower part of the Ameskhoud Formation is composed of sandy and silty dolomite alternating with red clay and marls. The siliciclastic fraction of the sediment increases rapidly upsection and well-sorted dolomitic siltstones and very fine to fine grained sandstones become dominant. These sandstones and siltstones contain wave-ripples, horizontal bedding and low angle cross-bedding. Some beds are extensively overprinted by bioturbation. Highly bioturbated beds have an upper surface dominated by *Ophiomorph*a and *Rhizocorallium*. Towards the upper part of this unit, the dominant ichnofauna is *Diplocraterion*. Some erosive irregular beds of coarse to very coarse sands also appear in the upper part of this unit, alternating with non-erosive fine and coarse sandstones become common.

Upper part of the Ameskhoud Formation (130 m):

The upper part of the formation is composed of thick red clay units with intercalations of silt and sandstone beds and rare nodular carbonate horizons. The sandstones are fine to coarse grained, well to poorly sorted and display horizontal bedding, cross-bedding, cross-sets and current ripples. Some of the medium and coarse grained sandstones have an erosive base and a lenticular shape, whereas others present flat top and base. These erosive beds fine-upwards and contain red clay mud-clasts and locally some angular quartz granules just above erosive surfaces, followed upsection by planar and trough cross bedding with mud-clasts. The upper part of the beds is generally bioturbated or presents current-ripples and occasional ripple-drift cross-lamination. More extensive beds of very fine to medium grained sandstones with thin horizontal bedding and cross-bedding have sharp but non erosive bases. Some of theses beds are also bioturbated and display roots traces. Beds with good sorting contain horizontal bedding, trough cross-bedding and ripple laminations with occasional mud drapes between the cross-beds and bidirectional currents. Some of these beds are heavily bioturbated, the most common trace fossil being *Thalassinoides* and *Arenicolites* on top of the beds. All individual sandstones beds are quite thin (up to 50 cm thick) while the clay horizons can be up to 5m thick.

Interpretation

Lower part of the Ameskhoud Formation

The sandy and silty dolomites of the lower Ameskhoud Formation are very similar to the upper part of the Tamarout Formation. The alternation with red mudstones indicates an increase in terrestrial influx. The well-sorted dolomitic sandstones and siltstones with wave ripples were deposited from oscillatory currents (Goldring and Bridges, 1973). *Ophiomorpha* is indicative of high energy marine conditions (Droser and Bottjer, 1989; Knaust, 2013). This association of sandstones and siltstones with some mud intervals are interpreted as a nearshore sequence (Howard and Reineck, 1981).

Upper part of the Ameskhoud Formation

Above the marine member, the main part of the Ameskhoud Formation is dominated by red mudstones. This marks the transition to a lower energy environment. The root traces are evidence for palaeosols and imply exposed conditions. The erosive fining-upward lenticular beds with cross-sets and current ripples have been interpreted as minor channels deposits. The cross-sets in sandstones are developed perpendicular to the channel axis, representing point-bar lateral accretion (Allen, 1963), while the ripple-drift cross laminations are interpreted as waning flows in the transition to flood plain red clays (Nanson, 1980). These isolated small channels, always separated by mud horizons, can be interpreted as ephemeral channels. The mud-dominated deposits with carbonate nodules and palaeosols are interpreted to have formed on a floodplain, subjected to ephemeral floods, depositing ripple and horizontal bedded sandstones (Turnbridge 1984, Muñoz et al., 1992). The association of well sorted sandstones with *Thalassinoides* and *Arenicolites* trace fossils is characteristic of tidal flat deposits (Gerard and Bromley, 2008). Occasional bidirectional currents and mud drapes between the cross-beds also indicate a tidal influence (Boersma & Terwindt, 1981). The upper part of the Ameskhoud Formation in the locality of Assif El Hade is an association of intertidal (tidal flat) and supratidal (tidal marsh, ephemeral channels) deposits (Terwindt, 1988).

4.7.2 Regional variations

4.7.2.1 Askouti

To the South of the Basin, in the locality of Askouti, the Amsekhoud Formation is thinner (100 m) and dominated by red mudstones and siltstones interbedded with lenticular bedded conglomerates and gypsum beds. The gypsum beds have a lateral extent of between 10 and 200 m, and a thickness up to 20 cm. The conglomerate beds have an erosive base and fine upward. They contain some cross beds and are poorly sorted. The red mudstones and siltstones locally form successions up to 8m thick and display abundant root traces. Evaporite horizons become less common upward and disappear near the top. At Askouti this formation contains continental deposits with paleosols and conglomerates, which have been interpreted as braided fluvial channels (Miall, 1977, 1996; Bridge, 2003), interbedded with playa evaporitic deposits. The gypsum beds indicate an arid environment that was occasionally flooded and later the sea water evaporated. These deposits present characteristics of playa lake evolving to fluvial dominated deposits (Handford, 1982).

4.7.2.2 Tikki

In the locality of Tikki, along the Argana Valley (Figure 4.2), the Ameskhoud Formation is more continental than in Assif El Hade. No tidal influence can be observed and the succession contains massive and horizontally bedded red mudstones and siltstones interbedded with sandstones and conglomerates. The conglomerate units are crudely bedded and tabular cross stratified. They are often part of fining upward units and topped by cross-bedded and horizontally bedded sandstones. The base of these units is often erosive. Channelized units can be observed and massive cross-sets of sandstones or conglomerates have been interpreted as stream-flow deposits, braided bar or lags on channel floor (Rust, 1972; Nemec and Postma, 1993; Lunt and Bridge, 2004). The mud and silt dominated facies, with occasional roots traces, are laterally associated to these fluvial channels and have been interpreted as flood plain deposits.

4.7.2.3 Amsittène and TMS1

Along the Amsittène anticline and in the wells ESS-1 and TMS-1, the Middle Jurassic deposits display a very different character from the rest of the basin. This interval is composed of dolomite beds alternating with oolitic grainstones, some anhydrite and gypsum horizons and some grey and red marls horizons. The facies of these wells and outcrops are very similar to the facies described in the Units T1 and T2 of the Tamarout Formation. The presence of red marls and evaporites can be linked to coastal salina deposits and the oolitic and dolomitic beds can be interpreted as peritidal to subtidal deposits.

4.8 Depositional environments

The continental deposits of the Amsittène Formation consists of facies associations FA1, FA2, FA3 and FA4 (Table 1), interpreted to be flood plain and braided river deposits (FA2 and FA3). The distribution of the deposits varies across the basin. In the Agadir sub-basin, the more proximal deposits are composed of braided river, flood plain and alluvial fan deposits. In the area of Tikki, the Amsittène Formation evolves vertically from a flood plain to alluvial fan deposits. This alluvial fan probably relates to activity along an ENE-WSW trending fault, parallel to the major Tizi N'Test fault (Laville and Petit, 1984; Haafid et al., 2000; Frizon de Lamotte et al., 2009) which can be traced from the Argana Valley to the East of the Imouzzer Anticline. This fault might be linked to the western termination of a Lower to Middle Liassic rifting phase of the Atlas Tethys (Frizon de Lamotte et al., 2009). Erosion of the footwall would have sourced the local alluvial fans in Tikki, while most of the sand-grade deposits in the Agadir Basin probably came from the erosion of older highs further afield, such as the Western Meseta and the Jebilet (Figure 4.12, A).

The vertical succession of floodplain to coastal plain deposits in the upper part of the Amsittène Formation records a transgression, where marine conditions were ultimately established (the shallow marine deposits of the Tamarout Formation). Units T1 and T3 of the Tamarout Fm record shallow marine open ramp oolitic and bioclastic WST to GST all over the basin, with locally oolitic shoals (FA6-S5). Sabkha and coastal plain deposits can be identified behind the shoreline, suggesting an important marine influence but disconnected from the platform and only received occasional marine incursions.The widespread development of evaporites, stromatolites and dissolution breccias in unit T2, reflects a more restricted environment with hypersaline conditions. This unit alternates between restricted (Figure 4.12, B2) and open marine deposits (Figure 4.12, B1). During restriction, sabkha or coastal salinas developed in a large saline tidal flat. Marine ingressions led to a more open marine system with development of higher energy facies.

The presence of siliciclastic deposits in the Tamarout formation can be linked to a major river system entering the Essaouira Basin, sourced from uplifting massifs to the east. Red sandstones can be identified, often associated to evaporites. The sand influx can also be linked to smaller rivers entering the Agadir Basin (Figure 4.12, B1).

The Tamarout Formation is overlain by the Ameskhoud Formation. This formation is strongly regressive, with a basal unit composed of intertidal siliciclastic deposits that grade upward to continental flood plain and braided river deposits.

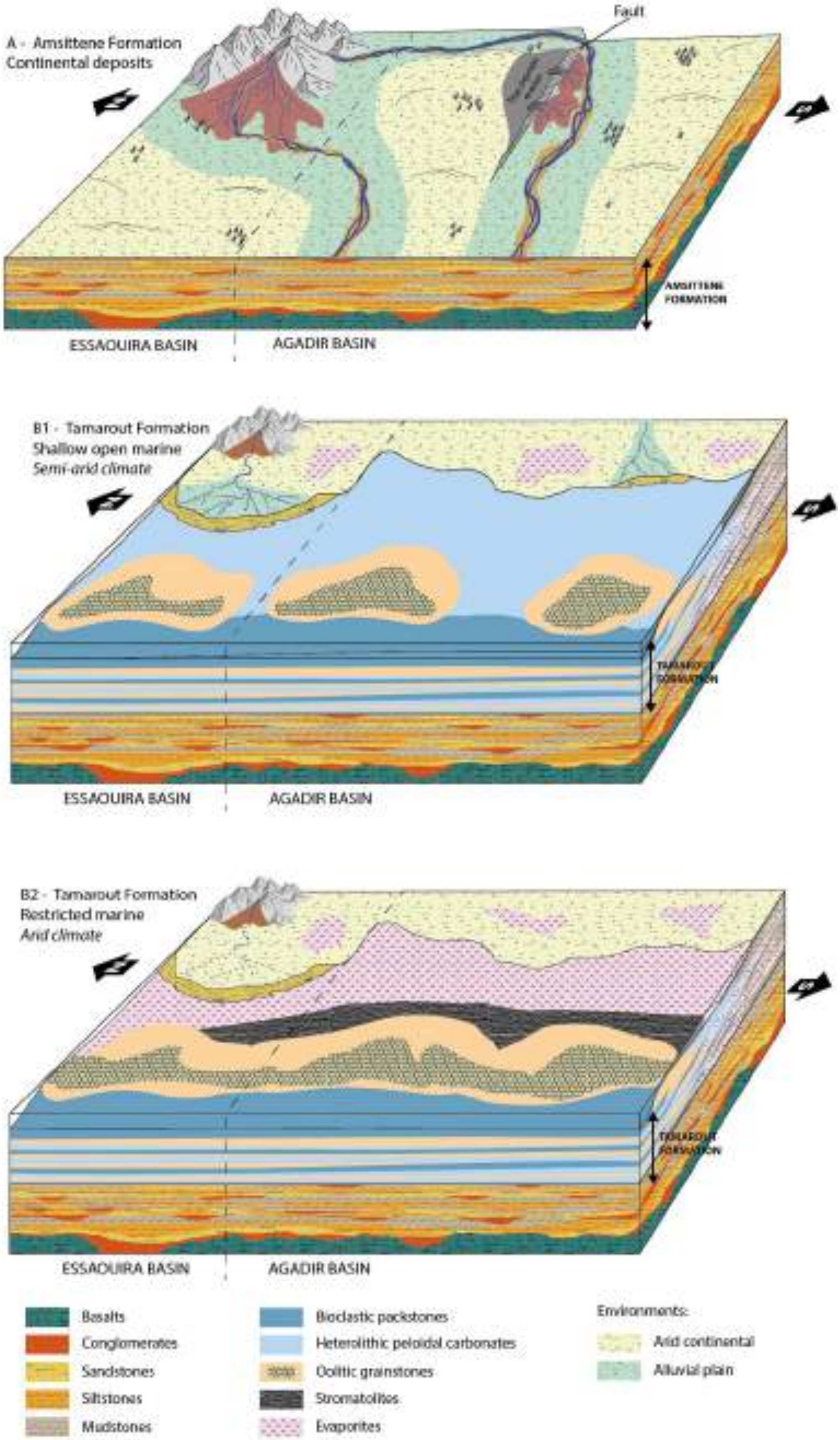

Figure 4.12: Depositional environments for the Amsittène Formation and Tamarout Formation.

4.9 Correlations

To understand the variations of facies and the lateral extent of the formations, the sedimentary sequences of 5 sections and a well have been studied in more depth. The well Essaouira-1 (ESS1) (Figure 4.3) was used to understand the extent and character of the Arich Ouzla Formation, a formation that otherwise only outcrops in the core of the Amsittène Anticline.

4.9.1 Sequence stratigraphic concepts

The nomenclature used is derived from the Depositional Sequence IV described by Catuneanu et al (2009, 2011), based on Hunt and Tucker (1992, 1995) and Helland-Hansen and Gjelberg (1994). Two major types of surfaces are used as correlation horizons in this paper.

The Maximum Flooding Surface (MFS) is defined as the point of maximum transgression of the succession in each sequence and marks the transition of depositional trend from retrogradation to progradation (Frazier 1974; Posamentier et al. 1988; Van Wagoner et al. 1988). The MFS lies above the Transgressive System Tract and in carbonate settings, it can be expressed by the deepest-water facies of a succession or by condensed sections or marine hardgrounds (Catuneanu et al., 2011).

The Sequence Boundary (SB) (Hunt and Tucker, 1992) represents the end of base-level fall; it is a compound type of boundary represented by a correlative conformity or subaerial unconformity. A subaerial unconformity forms as result of fluvial erosion, pedogenesis or karstification in subaerial settings and thus marks a period of relative sea-level fall (Posamentier et al., 1988; Aitken and Flint, 1996; Plint and Nummedal 2000). The correlative conformity of the subaerial unconformity can be traced basinward and marks the change from forced regression to lowstand normal regression in deeper marine settings (Hunt and Tucker, 1992).

4.9.2 Dating and Key Surfaces

Dating

The first datable surface of the sections is the CAMP basalts, which were active in Morocco around 199 Ma (Marzoli et al., 1999; Knight at al., 2004; Nomade et al., 2007; Davies et al., 2017). The top CAMP basalts can be used as a correlation surface.The first unit that has good biostratigraphic dating is the Arich-Ouzla Formation, where fauna is dated as Late Sinnemurian to Early Pliensbachian (Peybernes et al., 1987; Du Dresnay; 1988). The equivalent level in the DSDP borehole 547B (leg 79; Bernoulli and Kälin, 1984; Riegraf et al., 1984) (see 6.2) is also dated as Late Sinnemurian to Early Pliensbachian. The Toarcian age of the Tamarout Formation was established G. Dubar in Ambroggi (1963) based on brachiopod identification, including *Terebratula withakeri* (Oppel) and *Zeilleria anglica* (Walker), identifications confirmed by Peybernes et al. (1987) and supported by identification of Dacyclad *Sarfatiella dubari* in the Essaouira Basin (Bouaouda, 1987).

Age relationships based on correlation to High Atlas sections

In the Amsittѐne Formation, the fluvial facies is not dated, but the unconformity at the base of this formation has been correlated to its interpreted equivalent in the Central High Atlas, the Azilal Formation (see 9.1). The erosive base of the Azilal Formation has been dated as lower Toarcian (Lachkar, 2000; Wilmsen and Neuweiler, 2008; Malaval, 2016).

Sequence Stratigraphy (Figure 4.13)

The base of the red mudstones developed above the Triassic salt in the Essaouira sub-basin mark the last fully continental deposits before the onset of carbonate platform development during the Upper Sinemurian to Lower Pliensbachian. This marine flooding surface is recognised as a first transgressive surface (TS) and can be correlated between the Arich Ouzla section and the well ESS-1.

The base of the Amsittѐne Formation is defined by a strong erosion surface, cutting down to the CAMP basalts or the Trias in the Agadir sub-basin, and cutting into the upper part of the Arich Ouzla Formation along the Amsittѐne Anticline. In Arich Ouzla, the upper part of the Sinemurian is also marked by karstification and the contact to the Amsittène Formation present an angular unconformity, locally carving through the dolomites and filled by dolomitic and sandy breccias. This subaerial unconformity is defined as SB-1 that can be traced over the basin. In well ESS-1 its correlative conformity is marked by the abrupt transition from carbonate deposits to siliciclastics. Above SB1, the Amsittène Formation displays a retrogradational pattern from flood plain to coastal plain deposits in most of the basin and is followed by sabkha deposits in the Essaouira sub-basin.

The Tamarout Formation records the development of a marine succession. The maximum flooding surface MFS1 records the first appearance of intertidal to subtidal deposits in all the locations. It is followed in the south of the basin by a small regression topped by the second order sequence boundary SB1-1, directly followed by local aggradation leading to deposition of thick (>100m) supratidal carbonates. The deposits are then getting deeper again in the upper part of the Tmarout Formation, and a fourth order maximum flooding sucface (MFS1.1) is recorded by the onset of the development of a second succession dominated by subtidal deposits.The overlying Ameskhoud Formation records a strong regression in the south of the basin with development of fluvial deposits, but this is only represented by some sandy and silty levels in the Essaouira part of the basin. Fluvial erosion in the localities C, D, E and F (Figure 4.13) records the unconformity defines third sequence boundary (SB2) is represented by and its correlative conformity marked by shallower deposits and the influx of siliciclastics in the carbonate-dominated succession.

Towards the upper part of the Ameskhoud Formation, a small transgression recorded marine deposits up to the locality of Aqesri (Figure 4.13), in which locality the transgressive surface is combined with the maximum flooding surface (MFS2). In the Essaouira Basin, the expression of the MFS2 is linked to the transition from coastal plain or mixed marly deposits to carbonates-dominated marine deposits. The upper part of the Ameskhoud Formation presents a last regression in the Agadir part of the basin before becoming fully marine during the Late Middle Jurassic.

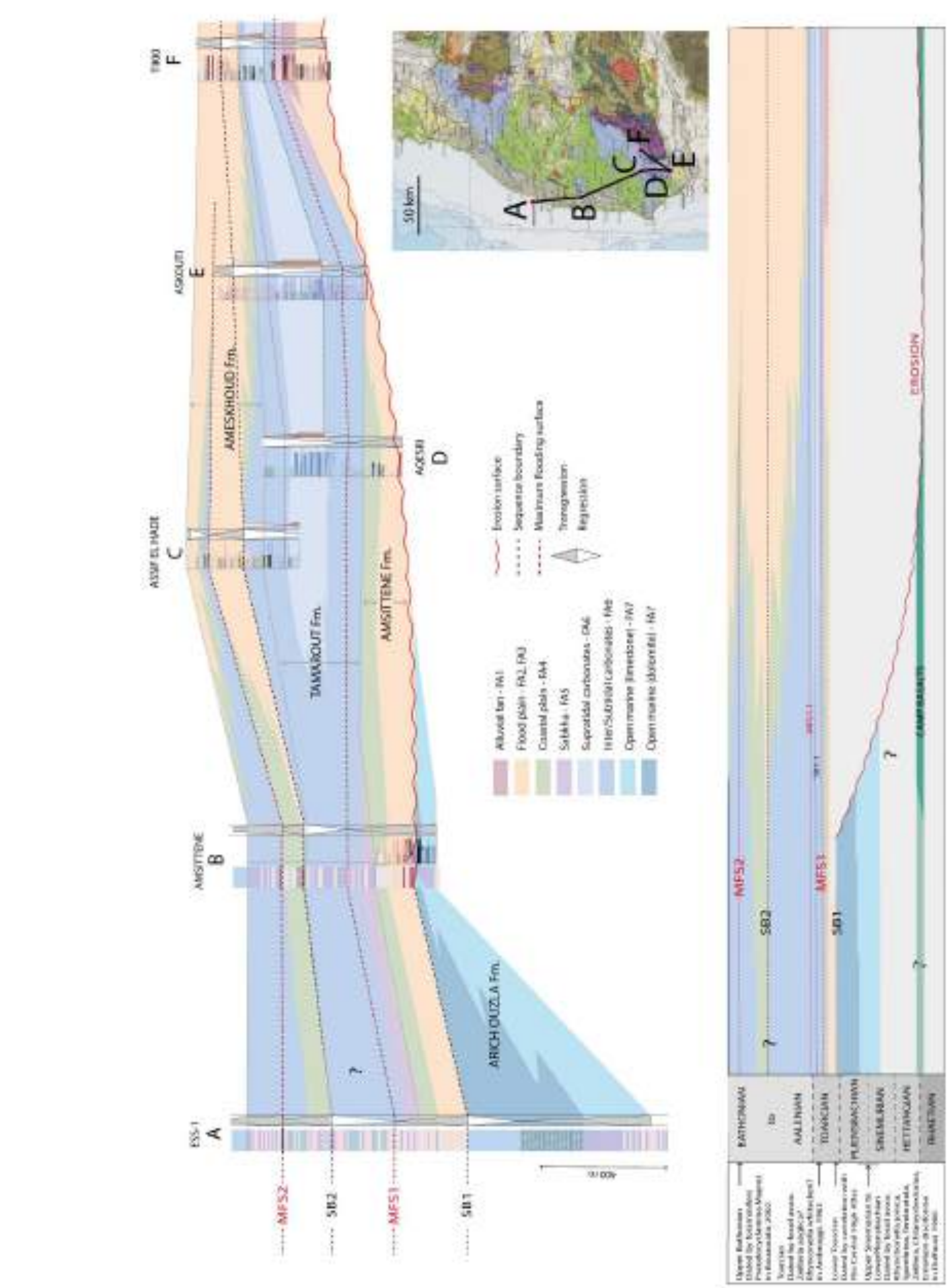

Figure 4.13: Lower and Middle Jurassic correlations between sections of the Essaouira-Agadir Basin.

4.10 Discussion

Following the end of rifting, the Lower Jurassic in the EAB records the first marine deposits transgressing into the Central Atlantic rift. A marine carbonate ramp developed during the Sinemurian-Pliensbachian (Arich Ouzla Fm and DSDP 547B leg 79 basinal equivalent) in the western and northern portions of the basin. The absence of Sinemurian-Pliensbachian deposits further east and south suggests that the earliest Jurassic marine incursion(s) did not reach these parts of the basin or were eroded during the basal Toarcian (see below). A second transgression initiated in the Toarcian, can be linked to the global early Toarcian transgressive event (Hallam, 1981). Shallow carbonate platform environments were established accross the study area (Amsittène and Tamarout Fms). This was followed by renewed regression in the Toarcian-Aalenian, leading to deposition of the continental Ameskhoud Fm. This broad succession is punctuated by phases of erosion and siliciclastic input that contrast with a general passive margin evolution. These are discussed in the following.

4.10.1 Toarcian erosion in the EAB and CHA

EAB stratigraphy

The contact between the Sinemurian Arich Ouzla Formation and the Amsittène Formation is erosive and locally filled by sandy dolomitic breccias (B, Figure 4.15). An angular unconformity can be observed that indicates tectonic activity (Figure 4.14). The top of the Arich Ouzla Formation is dolomitic and karstified, which indicates a period of emersion during the Pliensbachian or Early Toarcian.

Locally, breccias are found at the top of the Arich Ouzla Formation. The breccia can be mapped localy, they form a clear linear surface that separates the Arich Ouzla and Amsittène formations. It only extends laterally for a few metres. The breccias in contact with the two formations (Figure 4.14) are made of very angular autochthonous limestones boulders and pebbles. The limestone elements show very little transport and are in contact or floating in a fine to medium red sandstones matrix with occasional quartz granules and pebbles. This sedimentary feature prove that these breccias are not fault breccias putting in contact two different formations but rather related to the erosion of the limestones at the time of deposition of the continental sediments. Two different

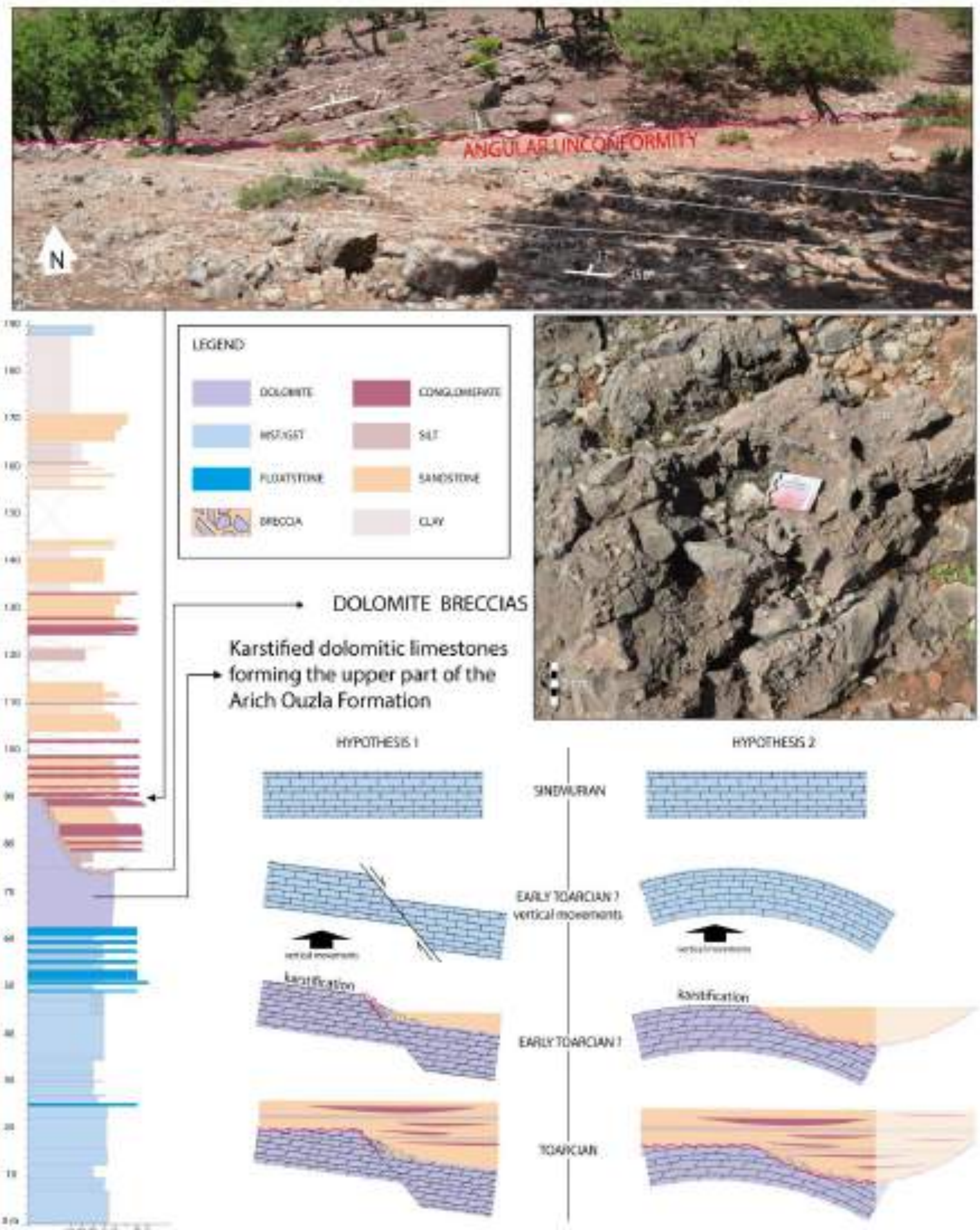

Figure 4.14 : Schematic evolution and hypothesis on the formation of the breccias at the contact between the Arich Ouzla Formation and the Amsittène Formation.

mechanisms can explain these deposits. Our first hypothesis (Figure 4.14) links the emersion and karstification of the upper Arich Ouzla Formation to vertical movements inducing some tilting and faulting of the formation. The faulting would have created relief in the landscape and the footwall would have been eroded and the elements deposited at the fault toe where it could be reworked by siliciclastic deposits. The second hypothesis implies vertical movements linked to the formation of the breccias with the erosion of the underlying carbonates by the overlying continental deposits.

Correlation with CHA

The Essaouira-Agadir Basin belongs to the western termination of the Atlas System, which is characterised by widespread Jurassic outcrops. In the Central High Atlas (CHA) these deposits have been recently studied to understand their stratigraphy and the impact of diapirism on the sedimentation (Saura et al., 2014; Martin-Martin et al., 2017; Moragas et al., 2016; Verges et al., 2017; Moragas et al., 2017; Teixell et al., 2017; Malaval, 2017; Joussiaume, 2017; Moragas, 2018). These deposits can be assigned to the Tethyan realm, wheras the EAB sections show affinity to the Atlantic margin of Morocco. A stratigraphic comparison of the Central High Atlas (CHA) and the Western High Atlas (WHA) has been undertaken to compare regional trends and analyse similarities and differences.

In the Central High Atlas, the Lower Jurassic deposits are composed of a thick open marine succession. Open platform carbonate sediments of the Bou Imoura Formation were deposited during the Sinemurian and followed by inner platform deposits of the Aganane Formation during the Early Pliensbachian. These deposits are the same age and of similar depositional environments as the Arich Ouzla Formation. This suggests the Central High Atlas and the EAB might have been in communication during this period, connected by an Atlasic sea, which allowed exchange between the Tethys and the Atlantic realm, supporting low-temperature thermochronology studies (Ghorbal et al., 2008; Ghorbal 2009; Saddiqi et al., 2009), which interpreted the Western Moroccan Arch area as a subsiding domain during the Late Triassic and Early Jurassic.

At Bin El Ouidane in the CHA (F, Fig 10), a major Toarcian erosive surface cuts down into the Aganane Formation. This surface is overlain by continental deposits and cuts down into the Sinemurian open platform and the CAMP basalts (Figure 4.15). The Toarcian unconformity recognised in the CHA has been here interpreted to be coeval with the major erosion surface at the base of the Amsittène Formation.

Provenance

The base of the Amsittène Formation across the EAB is an erosive surface that cuts down into the Sinemurian (Arich Ouzla Fm), the CAMP basalts (Aqesri) or into the Triassic red beds (Tixeront, 1974). It is the first Jurassic Unit in most of the onshore EAB and pinches out towards the NE. It records fluvial conditions of deposition and most paleoflow directions (Figure 4.15) suggest a sedimentary source from the Massif Ancien, presently bounding the basin to the east (Figure 4.15). The proximity of the source area is suggested by the pebbly to coarse grain size and poor overall sorting.

It is also possible that sediments from the Central Anti-Atlas could have been routed to the north in the remaining depression of the High Atlas rift basin, and then westward by local highs (e.g. salt diapirs in the Central High Atlas, (Moragas et al., 2016), the Massif Ancien acting as a basin divide (as already suggested during the Triassic (Domenech et al., 2018). The Massif Ancien is however not a documented topographic high in the Jurassic on the basis of low-temperature thermochronology results and time-temperature modelling studies (Charton, 2018).

The Toarcian siliciclastic deposits in the EAB and the CHA attest to a widespread erosive event and rejuvenation of the source area(s) during that time. In the EAB, the lower Jurassic continental deposits have paleoflow directions predominantly towards the W-SW, indicating a potential provenance area in the Meseta. Apatite fission track studies suggest that during the Late Triassic to Early Jurassic part of the West Moroccan Arch was exhumed during the Early Jurassic and was the most likely candidate source area, together with the Anti-Atlas (Ghorbal et al., 2008; Saddiqi et al., 2009, Charton, 2018).

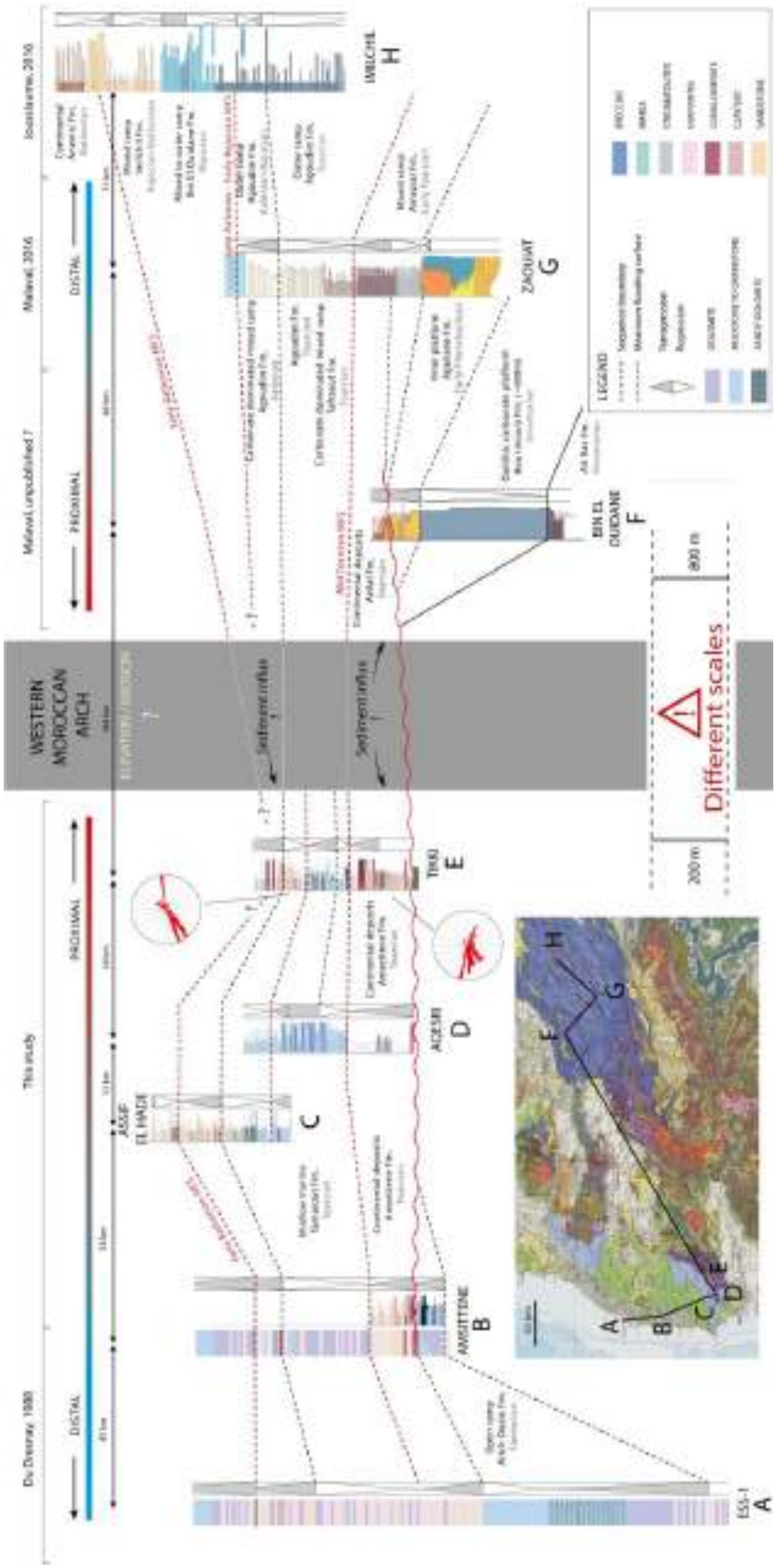

Figure 4.15: Western High Atlas to Central High Atlas time correlations for the Lower Jurassic. (See references for the logs therein).

Thus, the extensive Toarcian erosion, down to the Sinemurian and CAMP basalts, indicates a major regression probably linked to a local uplift of the Meseta between the Western and Central High Atlas.

4.10.2 Siliciclastic input: implications and potential provenance

The current study has outlined significant siliciclastic input to the EAB throughout the Lower and Middle Jurassic. Multiple source areas are possible. The Anti-Atlas and the Western Meseta (Zaer Massif, Rehamna, Jebilet and Massif Ancien de Marrakech (MAM)) are the closest potential sediment sources (Ghorbal et al., 2008; Charton et al., 2018). Three source area (Anti-Atlas, MAM and Jebilets) were exhuming during the Early Jurassic (Charton, 2018), while the North of the Meseta (Zaer Massif and Rehamna) were subsiding (Ghorbal et al., 2008). Lower and Middle Jurassic deposits in the Tikki section have paleocurrents towards the W-SW, which also indicates a source to the NE, such as the Jebilet or more likely the MAM. This is consistent with erosion of the West Moroccan Arch during the Toarcian discussed earlier.

Very strong exhumation rates in the Anti-Atlas during this period (Charton, 2018 after Gouiza et al., 2017) suggest that paleo-relief might have been created, leading to large drainage systems coming from the Anti-Atlas towards the north / northwest that could also possibly have delivered sediments to the EAB. There is however little available evidence to support the re-routing of sediments towards the west. Basalt clasts are found in the Amsittène Formation that may come from the basin itself at that time. It is however worth noting that the Anti-Atlas also records CAMP activity, with two 100 km long dykes (e.g. Touil et al., 2008; Silva et al., 2010), that were potentially sourcing basalts to the surface. Thus, the large Variscan belt remains a potential source of sediments for the EAB Toarcian braided river system.

In addition to regional sources of Amsittène Formation siliciclastics, the alluvial fan deposits near Tikki are potentially linked to activity of an ENE-WSW fault. This fault was subsequently reactivated during the Atlas orogeny and is now a reverse fault, but was probably a normal fault during the Toarcian, with an uplifted northern block. The well-rounded nature of the Toarcian conglomerates in the hanging wall is unlikely to be due to long distance transport, and our preferred interpretation is reworking of older Triassic conglomerates from the footwall. (Upper Carnian T6 unit; Mader et al., 2011). The Tamarout Foramation contains a siliciclastic fraction of silt to fine sand delivered into the carbonate-dominated system. Mixed siliciclastic sediments frequently occur in arid settings with a low input of sediments to the nearshore (Belperio et al., 1988; Zonneveld et al., 2001). Siliciclastic sediments in the peritidal environment can be derived from three mechanisms: aeolian input, fluvial input or longshore drift (Zonneveld et al., 2001). Sabkha and salina facies are recognised, indicative of an arid or semi-arid climate, under which both ephemeral streams and eolian processes operate. Development of coastal aeolian dunes are common in arid or semi-arid climate (Semeniuk, 1996) and aeolian sand could have been brought into the system by coastal winds, later reworked by the tidal currents and mixed with the carbonates deposits. Only the marine part of the system is observed at outcrop and no aeolian or fluvial deposits have been recorded in the field.

During the Middle Jurassic, the SE towards NE proximal-distal orientation of the system indicates more clearly the Anti-Atlas as a potential source. This period records siliciclastic sediments deposition in the south of the EAB only, while the north of the basin is too far from the source and still dominated by carbonate deposits. The important exhumation rates in the Anti-Atlas during this period (Charton, 2018) would explain the creation of a palaeorelief and the development of an extensive drainage system. The particularly high exhumation rates of the central part of the Anti-Atlas compared to the western part (Gouiza et al., 2017) could also explain the western direction of the flow.

4.10.3 Middle Jurassic regression

The evolution from the Tamarout to the Ameskhoud formations is transitional from marine carbonate to continental siliciclastics. The base of the Ameskhoud Formation is made of intertidal red siliciclastic deposits, dominated by marls and siltstones interbedded with thick sandstones with cross-stratification and stacked truncated wave ripples. This intertidal facies indicates a change in the composition of the sediments but no relative sea level variation compared to Unit T3 of the Tamarout Formation. The transition from carbonates to siliciclastics could be disconnected from eustatic change, but related to the exhumation of a regional source terrain. This evidence could support the hypothesis formulated by Stets (1992) and the later thermogeochronoly (Ghorbal et al., 2008; Saddiqi et al., 2009, Charton, 2018) that the West Moroccan Arch ("Terre des Almohades") was uplifting during this period. The MAM would be the principal potential source for the Middle Jurassic siliciclastic deposits in the EAB.

The regression observable in the EAB during the Middle Jurassic does not correlate with the global eustatic curve (Snedden and Liu, 2011; Haq, 2018). By contrast, the Middle Jurassic deposits are transgressive in the Central High Atlas (Teixell et al., 2017; Malaval, 2017; Joussiaume, 2017) to the East, and in the Aaiun-Tarfaya Basin, to the South (Arantegui, 2018).

This suggests the influx of siliciclastics into the EAB had to be significant, to overcome the global sea level rise. The regression from intertidal to continental environment during the Middle Jurassic is interpreted to be related to the tectonic exhumation of the hinterland that resulted in increased volume of sediments shed into the EAB, which triggered a forced regression.

4.11 Conclusions

The Early and Middle Jurassic succession of the EAB is a key sedimentary unit which records the sedimentary response along the passive margin during the opening of the Atlantic Ocean. The succession can also be compared with the equivalent units in the Central High Atlas, and as such elucidate local from global and tectonic from eustatic controls. By identifying the major stratigraphic units of the Western High Atlas and comparing them to the Central High Atlas within a broader framework, this study shows:

1. The refined sedimentological and stratigraphic study of the Lower Jurassic succession allowed identifying local and regional trends in the deposition of the Amsittène and Tamarout Formations and their characteristics and

variations across the basin. The Amsittène Formation has been identified as a strongly erosive fluvial unit deposited along all the basin and cutting down to the CAMP basalts and the Trias. The Tamarout Formation records a transgression in the upper part of the Toarcian which led to deposition of intertidal to supratidal carbonates dominated by oolitic limestones, dolomites and evaporites or dissolution breccias. Three units have been distinguished and linked to variations of the depositional environment, the lateral extent of this carbonate unit and its facies variations on the basin borders have been constrained.

2. Lateral facies variations from fluvial deposits to shallow marine carbonates of the Middle Jurassic across the basin have been highlighted and a SE-NW proximal-distal trend has been identified and linked to the uplift of the Anti-Atlas.
3. The bulk of the Middle Jurassic is marked by a major regression, which led to the establishment of a continental environment in the south of the basin. The input of siliciclastic sediments into the EAB also points towards a tectonic control, with rejuvenation of the hinterland source areas.
4. The oldest Jurassic deposits observed at outcrops in the EAB are open marine carbonates of the Arich Ouzla Formation. They were uplifted and tilted then partly eroded by continental deposits during the Toarcian. This unconformity can be correlated across the EAB and to the CHA (Malaval 2017; Joussiaume 2017) and records a phase of tectonic inversion of the margin.
5. Throughout the Lower and Middle Jurassic, the Jebilets, the MAM and the Anti-Atlas were uplifted and eroded (Ghorbal et al., 2008; Charton, 2018). The Jebilets and the MAM being the closer sources from the EAB, they are considered to be the main siliciclastic provenance area for the Toarcian continental and marine sediments. The Anti-Atlas was also exposed and extensively eroded during that period of time and was probably the main source of the Middle Jurassic siliciclastics and its source potential could also be tested for the Toarcian.

6. Between the Lower Pliensbachian and the Toarcian, the EAB has been subject to vertical movements leading to the emersion and tilting of the Arich Ouzla and Aganane formations. The elements presented here are supporting the recent developments in geothermochronology indicating the uplifts of the Meseta and Anti-Atlas as evidences that the Atlantic Passive Margin was actually experiencing vertical movements during the rift and post-rift.
7. This study highlighted numbers of key events and brought to light that pulses of siliciclastics triggering regressions on the passive margin can be linked to tectonic activity.
8. Controls of the accommodation on a passive margin can then be correlated with movements of the hinterland and do not depend only of the global eustasy.

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Chapter 5: The development and evolution of an Atlantic margin carbonate ramp during the Callovian to middle Oxfordian: Agadir-Essaouira Basin, Morocco.

Abstract

The Middle to Upper Jurassic is dominated by carbonate ramp deposits followed by coral buildups on both sides of the Central Atlantic, in the Essaouira-Agadir Basin (EAB) of Morocco and on the conjugate margin Scotian Basin of Nova Scotia. This study, carried-out in the Essaouira-Agadir Basin (EAB) on the Moroccan Atlantic Margin examines the stratigraphy and sedimentology of the Callovian-Oxfordian interval in order to get a better understanding of the mechanisms controlling their deposition and lateral variability. New data have revised the ammonite biostratigraphy, improving constraints on the ages of the different units and identified a potential hiatus. This allows improved regional correlation and quantification of sedimentation rates and identifies key surfaces that can be correlated across the basin to allow generation of depositional environment .models.

The Callovian is marked by a transgression in the EAB, where open marine deposits of the Ouanamane Formation follow deposition of the peritidal and continental deposits of the Ameskhoud Formation. The Ouanamane Fm. can be subdivided in three different members, bounded by major hard-ground surfaces that can be identified basinwide. The first member (Iggui-n'Tarhazout Oolite) is dominated by oolitic upper ramp deposits and could constitute a good reservoir. The second member (*Somalirhynchia* Limestone Mb.) is dominated by brachiopod-rich middle to outer ramp deposits with occasional small coral buildups. It provided the majority of the ammonites collected. The third Marl and Shale

Mb. consist of a thick package of outer ramp marls. Brachiopods are abundant in its lower part and scarce ammonites occur throughout.

Biostratigraphic evaluation of new data collected in the field and existing collections has improved the age constraint of the lithostratigraphic framework. The integrated study of the ammonite and brachiopod faunas allows the lower and middle members of the Ouanamane Fm. to be definitively dated as early to middle Callovian age, revising the previous proposed Late Bathonian age for the lower part of the Ouanamane Fm. The lower part of the upper member is dated as late middle Callovian age and the Callovian-Oxfordian transition is placed within the upper part of the third member and is affected by condensation and/or hiatuses. The uppermost part of the formation is of early middle Oxfordian age and a middle Oxfordian age is inferred for the onset of the buildups observed in the Lalla Oujja Fm. across the EAB.

The lateral variability of each unit is documented at the scale of the basin in order to identify the depositional environments and their evolution through time, attesting to a general transgression during the Callovian in the EAB.

5.1 Introduction

During the Middle to Upper Jurassic, carbonate platforms developed along the Central Atlantic Margin. The Callovian-Oxfordian interval has been studied on both sides of the conjugate margins (Davison, 2005; Weissenberger et al., 2006; Martin-Garin et al., 2007; Olivier et al., 2012; Eliuk, 2016). The main reservoir potential is in Oxfordian reefal buildups, which are a proven reservoir for the Panuke gas field in the Scotian Basin (Weissenberger et al., 2006), but this interval remains a challenging target, with many dry wells due to the difficulty in predicting reservoir quality. In order to improve the success rates and reduce the risk of petroleum exploration along the margin, it is crucial to get a better understanding of the system as a whole. Therefore, better constraints on the ages of the different units, the depositional controls and lateral facies variations need to be defined.

This study focuses on the Callovian to middle Oxfordian of the Essaouira-Agadir Basin (EAB) that contains a potential reservoir unit, e.g. the Iggui-n'Tarhazout Oolite Mb. of the Ouanamane Fm. This facies is also present on the Scotian Margin inner shelf deposits of the Abenaki Formation (Weissenberger et al., 2006; Eliuk, 2016). A revised biostratigraphy and sequence stratigraphic study of the Scotian Margin identified the top-Callovian maximum flooding surface as one of the main seismically mappable sequence-stratigraphic events (Weston et al., 2012). Constraining the age boundaries of this major event is important to extrapolate across the conjugate margins.

The Atlantic margin of Morocco offers the best outcrops to improve the understanding of the age constraints and vertical/lateral facies variations of the Callovian-Oxfordian interval. The aim of this study is to: (1) refine the litho- and biostratigraphic framework of the Callovian-Oxfordian successions of the EAB in order to understand the deposition of the carbonate units (carbonate fabric and geometry); (2) identify prospective hiatus or high sedimentation rates; (3) identify major correlateable surfaces across the basin and assess their significance in terms of environment of deposition and stratigraphic organisation; (4) reconstruct the depositional environment of each unit and evaluate their reservoir potential; (5) discuss the age and synchroneity of the onset of coral bioherms development at the scale of the basin.

5.2 Geological settings

5.2.1 Structural setting

The EAB is the southwestern extension of the Western High Atlas (WHA), location at the intersection of Atlas rifting with the opening of the Central Atlantic Ocean. The general N-S to NNE-SSW syn-rift structures of the Atlantic margin are linked to an E-W transfer faults system, the orientation suggesting some inheritance and reworking of older Hercynian faults (Medina, 1988, 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. The structure of these anticlines is believed to be linked to salt tectonics triggered by transtentional and transpressional movement in the Essaouira segment and fault reactivation during the inversion in the Agadir segment (Piqué et al., 1998). The anticlines are deeply eroded and expose the entire Jurassic stratigraphy (Figure 5.1).

5.2.2 Upper Triassic and Lower Jurassic

The initial fill of the Triassic extensional half-grabens are dominated by continental deposits (Courel et al., 2003; Mader et al., 2011; Leleu et al., 2016). The Triassic rift was followed by a phase of thermal subsidence and creation of a sag basin onshore during the Lower Lias (Hafid, 2000; Hafid et al., 2006; Frizon de Lamotte et al., 2008). Extensive salt deposition occurred in many of the syn-rift fault-controlled grabens, with the maximum extension of salt deposition corresponds to the latest syn-rift stage (Tari et al., 2003; 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 (Fiechtner et al., 1992; Marzoli et al., 1999; Knight at al., 2004; Verati et al., 2007). The beginning of the Early Jurassic marks the break-up of the Central Atlantic and the beginning of the post-rift stage dominated by the formation of a widespread carbonate platform (Morabet et al., 1998; Hafid et al., 2000; Le Roy and Piqué, 2001; Laville et al., 2004; Tari and Jabour, 2013).

The stratigraphic evolution of the EAB during the Lias is dominated by carbonate deposits. The Lower to Middle Lias is composed of open marine deposits, rarely preserved onshore (Roch, 1930; Duffaud, 1960; Du Dresnay, 1988). They are followed by continental siliciclastic deposits deepening upward to a thick succession of peritidal carbonates during the Toarcian (Ambroggi, 1963; Adams et al., 1980; Peybernès et al., 1987; Du Dresnay, 1988; Bouaouda, 2004). During most of the Dogger, the onshore southern part of the EAB was dominated by up to 300m of siliciclastic shallow marine to fluvial deposits while the northern part (Jbel Amsittène) and offshore deposits are dominated by shallow marine dolomitic deposits (Peybernès et al., 1987; Bouaouda, 2004; Chapter 3).

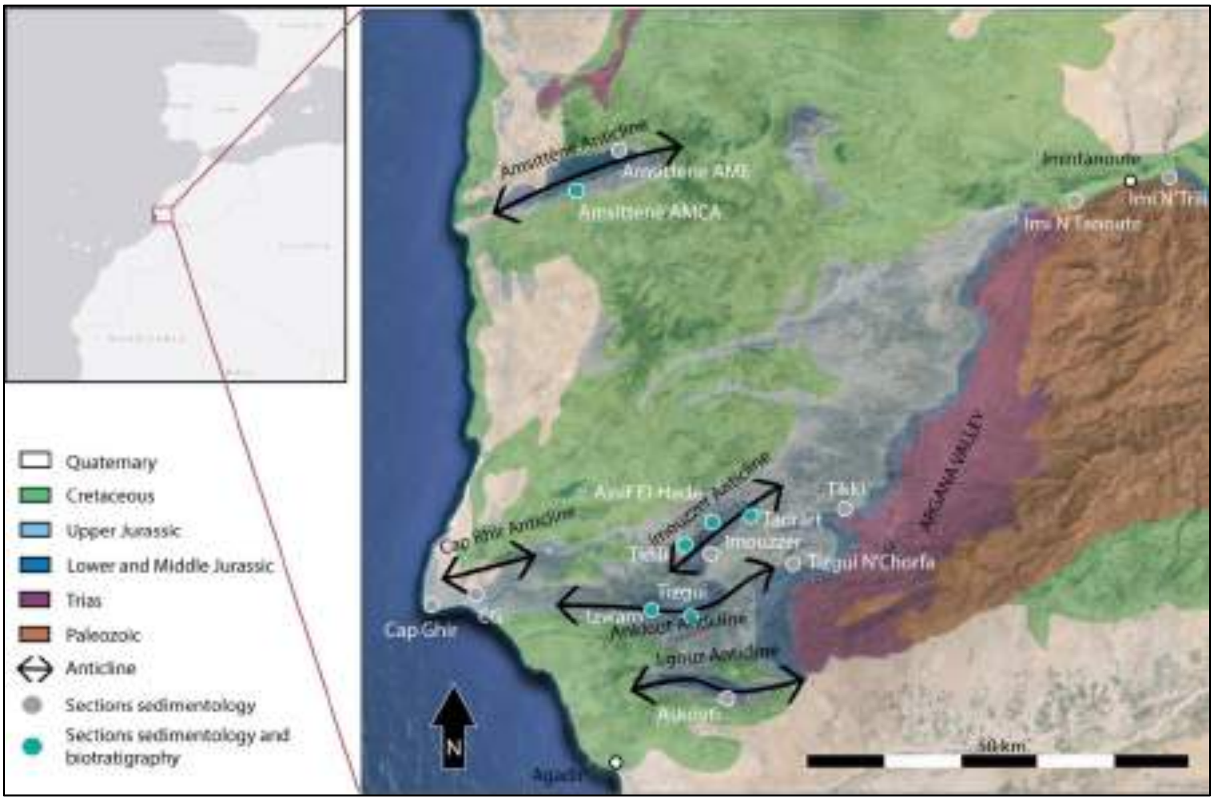

Figure 5.1: Localisation map of the study area and sections logged.
Biostratigraphic relevant sections are indicated in blue.

The Middle Jurassic of the EAB is dominated in the East and South by fluvial deposits, while in the Essaouira sub-basin, dolomitic limestones are present (Chapter 4). These dolomitic limestones are described as intertidal to supratidal ramp complex, evolving from sabkha deposits to oolitic marine limestones followed by oncolitic limestones (Peybernès et al., 1987). In the Agadir sub-basin, fluvial deposits are dominating during the Middle Jurassic and are transitioning to marine deposits towards the end of the Middle Jurassic (Ambroggi, 1963; Adams et al., 1980; Peybernès et al., 1987; Du Dresnay, 1988; Bouaouda, 2004).

5.2.3 The Callovian transgression on both sides of the Central Atlantic Margin

Callovian and more generally Middle Jurassic deposits are dominated by carbonate platform development along the Central Atlantic Margin (Davison, 2005 with references).

In Morroco, Adams (1989, 1990) was the first to address the timing and extent of the Middle to Late Jurassic transgression that affected the western High Atlas. In his opinion, the onset of the transgression took place in the Callovian and was marked by the deposition of oolite shoals in the lower part of the Ouanamane Fm. The transgression reached its maximum extent during the Oxfordian with the deposition of the argillaceous limestones of the upper part of the Ouanamane Fm. and the development of reefal systems in the lower part of the Lalla Oujja Formation. With minor changes, Adam's views, mostly based on the study of the Imouzzer area, were confirmed since by Peybernès et al. (1987) and Bouaouda (2004).

Adams (1989, 1990) also outlined the similarities of the Middle and Late Jurassic of the western High Atlas with the successions of the Scotia Shelf described by Pratt and Jansa (1989). It has been established since that on the conjugate margin of Nova Scotia, the Misaine Member, dominated by shales deposition, records a peak of transgression of Callovian age that can be identified as a seismically mappable maximum flooding surface (Weston et al., 2012, Weissenberger et al., 2006; Eliuk, 2016). The Callovian transgression is therefore a major regional feature on both sides of the Central Atlantic Margin.

5.3 Methods

5.3.1 Facies analysis and sequence stratigraphy

The Middle Jurassic of the EAB has been studied along East-West anticlines and the Argana Valley, which are the only outcropping localities in the basin. 11 sections have been logged, 2 in the Essaouira sub-bassin and 9 in the Agadir sub-basin (Figure 1). Detailed sedimentological analyses were conducted in the field, including facies and processes descriptions and geometry evolution. In addition 50 petrographic thin-sections were studied in detail for microfacies characterization. 2 local and 1 regional stratigraphic transects were built based on the recognitions of depositional sequences and biostratigraphy.

5.3.2 Biostratigraphy

In order to build a high resolution biostratigraphic framework, two field trips of a total of 15 days focused on bed by bed collection of ammonites, brachiopods and echinoids at 6 field sections (Figure 1). The new collections clearly show that the limited number of ammonites reported in the literature reflects the scarcity of this fossil group at the scale of the basin. Despite our extensive efforts only 21 identifiable specimens were collected.

Therefore, an extensive search for biostratigraphic 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 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 biostratigraphic scales established by the French Jurassic Group (Cariou and Hantzpergue Coords., 1997). Since ammonites are the prime markers for Jurassic biostratigraphy, the biostratigraphic interpretation of brachiopods, echinoids and benthic foraminiferas assemblages can only be considered as reliable when direct calibration to ammonite scales is available. The limitations of those groups were discussed at length by Bassoulet (1997),

This study	LITHOLOGIES	Ambroggi (1963)	Duffaud, et al., (1966)	Adams et al., (1980)	Peybernes et al., (1987) ESSAOUIRA BASIN	Peybernes et al., (1987) AGADIR BASIN	Bouaouda (2007)
IMOUZZER FORMATION		KIMMERIDGIAN	MARNES D'IMOUZZER	IMOUZZER FORMATION			IMOUZZER FORMATION
IGGUI EL BEHAR FORMATION		RAURACIEN - SEQUANIEN	CALCAIRES DE HADID	IGGUI EL BEHAR FORMATION	HADID FORMATION	IGGUI EL BEHAR FORMATION	IGGUI EL BEHAR FORMATION
LALLA OUJJA FORMATION		ARGOVIAN	RESERVOIR SIDI RHALEM	LALLA OUJJA FORMATION	SIDI RHALEM FORMATION	LALLA OUJJA FORMATION	TIDILI FORMATION
OUANAMANE FORMATION		OXFORDIEN CALLOVIEN	MARNES D'ANKLOUT CALCAIRES D'ANKLOUT	OUANAMANE FORMATION	ID BOU ADDI FORMATION	OUANAMANE FORMATION	OUANAMANE FORMATION
AMESKHOUD FORMATION		DOGGER	DOLOMIES DE L'AMSITTENE	AMESKHOUD FORMATION	ID OU MOULID FORMATION	AMESKHOUD FORMATION	AMESKHOUD FORMATION

Figure 5.2: Middle to Upper Jurassic lithostratigraphic subdivisions of the Essaouira-Agadir Basin.

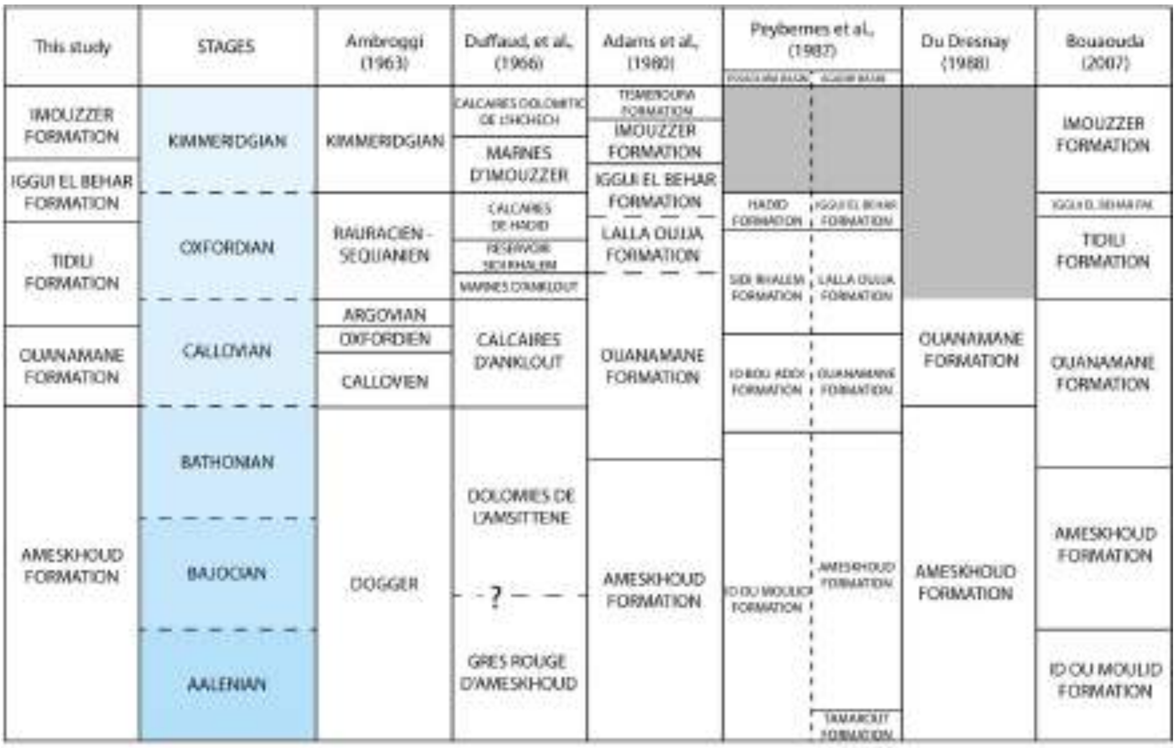

Figure 5.3: Middle to Upper Jurassic formations ages interpreted by the different authors in the Essaouira-Agadir Basin.Stratigraphy

5.3.3 Lithostratigraphic subdivisions

The lithostratigraphic framework retained herein is largely inspired by the scheme established by Adams (1979) and formalised by Adams et al. (1980). It was introduced to clarify the stratigraphic nomenclature introduced by Duffaud et al. (1966) that was not accurately defined, and therefore difficult to recognize with certainty. Subsequent additions by Peybernès et al. (1987) and Bouadoua (2007) are shown on Figure 5.2.

The Ouanamane Fm. was originally subdivided in four members by Adam et al. (1980). The lowermost member (Transition Dolomite Mb.) that marks the passage between the Ameskhoud and Ouanamane formations was only recognised 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 (Figure 3). From bottom to top the succession is as follows:

- Unit 1 (Iggui-n'Tarhazout Oolite Mb.) is dominated by oolitic grainstone and bioclastic packstone. Its base is marked by a firm ground that capes the Transition Dolomite Mb., or rests 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 and Shale Mb.) is dominated by thick marls with mudstones and floatstones.

These three units are homogeneous in the center of the EAB. They are associated with iron rich and very bioturbated firm-grounds and hard-grounds. The main hard-grounds coincide with the boundaries of the units and are associated to a major change in sedimentation (Figure 5.4).

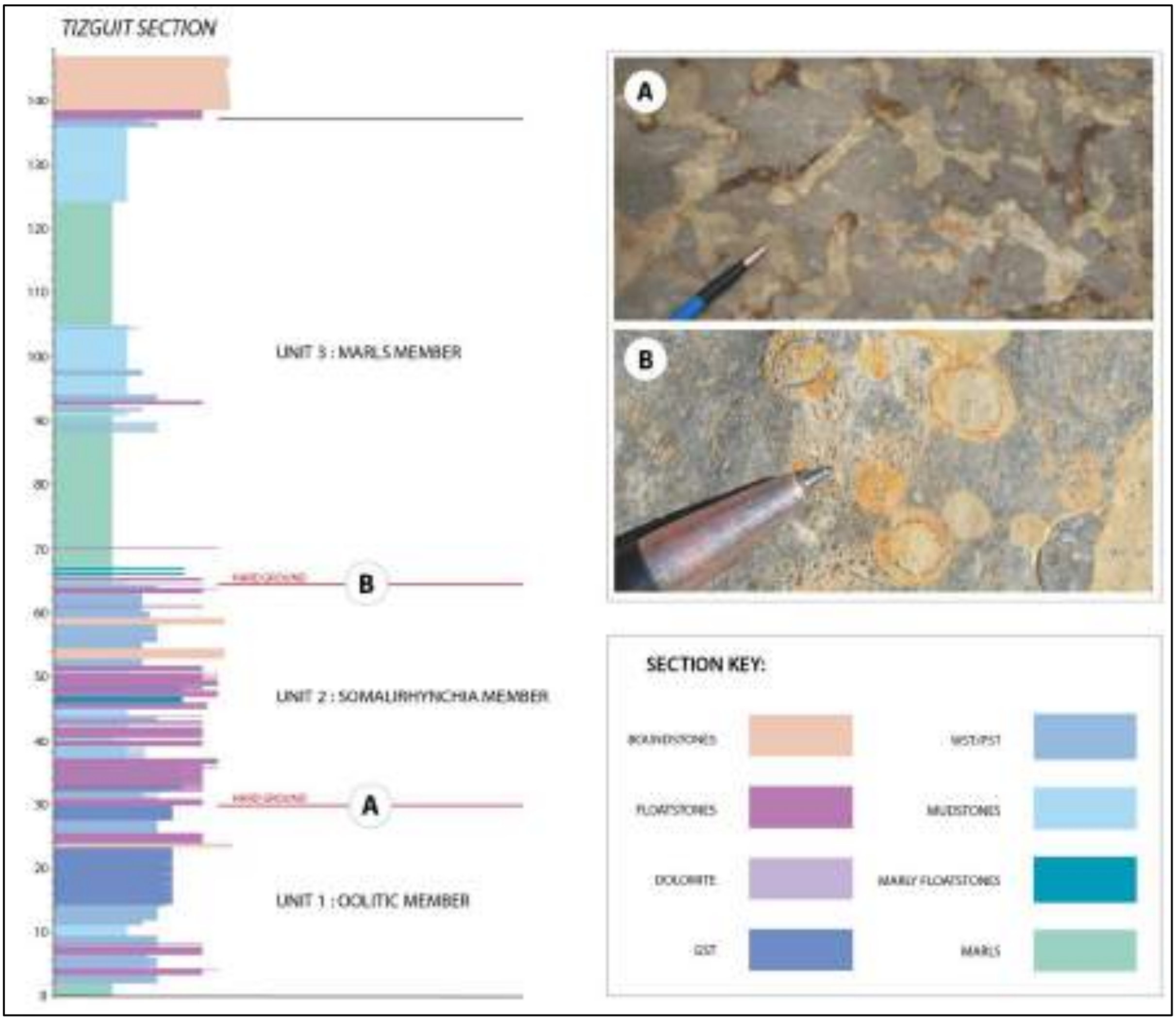

Figure 5.4: Example of simplified section of the Ouanamane Fm.
(Tizguit Section) and examples of hardgrounds bounding the different units (A and B).

5.3.4 Biostratigraphic framework

Over the past 25 years, Bouaouda and collaborators provided 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 with references). The data set available used and presented by these authors is largely heterogeneous.

The ammonite record was interpreted from a limited number of specimens listed in the literature (Gentil and Lemoine, 1906; Roch, 1930, 1950; Ambroggi, 1963) most of which were never formally documented and illustrated. The systematics of this historical material is in need of revision. Besides, brachiopods were illustrated and described by Gentil and Lemoine (1906), but the only modern study was performed by Yves Alméras in the late 1980's. A 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 one of those uncertainties is given by the single Jurassic ammonite ever illustrated from the EAB, e.g. *Perisphictes chavattensis* de Loriol *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 UPMC collections (Paris). Besides, the original data certainly do not allow the position of the specimen in the stratigraphic succession to be pinpointed even so a level close to the base of the Lalla Oujja Fm. was suggested by Ambroggi (1963).

The confusion derived from the stratigraphic position and age assignment of this specimen culminated with Bouaouda et al. (2004). Those authors report the three identifications of the same ammonite from two different stratigraphic levels, as if they were three different specimens. Moreover, the biostratigraphic interpretations of Bouaouda 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* 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 at 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 & Bird, 1828) and *Perisphinctes (Kranaosphinctes) pickeringius* (Young and Bird, 1822).

Based on the occurence of *E. (E.) phoenicium* alone, a middle Oxfordian age (Transversarium Zone) was accepted by Olivier et al. (2012). Besides, the same authors attributed a late 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 late Oxfordian, while the taxonomic content and stratigraphic distribution of *Orthosphinctes* s. str. still remains unclear (Malinowska, 1972; Brochwicz-Lewiński, 1975, 1980; Atrops, 1982; Atrops and Meléndez, 1994; Schlegelmilch, 1994; Meléndez et al., 2006).

To avoid such confusion, we concentrated our efforts on new bed by bed collections combined with rock sampling for micropalaeontological studies. We reinvestigated the material collected by Ambroggi (1963) from the Anklout anticline (Izwarn area) 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.

5.3.5 Ammonites

The material studied mainly originates from the Jbel Amsittène and Izwarn anticlines. Both areas proved to be more ammonite rich than the other localities investigated. The main 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 on Figure 5.5 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, 1984 (TZG-P-10B). This assemblage indicates a late early Callovian age (Gracilis Zone, Patina Subzone).

At Jbel Amsittène, the upper part of the *Somalirhynchia* Limestone Mb. (Unit 2) is fairly rich and includes: *Reinecke̦ia (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.

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* Gil 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, 1984 (MEM-AM-1427) and *Rehmannia (Loczyceras)* sp. (MEM-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 a late 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 early late 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 early late 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 latest Callovian (Lamberti Zone) (Pellenard et al., 2014);

- 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 Glowniak (2002). Those forms indicate the late early to early middle Oxfordian (see discussion in Glowniak, 2002).

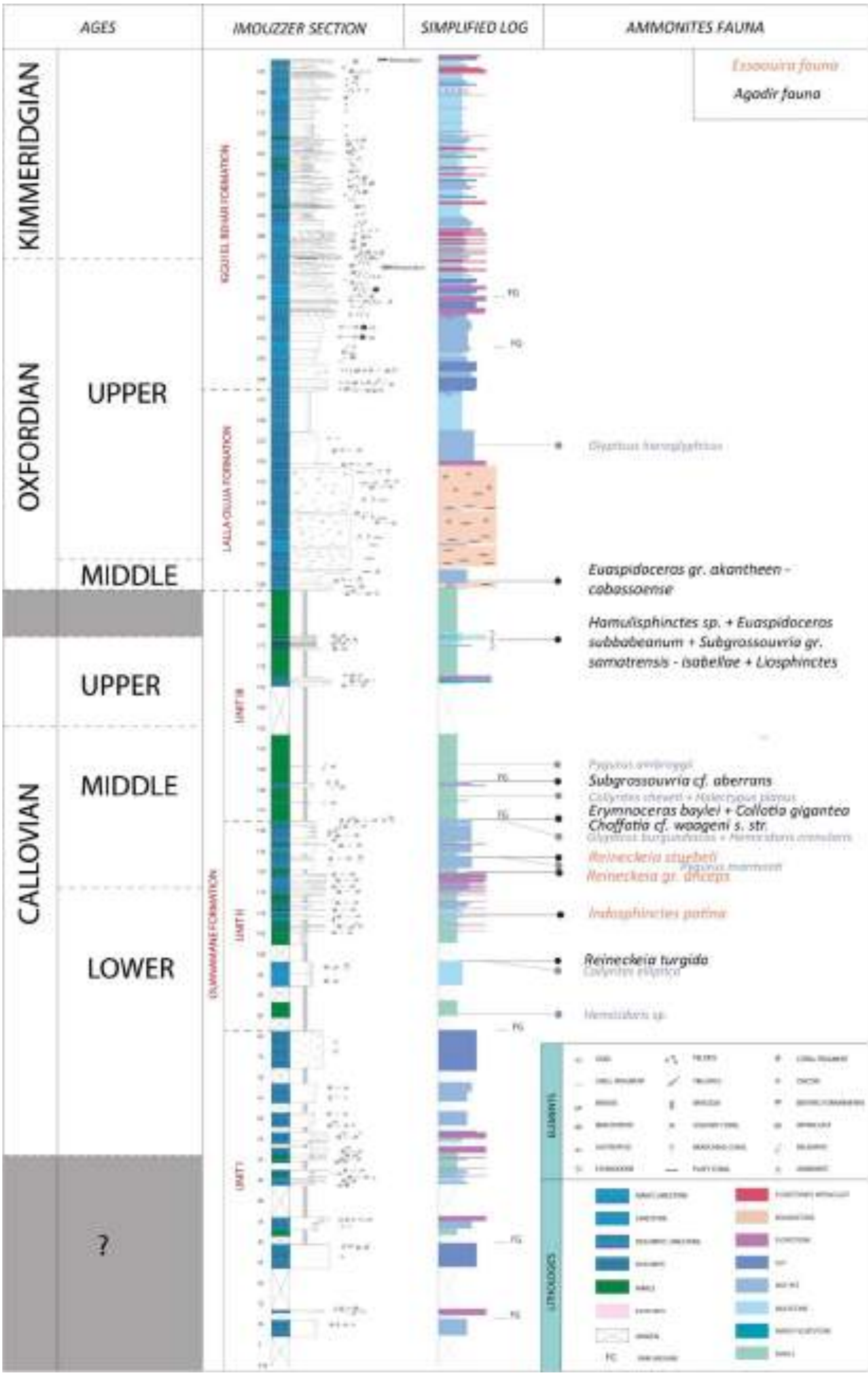

Figure 5.5: Callovian and Oxfordian ammonite distribution (EAB, Morocco)

The specimen originally identified as *Aspidoceras phoenicium* in Ambroggi (1963) is too worn to allow identification at the genus or species level. Even so the number of specimens is limited, three of them point to a late Callovian age (Athleta and Lamberti zones) while the fourth one indicates a late early to early 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 in the 3 meters 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 the 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). Even so, 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).

5.3.5.1 Brachiopods

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. Since this early work, brief discussion about the occurrence of *Somalirhynchia* and *Bihenithyris* by Ager (1974,a), Adams (1979) and Adams et al. (1980) were mostly concerned by the palaeobiogeographic affinities with the faunas that characterise the Jurassic Ethiopian Province (Ager, 1971, 1973; Hallam, 1977; Manceñido, 2002).

Based on 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; Peybernès et al. (1987) suggested that the Ouanamane Fm. is of early to late Callovian age. The same authors outlined that *Kutchithyris acutiplicata* (Kitchin, 1900), *K. aurata* (Kitchin, 1900) and *K. planiconvexa* (Kitchin, 1900) may indicate a late Bathonian age for the lower part of the formation. 11 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 taxa, e.g. *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, Figure 5.5).

For example, the association of *Kutchithyris acutiplicata*, *K. aurata* [a junior synonym of *K. propinqua* (Kitchin, 1900)] and *K. planiconvexa* indicates a Middle to Late 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.

To understand the origin of such discrepancies, a detailed revision of the brachiopod faunas from the Ouanamane Fm., based on new bed by bed collections calibratingtowards our ammonite findings, is currently in progress. The taxonomic study of the faunas collected is still to be completed but some important conclusions can already be reached based on the stratigraphic distribution of *A. bihinensis* morph *superstes*, *B. barringtoni, B. weiri, S. africana, S. orbignyana*, and *K. indica* in the EAB.

A. bihinensis morph *superstes* was recently revised by Alméras et al. (2010). In Saudi Arabia, this taxa first occurs in the late 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 so there is no solid evidence for this age.

Based on well constrained collections made in Israel, Algeria, Tunisia, Saudi Arabia, the early middle to early late Callovian age of *B. barringtoni* is now clearly established (Alméras et al. 2010, with references). The same is true for *B. weiri* that is known to range from the late middle Callovian into the late Callovian (Alméras et al., 2010, p. 69-70, with references). 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, with references), even so according to Feldman et al. (1991) it already occurs in the Lower Bajocian of Egypt (Gebel El-Maghara). 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 late 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 be also be noted that even if the Saudi Arabia occurrence is clearly dated to 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 late early Callovian, and that the species ranges throughout the middle Callovian.

S. orbignyana is a useful marker that ranges from the latest early Callovian (Gracilis Zone, Patina Subzone) to the early late Callovian (Athleta Zone) in Western Europe (France, UK) (Laurin, 1984, Alméras and Cougnon, 2013, with references).

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

The *Kutchithyris* species that were reported by Peybernès et al. (1987) from the lower part of the *Somalirhynchia* Limestone Mb. indicates a Late 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 Morrocan 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).

5.3.5.2 Benthic foraminiferas

The limitations of Jurassic benthic foraminiferas biostratigraphy have been discussed at length by Bassoulet (1996) as an introduction to the calibration towards ammonite zonal scales of the evolutive events recognised by Septfontaine et al. (1991).

Based on the co-occurence of *Praekurnubia crusei* Redmond, 1964; *Pseudoeggerella elongata* Septfontaine, 1988; *Andersenolina palastiniensis* (Henson, 1948); *Andersenolina minuta* (Derin and Reiss, 1966); *Pseudocyclammina maynci* Hottinger, 1967 and *Archaeosepta platierensis* Wernli, 1970; a late Bathonian age was assigned to the base of the Ouanamane Fm. by Bouaouda et al. (2004, p. 17). These authors also specify that *A. palastiniensis* and *A. platierensis* are strictly restricted to the western, e.g. the deeper, part of the EAB. The late Bathonian age of this association is far from being established since:

- Even so *Praekurnubia crusei* is most often assumed to range from the Upper Bathonian into the middle, and potentially, upper Callovian, this view should be tempered (Bassoulet, 1997). In central Saudi Arabia, *P. crusei* first occurs in the Hisyan Member of the Upper Dhruma Formation of middle Callovian age (Coronatum Zone) (Powers *et al.,* 1966; Enay *et al.,* 1987; Enay and Mangold, 2004). The genus *Praekurnubia* was also reported from the Lower Callovian (Bullatus Zone) of Burgundy (Thierry *in* Septfontaine *et* *al.,* 1991) but this occurrence remains to be fully documented. Similarly Upper Bathonian occurrences in Turkey are still to be confirmed (see discussion in Bassoulet, 1997). In recent literature (Peybernès et al., 1999), the first occurrence of the species is reported from close to the Bathonian-Callovian boundary.

- *Pseudoeggerella elongata* was originally described from the upper part of the middle Dhruma Formation of Central Saudi Arabia, for which a late Bathonian age is retained with doubt by Enay et al. (1987). Banner et al. (1991) transferred the species to the genus *Riyadhella* and considered it as a junior subjective synonym of *Riyadhella regularis* Redmond, 1965, a view that has been accepted since by Tasli et al. (2008). As understood by those authors the species ranges from the Bajocian into the Oxfordian.

- *Andersenolina palastiniensis* is most often treated as a *Kurnubia* in the literature. The known stratigraphic distribution of the species was discussed and summarised by Bassoulet (1997). At this type-locality (Kurnub anticline, Israel), it is known to range from the middle Callovian Coronatum Zone to the Upper Callovian Athleta Zone du Callovien moyen. Occurrence in the Upper Bathonian is still to be established. Closely allied forms (*Kurnubia variabilis* Redmond, 1964) were described from the Atash Mb. of the Upper Dhruma Fm. in Central Saudi Arabia for which an early Callovian age has been proposed by Almeras et al. (2010).

- *Pseudocyclammina maynci* is a poorly calibrated classical Middle to Late Jurassic species. Even so it was reported from the Bajocian, it likely does not appear until at a level close to the Bathonian-Callovian boundary (Bassoulet, 1997).

- *Archaeosepta platierensis* is known from the northern margin of the Tethys (Wernli, 1970; Septfontaine, 1978, 1981; Chiocchini and Mancinelli, 1996; Haas et al., 2013). A late Bajocian to late Bathonian age was retained by Bassoulet (1996, p. 296, table LII) based on the original findings of Wernli (1970). It should be noted that Septfontaine (1978, p. 2) pointed out that the precise range of the species is not established since it can be confused with *Protopeneroplis striata* Weynschenk, 1950. The illustration of the Moroccan specimens is present in the unpublished PhD thesis of Bouaouda (2007) that could not be accessed in the course of this study. Therefore, comparison with type-material could not be assessed. The species was not observed in the thin sections studied.

5.3.6 Biostratigraphic interpretations

Ammonite faunas clearly indicate a late early Callovian (Gracilis Zone, Patina Subzone) to middle Callovian (Coronatum Zone, Baylei Subzone) age for the middle and upper part of the *Somalirhynchia* Limestone Mb (Figure 5.6, A and B). This finding dismisses the early Callovian age accepted by Bouaouda et al. (2004, Figure 5.5) for this part of the succession.

Somalirhynchia africana already occurs 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 a late Bathonian age for this part of the Ouanamane Fm. As a consequence, we retain an early Callovian age for the lower part of the Ouanamane Fm. (Unit 1). Even so a slightly older age, e.g. Latest Bathonian cannot be excluded for the base of the formation.

The lower part of the Marls and Shale Mb. (Unit 3) is unequivocally dated from the late middle Callovian (upper part of the Coronatum Zone) by ammonites (Figure 5.6, 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 and significance of this bioevent remains to be tested.

Based on revision of Ambroggi’s collection Late Callovian ammonites (Athleta and Lamberti Zone) are identified from a 3m thick interval in the upper part of the Ouanamane Fm (Figure 5.6, D). From the same beds a single late early to early middle Oxfordian ammonite is reported. Even if this specimen was found loose, this fauna 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 the coral buildups that marks the base of the Lall Oujja Fm. falls within the middle Oxfordian (Figure 5.6, E).

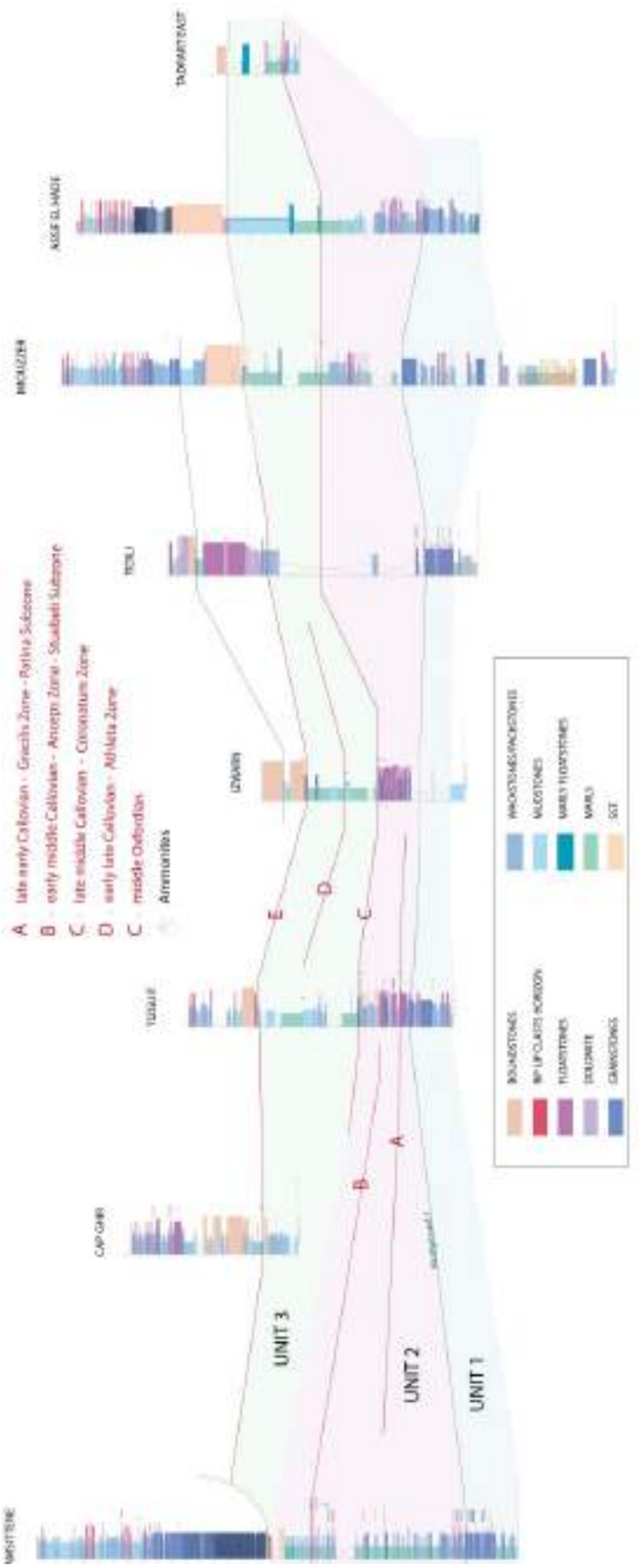

Figure 5.6: Biostratigraphic correlations across the Essaouira-Agadir Basin. Not to scale.

5.4 Lithofacies interpretations

The facies interpretation is a summary of different locations in the centre of the basin (Assif El Hade, Imouzzer, Tidili, Tizgui, and Izwarn). These locations present very similar facies associations and general trends can be extrapolated. The description of the different facies is synthetize in Figure 5.7.

5.4.1 Fluvio-marine transition

FO0: MF1, MF2 and MF3

The passage from the fluvial Ameskhoud Fm. to the marine Ouanamane Fm. is transitional. It is marked by red mudstone and dolomitic sandstone horizons passing up to mudstone and sandy dolomite (Figure 5.7, A). The marine or continental origin of these carbonates is subject to discussion as no definitive evidence has been found. The presence of bivalves and gastropods can be linked either to restricted marine conditions with little species diversity, or to fresh or brackish water in coastal lakes (Dodson et al., 1980; Hudson et al., 1995).

It is followed by a limestone with abundant benthic foraminifera, coated in micrite and associated with rounded echinoderm fragments, which indicates marine deposition. This unit is overlain by 30m of mudstone, siltstone and channelized sandstone. Marls and dolomitic sandstones succeed to this interval with a firmground located on top of a sandy dolomite bed marking the beginning of the Ouanamane Fm. This last sandy dolomite below the main carbonate succession was defined as the Transition Member by Adams et al. (1980). It follows the first marine unit and is part of a general transgressive trend with the establishment of fully marine conditions across the basin.

5.4.2 Unit 1: Oolitic Member

FO1: MF2, MF3, MF4, MF4d, MF5 and MF6

Unit 1 is dominated by oolitic GST (Figure 5.7, B) and oolitic and bioclastic PST, which are partially or totally dolomitised (Figure 5.7, C) and minor marl horizons, defined as facies association FO1. The dominant grains observed in this unit are mainly pseudopeloids, micritic and tangential ooids and grain aggregates (Table 1). Large (>400 µm) micritized tangential ooids with poorly recognisable tangential fabric and various nuclei, comprising benthic foraminifers, echinoderm spines and shell fragments, dominate. Some horizons of oolitic and peloidal PST and GST also contain larger elements such as bivalves or brachiopod shells and aggregates. The identification of thick micrite coatings and moderate to good sorting of the ooids indicates a high energy environment (Rao and Carozzi, 1971; Simone, 1980; Strasser, 1986). The aggregates, mostly composed of ooids bound together (grapestones), can indicate a change to lower water energy (Strasser, 1986; Steinhoff and Strohmenger, 1996). Their association with peloids and the low diversity of skeletal grains is a common feature in shallow-marine platforms (Flügel, 2010). The bioclastic and peloidal PST feature superficial ooids or non-coated grains, associated with some unbroken echinoderms and bivalves that often accumulate in muddy environments (micrite matrix) indicative of lower energy conditions. They are associated in some locations with horizons of gastropod-rich floatstones with minor unbroken *Trichites* bivalves at the base of the beds.

Porous dolomite bodies are also very common in this unit. The crystalline dolomite almost always contains relics of large (>300 µm) ooids or peloids (Figure 5.7, C). The restricted lateral extent of some of these dolomitic bodies is a good indicator for deposition within shoals or tidal bars. Most of this unit consists of extensive sharp-based mostly ungraded oolitic sand bodies with rare hummocky cross-stratification (Tikki location). Associated highly bioturbated (*Thalassinoides*) beds suggest oxygenated bottom conditions. These features indicate energetic shallow-water conditions and have been interpreted as middle to lower shoreface deposits (Palma et al., 2006). The more muddy peloidal, oolitic and bioclastic packstones, associated locally with floatstone lag at the base of the beds, are characteristic of a lower energy environment, possibly lower shoreface or sub-fairweather wave base deposits (Burchette et al., 1990; Tucker, 2002).

All these are interpreted as inner ramp deposits, that can be mapped out with a lateral extent of >10‘s of kilometres across the basin.

Transition to Unit 2:

The transition from Unit 1 to Unit 2 is marked by the disappearance of ooids and, in most localities, the development of an indurated surface, highly bioturbated by *Thalassinoide*s, rich in iron and perforated by small bivalves. This surface contains many reworked, fragmented fossils and the first occurrence of fauna including *Pygurus* echinoids, serpulids, gastropods and coral fragments.

The presence of iron nodules and the induration of the surface associated with a diverse fauna of variable preservation quality, including encrusters and boring bivalves (Figure 5.7, D), indicates a period of non-deposition (Fursich et al., 1992). This records a hard-ground overlying shallow-water sediments; a condensed section characteristic of sea-level rise (Kendall and Schlager, 1981), interpreted as being the expression of a transgression across the basin.

5.4.3 Unit 2: *Somalirhynchia* Limestone Member

FO2: MF6, MF7, MF8, MF9, MF11 and MF12

Unit 2 is composed of alternating thin marls and mudstones and highly bioturbated fossil rich wackestones, packstones, floatstones and rudstones. The macrofauna is dominated by brachiopods, in various proportions, with species such as *Rhynchonella* and *Terebratula,* associated with bivalves, including *Pholadomya*, *Lopha, Trigonia* and *Trichites*. Other fauna are present in lesser proportion and gastropods, benthic foraminifers, regular echinoderms and serpulids are also common. The faunal association is dominated by suspension-feeders and epifaunal grazers in the lower part of the unit, where large *Collyrites* and *Pygurus* echinoids are very common, associated with gastropods and coral debris. Towards the upper part of the unit, with the disappearance of gastropods and echinoderms, the suspension-feeders become the only visible macrofauna.

FACIES NAME	MAIN ELEMENTS	MATRIX	SEDIMENTARY FEATURES AND BIOTURBATIONS	LARGE SCALE OBSERVATIONS AND BEDDING	SIGNIFICANT PALAEO-ENVIRONMENTAL FEATURES	LEGEND
MF1 Clay	Red and grey claystones	-	Homogeneous, rare horizons with paleosoil	0.2-7m thick beds Laterally continuous	**Paleosoils**	
MF2 Sandstone	Very fine to medium sand, well sorted	Calcite cement	Massive, horizontal laminations Alternating with marls	1-3m thick beds Laterally continuous	**Well sorted sand** **Horizontal laminations**	
MF3 Sandy/Silty Dolomite	Silt to medium sand, well sorted 20-50% Shell fragments 0-5% Serpulids, wood fragments (occasional)	Euhedral dolomite Calcite veins	Massive, vugs and recristalisation (100-500 µm Ø) up to 30% Alternating with marls	0.2-2m thick beds Laterally continuous	**Shell fragments** **Wood fragments**	
MF4 Dolomite	Phantoms of ooids (200-400 µm ø) 30-40% Phantoms of peloids (200-400 µm ø) 20-30% Phantoms of shell fragments 5-15%	Anhedral & subhedral dolomite 60-80% Calcite sparite filling previous porosity 15%	Thalassinoides Vugs (dolomite dissolution: 100-1000 µm Ø) up to 30%	0.2-3m thick beds Laterally continuous	**Thalassinoides** **Ooids, peloids**	
MF5 Mudstone	Shell fragments (<5%) Silt-very fine quartz grains (0-3%)	Micrite	Massive or nodular	0.2-1m thick beds Laterally continuous	**Massive/nodular**	
MF6 Marly Mudstone	Silt 10-40%	Micrite	Chondrites	1-5 cm thick beds, very thin bedding	**Chondrites** **Silt**	
MF7 Bioclastic Wack-/Packstone	Shell fragments 10-25%, quartz grains (20-100 µm) 10-40%, glauconite <1%,echinoderm (plates, spines, crinoids) 4%, peloids 2-5%, brachiopods 5-10%, bivalves 2-5%, Annelids (Ytharthanella?) 1%, Bryozoa <1	Micrite Calcite sparite Fearoan calcite microspar Euhedral dolomite (3-65%)	Poorly sorted bioclasts (20µm-2mm ø) Sharp base Occasional wave ripples on top	0.3-2m thick beds Laterally continuous	**Wave ripples** **Micrite content** **Bioclasts** **Sharp base**	
MF8 Oolitic/bioclastic Packstone	Ooids (200-800 µm Ø) 20-40%, intraclasts (oolitic PST) 1-30%, peloids 10-15%, coated grains (superficial ooids) 0-10%, aggregates 1-5%, brachiopod fragments 0-10%, shell fragments (Ø<150µm) 0-10%, echinoderms (ursins plates, spines, crinoids) <1% - 25%, foraminifers <3%, bryozoa <1%, coral fragments <1%, annelids <1%	Micrite, locally sparite, euhedral dolomite cristals (up to 38%)	Poorly sorted bioclasts 50µm-1mm ø Thalassinoides, often more developped on top of the beds	0.5-2m thick beds Laterally continuous	**Thalassinoides** **Micrite content** **Bioclasts**	
MF9 Oolitic Grainstone	Ooids (100-500 µm Ø) 40%, aggregates 10%, Peloids 5%, coral fragment 0-5%, echinodrem fragments 2-5 %, foraminifers <1%	Drusy calcite and second generation Fe rich sparite 35%	Thalassinoides on top of the beds Well sorted ooids	Lot always continuous laterally, mounded shoal, firmgrounds on some top	**Mound geometry** **Good sorting** **Thalassinoides**	
MF10 Sandy Grain-/Floatstone	Ooids and coated grains(100-500 µm Ø) 10-40%, medium to granule size quartz grains (coated or not) (2-40%), peloids 0-20%, coral fragments (2-10%), gastropods (2-10%), shell fragments (5-25%), echinoderm fragments (2-8%), dacycladacean algae (0-3%)	Drusy and blocky calcite and second 20-30%	Cross-bedding Low angle cross-bedding Planar bedding Truncated sets Iron-rich sandy intervals	Thick units composed of multiple sets 2-35m thick Sets 15-25 cm thick	**Low angle cross-bedding, planar bedding and truncated sets** **Iron-rich sandy horizons** **Mixed heterolitic sets**	
MF11 Grain-/Floatstone	Coral fragments 20-35%, ooids and coated grains (100-500 µm Ø) 0-25%, gastropods 5-10%, peloids 15-25%, shell fragments 1-20%, bivalves 0-10%, echinoderms 5-10%	Grainstone (peloids, shell fragments, coral fragments)	Cross-bedding, Low angle cross-bedding Hummocky cross-stratification Skolithos and conichnus trace fossil	Units composed of multiple sets 2m thick Sets 10-30 cm thick	**Fauna fragments** **Grainstone matrix** **HCS**	
MF12 Brachiopod Float-/Rudstone	Brachiopods 30-50% (Somalirhynchia assemblage), bivalves 20-40% (Ostrea gregaria, pecten, Trichytes, Pholadomya...), shell fragments 5-10%, echinoderms 0-20% (Pygurus dominating)	MF7 - Bioclastic wackstone or packstone	Not to very bioturbated Common firm grounds top of beds Thalassinoides and top perforated Sharp base with common fossils accumulations, thinning upward	0.2-2.5m thin continuous beds	**Fauna** **Firm ground and perforations** **Sharp base and grading**	
MF13 Brachiopod Float-/Rudstone	Brachiopods 20-50% (Rhynconella Trilobata assemblage), bivalves 20-40% (Pholadomya dominating), shell fragments 5-10%, echinoderms 0-20%, bryozoa 0-2%, flat corals 0-3%, sponges 0-5%, coral fragments 0-2%, ammonites 0-2%	MF7 - Bioclastic wackstone or packstone	Not to very bioturbated Common firm grounds top of beds Thalassinoides and top perforated Sharp base	0.2-2.5m continuous beds Occasional low angle clinoforms ±100m long	**Fauna** **Firm ground and perforations** **Sharp base** **Low angle clinoforms**	
MF14 Marls	Grey and blue marls	-	Thin horizons to homogeneous units, very friable	0.1-30m units	**Absence of fauna**	
MF15 Branching coral Boundstone	Branching corals 5-25%, solitary corals 0-5%, bivalves 10-15%, trichites 5-10%, coral fragments 10-30%, shell fragments 10-20%, gastropods 10-15%, echinoderm fragments 5-10%, peloids 10-20%	Packstone and grainstone	Common firm grounds top of beds Thalassinoides	1.5-5m continuous beds	**Fauna** **Branching and solitary corals** **PST/GST matrix**	
MF16 Mud Boundstone	Platy coral 10-40%, sponges 0-10%, thrombolites 0-20%, coral fragments 0-8%, echinoderm fragments (spicules dominating) 2-5%, calcispheres 1-2%, brachiopods fragments 1-2%, bryozoae 0-1%, small shell fragments 0-1%, foraminifera <1%	Micrite	Locally sponges or thrombolites encrusting platy coral	0.5-4m thick mounts 6-100m large mounts	**Fauna** **Muddy matrix** **Mound shape**	

Figure 5.7: Micro and Macrofacies of the Ouanamane Formation.

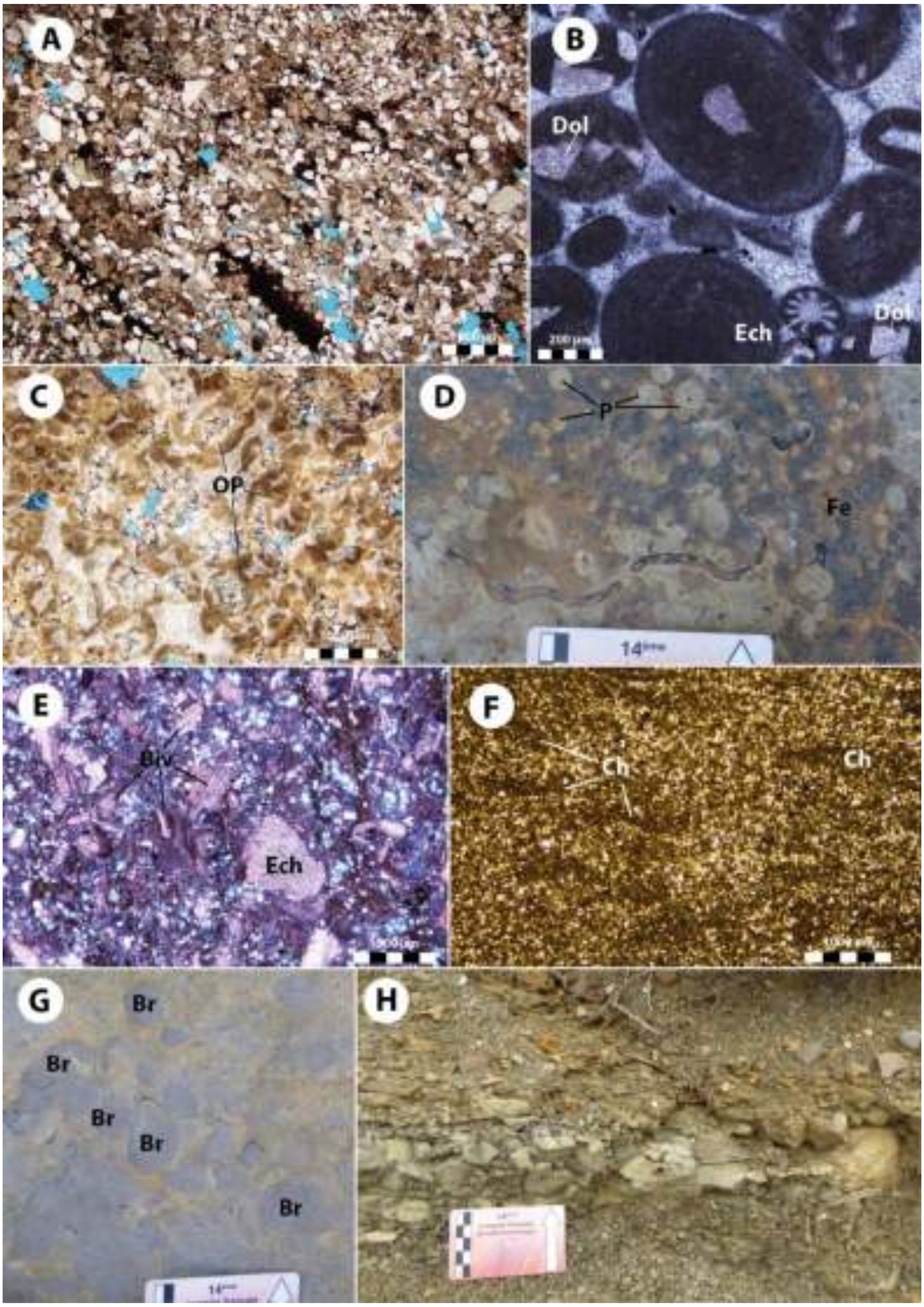

Figure 5.8: Micro- and macrofacies of inner to outer ramp.
(A) Sandy dolomite-MF3. (B) Oolitic GST-MF9. Euhedral dolomite (Dol) crystals partially replacing the ooids and pseudopeloids; micrite coated echinoderm spine (Ech). (C) Dolomite with ooids or peloids phantoms (OP)-MF4. (D) Hardground surface, iron-rich crust (Fe) perforated by bivalves (P). (E) Bioclastic PST- MF7. Silt quartz grains, echinoderm (Ech), bivalve (Biv) and brachiopod fragments in a micrite matrix. (F) Silty marls-MF6. Chondrites traces (Ch). (G) Brachiopod Rudstone-MF12. Terabratula and Rhynconella (Br). (H) Grey marl-MF14.

The microfacies of the fossiliferous units are dominated by bioclastic wackestones and packstones (MF7) (Figure 5.7, E). The bioclasts are mainly shell fragments of brachiopods and bivalves and echinoderm plates that have been broken up by the water energy, but are still subrounded to subangular, suggesting the bioclasts have not been transported too far by the currents. The bioclasts are derived from the same macro-elements present in the floatstones and rudstones units (Figure 5.7, G) and result from the reworking of these lithologies, either coming from shallower water and eroded by fair weather waves, or associated with storm-related currents. The amount of mud in the system and the many unbroken macrofossils in life position suggest the environment was quieter than Unit 1, but still subject to intermittent reworking.

In Unit 2 many bed-tops are reworked by *Thalassinoides* boxworks, which have often been later filled by coarser deposits. The top of these beds are generally partially indurated and sometimes encrusted and bored. These are interpreted as firm-grounds and hard-grounds and attest to a slow sedimentation rate and cementation and boring on the seafloor (Ekdale and Bromley, 1984; Fursich et al., 1992; Tucker, 2002).

Small buildups have been observed in different locations (consisting of an association of platy-corals growing-up in a muddy environment in the upper part of Unit 2. The background sedimentation is composed of mudstones and wackestones with few shell fragments, echinoderm spicules and bryozoa fragments. These geobodies are composed of mudstones, floatstones and rudstones. The facies are open marine, lower ramp deposits, with ammonites, bryozoan, terebratula and echinoderms. The corals are microsolenids (*Dimorpharaea*) and their flat plate-like habit is a common response to low light, low energy and low background sedimentation rates (Insalaco, 1996; Olivier et al., 2012). One of these small buildups observed at Tizguit has low angle clinoforms that prograde southward off the topographic high formed by the buildup (Figure 5.9).

Facies association FO2 had low energy but sufficient light for the macro-fauna, was relatively close to the shoreface to allow the presence of bioclasts fragments and quartz grains in the environment. This facies association is interpreted to have developed below the fair weather to storm wave base, in middle to outer ramp settings.

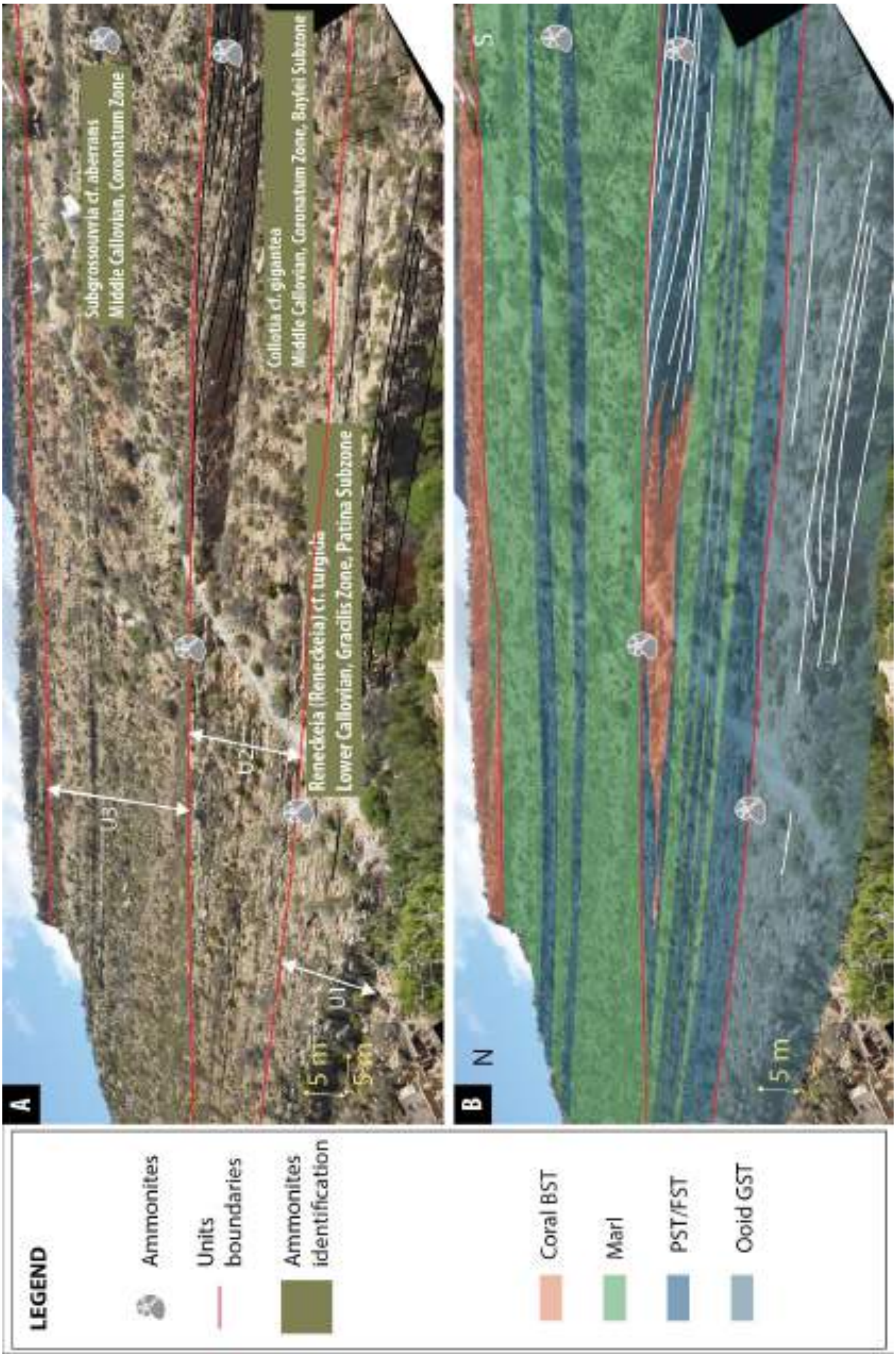

Figure 5.9: (A) Photomosaic of the Tizguit transect and ammonite fauna associated. (B) Simplified sedimentology and geometries associated.

Transition to Unit 3:

The transition from Unit 2 to Unit 3 is locally marked by firmgrounds and hardgrounds. The important bioturbation, iron accumulation, encrusting organisms (oysters, serpulids, sponges) and perforations indicate a time of little sedimentation, which can be linked to a major transgressive event. This horizon records a very diverse fauna, with bryozoan, crinoids, echinoids, massive, branching and flat corals, sponges, bivalves, brachiopods, gastropods and ammonites. The macro-fauna of this level indicates more open marine influence and record the change of environmental conditions towards deeper settings. These deposits mark a strong change in the sedimentation from a fossiliferous unit to a unit dominated by mud and marl.

5.4.4 Unit 3: Marls Member

FO3: MF5, MF6, MF12, MF13, and MF14

Facies Association FO3 is dominated by thick (up to 60 m) marls (Figure 5.7, H), with thin (0.2-1m) intercalations of brachiopod-rich floatstones and rudstones. The lower part of the section is dominated by nodular mudstones with occasional brachiopods floating in the matrix. The scarce fauna of small terebratula and the absence of broken shell fragments indicate a low energy environment, linked to muddy subtidal conditions (Ager, 1965).

The marls are interbedded with indurated levels of floatstones and rudstones. The fauna within these floatstones and rudstones is dominated by brachiopods, large gastropods, echinoderms, shell fragments and also ammonites and platy corals. The extent of these horizons is variable; some locally pinching out, while other horizons can be followed for kilometres. Compared to the marl horizons, where no fauna is to be observed, these fossiliferous horizons record a strong change in the environment of deposition. The base of these beds is often sharp and the floatstones become more marly towards the top or are terminated by a firm-ground.

The firmgrounds record periods of reduced sedimentation (Fürsich et al., 1992) and are dominated by boring bivalves (Figure 5.10B), the reduction of encrusters compared to Units I and II can be linked with the absence of fauna in the marls, suggesting a diminution of light and nutrients in deeper settings. The upward fining beds contain horizontal horizons of flat corals and unbroken *Hemicidaris* (regular echioid) in life position, often still retaining their spicules (Figure 5.10A and C). This is interpreted as evidence for a low to moderate energy environment, with little reworking. These floatstones and rudstones horizons are thus interpreted as recording deposition during regressive events in a general deeper environment.

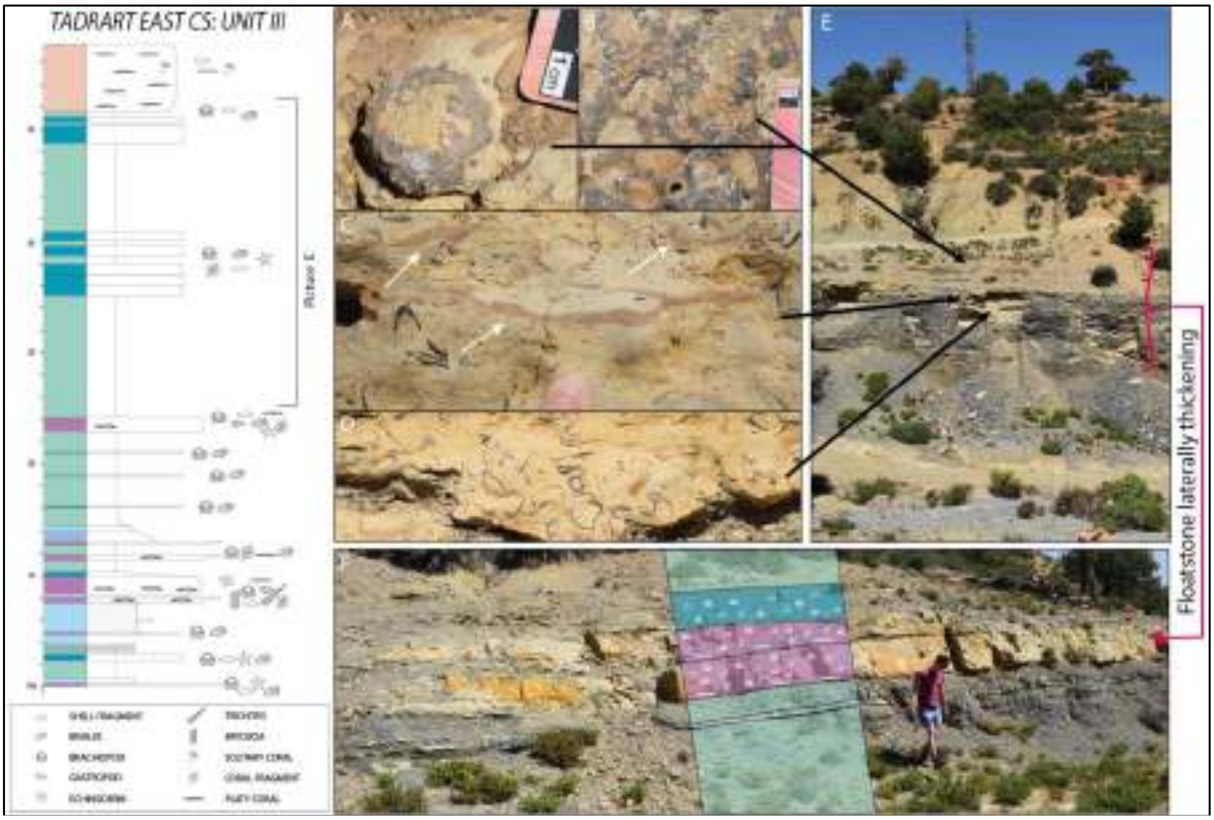

Figure 5.10: Composite section of the Tadrart East
(location Figure 5.1) area and pictures of key features. (A) Sea urchin with spicules in life position. (B) Firm ground; showing hardened and encrusted surface with bivalves. (C) Floatstone/boundstone facies; white arrow pointing to platy coral in life position. (D) Brachiopod floatstone. (E) Overview of the upper part of the section. (F) Close-up view showing how the indurated horizon thickens 200m northward from picture E.

Locally the upper part of the Marl Member is composed of up to 20 m of silty mudstones (40% silt) with abundant *Chondrites* burrows (Figure 5.7, F). *Chondrites* are typically interpreted as being a marine indicator and the presence of a monospecific trace fauna is an indicator of restricted 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).

Facies Association FO3 is characteristic of a quiet marine environment with low light. This facies association is interpreted as being deposited in a lower ramp environment.

5.5 Depositional environment

The integrated interpretation of the facies association and observations of the lateral variations across the basin allow a better understanding of the environments of deposition.

5.5.1 Facies variations across the basin

The three main stratigraphic units (Units 1, 2 and 3) have been correlated around the basin, with boundaries defined by major hardground surfaces. Facies variation is observed, interpreted to record lateral changes in environment and the control of proximal – distal, shallow - deeper conditions. In the more proximal localities (Tikki and Tizgui N'Chorfa) Unit 1 shows the highest facies variablilty, wherease Units 2 and 3 are similar. In the Essaouira sub-basin, along the Jbel Amsittène, variations are also visible from the northern to the southern part of the anticline between localities AME and AMCA. Variations and detailed description of facies observed are detailed below.

- **Askouti**

On the northern flank of the Lgouz Anticline, in the southern margin of the basin, the Askouti section records a very condensed Ouanamane Formation only 48 m thick. Unit 1 is thin silty dolomite alternating with peloidal and bioclastic packstones and grainstones. The upper part of this unit is composed of oolitic and peloidal grainstones, locally floatstones with gastropods and echinoderms. These facies are very similar to those observed in the centre of the basin, suggesting a similar environment of deposition, but it is condensed from 80m in the basin centre (Imouzzer section) to 5 to 10 m in this marginal location. Unit 2 is dominated by brachiopod-rich floatstones and rudstones alternating with marls horizons and Unit 3 comprises a 15m thick interval of marls with minor rudstones horizons dominated by brachiopods and bivalves. Overall the Ouanamane Fm. in this location is comparable in terms of facies association to that recorded from the center of the basin, but thinner.

- **Tikki**

At Tikki, located in the more proximal part of the basin, in the East, along the Argana Valley, the Unit 1 is dominated by oolitic grainstones in most localities but records lateral facies variations across the basin. In the Tikki section (TKK), the Unit 1 is composed of an alternation of oolitic-bioclastic grainstone with a coral and gastropod dominated floatstone to rudstone (MF11). The base of most sets is erosive and sedimentary structures visible in the rudstones and floatstones beds are restricted to low angle cross-stratification and planar bedding.

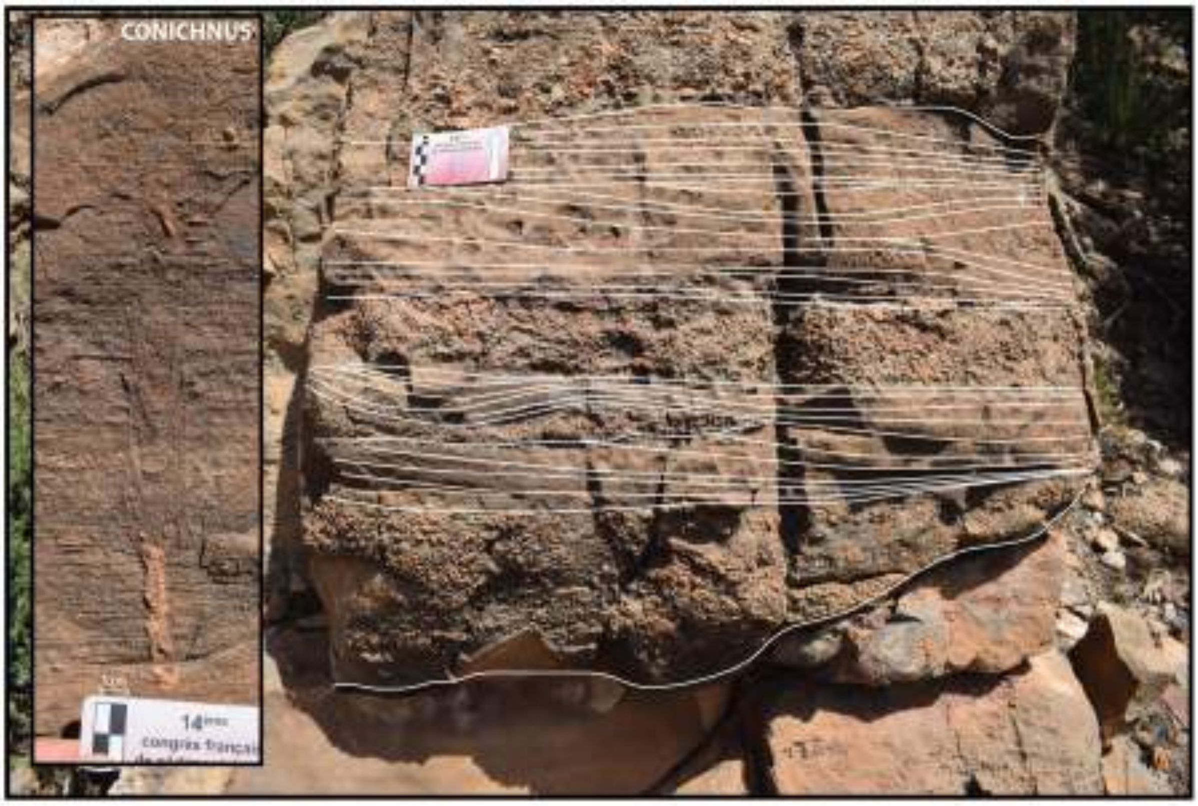

Figure 5.11: Upper shoreface deposits of the Tikki section
(Alternation of coraline rudstone and oolitic-bioclastic grainstone). Conichnus trace-fossil.

The presence of coral fragments, unbroken gastropods and echinoderm spines in the rudstone beds indicate an open marine environment with some reworking. In the grainstone horizons, the bedding is dominated by planar bedding, trough cross bedding and some hummocky cross stratification. These horizons contain vertical bioturbation and two types of trace fossils are observed, *Conichnus* and *Skolithos* (Gerard and Bromley, 2008). This association of sedimentary features and trace fossils are characteristic for shoreface deposits (MacEachen and Pemberton, 1992) (Figure 5.14, C).

This is followed by a brachiopod-rich floatstone, correlated as Unit 2. The main difference in this part of the basin is the development towards the top of branching coral colonies in Unit 2 (Figure 5.14E), which indicates good light availability and oxygenated conditions, suggesting shallower water with good circulation. This suggests shallower conditions at this locality than the center of the basin, where the upper part of this unit is dominated by increasingly marly deposits and deeper water fauna.

- **Tizgui N'Chorfa**

Tizgui N'Chorfa is a section located in the eastern part of the basin, in proximal settings; Unit 1 comprises a silty and sandy dolomitic member marking the transition between the fluvial Ameskhoud Fm. and the fully marine carbonates of the Ouanamane Fm. It is very thick (>20m) and is interpreted to record a lateral facies change from the "oolitic member" which dominates the center of the basin. It is contains alternating thin marl horizons (Figure 5.14, a) and is followed by a thicker marl unit before the development of Unit 2, which is dominated by a brachiopod-rich floatsone and rudstone. The bulk of Unit 2 is composed of grainstones, floatstones and rudstones (MF11 and MF12), with few marls (MF14) and the upper part of this unit is dominated by 2m of grainstone-floatstone with abundant branching corals (MF15) (Figure 5.14, e). The reduced proportion of marls compared to the middle of the basin tends to indicate higher energy conditions, where current winnowing transports finer grains away in suspension. The absence of fauna indicative of deeper water (e.g. ammonites and flat corals) in this part of the basin tends also to points towards shallower environments.

- **Amsittène Anticline**

AMCA

In the location of the section AMCA, in the South of the Amsittène Anticline, the facies are sensibly deeper than in the east and south of the basin and multiple ammonites have been collected. The age of the ammonites collected does not exceed the Middle Callovian, Anceps Zone (Stuebeli Subzone) (Figure 5.6).

Unit 1 in this section is quite thick (92 m) and composed of thinly bedded and highly bioturbated oolitic, peloidal and bioclastic wackestones, packstones and grainstones, (Figure 5.12, d). The top of the beds are often encrusted, very fossiliferous, floatstones to rudstones and are interpreted as firm-grounds and hard-grounds (Gruszczynski, 1979; Fursich et al., 1992). This unit is marked by a variety of fauna, comprising gastropods, echinoderms, branching and massive coral fragments and in situ small colonies (Figure

5.12, b and c), trichites, bivalves and brachiopods, as well as oncoid horizons (Figure 5.12, a). The lower part of the succession is dominated by oolitic and bioclastic grainstones beds, often followed by gastropod-rich or oncoidal floatstones in a wackestone to packstone matrix. These abundant lobate and concentric oncoids, up to 2 cm in diameter, associated with wackestones, packstones and floatstones, grow mainly on coral and shell fragments and are characteristic of a platform patch-reef environment (Flugel, 2010). The repetitive occurrence of the different facies and firm-ground surfaces shows cyclicity in the sediment deposition. This formation is interpreted to have been developed in shallow marine conditions in a protected environment, with fluctuations from periods of little agitation to periods with stronger currents. These have been interpreted as cyclic high-order transgressive-regressive cycles, alternating between a low energy inner ramp environment and oolitic shoals (Figure 5.14**,** b). The upper part of this unit is partly covered and dominated by silty oolitic and bioclastic grainstones and floatstones with horizontal and trough cross-bedding interbedded with some thin marl horizons (Figure 5.12, e). This is characteristic of a slightly deeper environment, and has been interpreted as the outer part of the inner ramp (Figure 5.14).

Figure 5.12: Facies of the Ouanamane Fm. in the section AMCA.
(A) Oncoidal Floatstones, dominated by oncoids (On), Coral fragments (Co) and Microbially coated shell (Me). (B) Scleratinian coral, bored by bivalve (Bo). (C) Larger colony of branching coral (arrow). (D) Bed heavily bioturbated by *Thalassinoides* (Th), with large *Trichites* bivalves (Tri). (E) Trough cross bedded horizon.

The transition to Unit 2 is characterized by the disappearance of the oolitic grainstones over the last 20m of the Unit 1, replaced by peloidal grainstones then peloidal packstones and marls, indicating a deepening and strong transgression,. The last few metres of this unit is made of marly wackestones with *Chondrites* bioturbation, followed by yellow marls topped by a peloidal packstone capped by a hard-ground. This transgressive surface has been taken as the limit between Unit 1 and Unit 2. Unit 2 in this section is very similar to that observed in the center of the basin, containing the same facies and fauna.

The main change in this part of the Jbel Amsittène is the proportion of marls, which is much higher, associated with a higher abundance of ammonites that indicates a slightly deeper environment.

It is uncertain if Unit 3 is present at this locality, it may be missing or represented by a 2m condensed interval in the marls at the top of the succession. Unit 2 is overlain by carbonate cement sandstones, consisting of very well sorted very fine quartz grains, with coated quartz grains horizons, alternating with gastropods of brachiopods rudstones. It contains horizontal bedding, trough cross-bedding and wave ripples. Large coral fragments or shell fragments are also common in this succession, which indicate the proximity of a carbonate factory. This succession has an erosive base. This unit become less silica-clastic rich towards the top and the coated grains are replaced by ooids with quartz nuclei, then the quartz influx disappear totally and the facies evolve to the shallow carbonate deposits of the Iggui El Behar Formation (Adams, 1980). This association suggests high energy conditions with alternating currents, and has been interpreted as a strandplain beach deposit (Dominguez and Wanless, 1991). The interpretation is this is the lateral equivalent of Unit 3 and the Lalla Oujja Fm., on the margin of the basin, or that Unit 3 has been eroded and replaced by these beach deposits.

AME

In AME, to the NE of the Anticline, the facies are quite different compared to the AMCA sections located to the south of the Anticline. The base of the succession in this locality is not accessible, but the first beds are composed of bioclastic packstones and grainstones, similar to Unit 1 as described from the centre of the basin. These beds are followed by a mixed unit, composed of well rounded quartz granules and medium sand mixed with small bioclasts and larger shell fragments and brachiopods. The siliciclastic content decreases rapidly upwards in the succession, and these mixed deposits are rapidly replaced by deposits characteristic of Unit 2. The base of Unit 2 is marked by a hard-ground level. Packstones, brachiopod-rich floatstones and rudstones are recorded, with intense *Thalassinoides* bioturbation and recurrent firmgrounds at the top of the beds, alternating with more marly horizons, rich in bivalves and brachiopods.

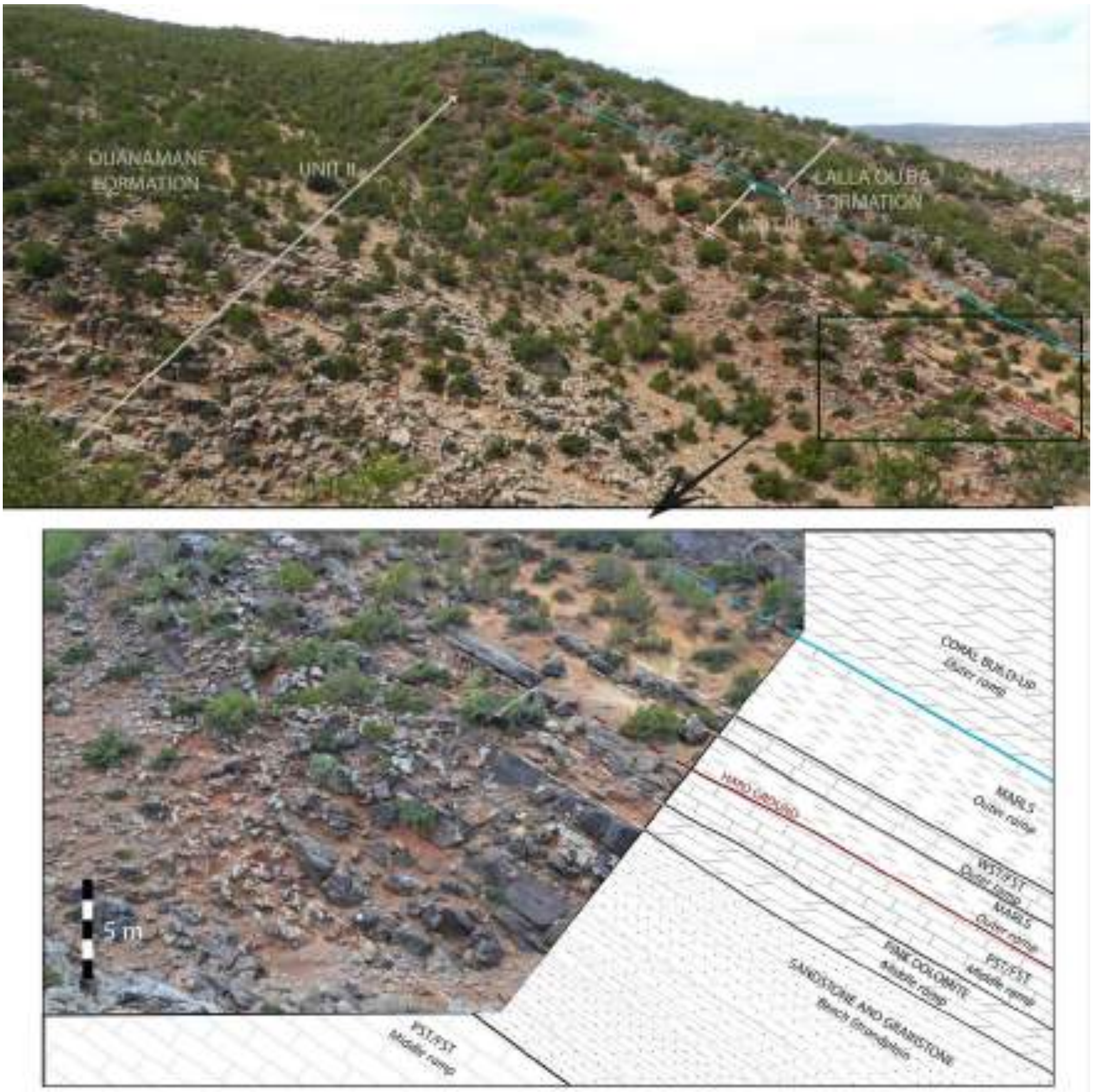

Figure 5.13: Succession of the Upper part of the Ouanamane Fm. in the South of the Jbel Amsittène, section AME.

This succession is interpreted to record middle ramp carbonates, later interrupted by a second flux of silica-clastic sediments into the basin (Figure 5.13). This sequence is composed of mixed deposits, organized in horizons with various proportions of quartz and bioclasts. These deposits are composed of well sorted fine and very fine sandstones with carbonate cement alternating with coated grains grainstones and fossiliferous horizons. Most of the carbonate-rich horizons are made of coated quartz grains or coated bioclastic elements, always well sorted fine grains. Occasional larger bioclasts (e.g. coral fragment, bivalves, and echinoderm spicules) appear in the succession along discrete horizons. The sedimentary features are dominated by horizontal bedding and low angle parallel laminations. Some horizons with cross-bedding can also be distinguished. The sandier horizons seem to concentrate iron deposition as they appear dark orange on the outcrop and can be easily distinguished. These deposits have been interpreted as beach strandplain (McCubbin 1981; Dominguez and Wanless, 1991) deposits (Figure 5.14, f) due to the open-marine faunal fragments associated to the good sorting of the deposits and the sedimentary structures typical of beach deposits (Thompson, 1937).

This sequence is followed by 4 m of sandy pink dolomite evolving to grey mudstones and capped by a highly bioturbated floatstone horizon with iron nodule and encrusting fauna interpreted as a transgressive hardground horizon. This horizon marks the transition to the Unit 3. Unit 3 at this locality is more comparable to that observed in the center of the basin. It is composed of 12m of yellow marls, with one horizon of bivalve and brachiopod rich floatstone. It has been interpreted as outer ramp deposits.

5.5.2 Depositional models

The detailed description of the stratigraphic succession and facies associations, together with analysis and correlation of facies variations across the basin has allowed three depositional environment models to be established for each unit of the Ouanamane Fm. They represent the evolution an open ramp system with gradual changes of facies from high energy to low energy, no evidence of break in the slope or protected lagoonal deposits. The transition from unit 1 to unit 2 is mainly a change in bathymetry or nutrient availability, while from unit 1 to unit 2, the depositional environment change more deeply, from oolite-dominated to macrofauna-dominated. The three units are each separated by a transgressive surface along a hard-ground and reflect an overall deepening of depositional environments.

The deposits of Unit 1 of the Ouanamane Fm. are similar to the ramp system of the Amellago transect, during Lower to Middle Jurassic in the Central High Atlas (Pierre et al., 2010). The Middle Jurassic carbonate ramp of the Lusitanian Basin also present equivalent facies distribution (Azerêdo, 1998). Both systems present inner ramp deposits, with large oncoids beds, ooids and peloids packstones, laterally evolving to middle ramp deposits dominated by oolitic packstones and grainstones shoal deposits. The Units 2 and 3 present a facies distribution closer to the Kimmeridgian of the Iberian Basin (Aurell et al., 1998), while less dominated by tempestites.

Unit 1: Iggui-n'Tarhazout Oolitic Member

Unit 1 is dominated by inner to middle ramp oolitic and bioclastic deposits. Inner ramp facies, partially protected by oolitic shoals, are dominated by platform patch-reef deposits, with recurrent oncolitic horizons. More proximal deposits comprise silty and sandy mudstones, potentially deposited in coastal lakes. Oolitic deposits dominate the outer inner ramp and the middle ramp deposits. Discrete beds of brachiopod-rich floatstones, which record deeper conditions, are rare in Unit 1 but dominate Unit 2.

Unit 2: *Somalirhychia* Limestone Member

Unit 2 records the middle to outer ramp facies. It is characterized by a basal transgression and the shift from packstones and floatstone-dominated carbonates to more marly deposits. Shallower conditions of upper middle ramp to inner ramp develop patch-reefs of thin branching corals in a packstone to floatstones matrix. The middle ramp is characterized by brachiopod-rich floatstones and rudstones, becoming more marly down-

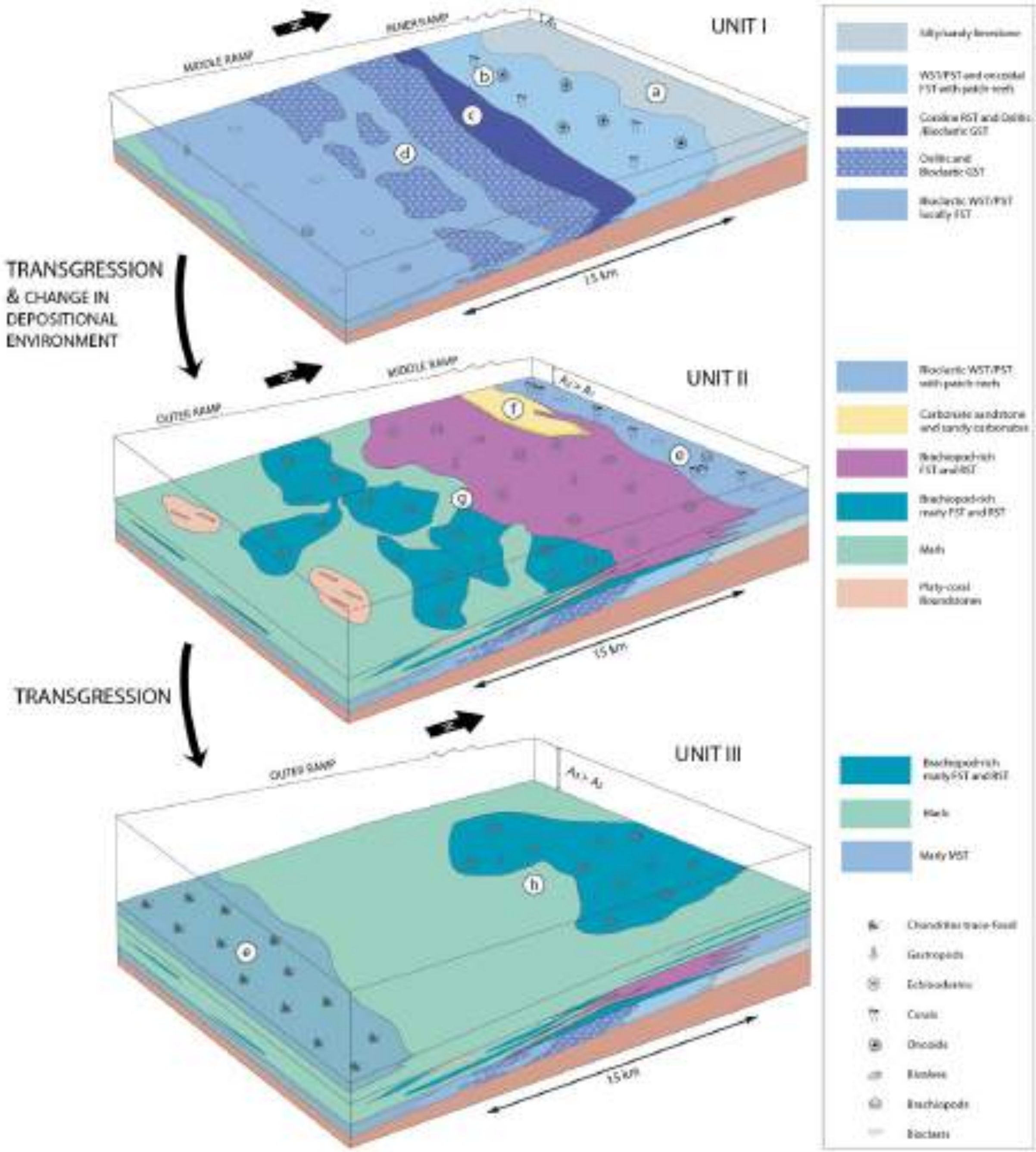

Figure 5.14: Depositional environments of the Ouanamane Fm. for the units 1, 2 and 3. Unit 1 depositional environments based on observations from Tizgui N'Chorfa (a), Amsittène AMCA (b), Tikki (c), facies associations from the center of the basin (d). Unit 2 environments based on observations from Tikki and Tizgui N'Chorfa (e), Amsittène AME (f) and facies associations from the center of the basin (g). Unit 3 environments based on Assif El Hade (i) and center of the basin facies association (h).

ramp. Beach strandplain deposits are recorded in the north. The outer ramp deposits are made of marls alternating with beds rich in brachiopods, bivalves, rare ammonites and local small-scale (10 to 40m wide) buildups composed of platy-coral dominated muddy bindstone.

Unit 3: Marl and Shale Member

Unit 3 records a deeper environment of deposition than Unit 2. It is characterized by extensive marls, with fossiliferous units composed of bivalves, brachiopods, and some ammonites and echinoderms. The deeper (more protected) part of the depositional environment is characterized by marly mudstones with intensive fine chondrites bioturbation.

5.6 Discussion

5.6.1 The Callovian-Oxfordian transition: potential hiatus and condensation

The ammonite data strongly suggest non deposition for the uppermost Callovian (Lamberti Zone pro parte) and lower Oxfordian (Mariae and Cordatum zones pro parte). Non deposition and hiatuses are a common feature along post-rift margins preservation during the late Callovian to early Oxfordian time interval (Aurell et al., 1994; Azerêdo et al, 2002; Meléndez et al., 2005; Ramajo and Aurell, 2008).

The western and southern margins of the Subalpine Basin do not record any late Callovian strata along the platforms margin (Elmi, 1990; Fortwengler et al., 2013). In Northern Switzerland, the latest Callovian Henrici Subzone (= Poculum Subzone) is affected by a hiatus, and the early 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 North Sea, the lower Callovian 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; Ramajo and Aurell, 2008).

All these examples, along the Atlantic and Tethyan margins, record a similar interval of sediments missing. This level has generally been linked by the authors to sea-level fall and more locally to tectonic movements, with erosion of the sea floor or creation of condensed horizons. In the Lusitanian Basin, which is geographically close to the EAB, the middle-late Jurassic discontinuity is associated to a forced regression (Azerêdo et al., 2002).

If the late Callovian – early Oxfordian has been long interpreted as a global sea-level rise (Haq et al., 1987, 1988; Norris and Hallam, 1995; Hardenbol et al., 1998; Hallam, 2001; Wierzbowski et al., 2009), multiple examples from the Tethys and the Central and North Atlantic tend to prove the opposite. In the Prebetic (Southern Spain), weathering of shelf deposits was induced by a relative sea-level fall during the Middle to Upper Jurassic times (Reolid et al., 2008; Reolid and Abad, 2018). Evidences of shallow to emerged episodes were recorded in the Lusitanian Basin (Azerêdo et al., 2002), the Iberian Basin (Aurell et al., 1994; Ramajo and Aurell, 2008), south Tunisia and northwest Libya (Walley, 1985, Mette, 1997), Algeria and Egypt (Carr, 2003).

The latest revised global sea-level curve (Haq, 2018) indicates a sea level rise from the Bathonian to the Middle Callovian followed by a sea level drop with a maximum of regression during the earliest Oxfordian, and a new sea level rise up to the Upper Kimmeridgian. Isotopic thermometric studies indicates that the late Callovian to early Oxfordian time was subject to a cooling and thus to lower concentrations of CO_2 which would have induced a drop in carbonate production (Dromart et al., 2003a, b; Cecca et al., 2005; Andrieu et al., 2016).

In the EAB, the thin interval that separates the upper Callovian from middle Oxfordian sediments indicates low sedimentation rates and/or potential erosion. This interval can be considered as a poorly constrained condensed interval. In the Atlantic Margin context cited above, this interval could be linked to a non-recorded regression-transgression event, but no field evidences were found to corroborate this hypothesis.

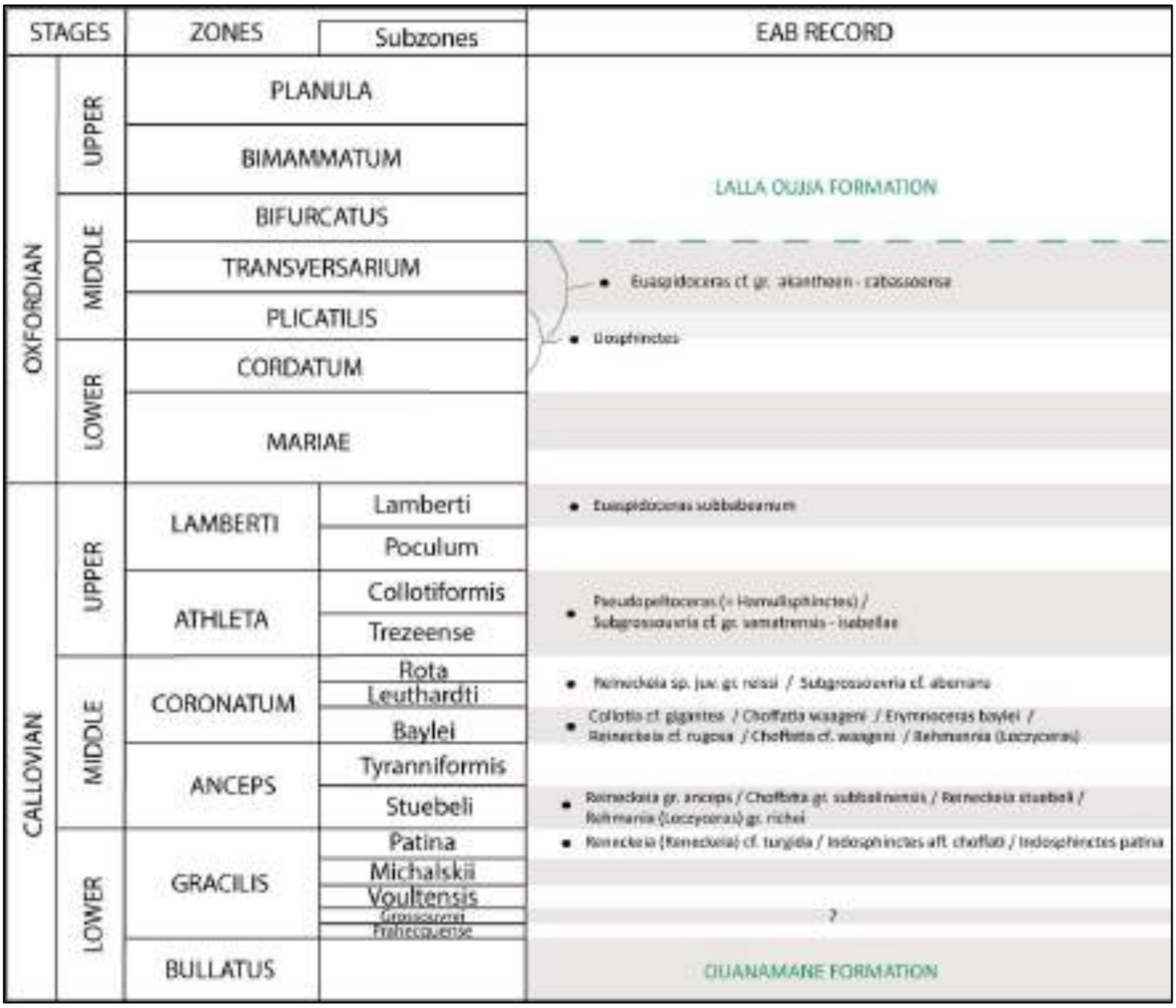

Figure 5.15: Synthetic ammonite biostratigraphic scale for the Callovian and Oxfordian of the EAB.

5.6.2 Buildups synchroneity

The biostratigraphic analysis has consistently found middle Oxfordian fauna underlying the Lalla Ouija Fm. buildups, indicating a middle Oxfordian age for the base of the main coral buildup in the Lalla Oujja Fornation. 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 middle Oxfordian. At Cap Rhir, small coral mud-mounds are observable below the middle Oxfordian ammonite found but do not represent major buildups. Different major buildup generations can be distinguished but the first is synchronous in the center of the basin. Therefore, the coral builds-up might be synchronous across the entire basin, up to Askouti, Tikki or the north of the Amsittène

Anticline (AME), where no ammonite fauna were found. The synchroneity of the coral buildups indicates basin-scale change in the environmental conditions.

The most recent global sea-level curves (Haq, 2018) indicate the middle Oxfordian as the onset of the Upper Jurassic transgression. Worldwide examples of early to middle Callovian transgression, late Callovian regression and middle Oxfordian regression, including the SE of France, the Lusitanian Basin (Portugal) and the Iberian Basin (Spain) (Dromart et al., 2003, with references) indicates these transgressive/regressive events to be global. As already mentioned, the latest Callovian to early Oxfordian interval was marked by a drop in temperatures, CO_2 level and sea-level, therefore the rise of temperatures, sea-level and CO_2 in the middle Oxfordian (Dromart et al., 2003) is very likely to have played a major role in the coral-reef development across Europe, North Africa, Middle East and Central Asia (Cecca et al., 2005). The fact that small platy-coral bioherms and biostromes already appeared during the Callovian but never extended over more than a few tenth of meters could be explained by the lower Callovian temperatures (Dromart et al., 2003), the environmental conditions being more favourable to other faunal assemblages.

5.6.3 Geometries

In Tizgui, the clinoforms linked to platy-coral buildup (Figure 5.9) are composed of mudstones, floatstones and rudstones. The facies are open marine, lower ramp deposits, with ammonites, bryozoan, terebratulids and echinoids. Two hypotheses can explain the occurrence of the low angle clinoforms.

- Hypothesis 1 (Figure 5.16)

An increase of marls towards the top of Unit 2 suggests a transgressive pattern. The limited fauna found in these horizons indicates that environmental conditions degraded up through the unit, with reduced light and muddier substrates possibly being important factors. This particularly affected the macrofauna that was dominant lower in the unit.

The growth of the coral buildup provides a local high. The shallower conditions offer light and a stable substrate allowing brachiopods, echinoderms and bivalves to anchor. In these conditions, bioclastic material produced at the surface of the buildup could be transported down-ramp to produce the low angle clinoforms.

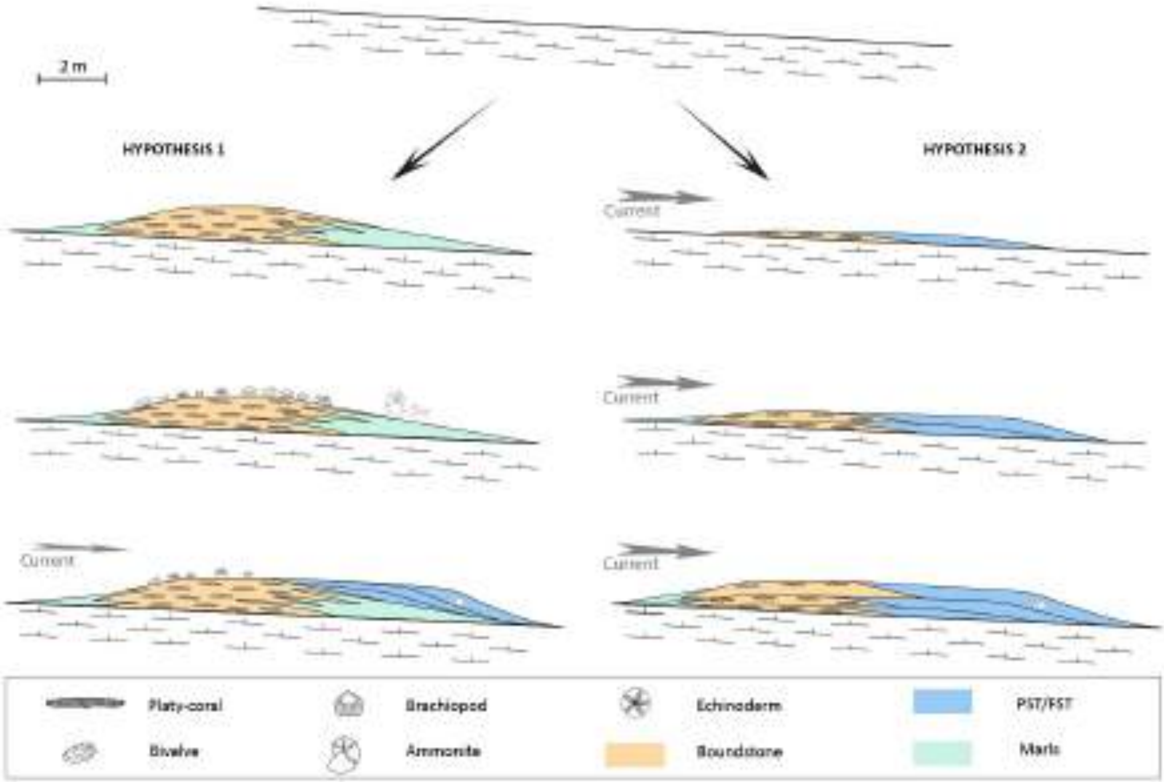

Figure 5.16: Schematic illustration of the hypothesis 1 and 2 about the link between the development of the coral mound and the low-angle clinoforms in the Ouanamane Fm. in the locality of Tizgui.

- Hypothesis 2 (Figure 5.16)

In a low energy environment with low light conditions, it could be questioned why the platy-corals began to develop. Three factors typically control faunal development; oxygen, light and nutrient level. Delivery of nutrients and presence of aerobic conditions are related to currents agitating the system. The growth of corals might occur linked to more energetic events bringing nutrients to the system, alternating with calmer periods when mud would be deposited. The lower energy fine deposits were then encrusted by the platy corals, which had the effect of stabilize the mud and prevent erosion/transport during higher energy periods. The higher energy currents transporting nutrients to the coral colony could also be favourable to the life of other fauna or even transport bioclasts. The clinoforms could then form downdip of the coralline mud-mound, which forms a cohesive structure, baffling or re-directing the current energy.

5.6.4 Comparison with the Scotian Basin and petroleum system implications

Oolites

The Iggui-n'Tarhazout Oolite Mb. (Unit 1) is a thick (30-60m) and porous stratigraphic body that extends across most of the basin. The oolites are generally well sorted, locally entirely replaced by crystalline dolomite, with good visible porosity. In Imouzzer, Assif El Hade and Tizgui the upper part of the thick oolitic units have a strong kerogenic smell, suggesting migration of hydrocarbons into open pore space during diagenesis. It is part of a major transgression, which led to the deposition of a thick marl unit (Unit 3) across the basin that might act as a seal. Therefore, the oolite member can be considered as a potential reservoir unit in the EAB.The facies of Unit 1 are very similar to the platform interior and lagoon facies reported of the Abenaki Fm. from the Scotian Margin (Weissenberger et al., 2006).

There, lagoonal deposits include common oncolitic facies and wackestones to packstones with oolitic and reefal material. The platform crest is dominated by oolitic grainstones.

Unit 1 can be correlated to the age equivalent Scatarie Mb. of the Abenaki Fm. (Ellis et al., 1985; Weissenberger et al., 2006; Eliuk, 2016). The Marl and Shale Mb. is the counterpart of the Misaine Mb. in the Scotian Basin and this regionally extensive shale unit could be a good seal for fluids migrating into oolites of the Scatarie Mb.

Buildups

In the EAB, the buildups of the Lalla Oujja Fm. (Ambroggi, 1963; Adams, 1979, 1980; Martin-Garrin et al., 2007; Olivier et al., 2012) have been locally extremely dolomitised and present very good porosity in outcrop. They are also capped by carbonate mudstones followed by thick red marls (Ambroggi, 1963), which are a potential seal. In the offshore Tarfaya Basin, further south from the EAB, the Cap Juby oil field proves the reservoir potential of the Upper Jurassic limestones (Morabet et al., 1998; Davison, 2005). The biostratigraphic revision of the EAB indicates a relatively synchronous appearance of the coral bioherms of the Lalla Oujja Fm. during the Middle Oxfordian. This implies a rapid development of the bioherms across the basin and a kilometre scale lateral extension of the bioherms unit.

In Nova Scotia, the Deep Panuke gas field is hosted in a reservoir unit equivalent to the Lalla Oujja Fm. and dated upper Jurassic (Ellis et al., 1985; Weissenberger et al., 2006; Eliuk, 2016). This reservoir unit is the upper part of the Abenaki Fm., the Baccaro Member, which is generally dated Kimmeridgian to Tithonian. If the assumption of a very similar stratigraphic development in Morocco and Canada is correct, the revised biostratigraphy in Morocco could indicate that the base of the Baccaro Mb. is slightly older than previously thought.

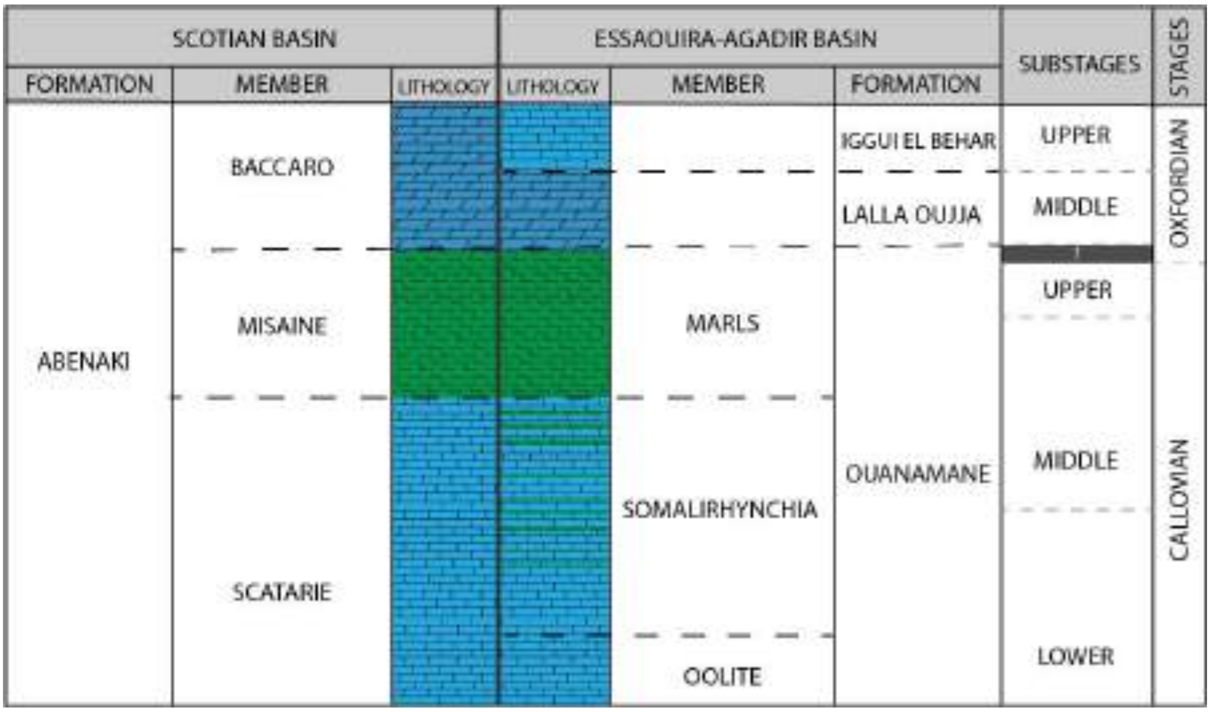

Figure 5.17: Stratigraphic chart comparing simplified lithologies of the Scotian Basin and the Essaouira-Agadir Basin for the Callovian-Oxfordian interval.
Scotian Basin stratigraphy adapted from Weston et al. (2012).

5.7 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 Essaouira-Agadir Basin:

1. A revised lithostratigraphic scheme is presented, with the identification of two regional hard-grounds interpreted to record major transgressions. They are used to define the boundary between the three members of the Ouanamane Fm.
2. Integrated biostratigraphic analysis points out to an early Callovian age for the base of the Ouanamane Fm., reassessing and discounting the late Bathonian age proposed by Bouaouda et al. (2004, 2009).
3. 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 and Shale 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).
4. In the Marl and Shale Mb. (Unit 3), a potential hiatus/condensation interval that spans the latest Callovian and most of the lower Oxfordian is highlighted.
5. The facies distribution indicates deposition dominated by an open ramp environment that can be correlated across the basin. The new high resolution correlation allows regional variations to be reconstructed for the depositional environment of the three members of the Ouanamane Fm.
6. The Iggui-n'Tarhazout Oolite Mb. is absent in the more proximal part of the basin (Tizgui N'Chorfa), where the equivalent interval is composed of sandy carbonates and transgressive shoreface deposits (Tikki section), and towards the north of the basin (Amsittène Anticline, AMCA section), the inner ramp deposits are platform patch reef and oncoidal deposits protected by oolitic shoals.
7. The *Somalirhynchia* Limestone Mb. is dominated by brachiopod-rich middle to open ramp deposits at the scale the basin.

8. The Marl Member consists of marls deposits with rare brachiopod-rich floatstones and deeper or more protected environments dominated by marly mudstone highly bioturbated by chondrites.
9. Small (up to 2 m thick) platy-coral buildups are identified in the *Somalirhynchia* Limestone Mb. These buildups are associated to the development of low angle fossiliferous clinoforms.
10. Synchroneity of the Lalla Oujja Fm. buildups was highlighted by the biostratigraphic analysis. Where dating by ammonites is possible, the base of the formation is always of middle Oxfordian age.
11. The late Middle to early Upper Jurassic successions of the EAB successions record the major mid-Callovian and mid-Oxfordian transgressive pulses that are recognised at the global scale (Haq, 2018 with references). There is no evidence of a late Callovian global sea-level rise such as the one interpreted by Wierzbowski et al. (2009). On the contrary, the EAB succession suggests a major regressive event close to the Callovian – Oxfordian boundary such as the one documented in the Lusitanian basin (Azerêdo et al., 2002).

Legend 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

Legend 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 Glowniak (2002) MEM-AM-997

4. *Hamulisphinctes* of the *hamulatus* group (Buckman, 1921) MEM-AM-993

5. *Erymnoceras baylei* Jeannet *sensu* Gil et al. (1985) MEM-AM-955- macroconch

6. *Erymnoceras baylei* Jeannet *sensu* Gil 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* Gil et al. (1985) MEM-AM-1425 - macroconch

12. *Erymnoceras baylei* Jeannet *sensu* Gil et al. (1985) MEM-AM-1425 - section

13. *Erymnoceras baylei* Jeannet *sensu* Gil et al. (1985) MEM-AM-1425 - macroconch

Legend Plate 3

1. *Euaspidoceras* TIDILI-2

2. *Euaspidoceras* TIDILI-2

3. *Euaspidoceras* TIDILI-2

4. *Euaspidoceras* TIDILI-2

5. *Euaspidoceras* TIDILI-2

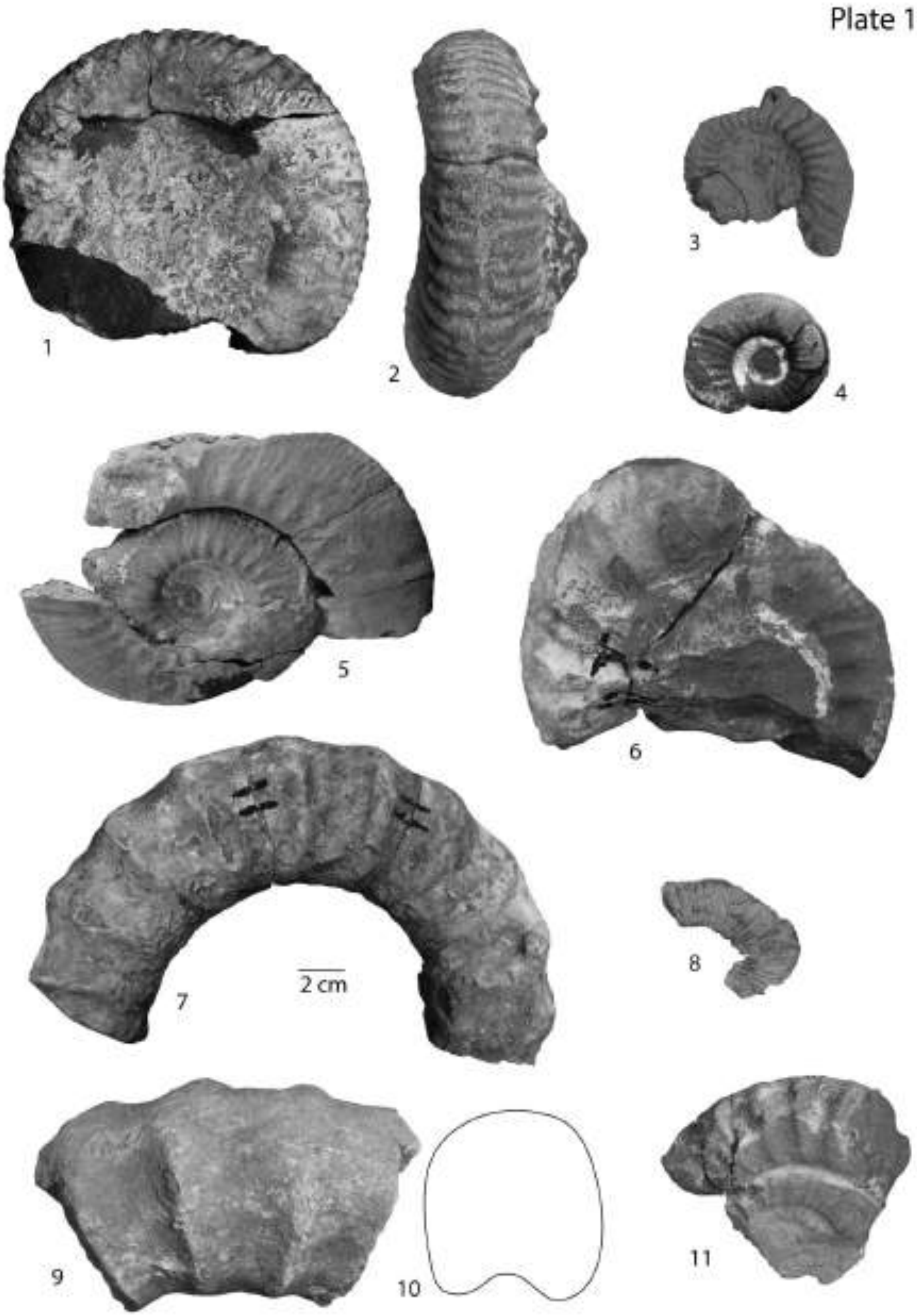
Plate 1
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Plate 2

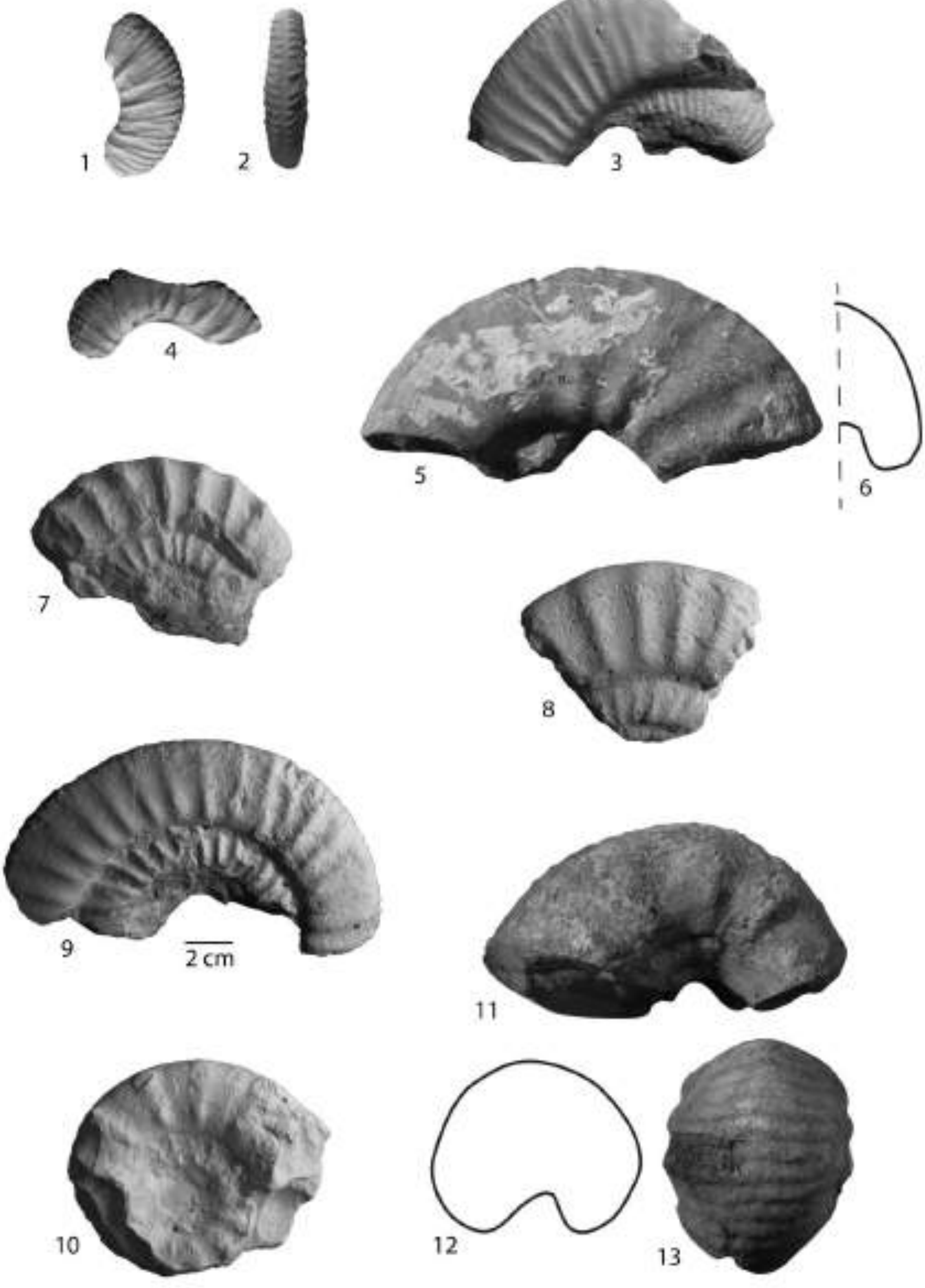
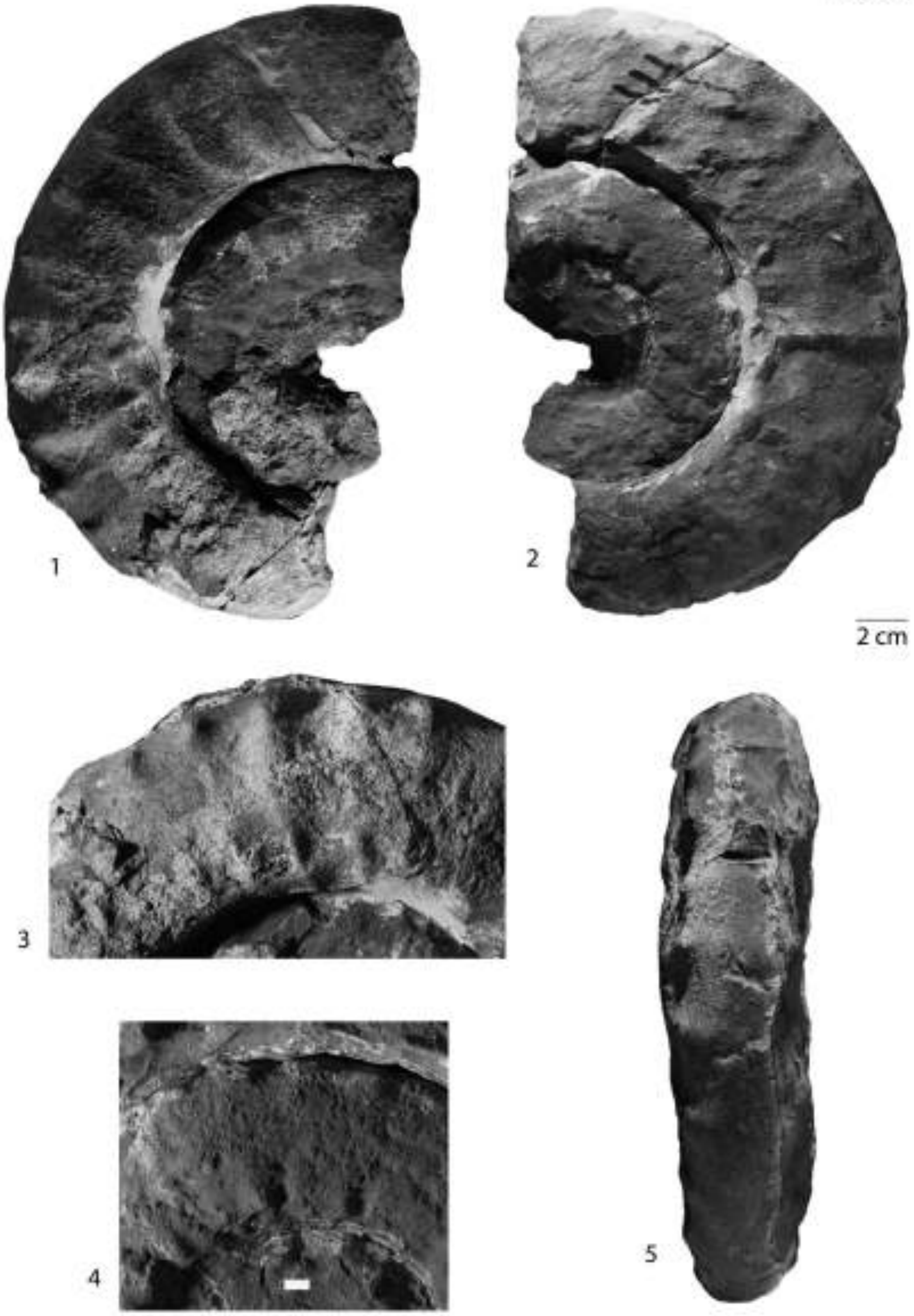


Plate 3



Chapter 6: Development and architecture of Oxfordian coral buildups along the Atlantic Margin. Essaouira-Agadir Basin, Morocco.

Abstract

The geometry, evolution and depositional setting of Oxfordian coral buildups dominated by microsolenid bioherms are documented across the Essaouira-Agadir Basin on the Atlantic Margin. The bioherms show a variety of geometries, with low-relief platy-coral colonies evolving into higher-relief more diversified coral buildups associated to coral-rubble and formation of large clinoforms on the sides of the structure. By comparison to similar environments that have beendescribed across Europe, the facies variations are interpreted to be controlled by an interplay of local changes in accommodation, sea water temperature and increase in CO_2 availability linked to the onset of the Middle Oxfordian global transgression. The coral buildups show a complex architecture, documented in a number of detailed transects. Diachronicity in the buildups is apparent, with younger coral bioherms growing in the depressions created between the initial bioherms. The zone of bioherms show a lateral transition into mid ramp deposits in the proximal part of the basin, and are interrupted by siliciclastic-rich shoreface deposits in the north of the basin.

Upper Jurassic carbonate platforms are important reservoir targets for oil and gas on the conjugate margins of Nova Scotia and Morocco. Understanding the controls on facies variability and geometry is important to get a better understanding of the Jurassic play. From a reservoir perspective, detailed correlation shows the complex lateral and vertical facies variations associated with the bioherms, each with potentially different poroperm characteristics. Dolomitisation is also controlled in part by the original facies. Dolomitised coral buildups offer the best reservoir quality. Individual bioherms are disconnected, but detailed logging suggests that the main facies found between them is made of coral-rich

floatstones and rudstones. This facies is also highly dolomitised and could potentially have similar if not better reservoir properties. The demise of the coral buildups in the EAB is linked to a regression, dated no later than Upper Oxfordian, and the establishment of peritidal carbonates across the whole basin, with continental deposition on the margin of the EAB.

This paper provides new data on the size range and lateral facies variation exhibited by the microsolenid biostromes and bioherms that can be extrapolated to the subsurface to give a better understanding of the buildup geometries in Offshore Morocco and the Scotian Basin. The reservoir potential of thebuildups is enhanced by extensive dolomitisation. Other facies including the coral-rubble deposits that flank the reefs are extensively dolomitised and porous and associated mid ramp and shoreface environments can also be considered as good potential reservoir units.

6.1 Introduction

During the Upper Jurassic, coral buildups prospered across the Tethyan and Atlantic domains (Insalaco, 1996; Kiessling and Flugel, 2002, Leinfelder at al., 2002). Tethyan microsolenid reefs have been extensively documented across Europe (Menot, 1980; Ali, 1983; Gygi, 1986; Insalaco, 1999; Chevalier et al., 2001; Lathuilière et al., 2005; Benito and Mas, 2006; Kołodziej, 2015). The environments and coral associations recorded in Europe are similar those observed in the Agadir-Essaouira Basin (EAB) and comparison is helpful to improve understanding of the bioherm environmental settings and controls. The Upper Jurassic carbonates were previously described in detail at Cap Ghir (Ourribane et al., 1999; Martin-Garin et al., 2007) and Izwarn (Olivier et al., 2012; Figure 6.1).

Upper Jurassic carbonates are proven reservoirs on the conjugate margins of Morocco (Cap Juby oil field (1984), Sidi Rhalem oil field: Morabet et al., 1998) and Nova Scotia (Deep Panuke gaz field (1998): Weissenberger et al., 2006), where only subsurface data is available. The main reservoir target in Nova-Scotia consists of Middle Oxfordian to Beriasian sponges and coral buildups. It is fundamental to understand the nature, extension and geometries of the coral buildups onshore Morocco, where they outcrop extensively , to reduce the risk of petroleum exploration offshore both margins.

This study extends the characteristion of local buildups across the onshore Essaouira-Agadir Basin to identify variation in facies, geometry and large scale architecture, and gain a better understanding on the controls of coral development and their ultimate demise. This work includes: (1) a detailed description of facies linked to the different buildups and associated deposits; (2) interpretation of the ramp facies organisation and depositional environments; (3) description of the evolution of the main framebuilders, bioherms architecture, geometries; (4) documentation of the lateral extent and spacial organisation and (5) discussion on the controls of the coral domination during the Middle Oxfordian and the demise of the buildups, and (6) a discussion on the reservoir potential of the coral-rich and bioclastic facies.

6.2 Geological setting

The WSW-ENE trending Atlas mountain chain extends from Tunisia to the Atlantic coast of Morocco. It was formed by Alpine compression of the late Palaeozoic to Lower Jurassic intracontinental Atlas rift (Choubert and Faure-Muret, 1962; Mattauer et al., 1977; Laville and Piqué, 1992; Frizon de Lamotte et al., 2000, 2009; Piqué et al., 2000; Domènech et al., 2015). The Essaouira-Agadir Basin (EAB) located on the eastern Atlantic passive margin and the conjugate margin of Nova Scotia (Pique et al., 1998), has been deformed and inverted as part of the Western High Atlas. The EAB began to form during the Late Palaeozoic rifting of the Atlantic, which resulted in the formation of N-S to NNE-SSW half grabens (Medina, 1988; Bouatmani et al., 2003). This rift-phase was followed by the formation of a sag basin during the Late Triassic (Baudon et al., 2012), and the onset of the passive margin drift phase is dated Sinemurian (Hafid, 2000; Sahabi et al., 2004). The Lower Jurassic records the first open-marine carbonate deposition along the margin during the Upper Sinemurian to Lower Pliensbachian (Duffaud, 1960; Ambroggi, 1963). This was followed by a return to fluvial conditions during the Lower Toarcian (Chap 4) and then a transgression during the Upper Toarcian, which led to development of extensive shallow-marine conditions across the basin (Ambroggi, 1963; Adams et al., 1980; Du Dresnay; 1988). The Middle Jurassic Ameskhoud Formation records continued shallow-marine carbonate deposition in the north of the EAB, but the south is marked by strong siliciclastic influx (Peybernes et al., 1987; Bouaouda, 2002).

In the Ouanamane Formation, the Bathonian to Lower Callovian record a significant transgression in the EAB and the development of a broad open ramp dominated by oolitic deposits. The following Middle Callovian deposits are dominated by brachiopod-rich floatstones, with increasing marl deposition upward in the succession. The Middle Oxfordian major transgression is marked by a transition to bioherms of the Lalla Oujja Formation (Ambroggi, 1963; Adams, 1979, 1980; Bouaouda, 2002) and followed by shallow water carbonates (Iggui El Behar Formation, Figure 6.2).

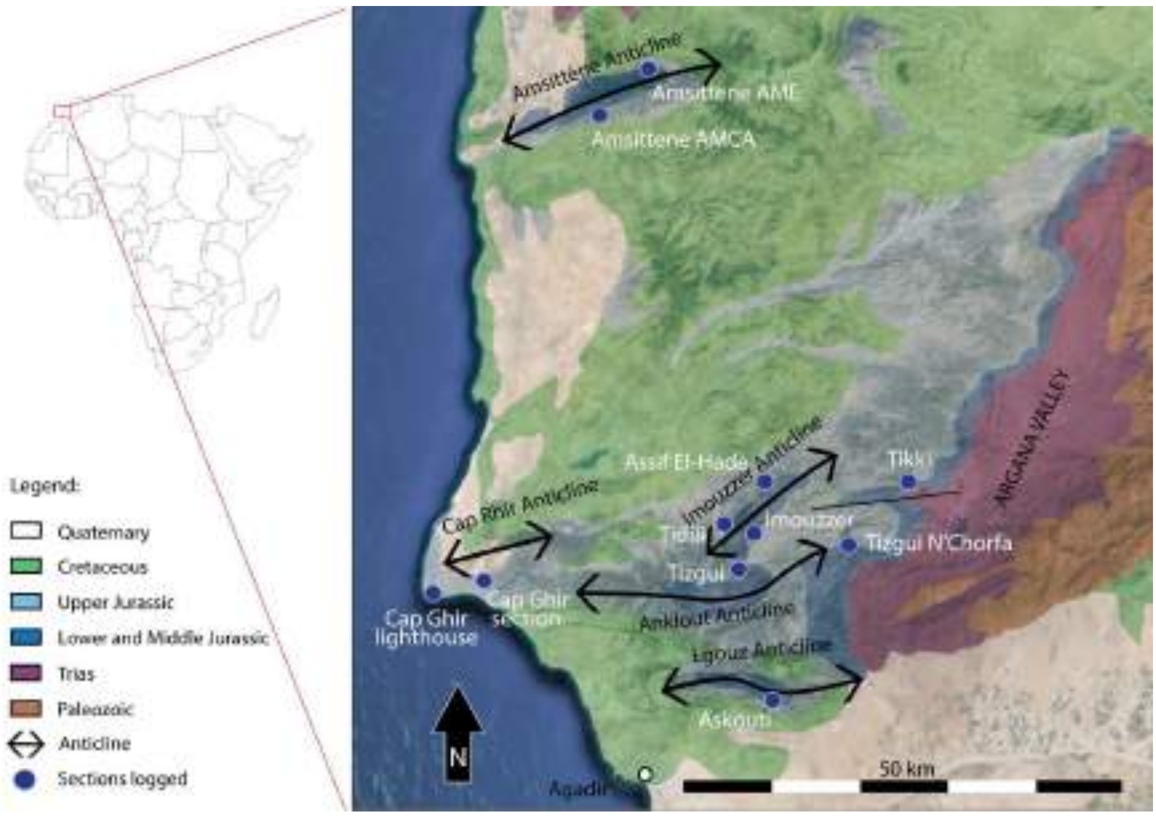


Figure 6.1: Geological map of the Essaouira-Agadir Basin with location of the sections logged for the Lalla Oujja Formation.

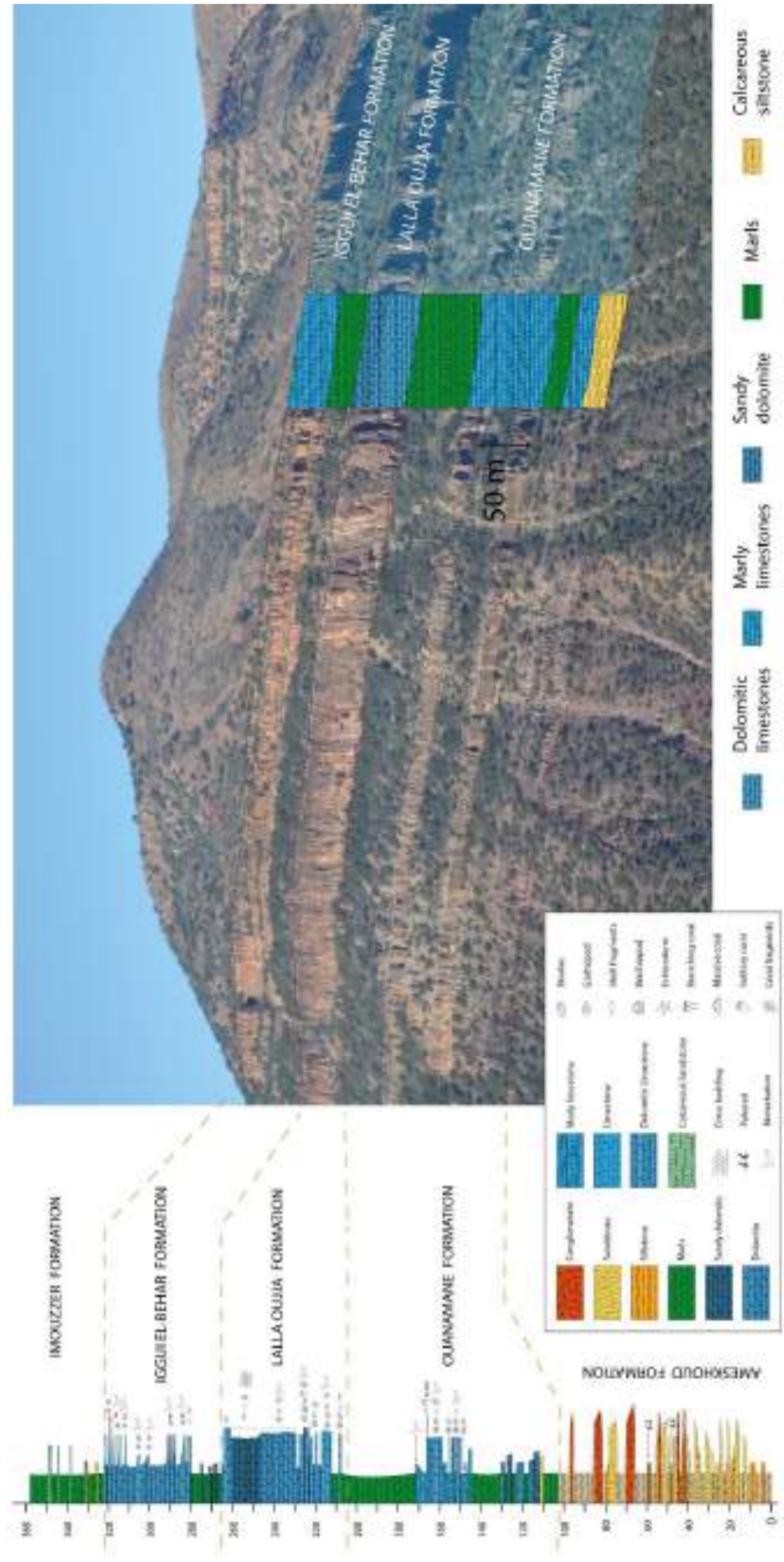


Figure 6.2: Formations overview and simplified section in the proximal part of the basin. Tizgui N'Chorfa location.

In the Central High Atlas, folding synchronous with opening of the Jurassic basins has been attributed to salt diapirism (Michard et al., 2011; Saura et al.,2014; Verges et al., 2017; Teixell et al., 2017; Malaval, 2017; Joussiaume, 2017). Onshore in the EAB, multiple NE-SW and E-W anticlines can be observed, but only the Tidzi and Jbel Hadid diapirs in the North of the Essaouira sub-basin have surface expression. Some of the anticlinal structures in the Agadir segment are also salt-cored, such as the Cap Ghir Anticline, and their offshore continuation can be observed in seismic, but many of the structures in the Western High Atlas (WHA) do not show significant evidence for salt movements and seem to have been rather formed during the orogeny of the Atlas. Onshore the EAB, the influence of salt tectonic on the development of reef structures is then restricted to the north of the basin.

6.3 Materials and methods

The data presented in this study come from 12 georeferenced sections logged at high resolution across the EAB. Six serial sections were also studied in the south of the Imouzzer Anticline (Figure 6.1); walked out and correlated to examine local-scale facies variations and buildup geometries. In addition, facies variations were mapped along a Plio-Pleistocene terrace which cut into the buildups and allowed observation of the lateral and horizontal changes in sedimentation. The main surfaces and geometries were examined and imaged using photomosaics. The outcrop-scale observations were linked to sample description and petrographic analyses conducted on 380 hand-samples and 55 thin-sections. Facies were described and classified based on lithology, biological elements, size, sorting and cement associated to macro-scale observations.

6.4 Depositional environments

The coral associations of Cap Ghir and Izwarn have been identified and ecological successions proposed by previous published work on Upper Jurassic coral buildups (Martin-Garin et al., 2007; Olivier et al., 2012). A number of Facies Associations can be defined, attributed to specific depositional environments.

6.4.1 Outer Ramp

Characteristic features

This facies association directly lies above marls and is dominated by two facies: Bioclastic wackestones to packstones (FO-1) and sponge boundstones (FO-2). The wackestones and packstones are locally slightly marly (Figure 6.3, D). They contain angular and poorly sorted bioclasts, including some complete bivalves and echinoderms. Horizons with crinoids contain complete stalks (Figure 6.3, C) and isolated columnals. Small coral fragments and rare ammonites (Chapter 5) are also visible in some horizons (Figure 6.3, B). A fraction of silt is always present (Figure 6.3, A) and bioturbation by *Thalassinoides* is frequent. In the sponge boundstones, the matrix is muddy with small bioclasts. Thrombolites and brachiopod shells are also common in this facies. The sponges are darker than the surrounding muddy matrix (Figure 6.3, E).

Table 2: Facies table for the Lalla Oujja and Iggui El Behar formations.

	Facies Name	**Main components**	**Matrix and mineralogy**	**Sedimentary features and bioturbation**
FO-1	WST/PST	Peloids 25-40% Shell fragments (bivalves)10-30% Crinoids 0-30% Brachiopods 10-25% Bivalves 5-25% Echinoderms 5-15% Silt 5-7% Foraminifers 2-5% Gastropods 0-3% Platy corals 0-2%	Micrite and euhedral dolomite crystals	Thinly bedded, 0.5 – 1.5 m thick
FO-2	Sponges BST	Sponges 20-30% Thrombolite 10-30% Shell fragments 5-10% Brachiopods 2-10% Echinodern 0-5% Silt 3-5%	Micrite	Beds 0.5-2 m thick

FO-3	Mud BST	Platy coral 10-40% Sponges 0-10% Thrombolites 0-20% Coral fragments 0-8% Echinoderm fragments (spicules dominating) 2-5% Calcispheres 1-2% Brachiopods fragments 1-2% Bryozoan 0-1% Very small shell fragments 0-1% Foraminifera <1%	Micrite Locally sponges or thrombolites encrusting platy coral	0.3-2m thick
FO-4	Dolomite	Pink and grey dolomite Coral ghosts Fractures and porosity: vugs up to 5cm large, intercrystaline porosity	Euhedral and saddle dolomite Some euhedral dolomite	Massives units up to several m
FO-5	MST/WST	Echinoderms 5-10% Brachiopods 5-8% Peloids 5-10% Bivalves 0-5% Undetermined bioclasts 0-2%	Micrite	0.1-0.8m thick
FO-6	Diverse coral BST	Corals (platy, branching, massive) 40-60% Gastropods 5-10%	FO-5 and FO-8	Massive units up to several m
FO-7	Coral-rich FST	Coral fragments (platy, branching, massive) 40-70% Echinoderm fragments 10-30% Gastropods 10-20% Shell fragments 2-15% Sponges fragments 0-5% Dacycladacean algea <1% foraminifers <1%	Blocky calcite	0.4-2 m thick
FO-8	Coral-rich GST	Coral fragments 40-60% Undifferentiated micritised grains 20-30% Echinoderm fragments 10-20% Gastropods fragments 0-5% Neogene quartz 0-2% Dacycladacean algea <1% foraminifers 0-5%	Blocky calcite	0.1-1m thick Lenticular Tabular
FO-9	Platy-coral FST	Platy-coral fragments 10-30% Branching coral fragments 0-8% Shell fragments 5-12% Brachiopods 0-5% gastropods 0-5%	Micrite	0.1-5m thick massive units
FO-10	Bioclastic WST/PST	Coral fragments 5-15% Shell fragments 5-10% Peloids 5-15% Brachiopods 0-5% gastropods 0-5%	Micrite	0.2-0.5 m thick beds

FO-11	Peloidal PST	Peloids 40-70% Sponges 7-15% Coral fragments 5-8% Shell fragments 5-10% Green algae 3-10% Foraminifera 2-10% Echinoderms 5-8% Autogenic quartz 5-7% Oncoids 0-10% Intraclasts 0-5%	Micrite and microspar	0.2-0.7 m thick beds
FO-12	Coral and nerinea FST/RST	Coral fragments 10-40% Large shell fragments (bivalves, gastropods) 20-50% In situ branching coral 0-20% Small shell fragments 20-60%	?	Horizontal bedding Alternation of horizons with different bioclast sizes
FO-13	Coral and shell FST/GST	Coral fragments 10-60% Large shell fragments (bivalves, gastropods) 10-40% Small shell fragments 20-60% Peloids 10-20% Echinoderm 5-10% Foraminifera 2-5% Sponges fragments 0-7%		Horizontal bedding Hummocky and Swaley cross-stratification Alternation of horizons with different bioclast sizes Trace-fossil: Skolithos and Conichnus Sets 0.5-2m thick
FO-14	Bioclastic sandstone	Medium to coarse quartz grains 40-60% (coated and non-coated) Gastropds 0-40% Coral fragments 5-30% Shell fragments 5-20% Isolated small massive coral 2%	Calcite cement; Locally only quartz grains and euhedral dolomite crystals	Swaley cross-stratification, horizontal bedding, through cross-stratification
FO-15	Megalodont BST	Megalodont 40% Peloids 30-45% Foraminifers 8-10% Small massive and branching coral colonies 5-10% Gastropods 5-10% Coral fragments 5-12%	Micrite; PST peloidal matrix	0.8-1 m thick beds Massive

FO-16	Nerinea and branching coral FST/RST	Branching coral 5-15% Branching coral fragments 10-35% Peloids 25-40% Neerinea 10-30% Shell fragments 25-30% Chaetetids 5-8% Bivalves 2-10% Brachiopods 5-8% Echinoderms 2-5% Sponges fragments 0-3%	Micrite, PST peloidal matrix	Massive Beds 2-4m thick
FO-17	Peloidal WST/PST	Peloids 10-30% Intraclasts (200-1500 µm Ø) 5-20% Foraminifers benthic and (3-4%) planktonic (0-1%) Ooids 0-7% Gastropods 0-3% Large oncoids (1000-1500 µm Ø) 3% Shell fragments 0-10% Dacyclad 0-2% Echinoderm 0-2% Coral fragment 0-1% Autogenic quartz 0-5% Ostracods <1%	Micrite and cement (drusy and blocky)	Massive Beds 0.2-0.8 m thick
FO-18	Mudstones	Ostracod (0-5%) Echinoderm fragment <1% Shell fragment <1% (50 µm Ø)	micrite	Massive Beds 0.4-1m thick
FO-19	Foraminifers WST/PST	Foraminifers 6-20% Intraclasts 5-10% Shell fragments 0-2%	micrite	Massive Beds 0.2-1m thick
FO-20	Rip-up clasts horizons	Rip-up clasts 10-70% Foraminifers 0-15% shell fragments 0-5% Echinoderms 0-1%	micrite	Horizontal and wavy bedding Beds 0.5-1.5m thick
FO-21	Oncoidal FST/RST	Oncoids 30-70% Foraminifers 0-5%	micrite	Massive Beds 0.4-1.5m thick
FO-22	Gypsum MST	gypsum crystals replaced by microspar 5-15% Ostracod 0-2% shell fragments <1%	micrite	Massive Beds 0.1-0.5 m thick
FO-23	Stromatolites	peloids 0-30%	micrite	subhorizontal micrite horizons
FO-24	Charophytes WST	Charophytes Ostracods	micrite	Thinly bedded

Depositional environment

Common peloids and the micrite matrix in the wackestones to packstones facies indicate a low energy environment. The bioclasts, made of angular shell fragments and echinoderm plates, and an association with complete brachiopods and bivalves, indicates a low level of reworking. Crinoids are suspension feeders and thus adapted to deep, low energy environments (Meyer and Macurda, 1977; Roux et al., 1988). The presence of ammonites also indicates open marine conditions. Calcareous sponges were common deeper-water bioherm builders during the Jurassic (Flugel and Steiqer, 1981; Crevello and Harris, 1984; Insalaco, 1996; Olivier and Boyet, 2006). This association of fauna indicates a relatively deep water environment, not significantly reworked by local currents, interpreted as an outer ramp setting.

6.4.2 Microsolenid bioherm (FO-3, FO-4, FO-5)

Characteristic features

The Microsolenoid bioherm facies association comprises three facies (FO-3 to FO-5). It is dominated by platy-corals in a mudstone matrix (FO-3), extensively replaced by dolomite (FO-4) and locally passing to mudstones and wackestones (FO-5). Where locally this facies is extremely dolomitised (FO-4), platy-coral colonies are only recognized as recrystallized horizontal features (Figure 6.4, B).

Locally, the platy-shaped corals are encrusted by microbialites or sponges, and in Cap Ghir and Izwarn, the development of a microbialitic crust on top of *the* platy-corals is particularly common (Figure 6.4, A).

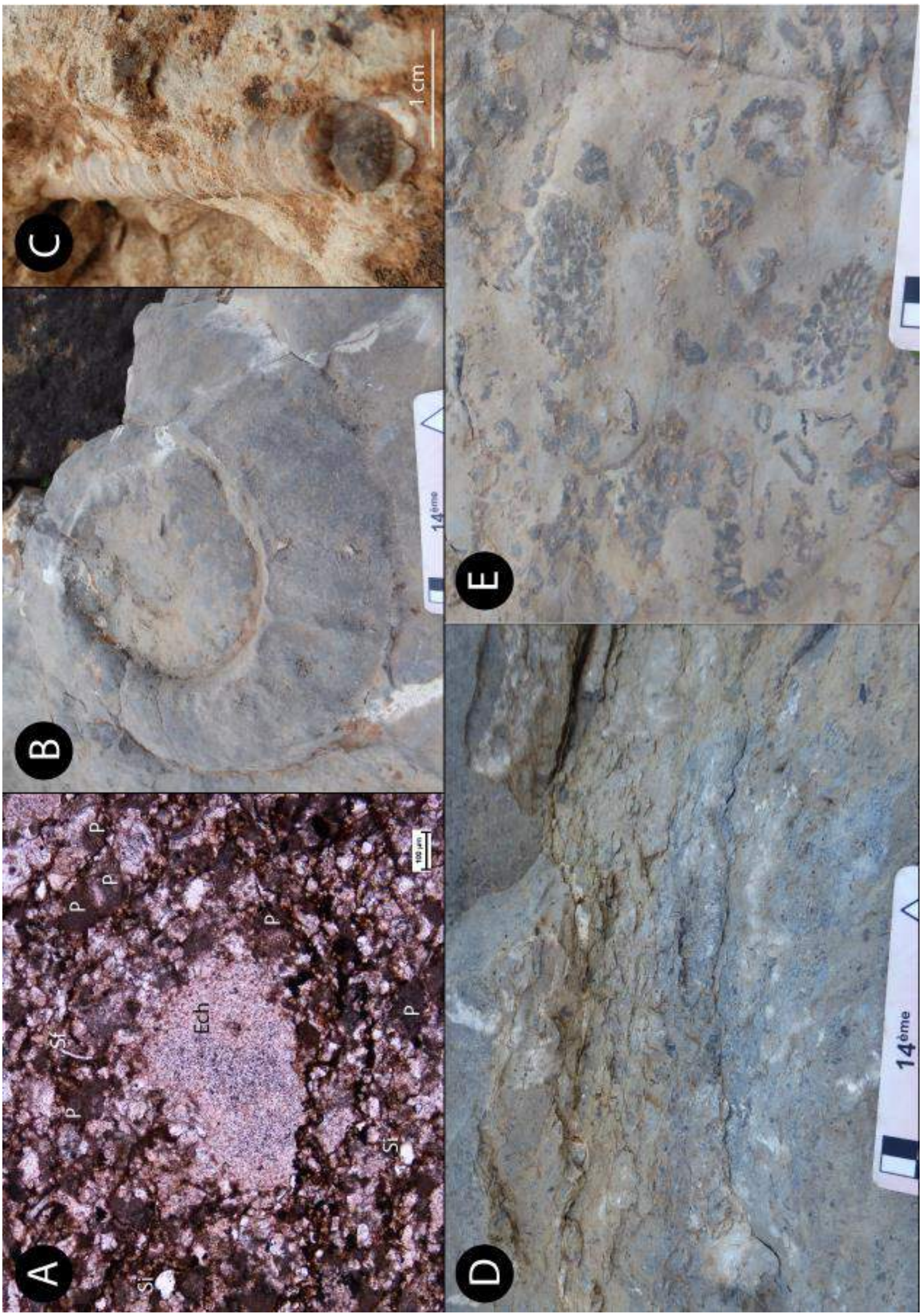


Figure 6.3: Outer ramp facies from Tidili location.

A – Peloidal and bioclastic packstone, partially dolomitised, with silt (Si), peloids (P), shell fragment (Sf) and echinoderm plate (Ech). **B** – Ammonite. **C** – Crinoid stalk. **D** – marly bioclastic horizon. **E** – Sponge boundstone.

Figure 6.4: Macrofacies of the Buildups of the Lalla Oujja Formation 1/2.
A - FO-3 *Dimorpharea* (black arrow) encrusted by microbialites (white arrow): Imouzzer; **B** - FO-4 Dolostone, black arrows pointing to phantoms of *Dimorpharea*: Tidili; **C** – FO-8 Thick branching coral: Imouzzer; **D** - FO-4 Dolostone with delicate branching coral: Assif El Hade.

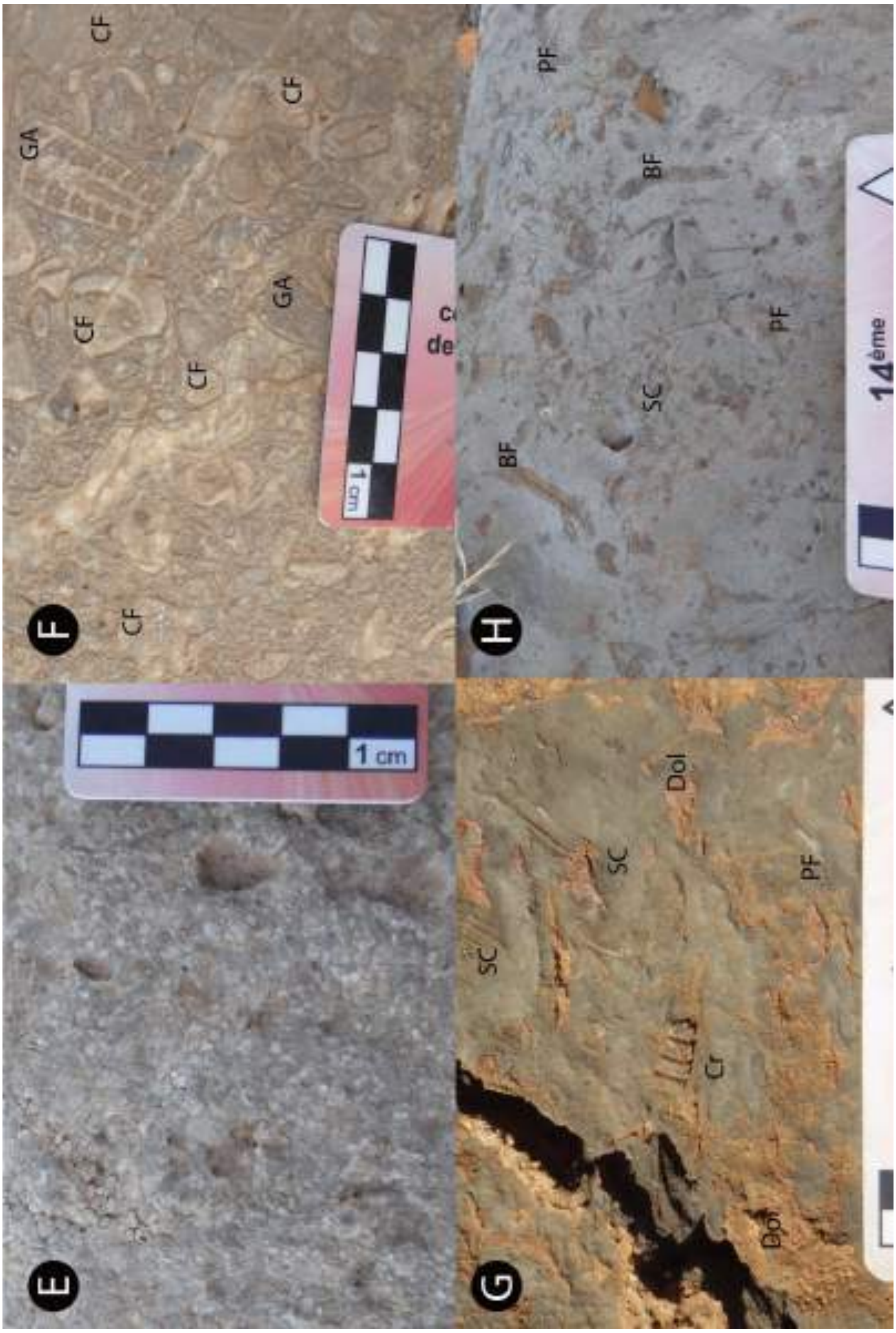


Figure 6.5: Macrofacies of the Buildups of the Lalla Oujja Formation 2/2.
E - FO-7 Coral-rich grainstone: Cap Ghir; F – Coral-rich floatstone, Nerinea (GA) and coral fragments (CF): Cap Ghir; G – FO-Coral-rich floatstone with fragments of crinoids (Cr), Platy-coral (PF), Solitary coral (SC) and partially dolomitised (Dol): Tidili; H - Coral-rich floatstone dominated by branching coral (BF), solitary coral (SC) and platy-coral (PF): Tidili.

Depositional environment

The scarcity of bioclasts in this facies association and the dominance of mud indicate a very low energy environment below fair-weather wave base. The dominance of microsolenid platy-coral *Dimorpharaea* is also an important environmental indicator. Insalaco (1996) established that microsolenids are generally subordinate to other corals if the light conditions are elevated. He recognised a growth window for microsolenid biostromes situated within the lower euphotic zone, below the highly diverse shallow-water coral communities. High turbidity water, resulting in suspension of mud in the water column, could also have influenced the fauna, by reducing the penetration of light in the environment, and thus favoured *Dimorpharea* growth over other corals (Hallock and Schlager, 1986; Insalaco, 1996). *Dimorpharea* colonies present a phenotypic plasticity and can modulate their morphology in response to their environment; and develop a platy shape from an estimated 30 to 70 m water depth (Lathuilière et al., 2005).

Encrustation by microbialites, both leolite and thrombolite (Riding, 2011), suggests a relatively low sedimentation rate, which allowed the microbial structures to grow on the pre-existing coral framework.

6.4.3 Higher diversity buildups (FO-4, FO-6, FO-7 and FO-8)

Characteristic features

This facies association contains a diverse coral assemblage that has been previously detailed in the Cap Ghir (Martin-Garin et al., 2007) and Izwarn (Olivier et al., 2012) localities. It is composed of coral boundstones, locally at Cap Ghir associated with coral-rich floatstones and grainstones (FO-6, FO-7 and FO-8) and extensively replaced by dolomite (FO-4). Facies FO-6 contains a wide variety of corals, and can be recognised across the basin, dominated by thin and thick branching, solitary and massive corals. These coral colonies differ in species diversity and dominance from one location to another. For example in Cap Ghir, 27 coral genera have been identified (Martin-Garin et al., 2007), while only 16 were identified in the Izwarn location (Olivier et al., 2012).

Facies FO-7 and FO-8 were observed in Cap Ghir only, at the top of the main reef unit (Figure 6.14). The coral-rich grainstones (FO-8) contain a very high proportion of coral material, up to 60%. This facies also has well-sorted rounded coral fragments, organised in lenticular beds up to 10m wide and 50 cm thick. The coral-rich floatstones (FO-7) are composed of coral debris associated with common gastropods and echinoderms, distributed on the sides of the main reef. The ground-mass of this facies is made of coral-rich grainstone (FO-8). The buildups with more diverse genera are generally very dolomitised (FO-4), and locally can only be identified by the differential dolomitisation between corals and the matrix. The corals tend to be recrystallized into large white crystals, the matrix is grey or pink with smaller dolomite crystals. These highly dolomitised facies contain large vugs (>5 cm) as well as visible intra- and inter-crystalline porosity.

Depositional environment

The diverse coral association is always found lying directly on top of microsolenid biostromes, and marked by an increase in branching coral colonies. This facies association comprises abundant thick branching ramose coral and a higher percentage of bioclastic material is interpreted to have developed in relatively high energy settings. The diverse and large colonies, and the reduction of microsolenids, suggest efficient coral growth where light was not a limiting factor. The growth of these colonies depends on higher water agitation, higher sedimentation rate and increased light than that which prevailed during deposition of the microsolenids colonies (Chappell, 1980; Insalaco, 1996; Insalaco et al., 1997; Lathuilière et al., 2005; Olivier et al., 2012).

The lenticular grainstones, dominated by coral bioclasts, also indicate high-energy currents, reworking the crest of the buildup. They have been interpreted as coral-debris filling shallow channels that developed on top of the buildup. The floatstones are also dominated by coral fragments and contain other species living on the buildup (gastropods and echinoderms). These are interpreted as debris-sheets, resulting from storm erosion of an exposed part of the buildup (Insalaco et al., 1997). The energy level of the Cap Ghir bioherm was potentially higher than similar facies seen in the other locations across the basin, suggesting it was deposited in a shallower environment.

This association has been interpreted as being formed above fair weather wave base (Lathuilière et al., 2005; Olivier et al., 2012).

6.4.4 Coral buildup rubble (FO-4, FO-9, FO-10, FO-11)

Characteristic features

This facies association is composed of coral-rich floatstone (FO-9), wackestones (FO-10), peloidal packstones (FO-11) and dolomite horizons (FO-4). Platy corals and solitary corals form the main types of coral fragments in the facies FO-9. Rare fragments of branching colonies are also present. The ground-mass of these platy-coral floatstones is generally wackestone, locally packstone, comprising smaller well-rounded fragments of coral, echinoderm, bivalves and brachiopods. These bioclasts have a micrite matrix. Many horizons of this facies are highly dolomitised. Large vugs and fractures filled with calcite are present as well as some intra- and inter-crystalline porosity. In the locality of Tidili, this facies is present within clinoforms prograding off a microsolenid biostrom.

Depositional environment

The large amount of coral fragments in this facies association indicates its proximity to the coral factory. The roundness of the bioclasts indicates that they have been transported from a higher energy environment and the micritic matrix suggests resedimention into a lower energy environment. The identification of clinoform geobodies suggests that these sediments are transported directly from the top of the bioherm. Floatstones are more common towards the topsets of the clinoforms, and the wackestones and packstones towards the bottomsets. These features indicate that some currents reworked the top of the bioherm and that the more unconsolidated elements were shed to the side of the buildup along with the sediments brought by the currents. Subsequently sponges and oncoids locally encrusted the deposited sediments. The high amount of mud in the system indicates that reworking by currents was not very frequent. This environment has been interpreted as being just below the storm wave base, in an outer ramp setting (Burchette and Wright, 1992; Wright and Burchette, 1998; Flügel, 2010).

6.4.5 Midramp

Characteristic features

In the eastern part of the basin, the bioherms are replaced by coral fragment and *Nerinea* floatstones and rudstones, containing isolated branching coral colonies (FO-12). This facies is organised in horizontal beds, separated by surfaces that are highly bioturbated by *Thalassinoides*. Some horizons show stratabound dolomitisation. The amount of bioclastic material varies laterally and vertically, and bioclasts are generally horizontally aligned. The Nerinea are mainly complete. In some horizons colonies of delicate branching corals are also found in situ, and some rare colonies of massive corals can be observed.

Depositional environment

This environment is marked by alternation of high and lower energy conditions. The horizontal alignment of the bioclasts in the rudstones beds indicates currents, while the horizons with delicate branching corals and intense bioturbation by *Thalassinoides* indicate lower energy conditions. These deposits records periods of turbulence, interpreted as being periodically reworking by storm waves, alternating with quieter periods. These features are characteristic of middle muddy ramp deposits (Christ et al., 2012).

Figure 6.6: Midramp deposits.
A – Gastropod floatstones; gastropods horizontally oriented. B – Overview of one unit: base dolomitised with branching corals, followed by bioclastic material, upper part bioturbated by *Thalassinoides*. C – Branching coral dolomitised followed by intense *Thalassinoides* bioturbation.

6.4.6 Inner ramp

Characteristic features

Two variations of this environment have been identified across the basin (FO-15 and FO-16). Both contain a peloidal matrix, with abundant shell fragments (bivalves and brachiopods) and foraminifers. The difference between them is based the proportion and type of contained macrofossils. In the western part of the basin, in the locality of Cap Ghir, facies FO-15 dominates. It is composed of large shells of *Megalodont*, with *Nerinea* and isolated branching and massive coral colonies. In the Eastern part of the basin, in

Imouzzer, this facies is dominated by smaller coral colonies and tabular chaetetid associated with gastropods and bivalves (FO-16). The two different facies variations both lack distinctive sedimentary features.

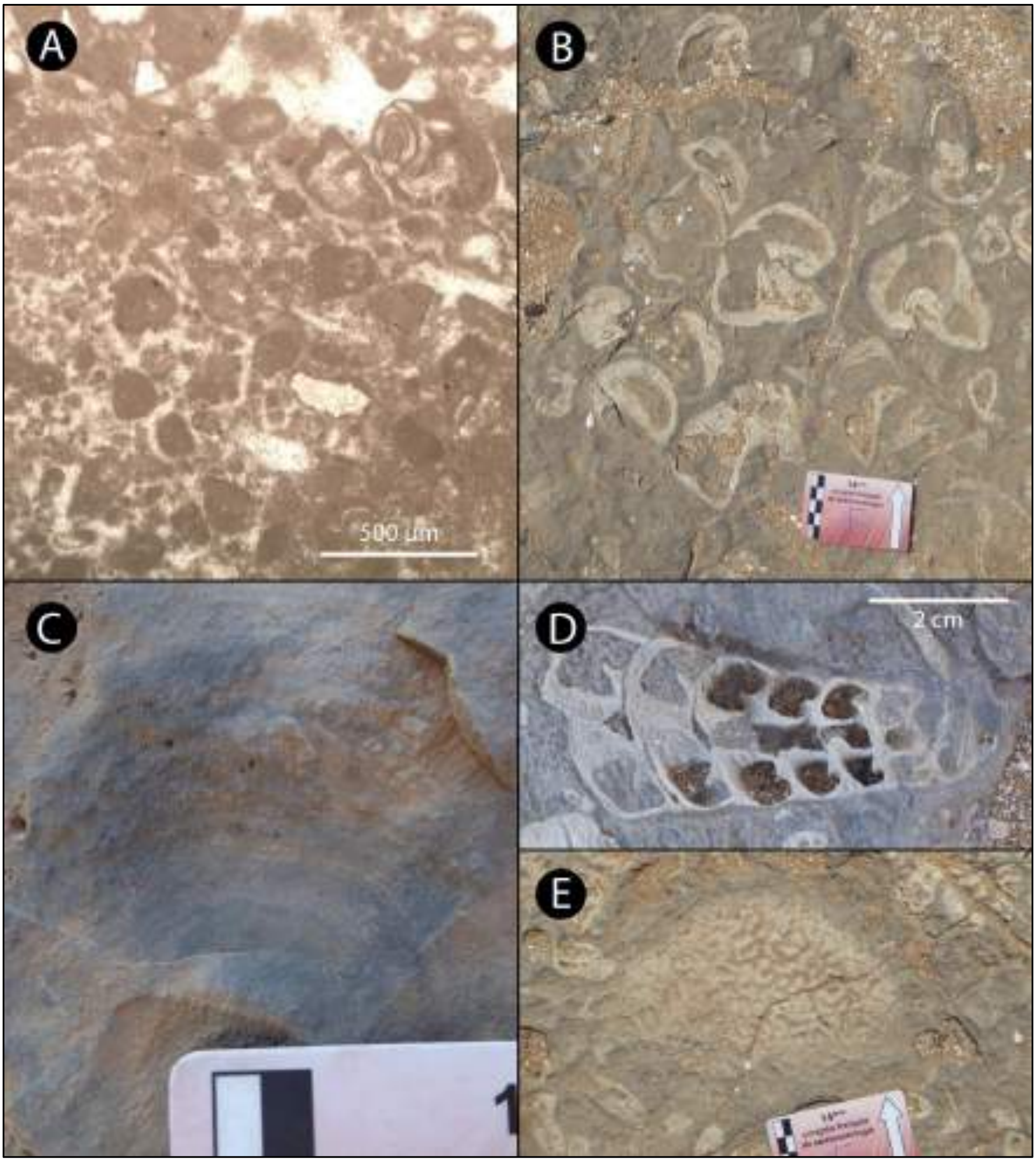


Figure 6.7: Inner ramp facies.
A – Peloidal packstone ground mass of the facies FO-15 and FO-16. B – Megalodont bivalves. C – Chaetetid. D – Nerinea gastropod. E- Coral colony in the Megalodont facies.

Depositional environment

The peloidal dominated packstones that contain a large amount of *Nerinea* gastropods indicates a good oxygenation and nutrient rates (Flügel, 2010) in both facies, associated to a quite low energy environment. The presence of similar coral colonies in both environments indicates equivalent environmental conditions. The presence of tabular colonies of chaetetids in FO-14 indicates shallow subtidal to intertidal (West and Kershaw, 2012). Megalodont are usually associated to shallow, peritidal waters (Todaro et al., 2012; Yümün et al, 2013). These two facies formed in similar environments, interpreted as inner ramp (subtidal to intertidal).

6.4.7 Foreshore to upper shoreface siliciclastics

Characteristic features

These deposits are siliciclastic rich (FO-14) in the north and east of the basin, and dominated by bioclastic grainstones and floatstones (FO-13) all across the basin. The more siliciclastic rich horizons contain up to 60% of medium to coarse quartz grains, partially coated by micrite, and few quartzite granules. Some of these horizons have been dolomitised and present anhedral dolomite crystals and intercrystalline porosity. The more fossiliferous horizons are grainstones rich in gastropods, coral fragments, rounded shell fragments and peloids (FO-13). These units locally present horizons with larger bioclastic material (Figure 6.8, D), essentially gastropods and coral fragments. Bioturbation is visible in this facies association, including *Skolithos* and *Conichnus* (Figure 6.8, B). The sedimentary features in these beds are horizontal bedding, trough cross-bedding, and some humocky cross-stratification and more recurrent swaley cross-stratifications (Figure 6.8, A) are present. Rare insitu small colonies of massive corals are also present (Figure 6.8, C). At the top of this facies association, gastropod floatstones or rudstones (Figure 6.8, E) are common, and locally followed by charophytes-rich marls and wackestones (Figure 6.8, F; Figure 6.12).

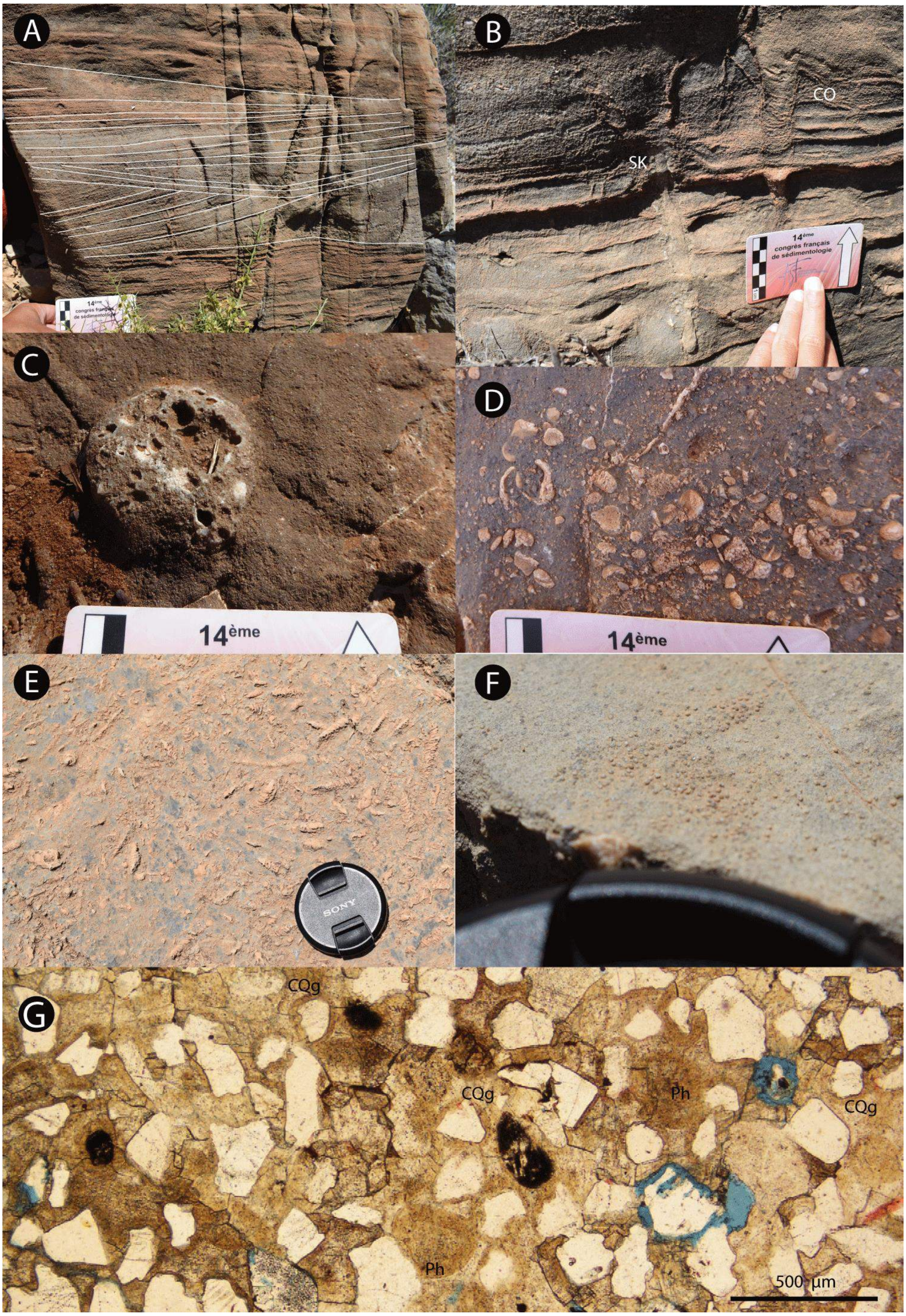


Figure 6.8: Facies and sedimentary structures of the upper shoreface.
A – Swaley cross stratification (Tikki). **B** – *Skolithos* (SK) and Conicnus? (CO) trace fossil. **C** – Small coral in sandstone. **D** – Coral, gastropods and bivalves floatstone to rudstone. **E** – Gastropod rudstone. **F** – Charophytes. **G** – Dolomitic sandstone, coated quartz grains (CQg), relics of peloids or ooids (Ph).

Depositional environment

This environment of facies dominated by grainstones and floatstones, without any carbonate mud indicates high energy settings, the mostly well rounded elements and coated quartz grains indicate some reworking. The presence of echinoderm fragments is significative of open marine conditions. The association of *Skolithos* and *Conichnus* trace-fossil is indicative of upper shoreface environment (Gerard and Bromley, 2008). The hummocky and swaley cross-stratifications are formed under storm-waves actions in the upper shoreface (Leckie and Walker, 1982; Dumas and Arnott, 2006). The gastropods and coral fragments rich horizons could also be linked to stronger energy events bringing elements from the midramp settings. This association of facies has been interpreted as upper shoreface deposits with local influx of siliciclastic.

6.4.8 Mudflat = Iggui El Behar Fm. (FO-17, FO-18, FO-19, FO-20, FO-21, FO-22, FO-23)

Characteristic features

This facies association, which typically forms the upper part of the carbonate succession, is dominated by 6 facies. It consist of a thinly bedded succession of peloidal packstones (FO-17), mudstones (FO-18), foraminifer-rich wackestones (FO-19), rip-up clast horizons (FO-20), oncoidal floatstones to rudstones (FO-21) and gypsum mudstones (FO-22). Recurrent horizons of rip-up clast "conglomerate" graded beds are abundant (Figure 6.9, C and D; Figure 6.10). Horizontal bioturbation, including *Thalassinoides*, *Rhizocorallium* (Figure 6.9, E) and *Cylindrichnus* (Figure 6.9, C) trace fossils is common. Oncoidal floatstones and rudstones are common in the lower part of this succession. Two types of oncoids have been identified. Large (up to 4 cm) porostromate oncoids (Figure 6.9, A), mostly forming rudstone horizons and encrusting gastropods and coral fragments, are always found at the base of the formation. Smaller (0.2 to 1 cm) oncoids (Figure 6.10, E) are present further up in the succession and occur in a wackestone matrix along with foraminifera and intraclasts. Horizons of foraminifer-rich (*Pseudocyclammina lituus* (Figure 6.10, B), *?Pseudocyclammina sphaeroidalis* (Figure 6.10, C), *Alveosepta jaccardi* (Figure 6.10, D), *Nautiloculina oolithica, Ammobaculites* sp.) wackestones and

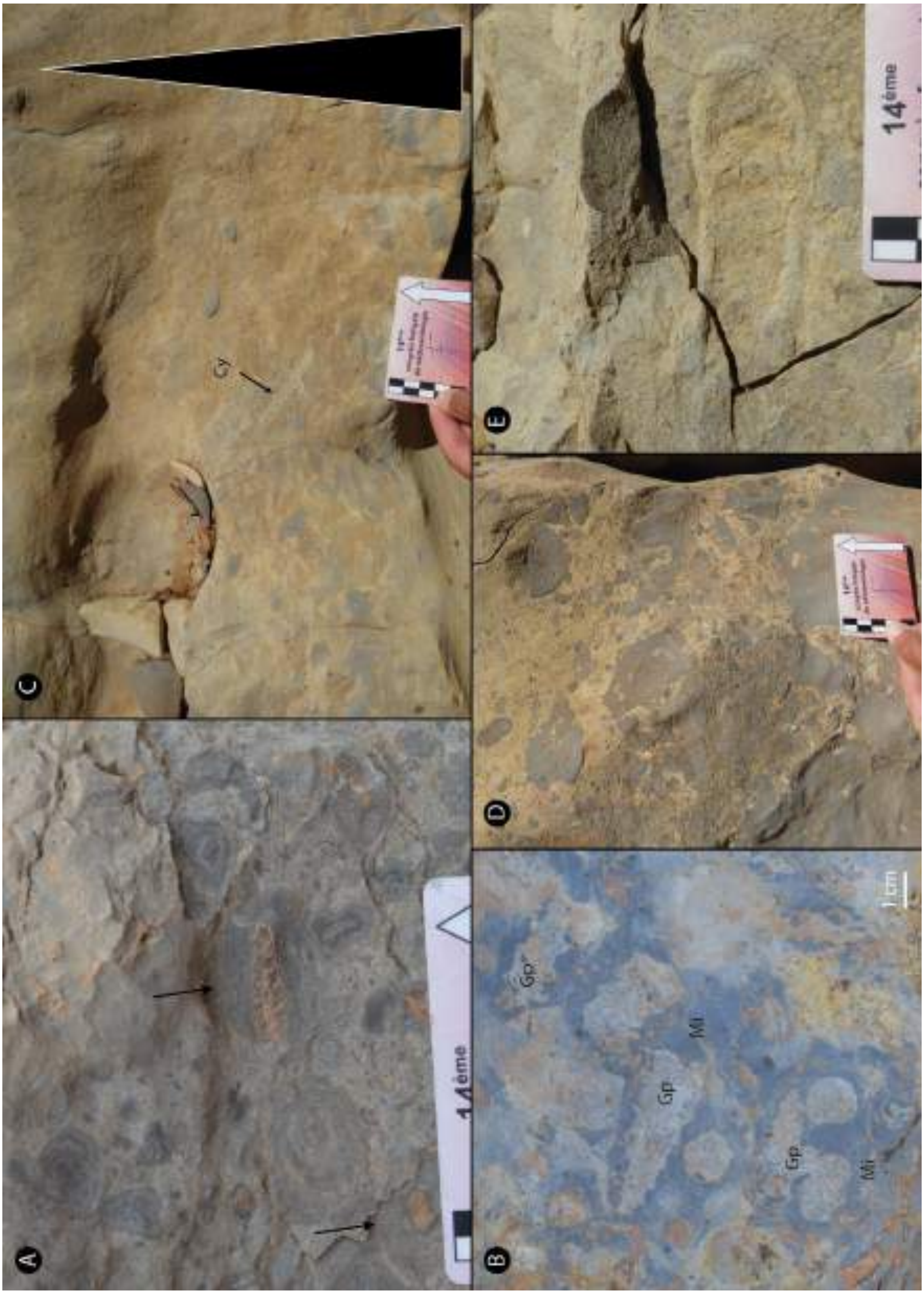


Figure 6.9: Facies of the Iggui El Behar Formation.
A – Rudstone of large porostromate oncoids, gastropod and a coral fragment as nuclei (arrows). **B** – Gastropod molds (Gp) in dark micrite (Mi). **C**, **D** – Rip-up clast horizons fining upward. *Cylindrichnus* trace fossil (Cy). **E** – *Rhizocoralium* trace fossil.

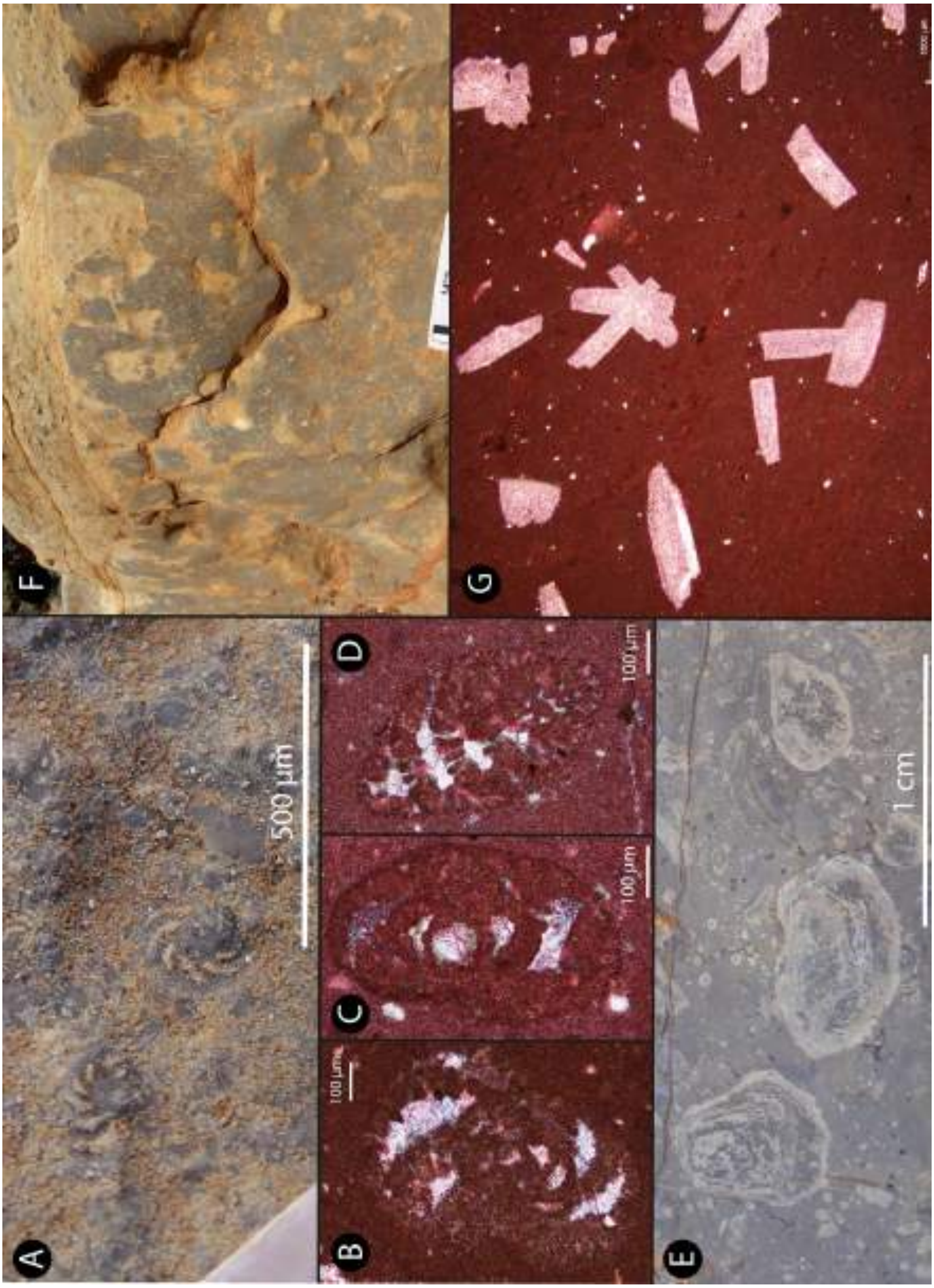


Figure 6.10: Facies of the Iggui El Behar Formation.
A – Foraminifera wackestone. **B** – *Pseudocyclamina lituus*. **C** – *Pseudocyclamina sphaeroidalis*. **D** – *Alveosepta jaccardi*. **E** – Oncoid floatstone. **F** – Microkarst surface in mudstone with gypsum molds. **G** – Gypsum moulds in mudstone.

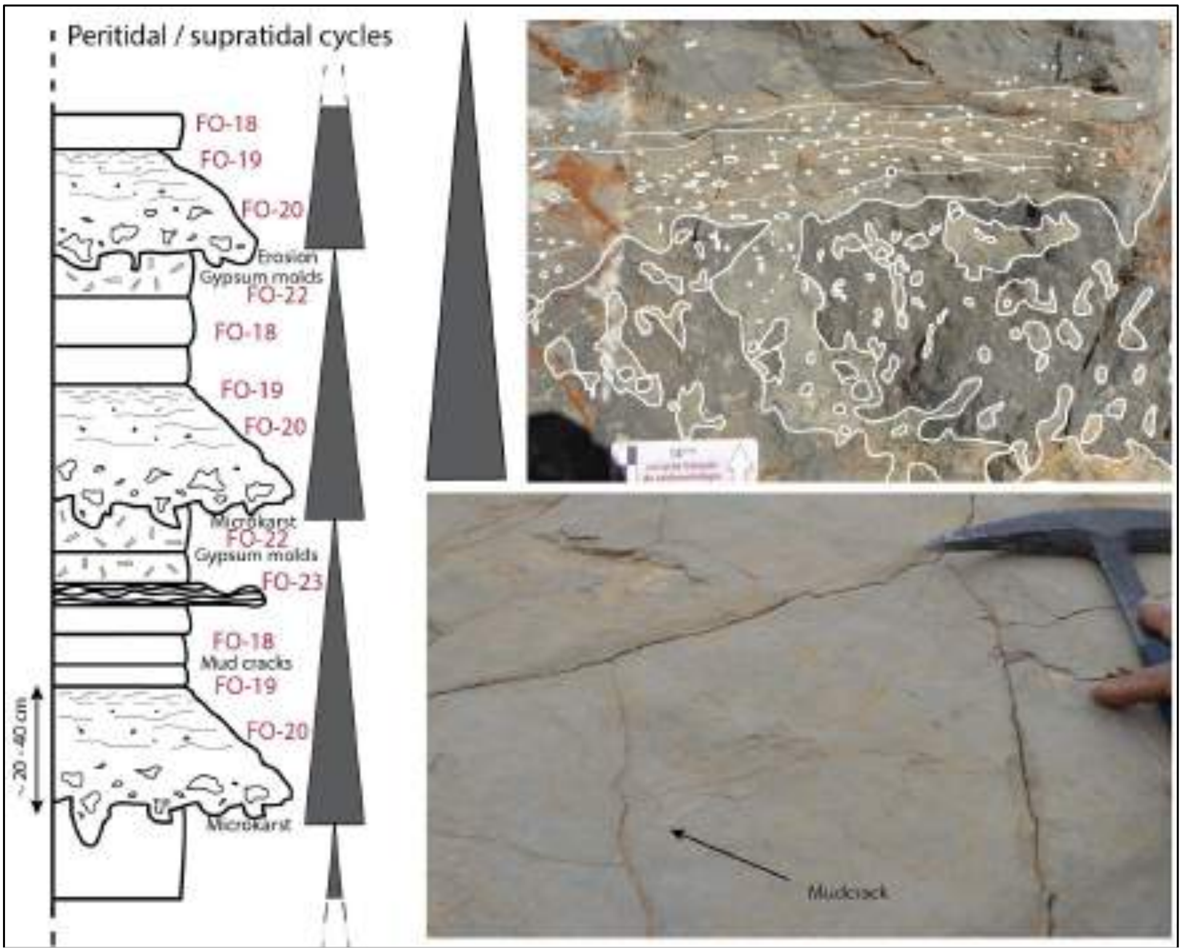


Figure 6.11: Peritidal to supratidal cycles in the upper Iggui El Behar Formation.

peloidal wackestones to packstones are very common and associated with a low-diversity fauna of scarce shell fragments, gastropods and ostracods, dacycladacea and ooids. The rip-up clasts horizons are normally graded with discontinuous wavy bedding towards the top and are commonly followed by foraminifer-rich wackestones and/or mudstones (Figure 6.11). The mudstones contain mud cracks (Figure 6.11) or are topped by highly bioturbated horizons, micro-karsts (Figure 6.10, F) or erosive surfaces linked to the next rip-up mudclast horizon. Towards the top of the formation, the oncoids disappear, some stromatolitic horizons appear, and gypsum moulds are visible in the mudstones.

Depositional environment

The presence of large oncoids in the lower part of the succession indicates open marine to semi-restricted environments with low energy (Gradzinski et al., 2004; Flugel, 2010). The association of *Rhizocorallium* and *Cylindrichnus* trace fossils in a muddy matrix also indicates a protected environment (Gerard and Bromley, 2008). The presence of rare coral fragments and dasycladacea, and abundant benthic foraminifera indicates aslo supports a shallow marine environment. *Pseudocyclammina lituus* are indicators for inner shelf, and give an age no older than Late Oxfordian, and *Pseudocyclammina sphaeroidalis*, if present, is typical of the Kimmeridgian in Morocco (Maync, 1967). The micro-karsts and the mudcracks indicate recurrent periods of emersion, the gypsum moulds indicate hypersalinity due to evaporation; features characteristic of supratidal mud flat conditions (Young et al. 1972; Demicco and Hardie 1994). The presence of stromatolites and evaporitic mudstones are also indicative of a low energy, restricted environment with frequent evaporation occurring and characteristic of higher tidal to supratidal flat environments (Park, 1977, Tucker, 2002). The rip-up clasts horizons have already been studied by Ager (1974), who interpreted them as storm deposits. Although the rip-up clasts are formed by a change in the water energy, they can be linked to shallow-water cycles and do not require storms to develop. This suggests that they are likely to have formed by tidal transgressions over very shallow to exposed units, indeed, these horizons often follow microkarsts, gypsum muds, mudcracks or very bioturbated horizons and rework the material of the beds they cover (Figure 6.11).

The facies associated following the rip-up clast units indicate an overall evolution from shallow (*Pseudocyclamina,* stromatolites) to restricted (gypsum mud) to emergent (microkarsts, mud-cracks) conditions. These shallowing upward successions are interrupted by the deposition of erosive or non rip-up clasts conglomerate horizons which mark the beginning of the next shallowing-upward cycle. The reworking that forms these horizons can be due to the currents formed by periodic fall in relative sea level in a supratidal mudflat (Young et al. 1972; Kwon et al., 2002). The already lithified underlying deposits are likely to have been locally fragmented prior to reworking by waves or currents (Chen et al., 2008). Different factors can be important processes in play that create fragmentation, such as bioturbation (Brodzikowski & van Loon, 1985), desication or karstification. Therefore, this facies association has been interpreted as an overall shallowing-upward succession, from subtidal at the base to intertidal to supratidal mud flat on the upper part.

In the eastern part of the basin, in the localities of Tikki and Tizgui N'Chorfa, the transition between the Lalla Oujja Formation and the Iggui El Behar Formation is observed to be a change to mudstones and blue and grey marls with ostracods and charophyte gyrogonites (Figure 6.12). These organisms can be found in a restricted or non-marine environment. These shallower conditions, potentially continental, highlight the sharp regression and the shoreline shift which mark the transition between the two formations. The upper part of The Iggui El Behar Formation becoming later dominated by thick tidal to supratidal mudflat deposits across the entire basin, which reflects similar subsidence rates in the center and on the margins of the basin.

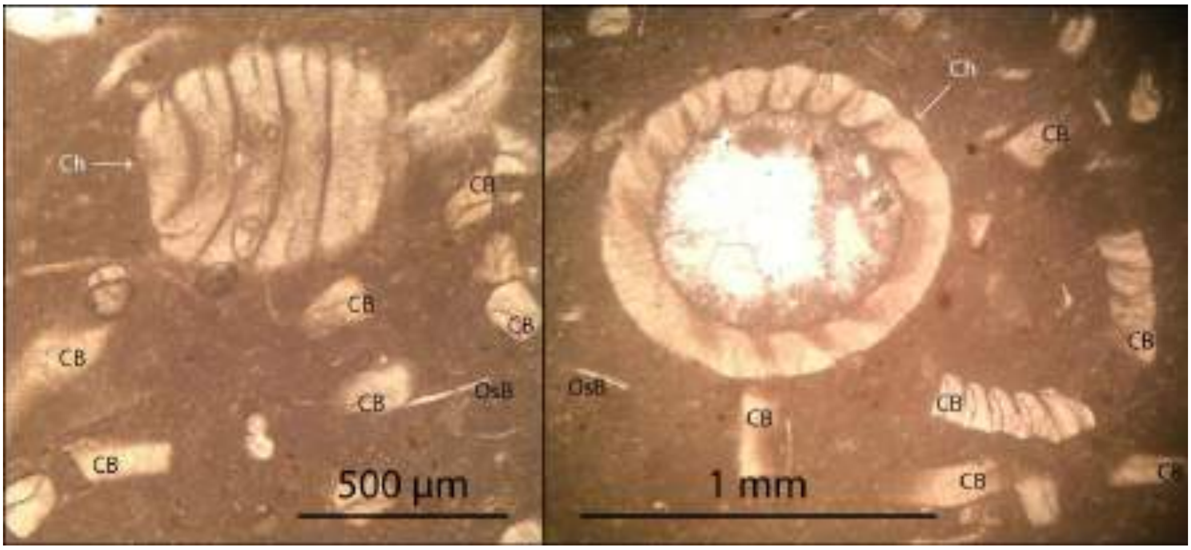


Figure 6.12: Charophyte gyrogonites, Tizgui N'Chorfa location. Charophytes gyrogonites (Ch), fragments charophyte gyrogonites (CB) and fragments of ostracods (OsB).

6.4.9 Depositional model

From the observation of the facies and their interactions with each other, a depositional model has been constructed (Figure 4.9: Example of lenticular breccia bed and associated syn-sedimentary ductile folding in the Tamarout Formation.).

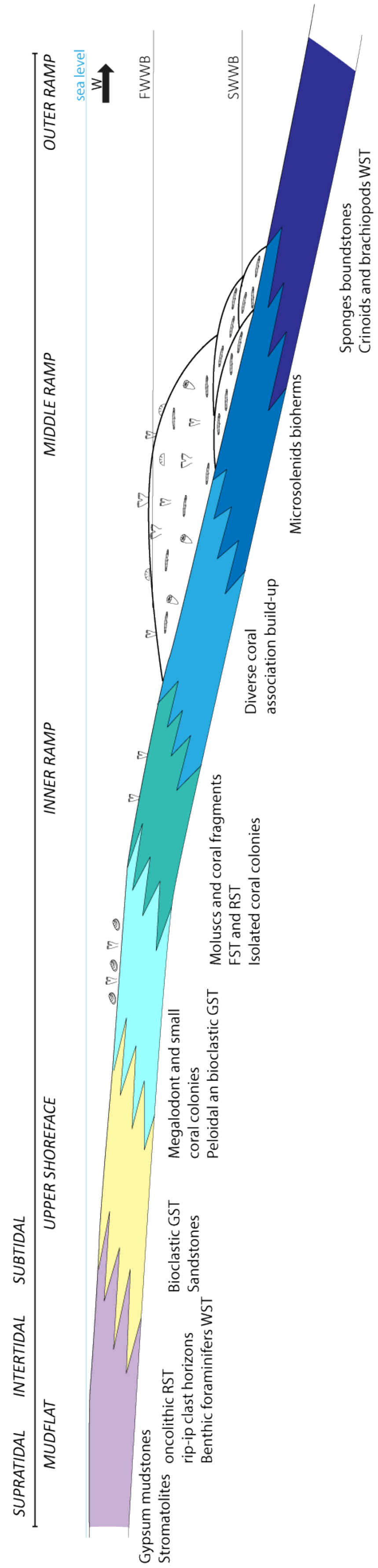


Figure 6.13: Lalla Oujja and Iggui El Behar formations depositional model.

6.5 Buildup geometries and lateral variations

In all localities across the basin, the base of the buildups corresponds to the development of platy-corals in a muddy environment, resting on outer ramp deposits. Some of these evolve towards more coral diversity, while others are capped by marls and die out. The size of the buildups can vary from 2 m to 700 m wide. The small buildups are solely platy-coral bioherms whereas the larger ones can either be exclusively platy-coral dominated or display an evolution from platy corals to a more diversified suite of coral species. No coral buildup were identified in the eastern part of the basin, suggesting a lateral transition towards a more energetic environment, where they are replaced by middle ramp deposits dominated by coral fragments and isolated colonies of delicate branching corals.

6.5.1 Cap Ghir

Lighthouse transect

Along the coast of Cap Ghir, a Plio-Pleistocene terrace cuts a horizontal surface through a large (700 m) buildup made of diverse coral associations, and allows exceptional exposure of lateral facies variations inside and surrounding a buildup. Five different environments have been identified along this transect (Figure 6.14). This transect is composed of 13 sections from 1m to 5m thick, each recording a few beds characteristic of a particular environment.

The transect sits on the top of Cap Ghir Anticline, which develops a low angle tilt to the beds. As a result we can observe not only lateral variations, but also vertical transitions, as the terrace cuts through the gently tilted horizons. Towards the south of the section, a shoreface association erodes into the inner ramp association, and finally oncoidal rudstones (mudflat environment) rest stratigraphically on top of the upper shoreface deposits. This stratigraphic evolution is common across the basin and illustrates the typical shallowing upward trend leading to the demise of the buildups.

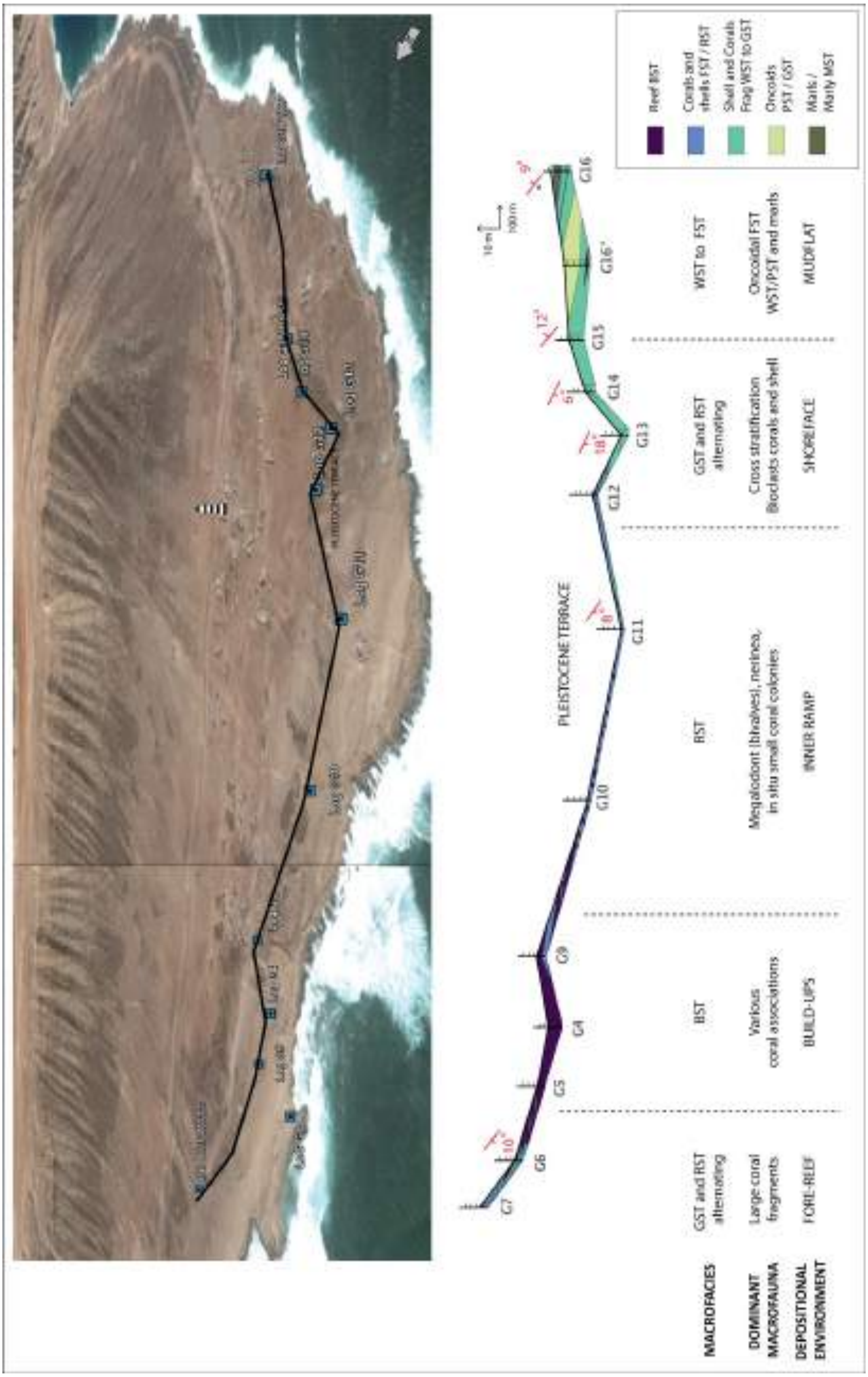


Figure 6.14: Facies variations along the coast of Cap Ghir.
Location of the transect on the map (Figure 6.15)

The variations of facies away from the core of the coral buildup are gradual. Towards the south of the exposure, the amount of bioclastic material decreases and the coral colonies become more isolated. The dominant fauna switch to large *Megalodont* bivalves and the proportion of coral fragments diminish, the ground-mass changes to peloidal PST and only isolated coral colonies are observed. This facies variation suggests a partially protected environment.

In a transect towards the north of the buildup, the amount of bioclastic material increases, the large coral colonies become more massive and then disappear, and the amount of bioclasts increases. These floatstones and grainstones are interpreted as coral-debris channels and the identification of storm induced debris-sheets indicates that this is the northern extremity of the buildup that was exposed and reworked by high energy currents.

The lateral facies variations show a W-E orientation of the reef, the north being open the sea and the south being an area protected by the reef.

Cap Ghir section and southern flank

On the southern flank of Cap Ghir Anticline, the Lalla Oujja Formation contains a succession of small platy-coral dominated buildups (Figure 6.15), locally encrusted by microbialites or serpulids. Some of the bioherms are up to 800 m wide and 20m thick, with multiple smaller biostromes that do not exceed 10 to 50 m in width. Between the buildups are dark marly mudstones and some wackestones, locally floatstones facies, rich in crinoids, coral fragments and bivalves. The facies between the main buildups are generally well-preserved and not dolomitised, while the coral-rich facies are partially or totally replaced by crystalline dolomite. The contact to the overlying Iggui El Behar Formation is a sharp surface, which places thick oncolitic grainstones on top of the coral build ups.

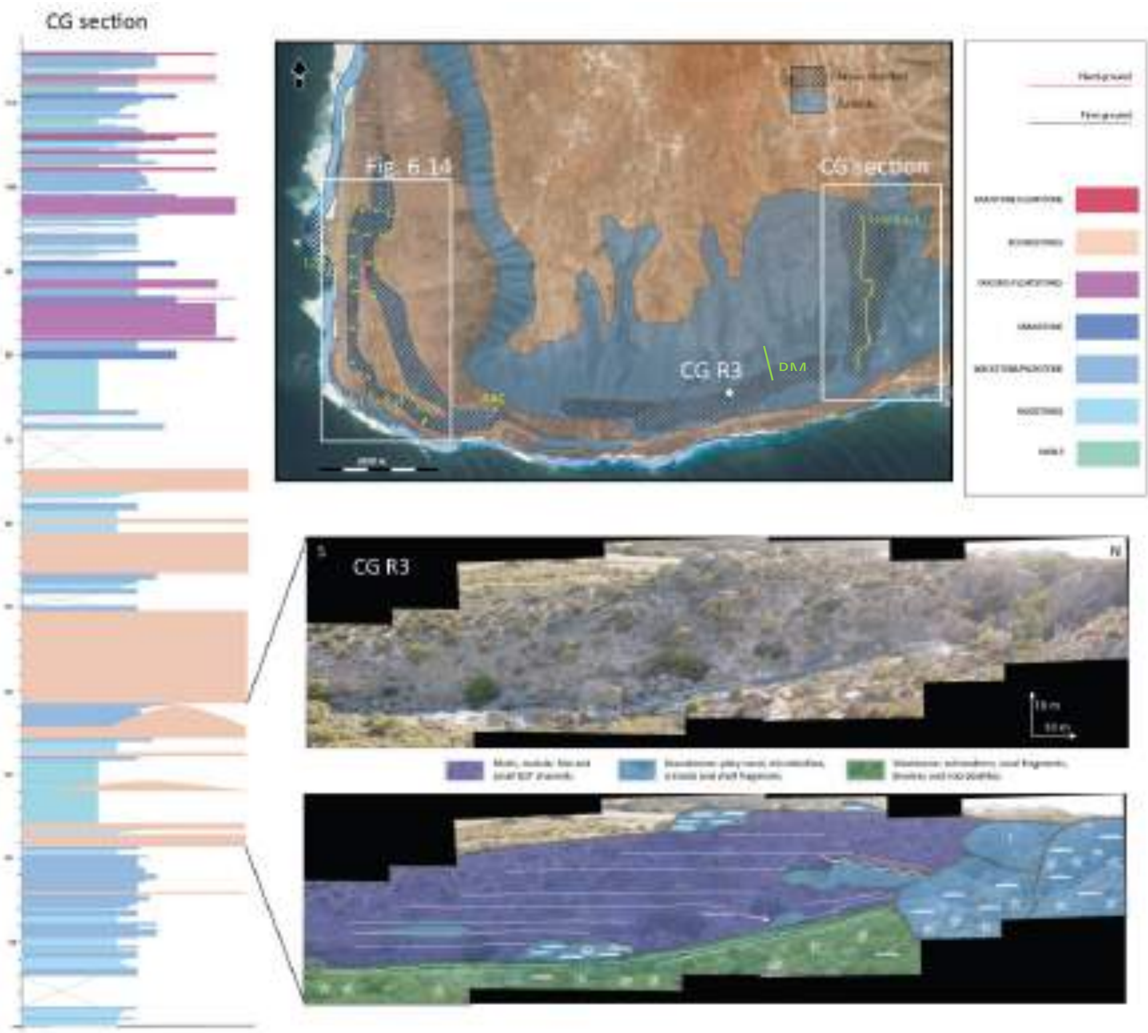


Figure 6.15: Log of the CG1 section, southern flank of the Cap Ghir Anticline. Photomosaic of the base of the Lalla Oujja Formation, location PM on the map. Map of the Jurassic of Cap Ghir, placed on a satellite view. (Google Earth and Geological Map of Taghazout 1:10000).

6.5.2 Tidili Transect

At Tidili, a biostrome around 400 m wide and >50 m thick with associated clinoforms is exposed, situated on the SW flank of the Imouzzer Anticline. A series of high resolution vertical and lateral logs have been constructed to allow identifcation of the facies composing the buildup structure, map out the different parts of the clinoforms, and distinguish the stacking pattern and geometric relationship to other buildups.

Build up and clinoform facies

The main buildup at Tidili is dominated by platy-coral communities growing in a mudstone to wackestone matrix. There is increasing diversity towards the top, dominated by branching colonies in a packstone matrix. Laterally to the west, the edge of the buildup can be observed, where clinoforms, about 80 m thick (Figure 6.16) prograde off the structure at a steep angle (35°).

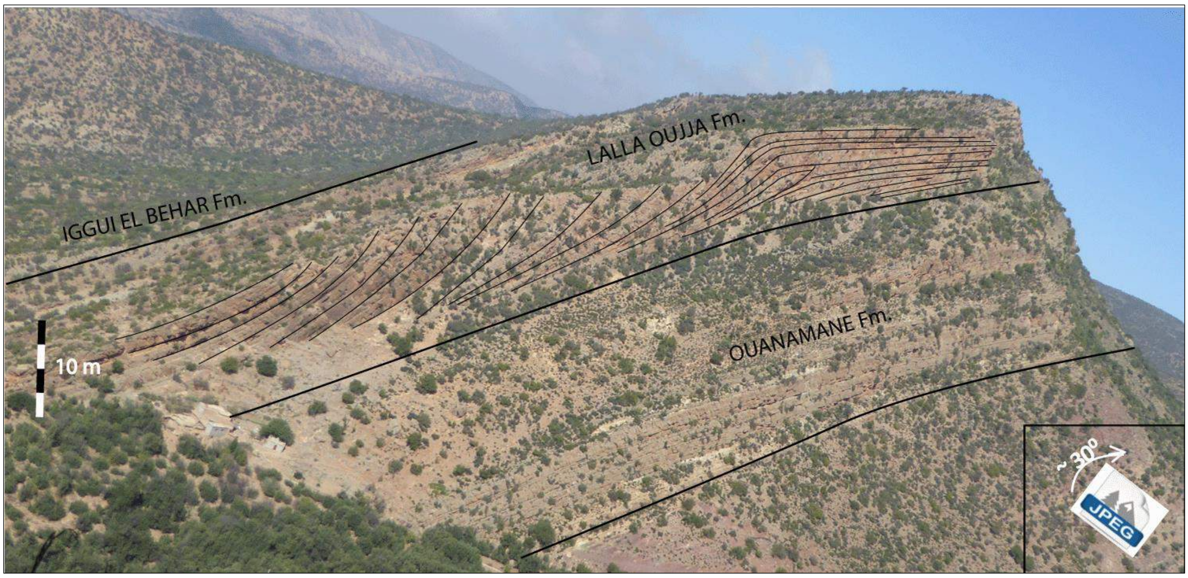


Figure 6.16: Clinoforms prograding off the buildup in the Lalla Oujja Formation
Locality of Tidili.

The clinoform foresets are made of coral fragment floatstones and rudstones in a wackestone to packstone matrix, and extend for 600 m away from the buildup, where they pass to bottomsets comprising wackestones and packstones. On the clinoform surfaces, small-scale (2-5 m wide) platy-coral build ups have been observed. These might have developed during a period of stabilisation of the main biostrome. The coral biostrome is partially dolomitised, the higher energy facies with branching corals being more dolomitised than the muddy platy-coral horizons. The clinoform foresets are also very dolomitised and contain large vugs, fractures and intergranular porosity. Dolomitisation created the vugs and most of the intergranular porosity, increasing the reservoir potential of this unit.

Buildups geometries

Six serial sections have beem logged in Tidili in order to identify the lateral facies variations from the basal build up along the clinoforms (Figure 6.17, 1) studied (Figure 6.17, right of the panel). Smaller buildups have been identified growing on top of the clinoforms, and are thus slightly diachronous to the main build up (build up 1). 300 m to the south of the buildup 1 , on top of the toeset of the last clinoform, a 200 m wide and 30 m thick buildup (Figure 6.17, 2) is visible on the other side of the valley. It also contains platy-corals at the base and a transition to increasingly diverse coral assemblage towards the top. It is followed by transgressive marly mudstones which pass laterally to a third buildup (Figure 6.17, 3), less than 2 m vertically above the buildup 2. This bioherm presents deeper-water facies, as it is dominated exclusively by platy-coral boundstones.

On the southern flank of the anticline, perpendicular to this section, other buildups are visible alond the crest. They have not been logged in detail, but two buildups are clearly visible, a larger one at the base, followed by a smaller one, which passes laterally to wackestones present at the base of the Buildup 3.

Clearly some of these buildups are diachronous, since they grow on top of each-other. The major buildups could be age equivalent (Buildup 1 and buildups on the southern flank) and might have created accommodation space between them, later filled by other buildups. Finally the entire buildup succession is overlain by the shallower facies of the Iggui El Behar Formation.

From a reservoir perspective, this correlation shows the complex lateral and vertical facies variations, each with potentially different poroperm characteristics. Dolomitisation is also controlled in part by the original facies. The sections allow measurement of typical dimensions of the buildups and relationships to the flanking clinoforms. Individual bioherms are disconnected, but the detailed logging suggests that the main facies found between them is made of coral-rich floatstones and rudstones. This facies is also highly dolomitised and potentially could have similar if not better reservoir properties.

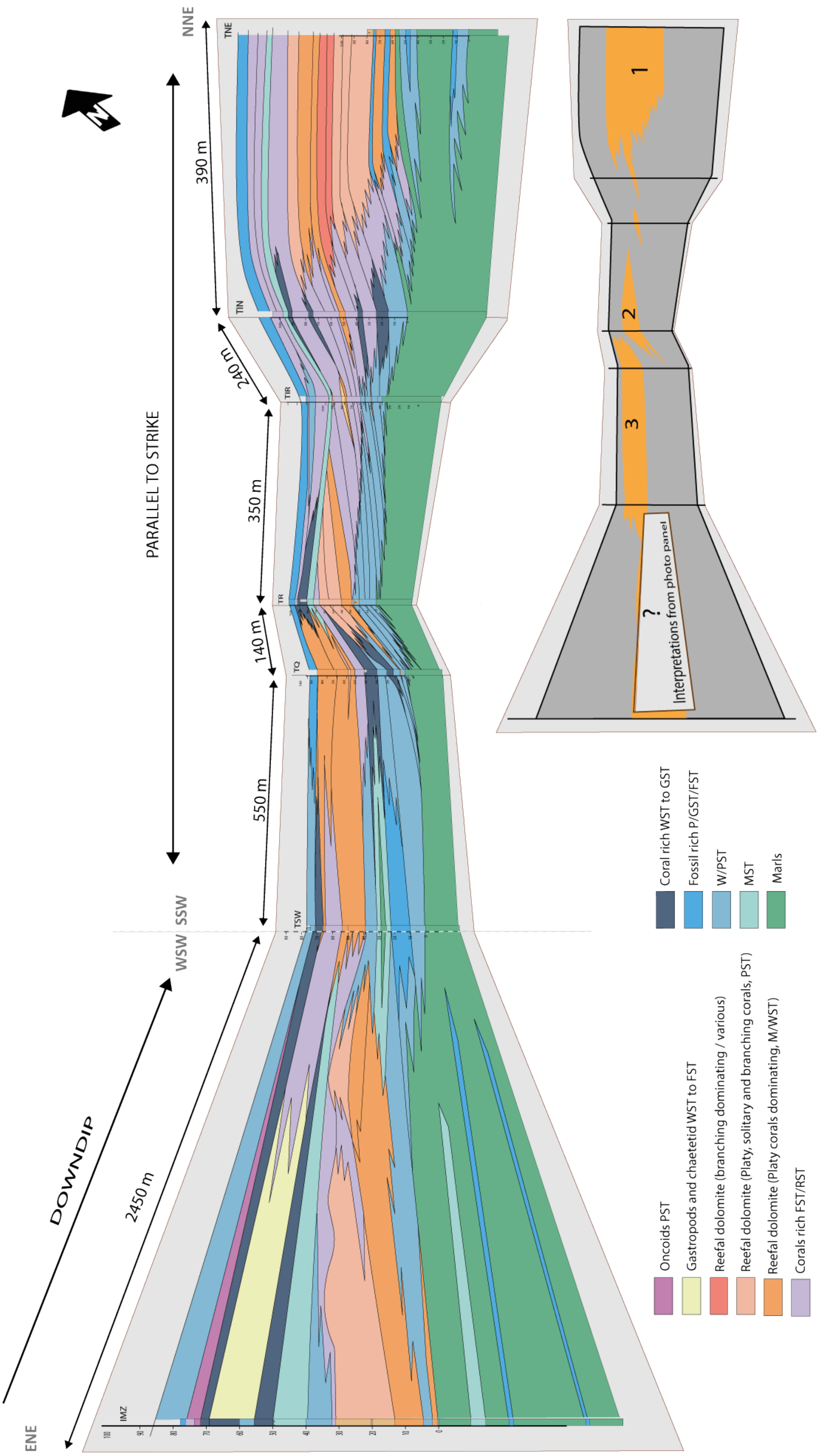


Figure 6.17: 3D facies correlations of serial sections in the locality of Tidili.

6.5.3 Tizgui

Further south from Tidili, on the southern flank of the Anklout anticline, the Tizgui section presents a transect that shows buildups with less relief. In this locality, the microsolenoids form a low relief horizontal biostrome, followed by small bioherms, still dominated by microsolenid corals but associated with small colonies of branching corals, solitary corals and thrombolites. These slightly more diverse buildups create some relief, filled with floatstones made of brachiopods, coral fragments in a muddy matrix.

In this locality, no thick diverse buildup have developed. This can be explained by the different growth rate between the platy-corals and the branching and massive corals. For the same time interval, a diverse coral buildup in an environment richer in nutrients and light would produce more material and therefore produce a thicker succession. While in Tizgui, no large buildups developed and the microsolenid corals expanded laterally. This could be explained by deeper water conditions, fewer nutrients or more turbidity in the water column compared to the rest of the basin.

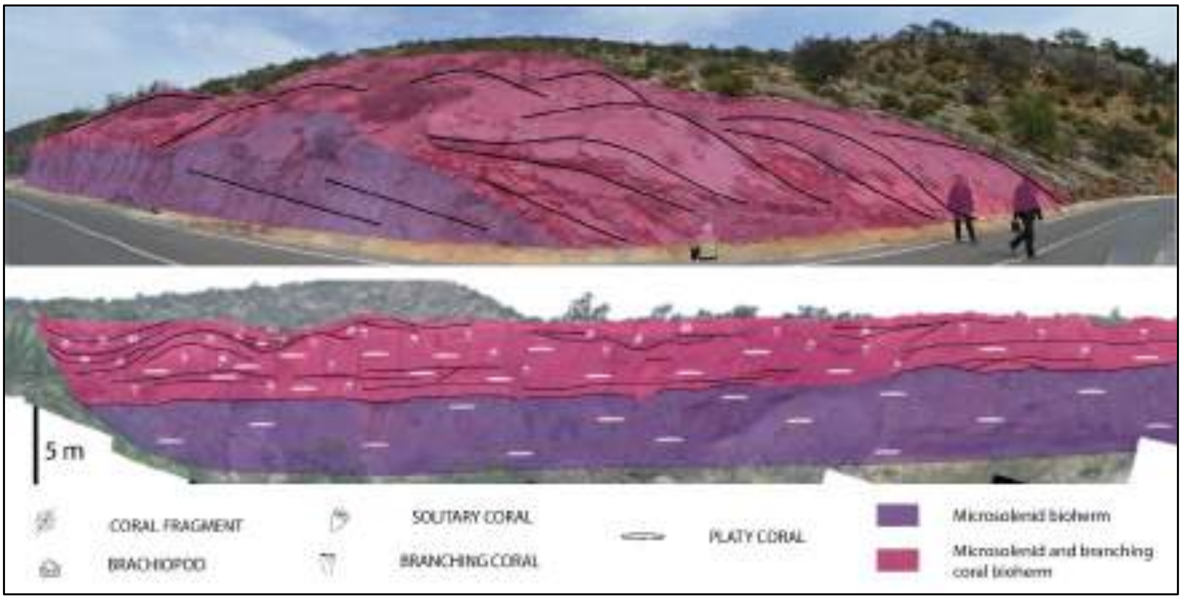


Figure 6.18: Overview on buildup evolution and geometries. Tizgui area.

6.6 Stratigraphic evolution

6.6.1 Correlations

Two correlations of the depositional environments were drawn across the central and southern part of the basin. These E-W and S-N correlations illustrate the lateral variation of the environments evolution across the basin.

E-W correlation (Figure 6.19, transect A)

The E-W correlation traverses between the localities of Tizgui N'Chorfa, Imouzzer, Tidili and Cap Ghir to highlight the proximal-distal facies distribution within the basin. It shows that the buildups are widely spread across the basin and displays the lateral shift to mid-ramp deposits. A shallowing upward trend is clearly visible to the east, with the development of inner ramp and shoreface deposits above the buildups. The upper part of all the sections are composed of peritidal deposits, first subtidal, then inter and supratidal mud-flat deposits. Locally in the eastern part of the basin, some very restricted or continental sediments were also deposited, which indicates maximum regression. In the western part of the basin, the buildup unit is constituted of multiple smaller buildups, separated from each other by marl and wackestones, while in the eastern part of the basin, the buildups are more massive and separated by coral-rich floatstones.

N-S correlation (Figure 6.19, transect B)

The N-S correlation between the locations of Askouti, Tizgui, Imouzzer and Assif El Hade emphasize the difference of thickness of the buildups units across the basin. In Tizgui, the buildups are smaller than in the rest of the basin, whereas in this locality the microsolenid coral colonies dominate. This suggests the growth of the buildup was not restricted by accommodation but rather by environmental conditions. From north to south, mid-ramp deposits mark the transition between the deeper-water buildup facies and the shallower mud-flat deposits. This correlation panel also illustrates the shoreface deposits developing in the north of the basin. The thick shoreface deposits visible in Assif El Hade, as well as in Tizgui N'Chorfa (transect A) are linked to siliciclastic influx in the basin.

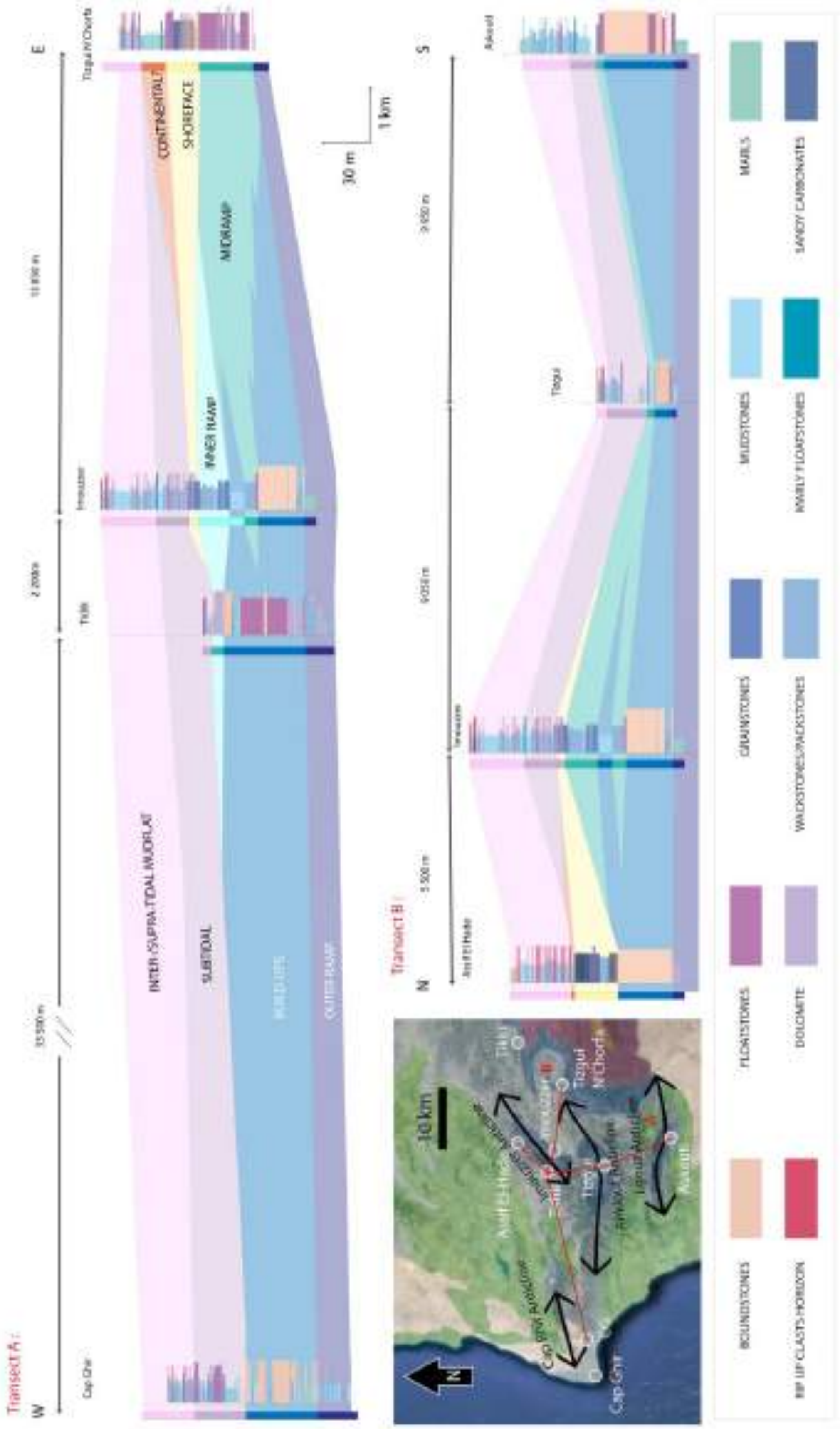


Figure 6.19: E-W and S-N cross sections across the Essaouira-Agadir Basin. Lateral variations of the depositional environments.

6.6.2 Stratigraphic evolution

The stratigraphic evolution during the Middle Jurassic in all the sections logged across the EAB begins with outer ramp deposits, followed by the development of microsolenid buildup. The major microsolenid bioherms evolve into more diverse coral assemblages while smaller microsolenid bioherms stay isolated or grow in the depression between the main buildups (Figure 6.20). Laterally the coral-dominated facies are replaced by midramp deposits dominated by coral fragments and Nerinea assemblages. The buildups disappear, most-likely as higher energy conditions prevail in the north and eastern part of the basin (shoreface, midramp), locally followed by very restricted to continental lacustrine deposits (charophytes mudstones). These proximal deposits are coeval with the shallower deposits (subtidal) that abruptly following the buildup units in the south and west of the basin. This appears to record a major basin-wide regressive event, leading to the development of extensive subtidal then inter and supratidal deposits during the Upper Oxfordian to Kimmeridgian.

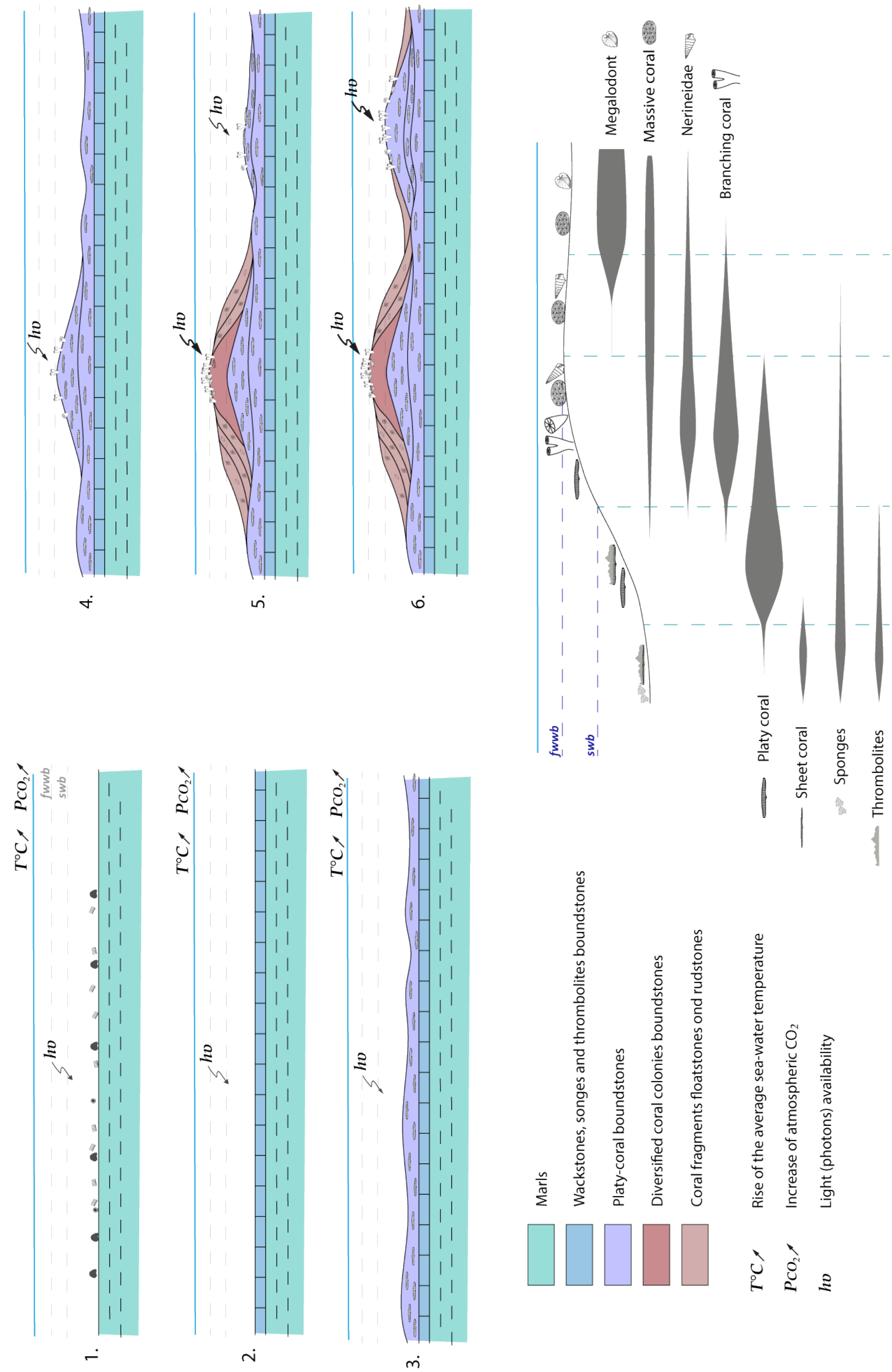


Figure 6.20: Buildup development and evolution (1 to 6) and biozonation of the main macrofauna during the Middle Jurassic in the Essaouira-Agadir Basin.

6.7 Discussion

6.7.1 Environmental transition to coral domination

Middle Oxfordian Bioherm Development

During the Middle Jurassic, large microsolenoid bioherms developed across the Essaouira-Agadir Basin. The fauna composing the framework of these buildups was already present during the Callovian, but only formed small bioherms (20m large) and biostromes (1 m thick). The development of large Upper Jurassic microsolenid biostromes is well documented from literature in French Juras (Beauvais and Bernier, 1981), Swiss Juras (Gygi, 1986; Dupraz, 1999, Lathuilière et al., 2005**),** Paris Basin (France) (Menot, 1980; Bertling and Insalaco, 1997; Chevalier et al., 2001), Poland (Kołodziej, 2015), England (Ali, 1983, Insalaco, 1999), Iberian Basin (Spain) (Benito and Mas, 2006).

This microsolenid biostrome development occurs across Tethys and the central Atlantic from Oxfordian time. In the EAB, buildup construction follows the Upper Callovian-Middle Oxfordian potential hiatus or condensed interval (Chapter 5). This widespread event has been associated to a significant shift in environmental conditions, which could in part explain the onset of Upper Jurassic coral growth. It is well documented that sea water chemistry has evolved through time with the ratio Mg/Ca fluctuating, favouring the growth of particular groups of organisms over others. In the mid-Jurassic, a shift to calcite seas, and associated high water temperatures, is interpreted to have favoured the growth of extensive coral colonies (Stanley and Hardie, 1998).

The Upper Callovian - Lower Oxfordian regressive event (Haq, 2018 and references), marked by a shift to cooler temperatures and low P_{CO_2} (Dromart et al., 2003; Cecca et al., 2005; Andrieu et al., 2016) was a time of reduced carbonate production. The onset of the Upper Jurassic global transgression (Haq, 2018), associated to higher temperatures and P_{CO_2} would have stimulated the carbonate factory and enabled the development of extensive coral colonies from the Middle Oxfordian onward.

In the EAB, the development of crinoids and calcite sponges encrusted horizons may also have played a role in the stabilisation of the sediment substrate allowing corals to develop.

Comparison to European microsolenid biostroms

Microsolenid biostroms are common during the Upper Jurassic, showing some variation in character depending on the environmental conditions. They are usually units dominated by in situ microsolenid corals development, forming a coherent framework and displaying a bedded sheet-like geometry (Insalaco, 1996). These biostromes are found at comparable stratigraphic positions within the reef successions and their fauna associations are quite similar. In many localities, the microsolenid biostromes develop on sponge-rich intervals (Insalaco, 1996; Chevalier et al., 2001; Eliuk, 2016). This has also been observed in the EAB in some locations. The microsolenid biostromes typically evolve towards more diverse coral associations and show development of a higher relief reef (Insalaco, 1996; Lathuilière et al., 2005; Benito and Mas, 2006).

Different ecozones have been identified by Lathuilière et al (2005), depending on the coral associations and their associated sediments. They identify the deeper facies as being dominated by non-encrusted *Dimorpharea*, in marls and wackestones matrix, and followed by the *Microsolena* ecozone, followed by the *Dendraraea* ecozone, where the morphology of the dominating corals shifts from sheet-like to branching coral. These authors distinguish the ecozones as being controlled by their supposed water-depth. Insalaco (1996) proposed another interpretation by looking at the siliciclastic content of the different sections, he identified that *Dimorpharea* dominates where the siliciclastic content is higher, and *Microsolena* dominates where the limestones are more pure, and linked this to the morphology differences between the two species.

The hypothesis of deeper facies and/or higher siliciclastic input could both be valid when considering the data from the EAB. *Dimorpharea* dominate at the base of the succession where the environment is silty, and they are present in what is interpreted to be the buildup facies in the deepest water environments. The evolution of the bioherms is also very similar to that recorded for European microsolenid biostromes, and therefore it is proposed that they probably developed under similar environmental conditions.

6.7.2 Influence of siliciclastics on coral disappearance

The demise of the coral buildup coincides with the development of siliciclastic rich upper shoreface sediments in the northern part of the basin. This rapid influx of siliciclastics into the environment could have triggered the disappearance of coral buildups locally. However, the demise of the coral buildups appears to be relatively synchronous across the basin (no later than Upper Oxfordian). In Assif El Hade, isolated corals are still present in situ within the shoreface deposits. This suggests conditions are still favourable to coral growth and their disappearance may not be only imputable to the sand influx, but rather to a combination of factors that changed the environmental conditions. All over the eastern and central part of the basin, shoreface deposits are followed by charophyte-rich wackestone and marl. This fauna indicates a change to restricted or non-marine conditions a result of a rapid regressive event across the basin. It is more likely it was this drastic change of bathymetry and associated change in water chemistry that triggered the final disappearance of the buildups rather than the siliciclastic influx, which affect only in the north of the basin.

The influence of the siliciclastic deposits can be observed south of the Amsittène Anticline (AMCA), as no major buildups have been identified in this area. The last ammonites collected are Middle Callovian in age (Chapter 5). In this locality, the Lalla Oujja Formation is likely to have been eroded away by siliciclastic beach deposits recorded directly on top of the Ouanamane Formation. The development of beach deposits may be linked to the growth of the Amsittène Anticline, which created shallower conditions, but the large influx of siliciclastics is still a trend more visible towards the North. This influx could then have prevented the growth of coral buildups or their record further North in the basin.

6.7.3 Reservoir potential

The potential reservoir units identified are primarily the buildups and the rubble that fringes the buildups and is sourced from the structures. These facies are highly dolomitised which results in development of large vugs and intercrystalline porosity. Other potential reservoir units are also present in the Lalla Oujja Formation. The midramp rudstones facies also have good intra and intergranular porosity and permeability in the grainstone matrix constituted of coral and shell fragments. This facies is also locally very dolomitised, which if not too severe, enhances the porosity of the unit. The shoreface deposits are also potential reservoir units, depending on the extent of cementation of the grains. These facies either directly follow the buildups or are laterally equivalent to them. They Lalla Oujja Formation reservoirs are overlain by muddy facies of the Iggui El Behar Formation and red Kimmeridgian marls (Ambroggi, 1963; Adams et al., 1980), which should form a basin-wide seal.

The Oxfordian reefal buildups are a proven reservoir interval on the conjugate margin of Nova Scotia, where the Deep Panuke gas field, located in the Scotian Basin, is currently producing. The reefal interval drilled in this field has been extensively studied (Ellis et al., 1985; Weissenberger et al., 2006; Weston et al., 2012; Eliuk, 2016). The buildups are described as being made of microbiolite mounds, microbial and siliceous sponge and microsolenid coral mounds. These facies are very similar to the facies observed in the deeper buildups in the EAB and are likely to be part of equivalent systems.

This paper provides new data on the size range and lateral facies variation exhibited by the microsolenid biostromes and bioherms, which can be extrapolated to the subsurface to give a better understanding of the buildup geometries in Offshore Morocco and the Scotian Basin.

6.8 Conclusions

The Oxfordian coral buildups facies developed in the Essaouira-Agadir Basin are dominated by microsolenid bioherms. They developed in a low light and low energy environment on a gently sloping ramp and are followed by shallower, more diverse coral associations dominated by branching coral morphologies (already described locally by Olivier et al, 2012). These observations have been extended across the basin and conclusions drawn:

- A number of geometries are formed by the coral buildups, including large coral-rich clinoforms extending from the bioherms and more horizontally extensive coral biostromes. The microsolenids bioherms do not grow into high relief structures, but the branching coral dominated structures make larger, more domeshaped structures.

- Coral dominated facies developed across the basin in isolated buildups, separated a few meters up to 400m from each other by reef-rubble and mudstone to wackestone horizons. Their size range from very small 2m wide and 0.5 m thick microsolenid colonies uo to 700 m wide and 80 m thick more diverse coral colonies.
- Laterally, in more proximal locations (to the east of the basin), the coral buildups are replaced by more energetic midramp deposits, dominated by coral fragments, delicate branching corals and gastropods.
- The buildups are generally topped by shoreface deposits. In the north of the basin, these deposits are siliciclastic-rich, with the siliciclastic proportion diminishing towards the south. This influx of siliciclastics into the basin might have initiated the coral buildup demise in the North, but it is not interpreted to be responsible for the overall coral disappearance, which is related to a widespread change in palaeoclimate.
- The Middle to Upper Oxfordian stratigraphic evolution is strongly regressive in the EAB, with the development of a inter- to supratidal environment during the Upper Oxfordian to Kimmeridgian.
- The onset of buildups is synchronous with the evolution of the environmental conditions that initiated carbonate buildups across the Tethys and Central Atlantic during Middle Oxfordian times; their demise is linked to basinal shallowing conditions during the Upper Oxfordian.
- The coral buildups exhibit good reservoir characteristics, with porosity development, and other facies also show potential, including midramp and shoreface deposits and the rubble facies on the margins of the buildup, which allow reservoir continuity between isolated buildups.

These conclusions can be extended to offshore Morocco, and to Nova Scotia on the conjugate margin where similar facies developed.

Chapter 7: Upper Jurassic stratigraphy and discussion

7.1 Upper Jurassic stratigraphy

The Imouzzer (Kimmeridgian) and Tismerroura (Tithonian) formations (Figure 7.1) have been logged in the center of the basin and in the borders of the basin. The comparison of these two areas gives important constraints on basin evolution.

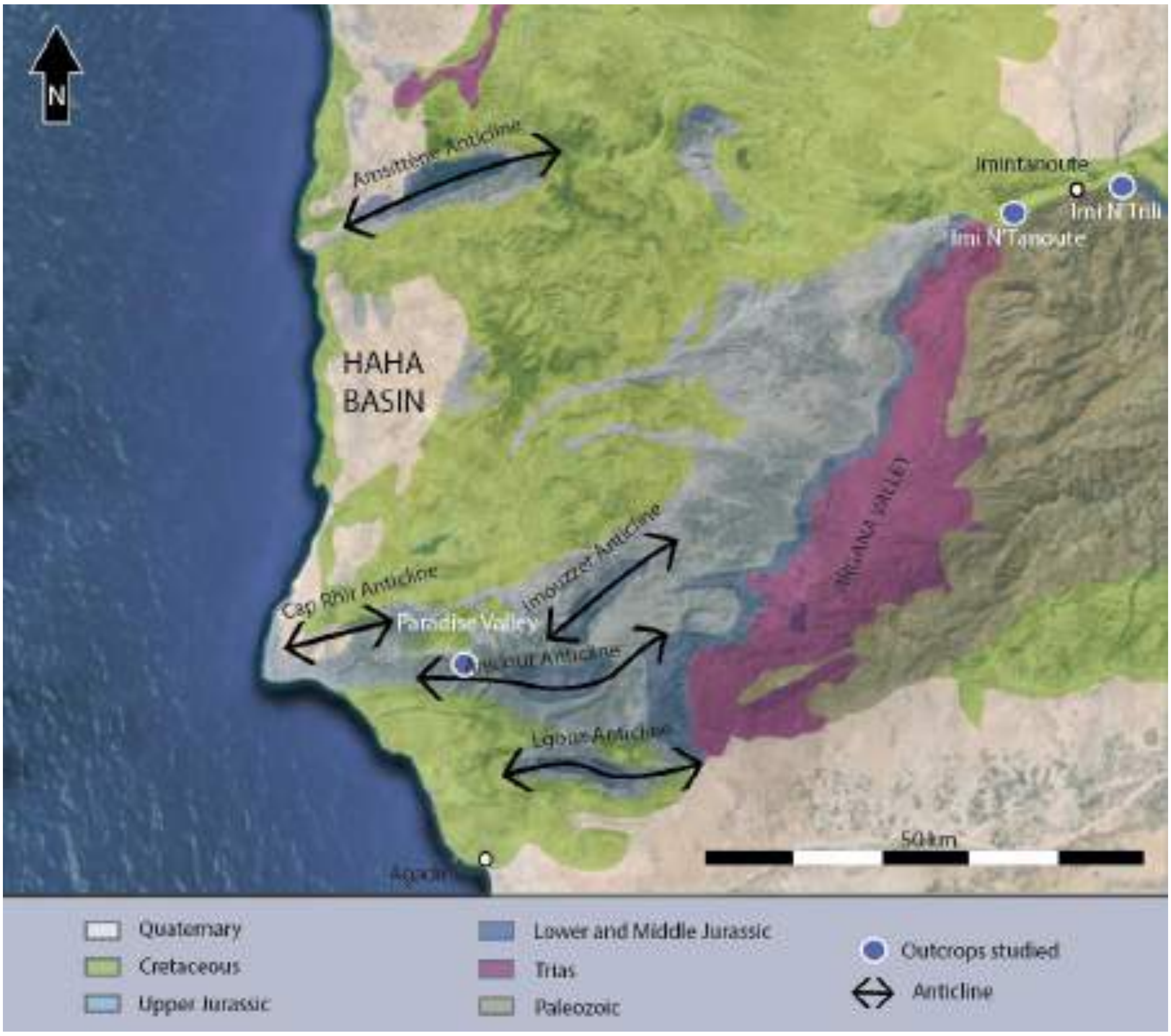


Figure 7.1: Basin map with location of the sections studied for the Imouzzer and Tissmeroura Formations.

7.1.1 Imouzzer Formation

Paradise Valley

The Paradise Valley outcrops are situated in the center of the basin. The stratigraphy and facies of this section are characteristic of the Imouzzer Formation as it appears in most localities across the basin.

Description

The Imouzzer Formation crops out almost without interruptions along the Paradise Valley in a 180 meters thick section. The transition from the Iggui El Behar Formation (see chapter 6) to the Imouzzer Formation is marked by an increase of siliciclastics in the system. The larger part of the section is made of repetitive thick (20-30 m) levels of red to chocolate-brown marls. These levels are separated by dolomitic units (20 cm to 2m thick) composed of marly limestones, marls and bioclastic wackestones to packstones (Figure 7.2, A). The basal dolomitic layer, just above the first three meters of red marls, is a peloidal and bioclastic packstone with synsedimentary folding (Figure 7.2, B). This unit contains peloids, horizontally oriented shell fragments, large benthic foraminifera and rare gastropods alternating with some horizons of medium sand-sized quartz grains. On top of it, five meters of red marls present calcite geodes with diameter around 5 cm, this level is followed by peloidal and bioclastic packstones with a large amount of thin bivalve shells, some echinoderm fragments, and quartz elements (Figure 7.2, C). After this first dolomitic unit, the formation is more monotonous, with four levels of reddish marls (20-30m thick) (Figure 7.2, D) separated by three dolomitic levels (5-7 m thick) composed of an alternation of wackestones or packstones with shelly fragments, echinoderm spines, peloids and ooids, sometimes burrowed by *Thalassinoides*; silty mudstones levels, and red and beige marls. Some of the dolomitic levels are brecciated and encrusted by iron, which appears to be recent weathering. Toward the top of the formation, the marl levels get thinner and are alternating with beds of dolomites with euhedral crystals and angular quartz grains; wackestones containing large benthic foraminifera and radial ooids; coarsening upward breccias levels and mudstones presenting horizontal lamination (Figure 7.2, E).

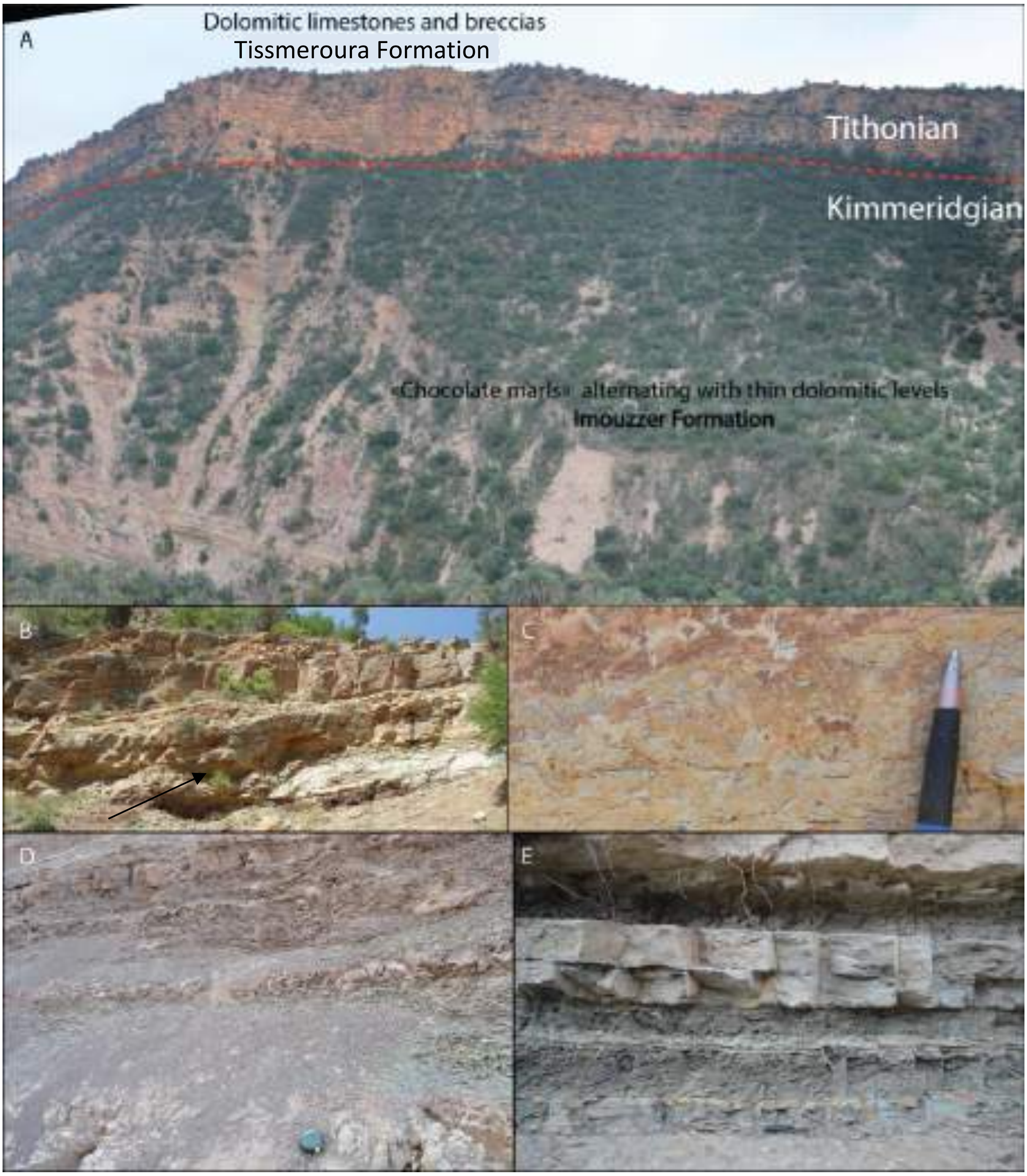


Figure 7.2: Imouzzer Formation, Paradise Valley.
(A) Overview of the Imouzzer Formation in the Paradise Valley; (B) Synsedimentary fold (the double arrow is one meter long); (C) Oolitic packstone with thin shell fragments (D) "Chocolate marls"; (E) Alternation of grey marls and mudstone layers.

Interpretation

The transition from the peritidal Iggui El Behar, composed of pure mudstones with foraminifera and horizons with gypsum molds to this quartz and clay-rich unit indicates an important change in the environmental conditions and new deliveries of siliciclastics into the basin

The presence of echinoderm plates and spines fragments, as well as large benthic foraminifers in most of the limestones horizons indicates a marine origin of the sediments. The preferential orientation of the elongated shell fragments suggests some current reworking, while the radial-fibrous structure of the ooids is indicative of low energy, shallow-marine to restricted settings (Loreau et al., 1973; Simone, 1980). The presence of *Thalassinoides* trace fossils is also indicative of a marine origin, as these burrows are often present in (but not restricted to) lagoonal and embayment settings (Richard and Bromley, 2008). The presence of angular quartz grains in the system indicates a continental origin of the siliciclastic input and a potentially close source. The mud-rich facies and shallow-marine elements indicate a protected environment, possibly lagoonal. Some rare heterolithic beds could also indicate tidal reworking at the base of the formation.

In the Central High Atlas, red marls beds are also widely described during the Middle Jurassic to Lower Cretaceous, and are particularly well developed during the Kimmeridgian, but they are indicative of lacustrine environments (Haddoumi et al., 1998, 2008; 2010; Mojon et al; 2009). One marine level has been identified in this succession in the Barremian and was interpreted as marine ingression and development of coastal lagoons connected to the Atlantic (Charrière et al., 2005).

Imi N'Tanoute and Imi N'Trili

In the NE border of the margin, the Imi N'Tanoute and Imi N'Trili formations records the most proximal deposits of the Imouzzer Formation.

Description

In the Locality of Imi N'Tanoute, this formation is 70m thick, its base is dominated by red siliciclastic mudstones and silt, alternating with lenticular horizons of evaporites, 10 to 40 cm thick. It is evolving upward to red siliciclastic mudstones alternating with fine to medium sandstones and some erosive fining upward beds of microconglomerates and sandstones. The fine to medium sandstones present well-developed wave ripples, climbing ripples (Figure 7.3, A), cross lamination and horizontal lamination. Some microconglomerates and coarse sandstones have an erosive base and are cross-stratified, while others present tabular cross-bedding and are organised in units stacked on top of each other. Paleoflow measured on current ripples and sandsones cross beds is to the west.

In the locality of Imi N'Trili, the formation is 40 m thick, its base is dominated by red siliciclastic mudstones and rare horizons of well sorted fine sandstones, and the upper part is dominated by conglomerates and sandstones units, organised in finning upward lenticular units. The conglomerates and sandstones units are up to 3 m thick, the pebbles are dominantly quartzite, and the units present cross-bedding indicating a NW direction. In the poorly sorted conglomerates horizons, lenses of cross-bedded sandstones (Figure 7.3, B) are common. They are fining upward to medium and fine sandstones, and to red mudstones with horizontal laminations.

Interpretation

The basal alternation of red mudstones and rare evaporite horizons indicates a restricted environment with some marine influence linked to the regressive context. The red mudstones and siltstones are marking a change of sedimentation from the carbonates of the Tadlat Oudman Formation (Equivalent to the Callovian to Lower Kimmeridgian interval in this part of the basin, Figure 7.3, C) (Medina, 1989) to a siliciclastic dominated formation. The facies association at the transition between marine and continental deposits suggests an evaporitic coastal environment. It has been interpreted as a coastal sabkha, potentially linked to supratidal water flooding.

Figure 7.3: Imouzzer formation facies
A- Climbing ripples. B- cross-bedded sandstones and conglomerates. C- Overview of the stratigraphy in the locality of Imi N'Trili.

Further up in the formation, the mudstone and siltstones are interbedded with the fine sandstones and are interpreted to have been deposited on floodplains in inter-fluve areas. The cross-laminated and climbing-ripples sandstones likely represent overbank splay deposits (Lunt et al., 2004). The stacked cross bedded sandstones and microconglomerates were interpreted as migrating dunes forming a compound channel bar (Reading, 1996). The small erosive and lenticular microconglomerates and sandstones represent minor channels.

In Imi N'Trili, the thick lenticular conglomerates and sandstones units with large tabular cross-beds have been interpreted as braided river channels. The presence of interbedded sandstones lenses indicates a compound braid bar system, where the fining upward trend is due to waning flows.

The upper part of the formation in the western part of the area (Imi N'Tanoute) is dominated by red clays and minor conglomerates and sanstones units, while the eastern part of the area is dominated by thick conglomerates units. This can be explained by the observation of the lateral variation of the formation in Imi N'Trili (Figure 7.3, C), the thick sandstones and conglomerated disappear towards the East because the channel belt was localised along the actual river bed (left of the picture).

Conclusions

The Imouzzer Formation records shallow marine conditions in the center part of the basin and is composed of continental deposits towards the NE, along the margins of the basin. The development of these fluvial environments with W-NW flow direction can explain the increased proposition of siliciclastics in the center of the basin.

7.1.2 Tismerroura Formation

Paradise Valley

Description

Along Paradise Valley, the Tismerroura Formation is made of two thick dolomitic units separated by a unit of dolomites alternating with marls. The first limestone unit is 27 meters thick and composed of an alternation of oolithic grainstones, mudstones with horizontal lamination, mudstones with mudcracks, few thin marls beds, breccias (mostly coarsening upward breccias) and massive dolomite beds. The middle unit is 52 meters thick and is composed of an alternation of grey and red marls, with, from base to top: Micritic dolomite beds, thick mudstones with mudcracks level (with a small fraction of siliciclastic silt), some vuggy dolomitic micrite with few mudcracks, micrite, dolomite with wave ripples, dolomitic mudstone with horizontal laminations, microbial mats and dolomitic mudstones laterally evolving to breccias. At the base of the upper unit, a bed of mudstone rich in wood debris is included between two marls layers. A limestone level with root traces has also been found six meters above this level. Generally, the upper unit is composed of more cyclic facies variations. A cycle begins with a dolomitic oolitic grainstone to mudstone level, locally with wave ripples or followed by a mudcrack level.

Above it lies a stromatolitic unit, followed by coarsening-upward breccias (Figure 7.4). In a few dolomitic layers, some bird-eyes and square mineral molds can be identified. Towards the top, a level highly bioturbated by T*halassinoides* is visible. Some dolomitic layers, and in particular the stromatolites layers present a strong kerogenic smell.

The Tismerroura Formation in the Paradise Valley contains also small-scale syn-sedimentary folds and faults striking E-W, consistent with the axial orientation of the Anklout anticline.

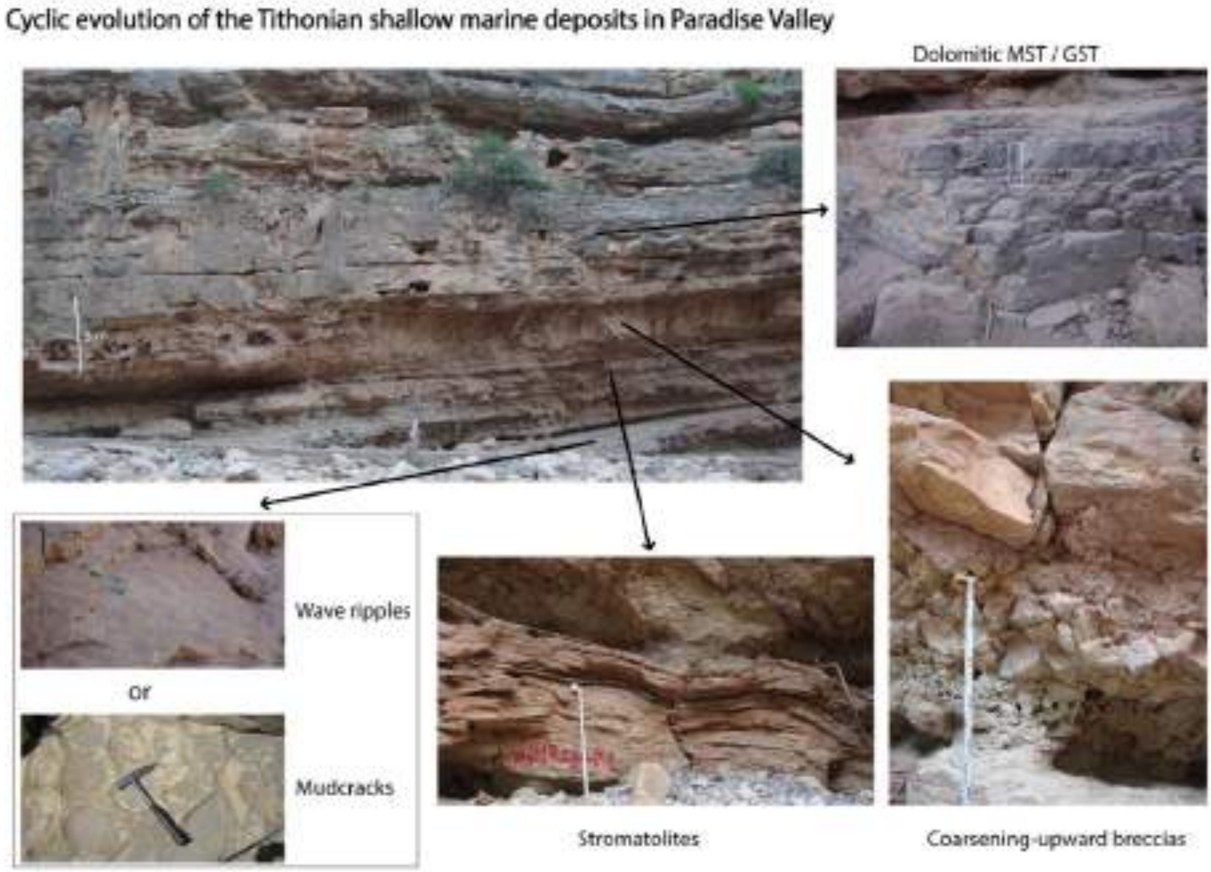


Figure 7.4: Cyclic evolution of the Tithonian peritidal deposits in Paradise Valley.
(Staff is 1.10 m long)

Interpretation

The mud-cracks and roots traces indicate periods of emersion, while the laterally extensive stromatolites indicate colonisation of the sea bed by microbial mats without current agitation. The inversely gradded breccias, with no preferential orientation of the clasts in the breccias beds are evidences for the collapse-dissolution origin of the breccias (Friedman, 1997). The presence of evaporites in the eastern part of the basin will be discussed thereafter. All these facies are characteristics of supratidal conditions. The association of dolomites, oolitic and peloidal packstones, grainstones and dolomite horizons is slightly more energetic than the other facies present in the formation. They were interpreted as intertidal to subtidal deposits. The extensive dolomitisation and micrite replacement of the ooids and bioclasts does not allow further discussion on the ooids formation. The presence of *Thalassinoides* and wave ripples suggests peritidal conditions (Richard and Broomley, 2008). This facies association have been interpreted as peritidal environment with shallowing upward cycles.

Immi N'Tanoute and Imi N'Trili

Description

This Formation is 230 m thick in the locality of the section Imi N'Tanoute, and its thickness is greater than 180 m in Imi N'Trili, but its full thickness cannot be measured because the upper part of the succession is extremely folded. In these two localities, the formation is largely dominated by thick evaporites packages (Figure 7.5, B), up to 5m thick. The evaporites beds present different textures, their color varies from white to grey or orange, and they can be massive, thinly bedded, present a chicken-wire structure or be an association of evaporites and/or mudstones clasts. They alternate with mudstones horizons (30 cm to 4 m thick) presenting birdeyes and thin microbial units (Figure 7.5, C and D). More rarely some peloidal and oolitic packstones with wave ripples or flaser bessing (Figure 7.5, A), and some blue and red siliciclastic mudstones are present.

Interpretation

The lateral continuity of the thick evaporites units and their association to microbial horizons such as mudstones with birdeyes and microbial mats indicates their deposition in subaqueous conditions (Rouchy and Caruso, 2006; Warren, 2016). The evaporite horizons constituted of mudstones and evaporite pebbles are also indicative of some reworking and in-situ re-deposition. The dolomitic mudstones with birdeyes are characteristic of peritidal environments (Shinn, 1983). The peloidal and oolitic heterolitic horizons are rare but the show evidences of tidal currents in the system. The overall facies association is characteristic of peritidal conditions. The red and blue mudstones indicate small siliciclastic influx in the system and they might be linked to periods of emersion. The rare evidence for reworking and the extensive evaporites indicate a restricted environment. This formation has been interpreted as inter to supratidal and is characteristic of a coastal salina or sabkha environment with rare more open marine incursions leading to inter to subtidal environments.

Figure 7.5: Facies of the Tismeroura Formation.
A - heterolitic unit with flaser bedding and wave ripples. B – Gypsum bed with thin mudstones horizons. C – Mudstones with birdeyes overlaid by massive mudstone. D – Microbial structure presenting bird eyes along the top of the bed and followed by evaporites.

Conclusion

The Tismeroura Formation is dominated by sabkha conditions in the eastern part of the basin, with development of very thick evaporites succession, and by peritidal deposits in the center of the basin. The wood debris and the kerogenic smell found in the upper part of the Tismerroura Formation, particularly in the extensive stromatolites levels could also raise the question of source rock potential of the formation, if these horizons are more developed further offshore.

7.2 Palaeogeography - Discussion

In order to construct palaeogeographic maps, multiple well data were associated to the outcrops observations and interpretations (Figure 7.6). The palaeo-shelf has been redrawn after the interpretations of Hafid and co-authors (2000, 2006). The interpretations taken from the wells are subject to discussion, as the biostratigraphic framework is far from complete. The ages of the units have been used as given in the well reports, but these retain some uncertainty.

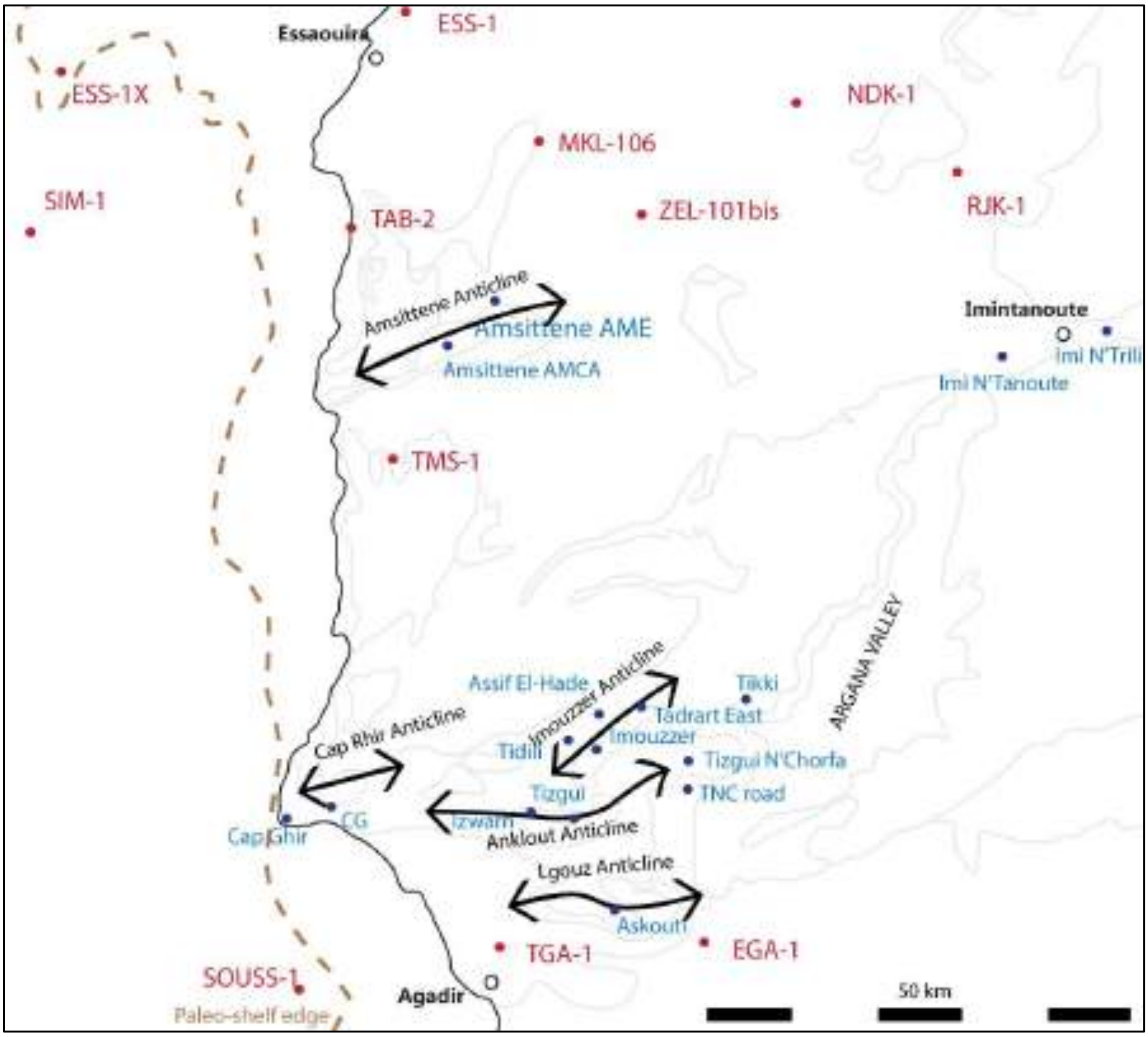


Figure 7.6: Basin Map and location of the sections (blue) and wells (red) used to build the palaeogeographic maps. Jurassic Shelf-edge from Hafid et al. (2006).

TOARCIAN – AMSITTÈNE FORMATION (Figure 7.7)

The first Toarcian deposits consist of erosive continental deposits observed at the outcrop, with development of alluvial fans in the area of Tikki. The upper part of the formation presents coastal plain deposits in Akesri, and this facies association has been extrapolated to TGA-1, TAB-2 and TMS-1 according to the mixed siliciclastic and carbonate lithologies observed in wells for this interval. Lower Jurassic carbonates seem to be directly following the Trias in the well ESS-1X, and this has been drawn as peritidal deposits by analogy to the Tamarout formation. The erosive event have been correlated with the CHA (chapter 4) and this unit records a regional regression, following the Arich-Ouzla open marine formation.

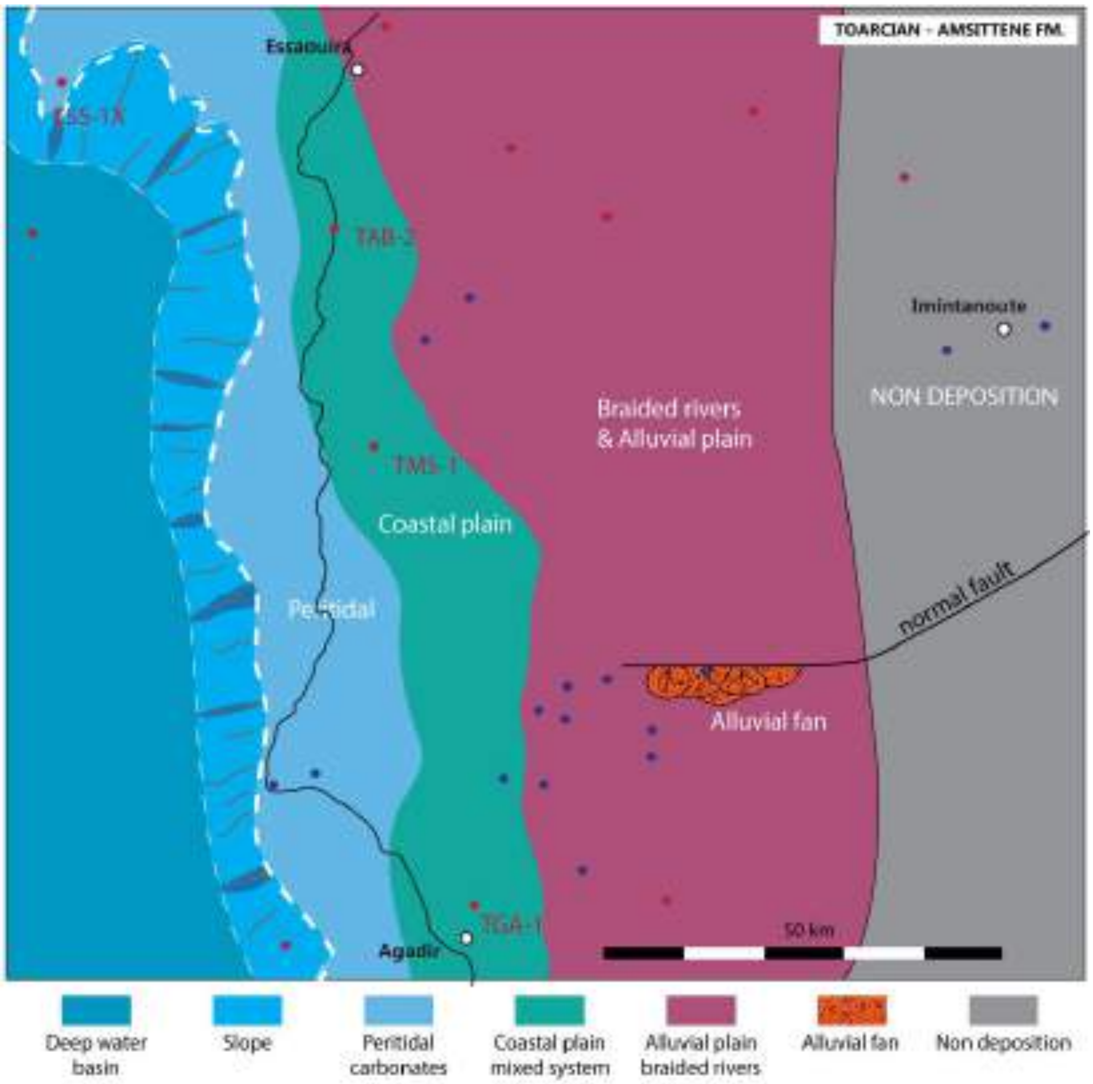


Figure 7.7: Toarcian palaeomap: Amsittène Formation.

TOARCIAN – LOWER TAMAROUT FORMATION (Figure 7.8)

This palaeomap is indicative of the environment during the deposition of the Unit T2 of the Tamarout Formation (Chapter 4). Continental deposits were still present in the eastern part of the basin, like for the Amsittène Formation. Non deposition occurred in the eastern margin of the basin (Imi N'Tanoute). In the locality of Tikki, it is dominated by sabkha deposits, and evaporitic horizons are also present in the wells TMS-1 and TAB-2. The center of the basin was still dominated by peritidal to sabkha deposits and a distinction between the localities where the evaporites units are still present and where they have been dissolved and where dissolution-colapse breccias can be made.

The data indicates that this initial transgression seems to be restricted to the Agadir segment.

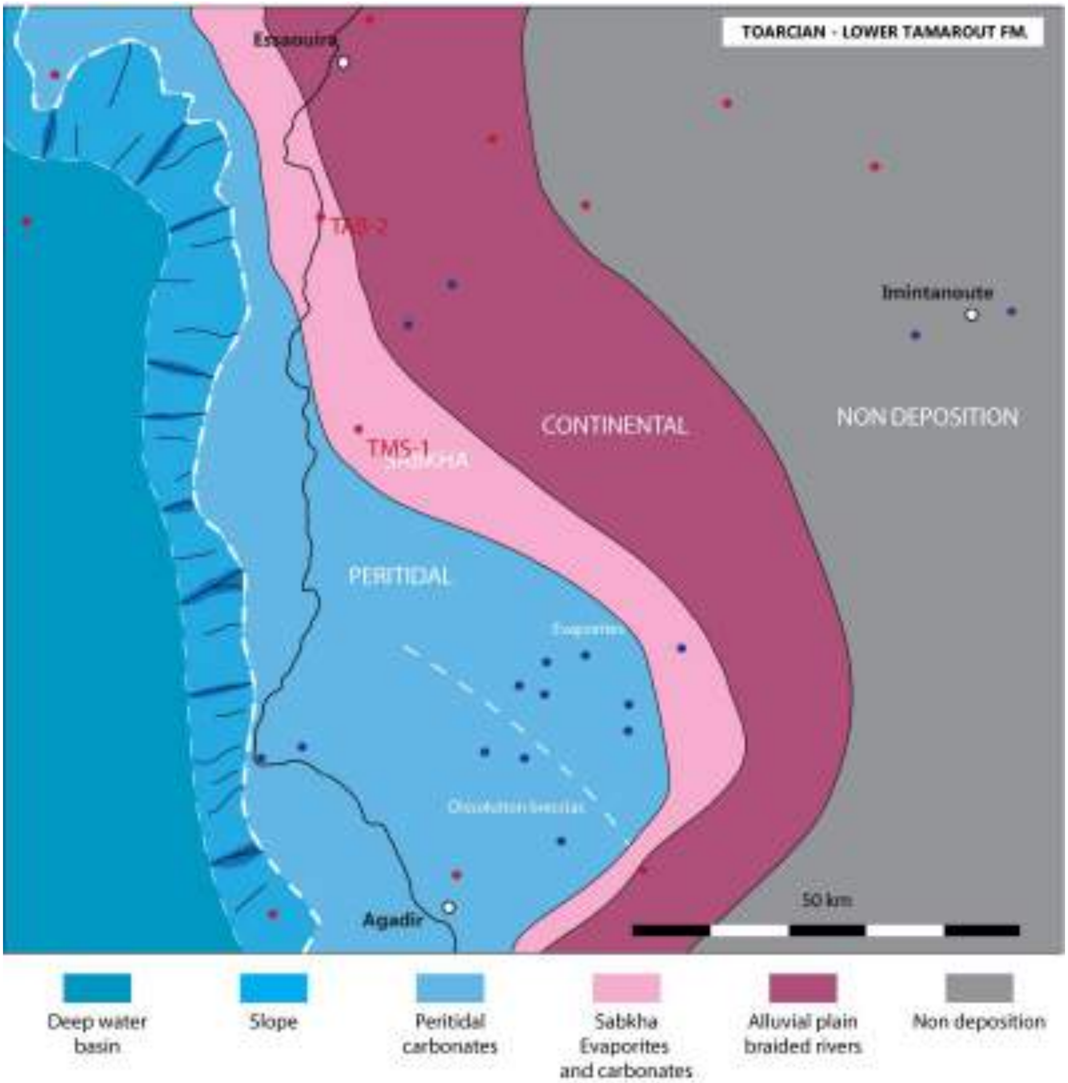


Figure 7.8: Toarcian palaeomap: Lower Tamarout Formation.

TOARCIAN –UPPER TAMAROUT FORMATION (Figure 7.9)

The Unit T3 of the Tamarout Formation is dominated by inter- to subtidal oolitic carbonates, it marks a generalised transgression across the basin and the disappearance of evaporites in the central part of the basin. The eastern part of the basin still did not record any deposition (Imi N;Tanoute, well RJK-1) and the rare evaporitic deposits are only found in the east of the area. This unit records the opening and extension of the basin and the development of more marine deposits in the north.

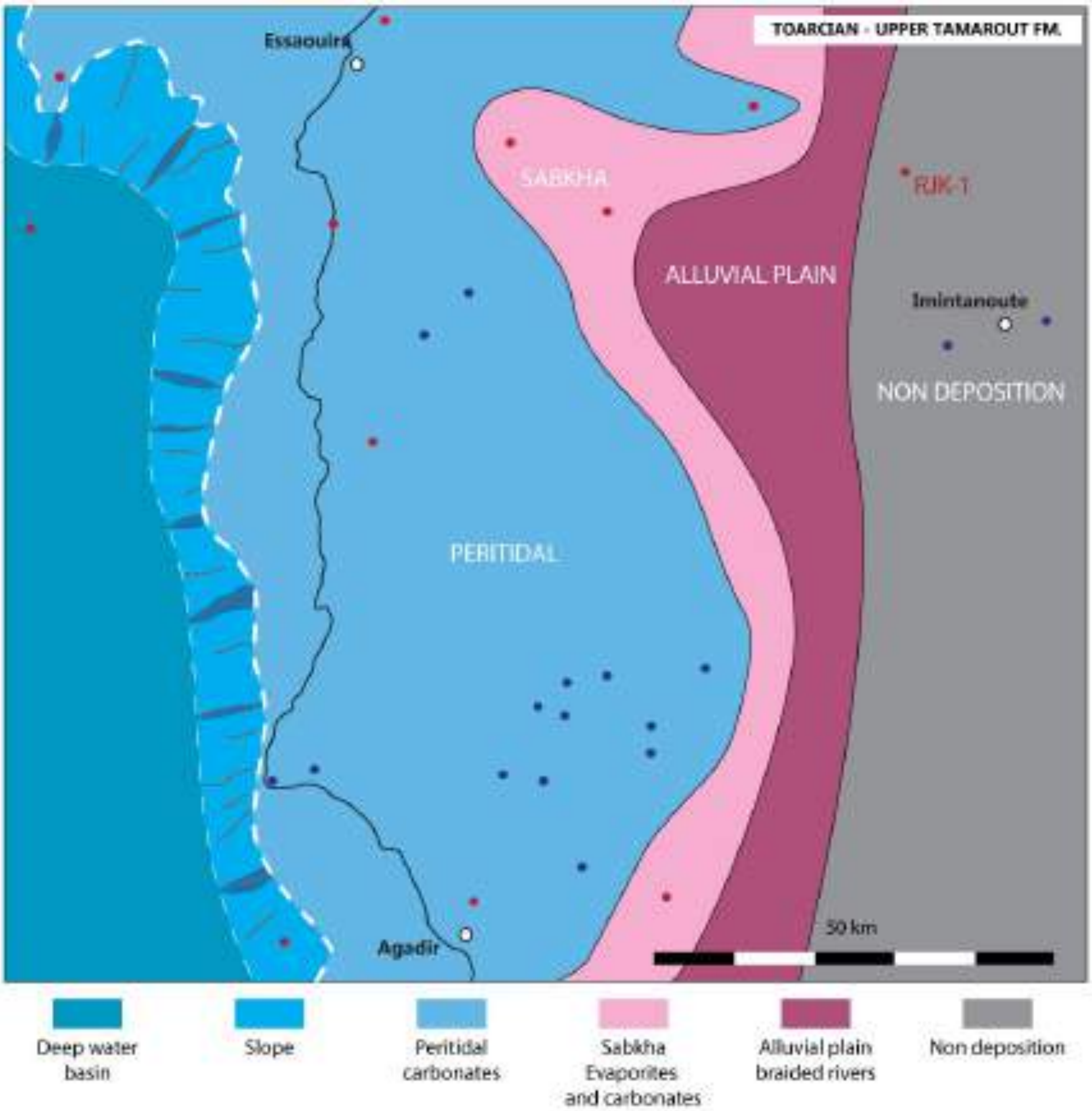


Figure 7.9: Toarcian palaeomap: Upper Tamarout Formation.

MIDDLE JURASSIC – AMESKHOUD FORMATION (Figure 7.10)

During the Middle Jurassic, the east and the south of the basin were dominated by continental deposits. The western part of the Agadir segment presented a peritidal environment, followed by fluvial deposits. The northern part of the basin recorded peritidal carbonates similar to the Tamarout Fm. All these trends are confirmed by the well data.

The siliciclastic influx recorded in this interval has been linked to the middle Jurassic movements of the Anti-Atlas (Charton, 2018; Chapter 4). The pulse of siliciclastic sediments, likely coming from the central part of the Anti-Atlas (Gouiza et al., 2017) has created a regression in the south of the basin.

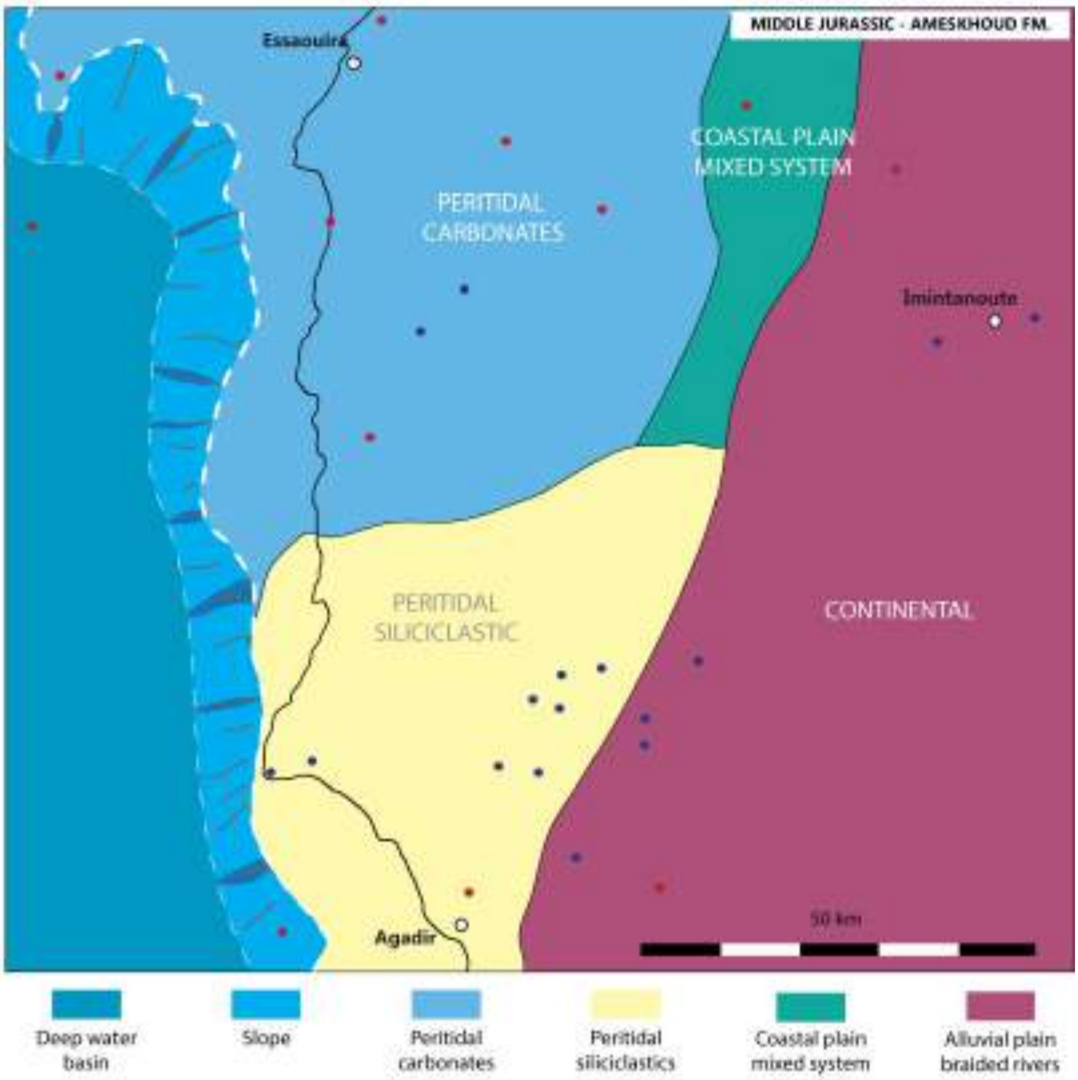


Figure 7.10: Middle Jurassic palaeomap: Ameskhoud Formation.

CALLOVIAN – OUANAMANE FORMATION (Figure 7.11)

During the Callovian a general transgression led to the successive development of inner, middle and outer ramp deposits across the basin. In the Unit 2 of the Ouanamane Formation, the deeper facies present in the Amsittène Anticline and shallower facies present in the south-eastern part of the basin indicate a SE-NW orientation of the basin. This orientation is probably inherited from the siliciclastic influx building rapidly some topography in the south during the Aalenian to Bathonian time. Inner ramp deposits were identified in the eastern part of the basin and in the locality of Imi N'Tanoute. No evidences for sabkha or continental deposits have been found, but they are inferred in the SE of the basin because the Callovian unit is thinning dramatically towards the east.

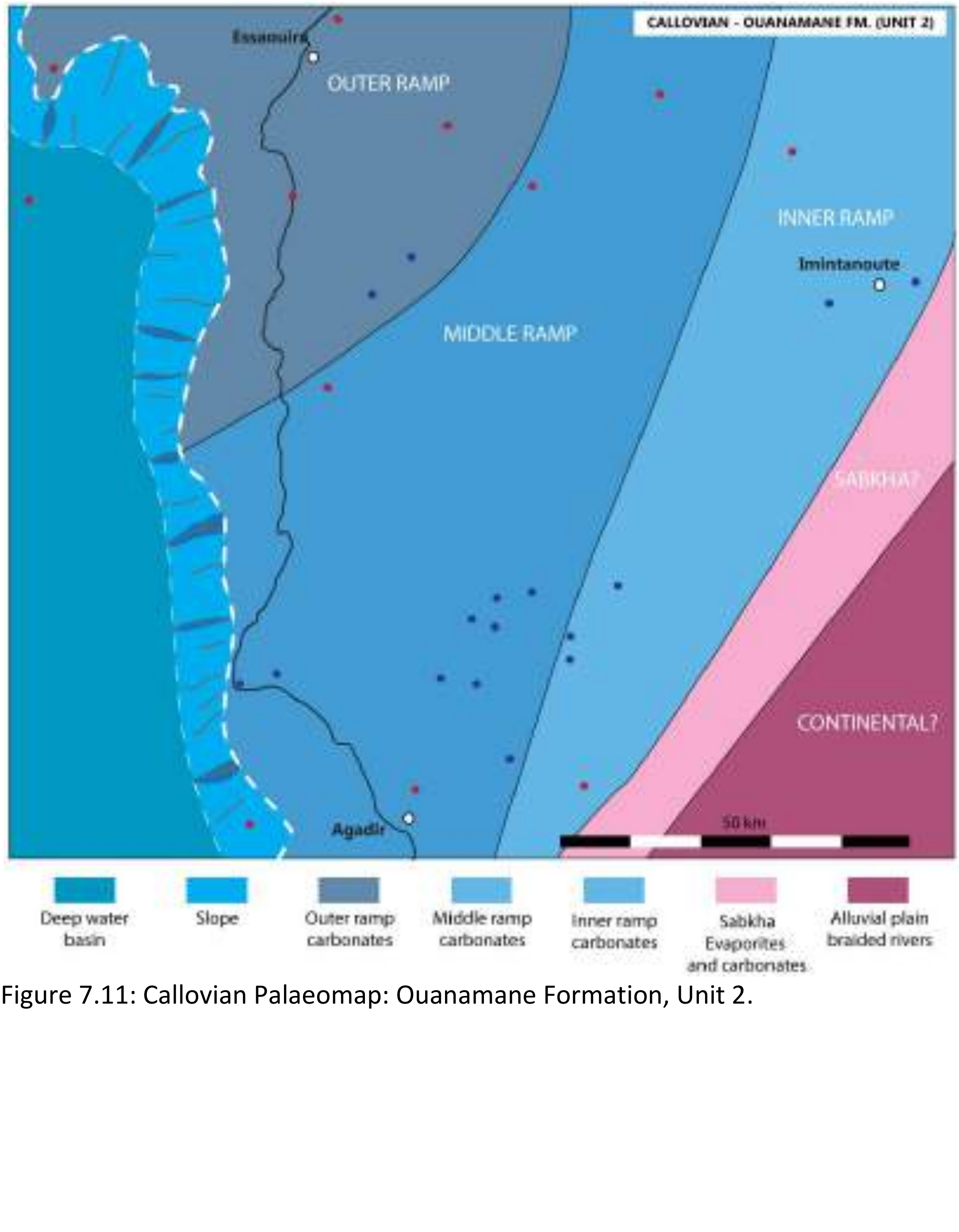


Figure 7.11: Callovian Palaeomap: Ouanamane Formation, Unit 2.

OXFORDIAN – LOWER LALLA OUJJA FORMATION (Figure 7.12)

The Lalla Oujja Formation is dominated by coral buildups in a large part of the basin, but no coral buildups were identified in the well ESS-1X, which have been interpreted as outer ramp conditions, and in the localities of Imi N'Tanoute and in the well EGA-1, some evaporites horizons led to a potential sabkha environment. Middle-ramp deposits were also identified along the Argana Valley, and inner ramp deposits are probable for the well RJK-1, which records dolomitic sandstones.

In the locality of Amsittène, no deposits were recorded for this interval, which led to suspect the onset of salt diapirism forming a palaeorelief in this locality.

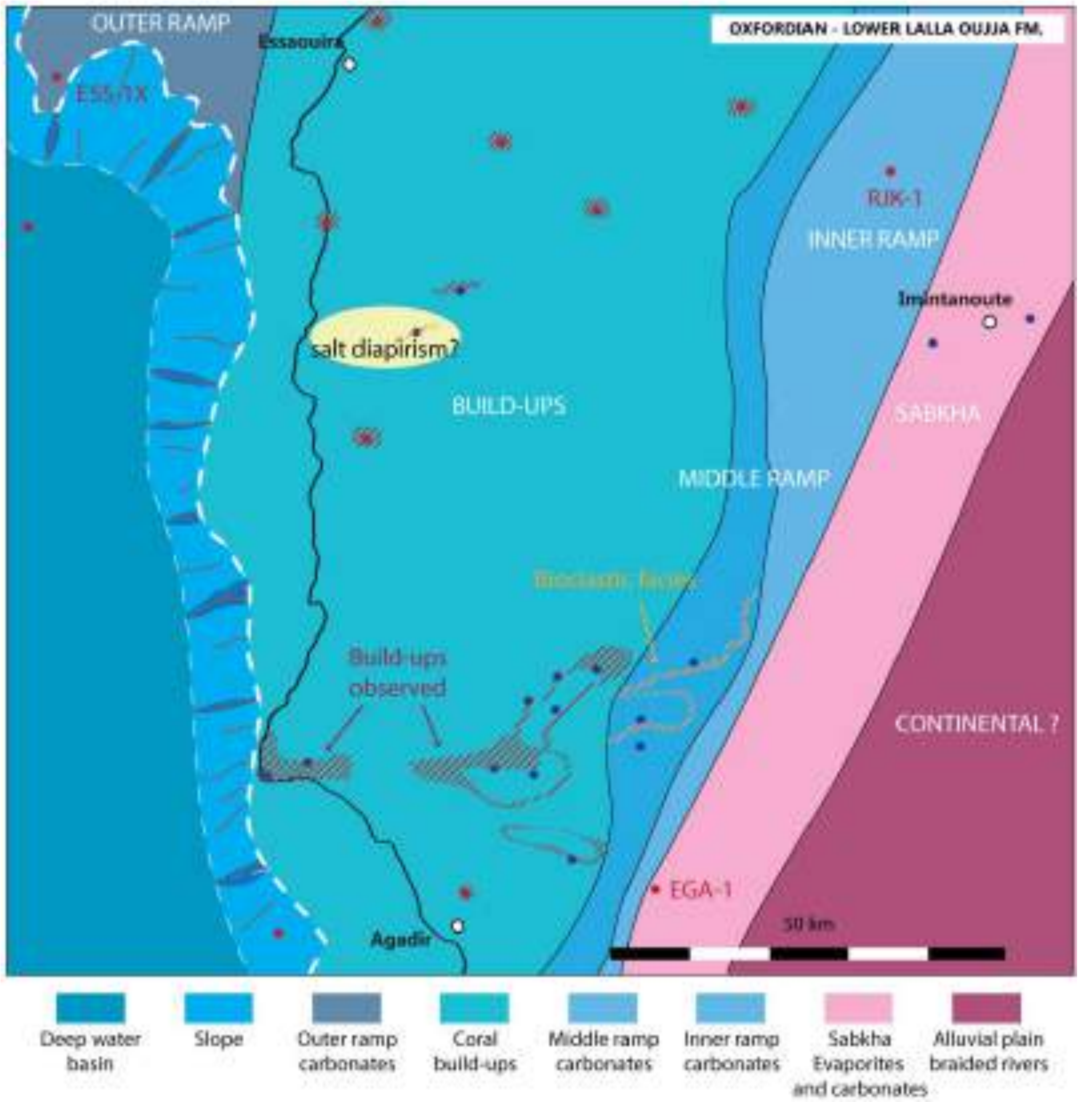


Figure 7.12: Middle Oxfordian Palaeomaps: Lower Lalla Oujja Fm.
(Red rays filling represent the outcrops and wells where the coral buildups were identified).

OXFORDIAN – UPPER LALLA OUJJA FORMATION (Figure 7.13)

The Upper part of the Lalla Oujja Formation records a siliciclastic influx coming from the north of the basin which led to the development of mixed siliciclastic and carbonate shoreface deposits. Some wells in the NE and NW of the basin, and outcrops in the SW do not record this siliciclastic influx, which constrains these deposits to the center and north of the basin. The extent and organisation of these facies indicate that they compose a N-S drawn barrier island or strand plain complex situated in the middle of the inner ramp.

This interval represents the onset of a basin regression which cannot be explained by the global eustasy (Haq, 2018) and could be linked to the upper Jurassic movements of the Western Messeta (Charton, 2018). The pulse of siliciclastic, likely coming from the north would also be explained by erosion of the Western Messeta.

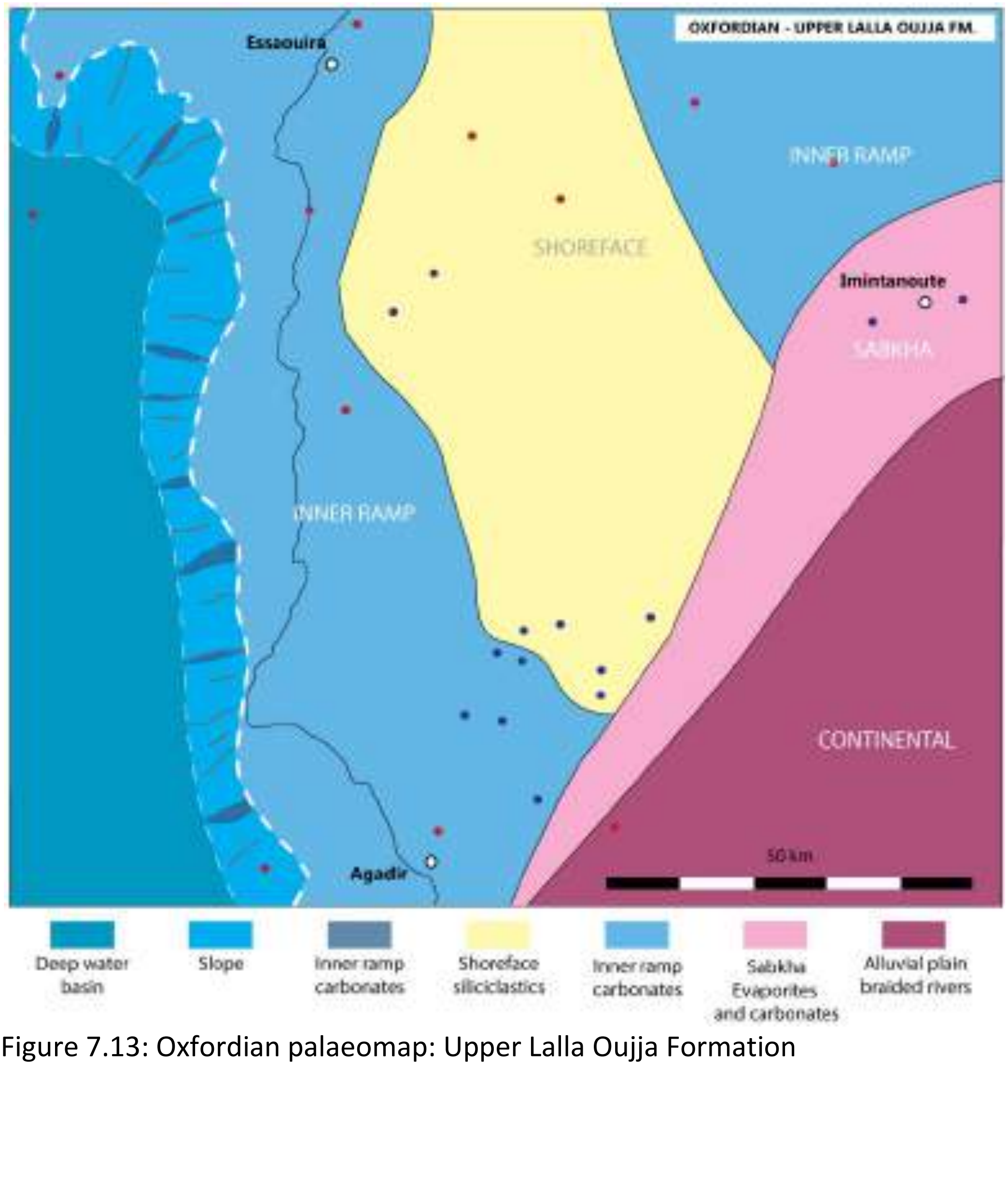


Figure 7.13: Oxfordian palaeomap: Upper Lalla Oujja Formation

OXFORDIAN – IGGUI EL BEHAR FORMATION (Figure 7.14)

The Iggui El Behar Formation records mudflat deposits with shallow inter to supratidal deposits over most of the basin. The north and the south of the basin is richer in evaporites, and a wide sabkha environment has been hypothesized. During the entire Oxfordian, the well EGA-1 presented proximal facies, with deposition of red siliciclastic mudstones and rare anhydrite horizons.

This onset of this formation characterizes a strong regressive event followed by stable conditions during the upper Oxfordian to Kimmeridgian.

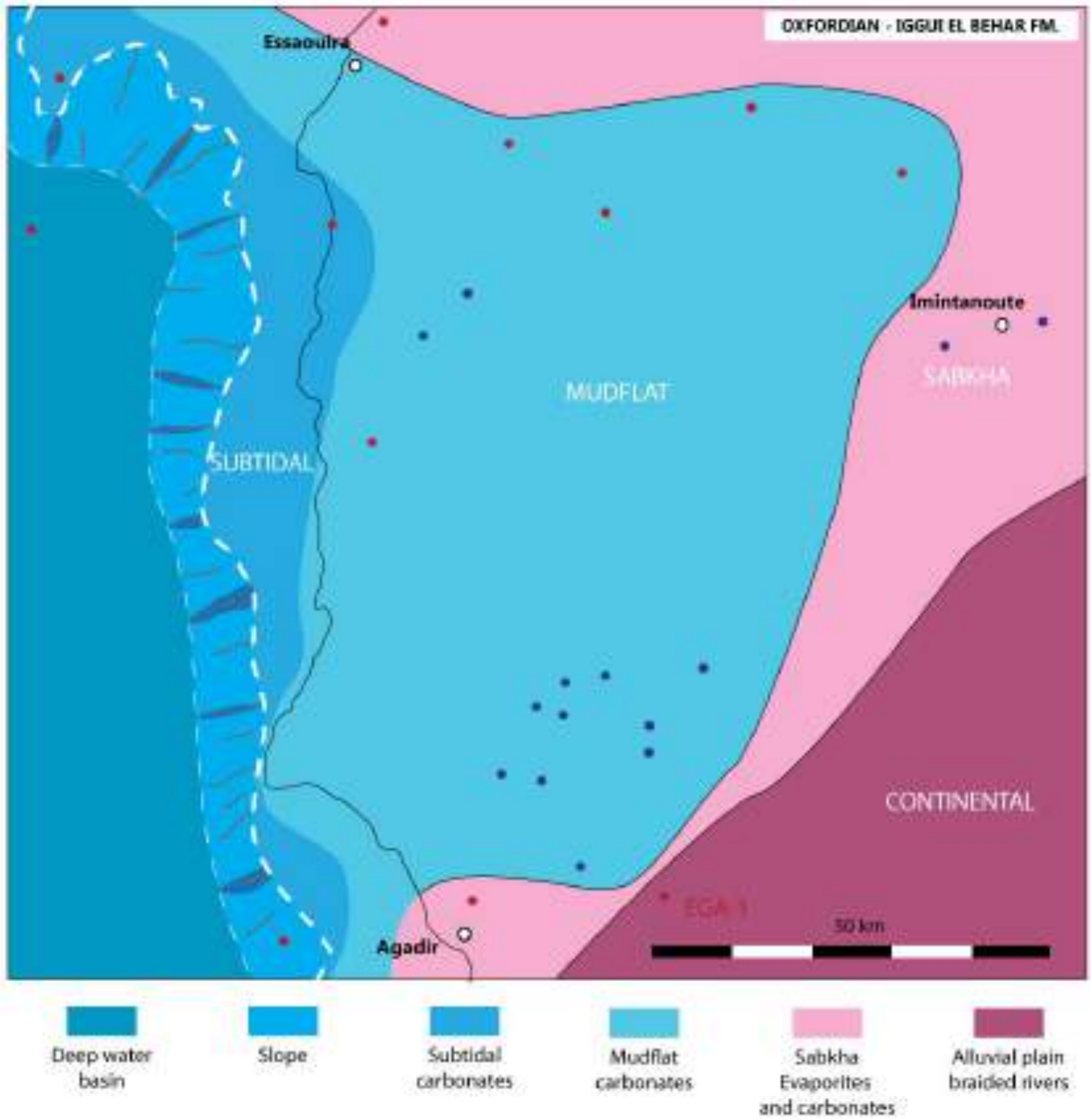


Figure 7.14: Oxfordian Palaeomap: Iggui El Behar Formation

KIMMERIDGIAN – IMOUZZER FORMATION (Figure 7.15)

The Kimmeridgian is dominated by continental siliciclastic in Imi N'Tanoute, but all the north of the basin records important evaporites, as well as the very south. The center of the basin records thick red marls and some dolomitic horizons interpreted as proximal marine peritidal carbonates. The NW of the basin do no record much siliciclastic and are dominated by carbonate deposition (Well TAB-2 and ESS-1X).

This new siliciclastic influx in the basin can also be linked to upper Jurassic movements of the Western Meseta (Charton, 2018). The fluvial systems in the east of the basin brought into the basin the siliciclastic elements that composed the red marls and are present in the carbonate horizons.

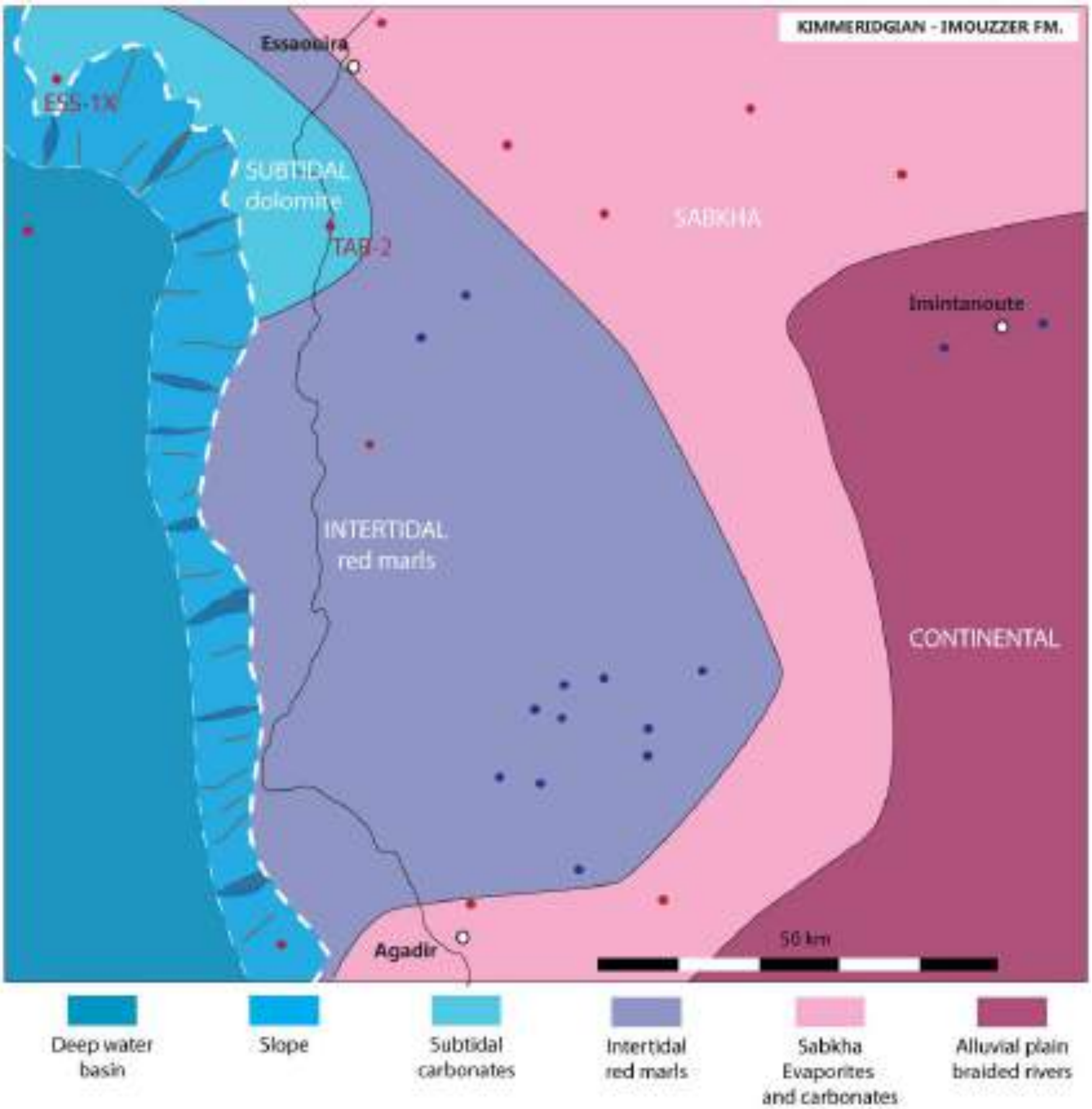


Figure 7.15: Kimmeridgian palaeomap: Imouzzer Formation

TITHONIAN – TISMEROURA FORMATION (Figure 7.16)

The Tismeroura Formation records very thick evaporitic (gypsum units separated by thin mudstones beds) succession in all the NE of the basin, and in the rest of the basin, it passes to peritidal carbonates. It can be noticed that the top of this Formation records a strong transgression, which marks the transition to the Cretaceous.

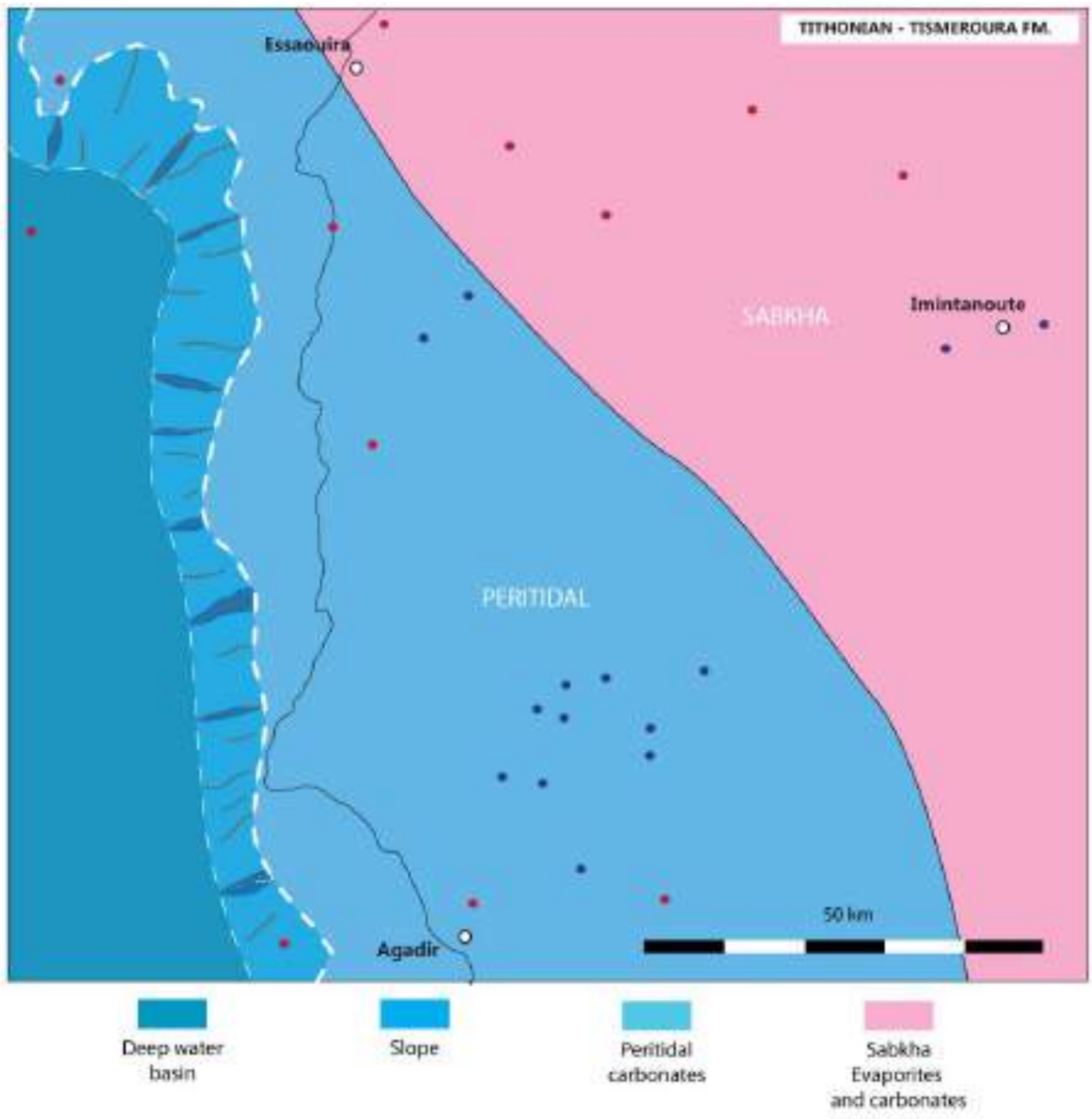


Figure 7.16: Tithonian palaeomap: Tismeroura Formation

7.3 Salt movements - Discussion

In the Central High Atlas, salt diapirism played a crucial role in the carbonate development by creating palaeohighs during the opening of the Jurassic basins (Michard et al., 2011; Saura et al.,2014; Verges et al., 2017; Teixell et al., 2017; Malaval, 2017; Joussiaume, 2017). Onshore in the EAB, multiple NE-SW and E-W anticlines can be observed, and the Tidzi, Jbel Hadid and to some extent the Jbel Amsittène in the North of the basin are also known to be the result of salt diapirs. Some of the anticlinal structures in the Agadir segment are also salt-cored, such as the Cap Ghir Anticline, which has an offshore continuation that can be observed in seismic. Further offshore the same trends are observable, with different salt-tectonic geometries developed in the Essaouira and Agadir segments (Muniz-Pichel et al, submitted).

Towards the south of the basin, the thickness of the Triassic salt decreases (Tari and Jabour, 2013, Muniz-Pichel et al, submitted) which prevented the formation of salt diapirs. Offshore the EAB halokinesis structures significantly influence the Mesozoic stratigraphy. For example, the well CAP SIM 1 (Figure 7.6) was drilled on a structure created by halokinesis and the Jurassic is absent, either not deposited or eroded during the Cretaceous.

Local uplift - Amsittène

The Amsittène Anticline is one of the rare areas where the Jurassic outcrops in the Northern part of the EAB. Observations North to South along the anticline provide valuable information on the effect of local paleogeography.

Across the basin, Unit 3 of the Ouanamane Formation records the deepest facies observed during the Callovian and Oxfordian times, and the base of the following Lalla Oujja Formation is dominated by platy-coral associations, indicative of below storm wave base conditions (Olivier et al., 2012). In the NE of the Amsittène anticline, (section AME), both the marly Unit 3 and the Lalla Oujja coral buildups are present, but to the SW, in the AMCA section (Figure 5.1), a marine siliciclastic unit lies directly on top of Unit 2, and the marly Unit 3 and the Lalla Oujja Formation are absent. The development of this siliciclastic unit, interpreted as a beach strandplain deposit, indicates shallower high energy conditions. This can be explained either by an erosion of the Unit 3 (and the Lalla Oujja Formation), or by lateral facies changes. Both interpretations require local shallow conditions in the SW of the Amsittène Anticline, and suggest the anticline was active at this time.

Erosion hypothesis:

The strandplain deposits of the AMCA section could be equivalent to other high energy siliciclastic deposits recorded elsewhere in the basin just on top of the Lalla Oujja Formation (AME, Assif El Hade, Tikki, and Tizgui N'Chorfa). If this interpretation is correct, it would require a significant reduction in accommodation to erode all the Lalla Oujja Formation and the upper part of the Ouanamane Formation, presumably due to a local uplift.

Strandplain equivalent hypothesis:

The siliciclastic unit of the AMCA section could be correlated to the upper part of the Unit 2, the Unit 3 and the Lalla Oujja Formation of the AME section. This would imply some sort of drainage system separated the two sections, with continuous siliciclastic input in the western section (AMCA) and discontinuous pulses in the eastern section (AME) with establishment of deeper carbonate rich facies between the pulses. This would indicate a siliciclastic source from the NW of the basin and that the AMCA section, after the Middle Callovian records shallower environment than the AME section.

The erosion theory seems more likely, because it is linked to a more regional siliciclastic influx event. Both hypotheses imply strong local accommodation variations. These variations could be linked to vertical movements induced by the growth of the salt diapir, which today outcrops in the Arich Ouzla salt mine, in the western part of Amsittène Anticline, close to the AMCA section. This interpretation would support the strong facies differences between of the AME and AMCA sections, which would then have been partially separated by the growing anticline induced by salt movements. North of the Jbel

Amsittène, the Tidsi diapir also shows evidence of activity during the Jurassic to Lower Cretaceous (Medina et al., 2011).

Conclusions

The comparison of the Amsittène Anticline section AMCA with the south of the basin and the other Amsittène Anticline Formation AME revealed synchronous tectonic activity associatied with underlying salt diapirism at the time of deposition in the locality of the AMCA section between the Middle Callovian and the Upper Oxfordian. This has locally controlled facies distribution. At this time a significant siliciclastic input to the basin from the North may also be associated with more regional inversion of the source areas, as indicated by recent thermogeochronology studies.

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Chapter 8: Conclusions

This final chapter summarises the observations and findings of this project, and states their implications for the carbonate development along the Atlantic margin.

8.1 Synthesis

This PhD project aims to give a better understanding of the different controls on carbonate development along the Moroccan Atlantic Margin during the Jurassic. It relies on the construction and analysis of a coherent stratigraphic framework across the EAB to identify how the sea level, climate, tectonic and environmental variations affect the growth and the evolution of the carbonate platform. By identifying the general stratigraphic trends, the lateral facies variations of different depositional environments along the basin and by refining the stratigraphic and biostratigraphic framework, this project distinguishes the main controls on the facies distribution for different periods of the Jurassic.

Facies associations and depositional environments

32 stratigraphic sections were logged onshore the EAB and the facies described on a centimetre scale. This refined facies identification and the observation of their development and distribution allowed the interpretation of depositional environments. Furthermore, the different factors influencing lateral and vertical facies evolution characterise the environmental changes taking place in the basin.

During the Toarcian, the facies evolved from red siliciclastic mudstones and siltstones eroded by channelized sandstones and conglomerates to limestones with cycles of oolitic grainstones, dolomites, stromatolites and evaporites or dissolution breccias. This marks the change from a continental domain with braided rivers and alluvial plains to a shallow marine domain with less siliciclastic input and more subject to climatic variations and sea level. The Toarcian limestones homogeneity and thickness indicate a period of stable conditions with regular low siliciclastic input and relatively dry weather conditions.

Variations of facies proportions in the limestone facies association suggest shallowing trends marked by thicker evaporites or dissolution breccias after firstly a period of marine carbonates dominated by dolomite units. The upper part of this formation is influenced by higher hydrologic conditions illustrated by cross-bedded oolitic grainstones predomination and deeper environmental conditions exemplified by the disappearance of evaporites and stromatolites. These changes interpreted as the evolution from inter-/supra-tidal to subtidal conditions are the expression of a minor transgression across the EAB.

The peritidal limestones persisted throughout the Middle Jurassic in the north of the basin, in comparison to the south, where siliciclastic deposits dominate during this time period. These deposits are composed of red mudstones, siltstones and sandstones organised in horizontal units where wave ripples prevail until in the upper part of the Middle Jurassic formation. Root traces and incised channels appear, which indicate a shift to continental conditions. Both carbonate and siliciclastic facies associations are shallow marine for the lower and middle part of the formation, but the south of the basin differs as the upper part of the Middle Jurassic formation is characterised by the local regression to continental deposits.

During the Callovian, the facies associations evolve from being dominated by oolitic grainstones, to brachiopod-rich floatstones and rudstones with multiple firmground horizons, up to a marls unit. Each of these facies transitions are highlighted by hardgrounds and mark the evolution to lower energy and deeper environments. The depositional environments identified are inner ramp for the oolites, middle ramp for the brachiopod-rich units alternating with some marly horizons, and open ramp for the marl unit, where deeper fauna such as ammonites and platy-shape corals are more frequent. This Callovian succession is a transgressive sequence which can be followed across the entire basin.

The middle Oxfordian records the development of coral colonies in the EAB. The coral build-ups present an important variability in their sizes, from a few meters in diameter and cm-thick up to 800 m in diameter and 50 m thick. Two main categories of build-ups were identified, 1.platy-coral buildups and 2. diversified buildups. The platy-coral buildups have a large spatial extent but are not very thick and hosted in muddy matrix. The diversified buildups show more relief and are usually the biggest build-ups. These buildups have a platy-coral dominated base, but as the build-ups gain more relief, different species of corals appear and the matrix is enriched in bioclastic material. The facies present between the build-ups are vario-lithic from wackstones to rudstones. Between the larger build-ups, grainstones, floatstones and rudstones are common and composed of reworked coral material derived from the buildups. The middle Oxfordian environment was relatively deep (up to 50 m below sea level) with platy-coral colonies which formed below fair weather wave base, while the diverse colonies formed in shallower waters.

In the north and middle of the EAB, the upper Oxfordian records siliciclastic input directly overlying the coral-rich formation. These coarse sandstones are organised in bed presenting planar horizontal stratification, trough cross stratification and less common HCS. This unit is typically up to 30m thick and becomes progressively enriched in rounded carbonate bioclasts. It is interpreted as foreshore to upper shoreface deposits. Above this unit in the north, or directly above the coral-rich formation in the south of the EAB, a mud-dominated formation develops in the entire basin during the upper Oxfordian to lower Kimmeridgian. The facies association is composed of wackstones to packstones with numerous benthic foraminifera, oncoids, peloids, gastropods,and autogenic intraclasts. The top of many beds are reworked and form mudclasts at the base of the next bed. Desiccation cracks and microkarsts are common features, and the upped part of the formation shows common gypsum crystals growth in a mudstone matrix. These features are characteristic of an intertidal to supratidal mudflat environment, where the tidal transgressions and regressions shape the facies organisation. This formation records a small regression and stable conditions in the EAB.

Age constraints and limitations

In order to identify the stratigraphic relationships between the different formations and correlate these across the basin, dating the geological units is crucial. The dating used in this study originates from three types of sources: literature biostratigraphy and chemostratigraphy (CAMP basalts); new biostratigraphy (review of material collected in previous studies and new collections); relative dating (dating by bracketing and by comparison to the CHA).

In the biostratigraphy review, three main genders have been studied. The ammonite identification provides precise zone and subzone dating for the late early Callovian to the middle Oxfordian. They allow the identification of a potential sedimentary hiatus between the upper Callovian and middle Oxfordian. The brachiopods fauna does not allow us to give precise ages, but the brachiopods associations are compared to the ammonite fauna and the changes in associations are similar and easy to identify across the basin. They constitute a good reference point for correlations. The foraminifera are mainly useful in identifying the upper Oxfordian and lower Kimmeridgian. This time period is rich in large benthic foraminifera which are good stratigraphic markers.

The continental Toarcian formation is dated by correlation with the CHA. The Toarcian in the CHA is characterized by a prominent erosive surface and the development of continental deposits on top of a marine formation. The same succession of environments is observable in the WHA and the marine formation below the continental deposits is dated Sinemurian to Pliensbachian in the literature while the carbonate formation above is dated Toarcian in the literature (Duffaud, 1960; Ambroggi, 1963; Du Dresnay, 1985; Bouaouda, 2007). A Toarcian age is then assigned to this unit as it correlates well with the CHA and is coherent with the bracketed age provided in the literature. The marine and continental siliciclastic of the Middle Jurassic is also dated by bracketing between the Toarcian carbonates and the Callovian carbonates.

In the EAB, the continental deposits do not present any bio-markers to allow dating and correlating them with lateral units or bracketing their ages between two units of known age is the approached followed. This last method only provides an age range for the succession and does not allow us to identify any potential condensed levels or hiatus. Dating by correlation is also a limited approach, as the dating is also used to correlate the different units; this can lead to circular reasoning and should be checked by identifying the main sequences.

Seal level and tectonic controls

The biogenic carbonate factory is dependent, inter alia, of the light accessing the environment, and the non-biogenic part of the carbonate factory depends essentially on the energy of the system. Both light availability (function of the depth and the turbidity) and energy of the environment (function of the depth and the geometry of the system) are factors influenced by the interaction between the eustatic sea level variations and the local tectonic movements.

The late Sinemurian to early Pliensbachian carbonates deposits present an angular unconformity with the overlying Toarcian siliciclastic unit. This unconformity indicates that the pre-Toarcian deposits were tilted prior to the Toarcian erosion which marks the base of the Jurassic in most of the basin. A tectonic uplift brought this unit to the surface and tilted it between the Pliensbachian and the Toarcian. This local uplift is potentially linked to early salt movement as this succession of tilted carbonates and erosive Toarcian siliciclastics is only observed along the Amsittène Anticline which is a salt-cored structure. A hiatus of the middle Callovian to upper Oxfordian in the north of the anticline is also evidence for different periods of salt diapirism and reactivation along this anticline.

During the Toarcian, the evolution from continental deposits to an epiric carbonate platform (Id Ou Moulid Fm.) is due to a deepening of the basin (increase of accommodation space), which has been linked to the global lower Toarcian regression followed by a sea-level rise (Haq, 2017). Evidence of potential Hercynian fault reactivation (Tikki location) during the Toarcian, combined with evidence in the literature (Gouiza et al., 2017; Charton, 2018) of exhumation of the Anti-Atlas during the Lower and Middle Jurassic indicate that the basin-wide erosion followed by deposition of continental siliciclastics during the (lower?) Toarcian is the result of a combination of sea-level fall and active tectonics. The sea-level fall exposes the Lower Jurassic deposits and leads tobasin-wide erosion, while the Anti-Atlas movements and local fault movements provide silici-clastic sediments in the basin and induce local depocenters and reworking of the fault footwalls.

The thickness of the Toarcian carbonate platform and the homogeneity of facies until the Bathonian in the north of the basin indicate stable tectonic conditions in the EAB during this period. The siliciclastic deposits in the south of the basin travelled likely from the Anti-Atlas and were deposited in the Agadir area in a shallow water environment. During the Middle Jurassic, the Ameskhoud Formation evolves from shallow marine to continental deposits. This evolution is interpreted to be due to the accumulation of siliciclastic deposits which reduced progressively the accommodation space in the south of the basin, while in the north of the EAB, farther from the source of the sediments, the same shallow water carbonates deposits lasted until the Bathonian.

The Callovian records a transgression across the entire Basin, with deposition of inner ramp deposits during the lower (?) Callovian, middle ramp deposits during the middle Callovian, and outer ramp deposit during the upper Callovian and until the middle Oxfordian. The lower Callovian transgressive trend is in agreement with the global sea level curve (Haq, 2017), but the persistence of this transgression during the middle and upper Callovian is not. The continuous basin deepening should then be explained by an alternative mechanism than global sea-level rise. Local subsidence of the EAB during the Callovian could explain the continuation of this transgressive trend during the entire Callovian to middle Oxfordian.

The middle Oxfordian coral build-ups are followed by a shallow marine succession during the upper Oxfordian and Kimmeridgian. This regression is against the global sea-level transgression for this period, and can be linked to regional Upper Jurassic movements. Uplift of the Jebilet and Western Meseta are highlighted in the literature (Charton, 2018 and references therein). The uplift of these highs provided part of the siliciclastic elements entering the EAB during this period. Furthermore, the uplift movement was likely not strictly restricted to these massifs but also influenced the EAB and created shallow marine and restricted conditions during this period by reducing the accommodation space.

Climate and environmental controls

The variation in environmental conditions; temperature, precipitations, availability of CO_2 and O_2 are important controls of the biogenic and non-biogenic carbonate factory. During the Jurassic in Morocco, periods of dry climate dominated, which increased the evaporation in restricted environments and induced periods of extensive salt deposition, mainly during the Toarcian and Tithonian.

Periods of uplift are associated with the creation of positive topography and increased sediment delivery to the system. This increase of sediment delivery is a function of the erosion of the relief created, but is also a good illustration of the back-loop influence of highs on climate and erosion. The tectonic movements control locally the climate by creating topographic highs where atmospheric water will condense and induce precipitations, which increases the erosion and the sediments transport.

The Callovian-Oxfordian period shows significant variations in environments of deposition and associated fauna. The evolution during the Callovian of the different fauna can be largely attributed to the variations of sea level. However, the synchronicity of coral build-ups formation during the middle Oxfordian in a similar depth-range than the Upper Callovian requires the development of different environmental conditions. The latest Callovian to early Oxfordian is marked by low temperatures and decrease of P_{CO_2}, while the middle Oxfordian records a rise of the temperature and P_{CO_2} (Dromart et al., 2003). This evolution to more favourable environmental conditions for the coral species allowed them to develop and replace the brachiopods as the dominant species, while they were inhibited by the low temperatures and CO_2 availability during the Callovian.

Implications

The understanding of the different controls affecting the evolution of the Jurassic carbonates is important for understanding the Moroccan Atlantic margin as part of the Atlantic margin, but also consider the influence of the Atlas System andhinterland movements.

The link between the hinterland uplifts and the siliciclastic influx in the basin can give the extent of the potential siliciclastic reservoir. For example, during the middle Jurassic, the siliciclastics coming from the Anti-Atlas did not reach the Amsittène area. Therefore this potential reservoir unit can be constrained to the Agadir area or tracked towards the south along the margin, but should not be expected further north. Similarly, the upper Oxfordian and Kimmeridgian siliciclastics are entering the basin from the NE and the quality of any potential siliciclastic reservoir for this period is likely to decrease towards the south of the basin.

The variation of environmental conditions during the Callovian-Oxfordian indicates that the optimal period for coral build-ups or coral reef growth is the middle to late Oxfordian. This optimal period is the same as in Central Europe and can likely be extended to Nova Scotia which was close and therefore in similar climatic? environmental conditions. The observation of synchronous coral build-ups over a large area can also be transferred to the Scotian Basin, where the build-ups can be explored for not only at the edge of the margin, but generally immediately above the marls of the Misaine Member.

8.2 Conclusions

This study has investigated the Jurassic succession in the Essaouira-Agadir Basin (EAB). Jurassic carbonates are proven but poorly constrained reservoirs offshore Morocco and along the conjugate margin of Nova Scotia. In order to understand the nature, extent, geometries and capture the reservoir potential of the different formations, this study has surveyed and interpreted multiple outcrop sections and 3D geometries in the center and the margin of the basin. These observations were supplemented by onshore and offshore well data leading to the development of palaeogeographic maps to illustrate the lateral and temporal variations. The research objectives were addressed and the following conclusions can be drawn from the results:

Objective 1: Reinterpret and refine the basin stratigraphy in order to gain a better understanding of the depositional environments.

This study revised and refined the general Jurassic stratigraphy across the EAB, by reviewing the previous studies conducted on the basin and completing extensive field work and sampling, supported by petrographic and biostratigraphic analysis. The aim of these observations was to assess the lateral and vertical facies variations, their extent across the basin, and to use these new considerations to reinterpret the depositional environments at basin scale. The following points have been clarified:

- The open-marine carbonates of the Upper Sinnemurian to Lower Pliensbachian Arich Ouzla Formation have been uplifted, tilted then partially eroded during the Toarcian, recording a phase of tectonic inversion in the basin.
- The Toarcian Amsittène Formation deposits are erosive across the entire basin, cutting down to the CAMP basalts and the underlying Trias. They are dominated by fluvial deposits, and alluvial fans controlled by local fault reactivation.
- The Toarcian Tamarout Formation records cyclic peritidal deposits, with open and more restricted phases during which evaporitic units developed across the basin. They are locally replaced by dissolution breccias. It has been divided into three units, the first unit record the initial transgression, the second unit is more proximal implying a regression, and the third unit records a small scale transgression.
- The Aalenian to Bathonian Ameskhoud Formation presents a strong S-N proximal-distal orientation, with deposition of fluvial deposits in the south, siliciclastic peritidal later replaced by fluvial in the center of the basin, and peritidal carbonates in the north. These observations wave been linked to recent thermochronology studies and the Anti-Atlas, which was uplifted during the middle Jurassic is believed to be the main sediment source for these deposits which would explain the proximal to distal facies variations.
- The Callovian Ouanamane Formation records a global transgression and the development of ramp deposits. Three units can be distinguished, separated by hard-ground levels recording periods of transgression across the basin. The formation contains inner ramp oolitic deposits replaced by middle ramp brachiopods rich beds and by a thick unit of open ramp marls. The presence of

small platy-coral buildups and biostromes have also been observed and described in this unit for the first time.

- The Middle Oxfordian Lalla Oujja Formation is dominated by coral buildups deposits. Different bioherms sizes, geometries, and facies variations have been identified. The larger buildups (400-700 m large) show a typical evolution from low relief platy coral colonies to higher relief diversified colonies and are fringed by rubble facies organised in clinoforms coming down the structure. Smaller sponges, microbialite and platy-coral biostromes are also present. Towards the east of the basin, the coral buildups pass into mid-ramp, higher energy deposits. The upper part of this formation records a regression with the deposition of inner ramp deposits in the south and upper shoreface deposits with increased silicilastic input in the north of the basin.
- The Upper Oxfordian to Kimmeridgian Iggui El Behar Formation records a sharp regression on top of the Lalla Oujja Fm., with deposition of restricted to continental deposits in the east and development of peritidal mudflat across the rest of the basin. The top of this formation records a sequence boundary with the development of supratidal conditions.
- The Kimmeridgian Imouzzer Formation records fluvial deposits on the eastern margin feeding thick red marls which alternate with sandy, peritidal dolomites.
- The Tithonian Tismerroura is dominated by sabkha conditions in the eastern part of the basin with thick evaporites deposited, whereas the center of the basin records more peritidal conditions and development of thick extensive stromatolites horizons.

Based on these observations and conclusions, and their correlation to onshore and offshore well data, 10 paleogeographic maps have been created for the key intervals during the Jurassic.

Objective 2: Collect new biostratigraphic material and evaluate the dating of the succession to attempt to improve precision and regional correlation

New biostratigraphic material was collected in the field, including 21 identifiable ammonites, 15 echinoderms, and over 500 brachiopods. This assemblage was completed by new identification of ammonites collections from previous studies alongside foraminifera assemblages identifications. The integrated study of the ammonite, brachiopod and foraminifer faunas indicates has led to the development of a refined biostratigraphic framework for the Moroccan Atlantic Margin:

- The Late Bathonian age formerly proposed for the lower part of the Ouanamane Fm. is dismissed.
- The lower and middle members of the Ouanamane Fm. are of early to middle Callovian age.
- The lower part of the upper member is of late middle Callovian age.
- The Callovian-Oxfordian transition is situated in the upper part of the third member and is affected by condensation and/or hiatuses.
- The uppermost part of the formation is of early middle Oxfordian age, which has been inferred for the onset of the buildups of the Lalla Oujja Fm. across the basin.
- The Oxfordian-Kimmeridgian transition occurs in the middle of the Iggui El Behar Formation.

Objective 3: Identify any reservoir units, their extent, and lateral facies variation.

In the middle to upper Jurassic, two different reservoir units were identified. Their ages have also been better constrained by the biostratigraphic framework:

- The lower Callovian Iggui N'Tarhazout Oolite Member of the Ouanamane Formation is a thick (up to 60 m), porous and laterally continuous body and forms the first potential reservoir interval. It extends across all the central, northern and southern parts of the basin, before disappearing towards the east along the boundary of the Argana Valley. It is part of a basinal transgression topped by a thick marls unit which can act as a seal.
- The middle Oxfordian Lalla Oujja Formation is dominated by highly dolomitizied coral bioherms, with good inter and intracristaline porosity. The coral-rich

floatstone facies occurring between the bioherms is also extensively dolomitised and extend the reservoir potential of this formation. In the North of the basin, this formation is overlain by siliciclastic-rich shoreface deposits, which could also constitute an extension of the reservoir formation. This Formation is overlain by a potential seal unit composed of tight mudstones and red marls of the Iggui El Bahar and Imouzzer formations.

Objective 4: Establish the timing of siliciclastic influx into the basin and constrain the controls on their origin and their influence on the carbonate succession.

Multiple periods of siliciclastic influx have been identified across the EAB. The carbonate facies are mostly presenting low quartz content, and alternate with fluvial and marine siliciclastic units. The timing, origin and controls of the detrital elements have been studied by identifying the lateral facies variations across the basin and linking them to recent thermochronology studies showing that throughout the Lower and Middle Jurassic, the Jebilets, the Massif Ancien de Marrakech (MAM) and the Anti-Atlas were uplifted and eroded.

- The Toarcian records the first strong siliciclastic input in the basin, with deposition of an erosive fluvial succession in most localities. This unit has been tentatively dated by correlation with the Central High Atlas. The MAM and Jebillet are the more likely source terrains for this interval.
- The middle Jurassic records a second important siliciclastic influx into the southern part of the basin. The orientation of the depositional environments compared to the movements of the hinterland indicates that the source of these deposits is the Anti-Atlas. This siliciclastic influx is affecting the south of the basin only, preventing carbonate development, but in the north and offshore, the peritidal carbonates were still dominating.
- The middle to Upper Oxfordian records a siliciclastic influx from the north, leading to the development of mixed shoreface deposits. A second stronger siliciclastic pulse occurred during the Kimmeridgian and records a strong east-west proximal-

distal component. These elements indicate the Western Meseta as a potential source.

Objective 5: Compare the Jurassic stratigraphy and depositional sequences in EAB with Nova Scotia and the Central High Atlas

During the Jurassic, the EAB was at the junction between two systems; the Atlas and the Central Atlantic. The EAB belongs to the Atlantic system, but share a common history with the CHA, part of the Atlas system.

- By comparing the EAB and the CHA, similar trends during the lower Jurassic were observed. Carbonate development occured in both basins during the Sinnemurian and Pliensbachian, before an erosive event recorded by the deposition of fluvial siliciclastics during the Toarcian. This erosive phase has been dated as Toarcian in the CHA and by comparison and bracketing, the same age is inferred for the EAB Amsittène Fm. This uplift and erosion event is believed to be a major event for the Western High Atlas.
- The stratigraphies of the Scotian Basin and the Moroccan Atlantic Basin have been recognized to be remarkably similar during the middle to upper Jurassic. The Ouanamane Formation is the equivalent to the Scatarie Members of the Abenaki Formation in Nova Scotia; and the Lalla Oujja Formation is equivalent to the Baccaro Member. For all these comparisons, the facies and the dating correlates closely, and the observations made in the EAB can be used to predict the reservoir extension offshore Nova Scatia.

Objective 6: Resolve the relative controls of climate, tectonic and eustasy on the development of a carbonate-dominated system that develops along a passive margin

During the Jurassic, the Essaouira-Agadir Basin recorded various carbonate and siliciclastic-dominated environments. The influence of hinterland movements, climatic variations, and eustacy has been addressed here:

- The unconformity between the Arich-Ouzla open marine carbonates and the Toarcian fluvial deposits of the Amsittène Formation indicates that the lower Jurassic sedimentation was strongly influenced by local uplifts.
- The South-North proximal-distal trend of the siliciclastic deposits during the middle Jurassic indicates the Anti-Atlas as being the preferential source of the siliciclastics during this period. By comparing the south of the basin to the north, where the carbonate sedimentation continued during all the middle Jurassic, it appears that the controls of the accommodation along passive margins can be strongly influenced by movements of the hinterland and do not depend solely on global eustatic changes.
- During the Callovian, the global sea-level rise induced a transgression in the basin and more generally along the Atlantic Margin.
- During the late Callovian to middle Oxfordian, climatic and eustatic variations were controlling the carbonate sedimentation along all the Tethys and Central Atlantic domain.
- The evolution of the stratigraphy along the Amsittène Anticline also shows local controls on the sedimentation linked to salt diapirism.

The general climatic and eustatic trends are the primary control on development of carbonates on the passive margin, but local uplifts, salt-tectonic and hinterland movements can also control the accommodation by creating highs and increasing the rate of clastic input into the system.

8.3 Further investigations

8.3.1 Provenance

Provenance areas for the multiple fluvial deposits have been proposed in this study. To constrain and identify the source of these deposits would be important to build a more robust framework and constrain the wider palaeogeography of the area. A complete source-to-sink study should also be carried out for the Toarcian deposits as they could constitute a potential reservoir further offshore. Samples were collected in numerous locations for the Toarcian, middle Jurassic and Kimmeridgian intervals to undergo provenance analysis

8.3.2 Dolomitization

The two Jurassic carbonates identified as reservoir units are widely dolomitised. But dolomitization processes are complex and understanding their influence on the reservoir properties of the formations studied is very important to assess the reservoir potential of the units. Such a study is being undertaken at the University of Manchester by NARG student Nawwar Al Sinawi.

8.3.3 Source rock

During the multiple field seasons, two intervals have been identified with a potential for source rock development. The Middle Oxfordian interval, just before the development of the coral buildups is in Cap Ghir and Assif El Hade, is dominated by thick black mudstones. TOC analyses should be run on this interval for further investigations. The second potential source rock interval is the Tithonian. The stromatolites units are very kerogenic along the Paradise Valley, and up to 1.5 m thick. The stromatolite units should be mapped out across the EAB, with TOC measurements undertaken to quantify the source potential of this unit.

8.3.4 Biostratigraphy

Despite multiple biostratigraphic field seasons, the collection of ammonites turned out to be scarce. New road works have been engaged in the Anklout Anticline during summer 2018, which gives access to new sections located in the area where the collection of ammonites from the literature, that have been revised in this study, are originating. Collecting new ammonites in this locality, as well as in the North of the Imouzzer Anticline, which has been under-explored could help to refine further the biotratigraphy and confirm of infirm the hypothesis of a hiatus or condensed section in the late Callovian to middle Oxfordian.

8.3.5 Charophytes

Charophytes have been found in two locations and have never been reported or studied in this basin before. Identification of the charophyte fauna can be a good indicator to distinguish between continental and restricted marine environments. These fauna could also be compared to the charophytes fauna observed in the Lusitanian Basin (Pereira at al., 2003) and give good indications on the palaeogeographic implications and the age of the transition between the Lalla Oujja and Iggui El Behar Formations

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