

Diapiric ridges and minibasins in the Central High Atlas: impact on geometries and facies distribution (Lower-Middle Jurassic, Morocco)

Rémi Joussiaume^{1,*}, Manon Malaval¹, Philippe Razin¹, Carine Grélaud¹, Juan Diego Martín-Martín², Eduard Saura^{3,4}, Mar Moragas⁴, Jaume Vergés⁴, Grégoire Messenger⁵ and David Hunt⁵

¹ Univ. Bordeaux, CNRS, Bordeaux INP, EPOC, UMR 5805, F-33600 Pessac, France

² Departament de Mineralogia, Petrologia i Geologia Aplicada, Universitat de Barcelona, C/ Martí i Franquès s/n, 08028 Barcelona, Spain

³ Departament de Geologia, Universitat Autònoma de Barcelona

⁴ Group of Dynamics of the Lithosphere, Geosciences Barcelona, Geo3Bcn-CSIC, Barcelona, Spain

⁵ TPD RDI, Equinor, Bergen, Norway

Received: 16 January 2025 / Accepted: 29 September 2025 / Publishing online: 21 January 2026

Abstract – The Central High Atlas of Morocco is the result of the Alpine tectonic inversion of an intracontinental Mesozoic basin, the infill of which started with thick Upper Triassic clay and evaporite deposition. This slightly deformed mountain range is characterized by narrow ENE-WSW ridges that expose a Triassic core. These ridges bound unique synclines filled by thick Jurassic successions showing well-preserved halokinetic depositional sequences, interpreted as salt-related minibasins. This paper aims to describe and analyze the evolution of diapiric movements and their interactions with Lower and Middle Jurassic carbonate dominated sedimentation in the Imilchil area, which corresponds to the axial part of the Atlas basin. Four diapiric ridges have been identified in this region: the Tasraft, Tassent, Ikkou, and Amagmag, which expose Triassic shales and are intruded by Triassic and Jurassic magmatic rocks. These ridges delineate three minibasins: Ikassene, Lakes (Lake Plateau), and Tilmi (Ikkou), each 5 to 10 km wide, preserving over 6,000 m of Upper Liassic and Dogger deposits. The Atlas basin is filled with five third-order transgression-regression cycles organized as one regressive megasequence. This long regressive trend is expressed by the succession of three depositional systems: i) a distal to proximal carbonate ramp system from the Toarcian to Late Bajocian; ii) a shallow mixed carbonate-siliciclastic system from the Late Bajocian to Early Bathonian; and iii) a fluvial system from the Bathonian to Early Callovian. Throughout this period, the development of coeval diapiric ridges significantly influenced these sedimentary systems, though the intensity of the diapiric activity varied over time. During the Late Bajocian, ridge growth slowed considerably amid the peak progradation of the carbonate systems. It then increased significantly from the latest Bajocian onward, due to a drastic rise in siliciclastic flux and sedimentation rate in the Atlas basin. The effects of diapirism on sedimentation are observed regionally (several kilometers), as minibasins evolve as depocenters, each exhibiting distinct subsidence rates and sedimentary thicknesses. Despite these local variations, the regional scale paleogeography and the sequential organization of the deposits remained largely unaffected. On a smaller scale (hectometers to kilometers), diapiric structures are always characterized by halokinetic depositional patterns. Internal angular unconformities only occur in very shallow-marine to continental deposits during periods of maximum ridge growth and subaerial exposure (notably in the Bathonian). At this scale, diapiric movements lead to local facies variations within shallow-marine deposits, fostering the development of bioconstructed or grainy carbonate sediment bodies along the ridges. However, in outer ramp environments, the submarine topographies created by ridge growth were generally insufficient to induce similar facies variations.

Keywords: Diapirism / minibasin / carbonate system / mixed system / Morocco / Jurassic

* e-mail: remi.joussiaume@gmail.com

1 Introduction

Salt tectonics include all types of vertical, horizontal, extensive and compressive movements and gathers all deformations due to salt displacement (Brun and Fort, 2008), at the basin scale as well as the local scale around isolated structures. Salt tectonics determine the structural style of many sedimentary basins throughout the world (Hudec and Jackson, 2007). These basins record the rise of diapirs and, when the layer of salt is thick enough, the development of salt withdrawal minibasins, surrounded by salt walls (Jackson and Talbot, 1991; Hudec and Jackson, 2009). The formation of these structures and their impact on neighboring sedimentary geometries have been widely studied through analog models (e.g., Faugères and Brun, 1984; Vendeville and Jackson, 1992 (a), 1992(b); Jackson and Vendeville, 1994), often compared to seismic profiles (e.g., Fort *et al.*, 2004). Several field-based studies have also handled these problematics, mainly focusing on the geometries resulting from interactions between diapirism and deposition (e.g., Alsop *et al.*, 2000; Rowan *et al.*, 2003; Giles and Rowan, 2012). In addition to the geometry of the flanking sediments, few studies analyzed the influence of diapirism on sedimentary facies distribution within siliciclastic systems (e.g., Shelley and Lawton, 2005; Banham and Mountney, 2013; Ribes *et al.*, 2015), or carbonate and mixed systems (e.g., Orszag-Sperber *et al.*, 1998; Bosence, 1998, 2005; Giles *et al.*, 2008; Purkis *et al.*, 2012; Rowlands *et al.*, 2014; Poprawski *et al.*, 2014). However, the diversity of sedimentary systems, salt-related structures, and tectonic, stratigraphic and paleogeographic settings where the interactions between diapirism and sedimentation take place, requires a large number of case studies. Therefore, the analysis of the interactions between diapirism and sedimentation remains a research topic to further develop. During the Late Triassic the Atlas intracontinental basin recorded the deposition of over 1 km of clays, siltstones and evaporites (Benaouiss *et al.*, 1996; Courel *et al.*, 2003; El Arabi, 2007). These deposits were mobilized during Early and Middle Jurassic to form diapiric structures in the Atlas basin, from the Atlantic margin (Cochet *et al.*, 1970; Hafid *et al.*, 2006) to eastern Maghreb (Tunisia and Algeria) (Perthuisot *et al.*, 1999; Bracene *et al.*, 2003; Hlaïem, 1999; Zouaghi *et al.*, 2013; Martín-Martín *et al.*, 2017; Khomsi *et al.*, 2022) (Fig. 1a).

In the Central High Atlas of Morocco, syn-rift and post-rift diapir growth is responsible for the formation of elongated diapiric ridges separating large minibasins (Saura *et al.*, 2014; Malaval, 2016; Joussiaume, 2016; Vergés *et al.*, 2017; Martín-Martín *et al.*, 2017; Moragas *et al.*, 2018; Teixell *et al.*, 2017, 2024) (Fig. 1b). The intricate salt-related Jurassic tectono-sedimentary evolution was analyzed in Saura *et al.* (2014) for the first time. A series of following papers were published showing different aspects of the amazing Jurassic tectono-sedimentary interplay between halokinetic depositional sequences infilling unique minibasins bounded by long salt walls somehow controlled by extensive basement fault systems. Their state of preservation is exceptional due to the very low shortening rate of the Atlas Mountains during the alpine orogeny (Froitzheim *et al.*, 1988; Brede, 1992; Laville and Piqué, 1992; Teixell *et al.*, 2003).

Salt diapirs and related structures are dramatically exposed in the heart of the Central High Atlas basin in the Imilchil region (Michard *et al.*, 2011; Saura *et al.*, 2014). In this area, four narrow and elongated diapiric ridges, several tens of kilometers long, oriented NW-SE to WNW-ESE, delimit three minibasins 5 to 10 kilometers wide. The development of these ridges and minibasins occurred during the Early and Middle Jurassic. The minibasins were filled up with a thick succession of carbonate, mixed and finally siliciclastic deposits during the large-scale tectonic and sedimentary evolution of the Atlas basin.

The aim of this paper is to study the evolution of the diapiric ridges and associated minibasins of the Imilchil area (Central High Atlas, Morocco), and to determine the interactions between diapirism and sedimentation at various scales. Specifically, the objectives of this study are: 1) to characterize the evolution of diapiric movements over space and time through a detailed study of synchronous sedimentary facies and thicknesses, and 2) to determine how and at which scale these movements were recorded by sedimentary dynamics (sedimentation types and rates, relative sea level variations), and by diapiric dynamics (intensity of diapir movements, diapir location).

2 Geological settings

2.1 Regional setting

The Atlas system is a WSW-ENE intracontinental mountain range, extending over 2,000 km in northwest Africa between the Atlantic Ocean and the Mediterranean Sea (Michard, 1976; Mattauer *et al.*, 1977), including the Western High Atlas, the Central High Atlas, the Saharan Atlas and the Tunisian Atlas. It is the result of the inversion of a Mesozoic intracontinental tectonic basin initiated during the Triassic rifting (Brede *et al.*, 1992; Laville and Piqué, 1992; Beauchamp *et al.*, 1996; Teixell *et al.*, 2003, 2024).

The Atlas system of Morocco is made of two distinct segments, the WSW-ENE trending High-Atlas and the SW-NE trending Middle Atlas, both characterized by the outcrop of gently folded Triassic and Jurassic sedimentary sequences. These strata thin out progressively westward, favoring the exposure of Variscan domains in the western part of the Moroccan Atlas. This western region called the *West Moroccan Arch (WMA)* was not subsiding during the Mesozoic, thus separating the actual Atlas basin from the Essaouira basin located on the Atlantic margin (Fig. 2). The Central High Atlas is now tectonically inverted and uplifted, structured in wide ENE-WSW minibasins, separated by narrow diapiric ridges, which control the emplacement of Triassic and Jurassic magmatic rocks. It is bounded by two stable domains where Jurassic series are not preserved: the Moroccan Meseta domain to the northwest, part of the “West Moroccan Arch” (Jabour *et al.*, 2004), and the Anti-Atlas domain to the south. The Jurassic Atlas basin was connected to the Tethys Ocean, displaying present-day a WSW to ENE polarity and forming a tectonic rifted system and a geographic gulf. This relatively narrow gulf closed westward on the West Moroccan Arch, and was therefore disconnected from the

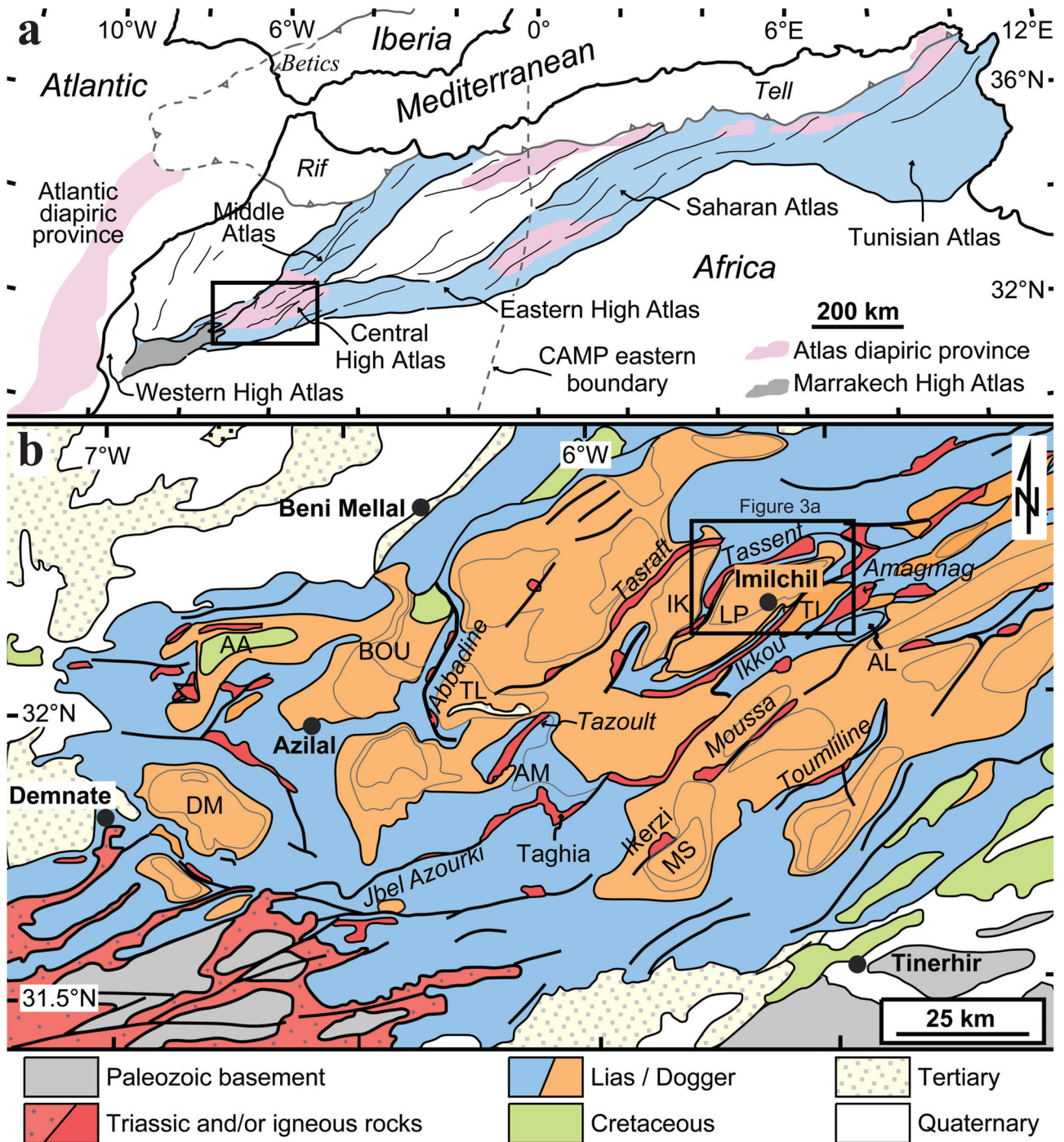


Fig. 1. a) Location of the Central High Atlas (box) and of the main diapiric provinces (pink) in the Atlas system (blue) (Saura *et al.*, 2014). b) Location of the studied Imilchil area (box) and of the main diapiric structures of the central High Atlas (e.g., Tasrafi) separating numerous mini-basins [Almghou (AL); Amezraï (AM); Demnate (DM); Ikassene (IK); Tilmi (TI); Lake Plateau (LP); Bin El Ouidane (BOU); Tillouguit (TL)] (modified from Saura *et al.*, 2014).

passive Atlantic margin, at least partially. The Atlas basin developed as the result of several rifting phases between the middle to late Triassic and the lower and middle Jurassic prior to the coeval opening of the Central Atlantic and Ligurian Tethys oceans (Ellouz *et al.*, 2003; Laville *et al.*, 2004; Wilmsen and Neuweiler, 2008; Lachkar, 2000; Moragas *et al.*, 2016).

During the Triassic, a first rifting phase triggered the formation of half-grabens along ENE- WSW normal faults (Piqué *et al.*, 2000; Domènech *et al.*, 2015). Triassic syn-rift sediments are made of continental clays and sandstone, overlies by clays and evaporites (Michard, 1976; El Arabi, 2007; Frizon de Lamotte *et al.*, 2008). The Triassic filling of the Atlas basin

EARLY-MIDDLE LIASSIC

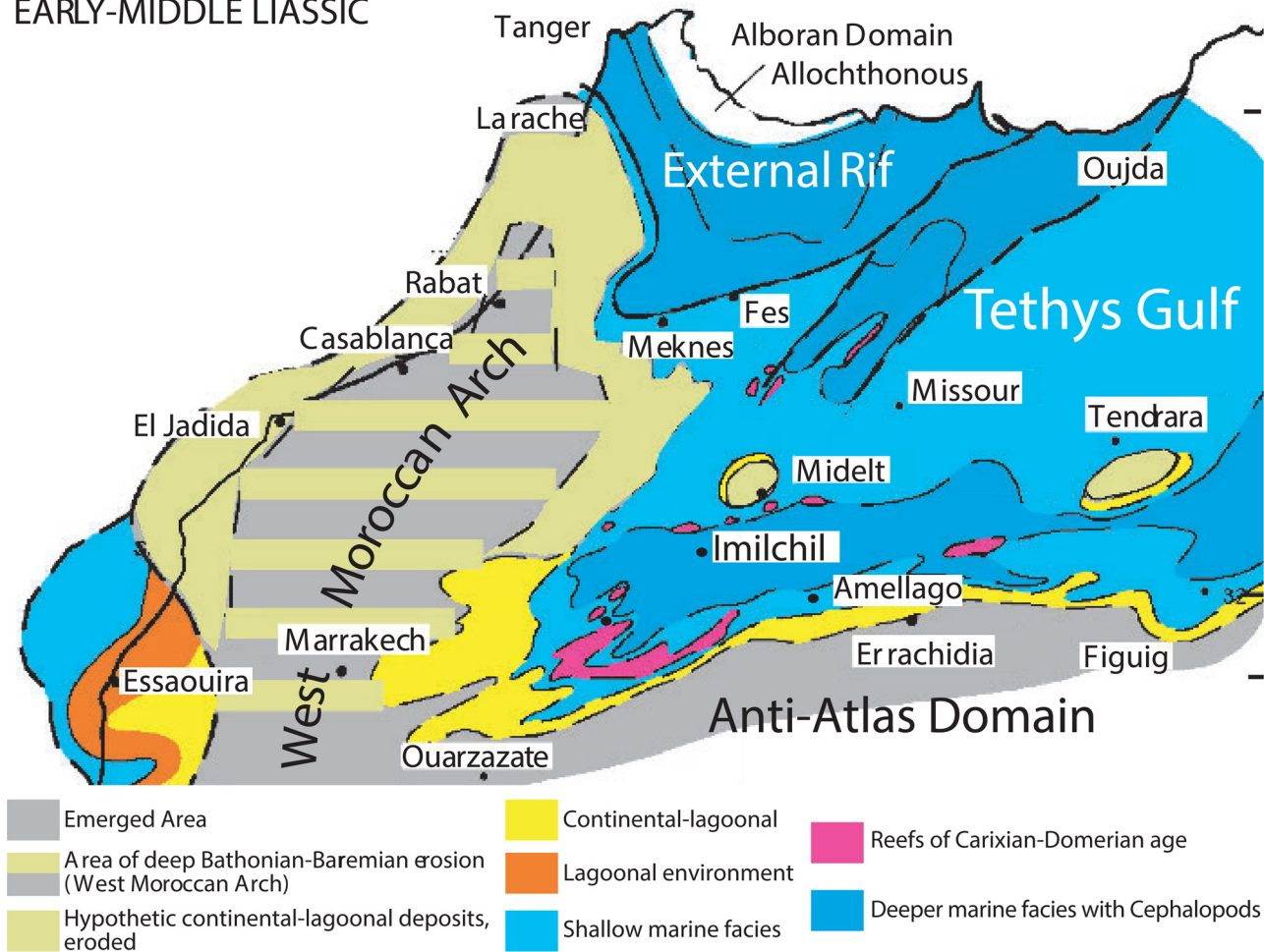


Fig. 2. Paleogeographic map for the Lias epoch, modified in [Frizon de Lamotte *et al.* \(2008\)](#) after [Jabour *et al.* \(2003-2004\)](#). The most important modification from the original figure concerns the West Moroccan Arch, which is no longer regarded as a Liassic emergent land, but as a shallow platform eroded during the late Middle Jurassic-Early Cretaceous interval.

ended with a major volcanic event, linked to the Central Atlantic Magmatic Province (CAMP). It is expressed by the accumulation of a thick volcano-sedimentary unit, including basalt flows intercalated in silty-clays ([Marzoli *et al.*, 2004; 2011](#)).

Above the Triassic, the Jurassic transgression was first marked by dolomitic clays and evaporites during the Hettangian. From the Sinemurian, the development of a thick carbonate platform succession (several hundreds of meters) attests a significant increase of the subsidence rate in the basin during the post-rift phase ([Laville *et al.*, 2004; Moragas *et al.*, 2018](#)).

During the Pliensbachian, normal faulting and block tilting increased notably during Pliensbachian times resulting compartmentalization of the Atlasic Basin ([Elmi *et al.*, 1999](#)) with the individualization of hemipelagic basins surrounded by carbonate ([Wilmsen and Neuweiler, 2008](#)). Between the latest Pliensbachian and the earliest Toarcian, the Lower Jurassic carbonate platforms disappeared and made place for terrigenous dominated sedimentation. This change of

depositional system is considered to be triggered by global anoxia during the latest Pliensbachian and the earliest Toarcian occurred in close association with regional drowning of the lower Liassic platforms and localized deposition of basinal marls ([Kenter and Campbell, 1991; Ettaki *et al.*, 2000; Wilmsen and Neuweiler, 2008; Lachkar *et al.*, 2009; Merino-Tomé *et al.* 2012](#)).

During the Toarcian, continental clays and sandstones were deposited in the western part of the Atlas basin, probably sourced by the *WMA* erosion. The basin was gradually deeper eastward, where shallow-marine mixed carbonate-siliciclastic platforms passed laterally to hemipelagic deposits ([Sadki, 1992; Souhel, 1996](#)). Between the end of the Liassic and the beginning of the Dogger, a carbonate-dominated sedimentation progressively resumed in a globally transgressive context. This evolution led to the development of a wide Bajocian carbonate platform over most of the Atlas domain ([Du Dresnay, 1971; 1972; Rebouillat, 1983; Hadri, 1993; Aït Addi *et al.*, 1998; Pierre *et al.*, 2010; Aït Addi and Chafiki, 2013](#)). The carbonate platform, first located in the shallow

areas, finally prograded and filled hemipelagic domains during the Late Bajocian, resulting in the homogenization of the Atlas basin (Stüder, 1980; Aït Addi and Chafiki, 2013). The sedimentation of platform carbonate lasted until the latest Bajocian and was then replaced first by mixed platform deposits mostly siliciclastic in the southwest, calcareous in the northeast, followed by Bajocian to Early Callovian continental deposits (Jenny *et al.*, 1981; Haddoumi *et al.*, 2010). These continental deposits marked the final infill of the intra-continental Atlas basin, which ceased to subside during the Callovian as shown by the lack of Upper Jurassic and part of Lower Cretaceous deposits in the Atlas basin. Sedimentation only resumed during the Hauterivian in endoreic basins of limited extent (*e.g.*, Moragas *et al.*, 2018). However, from the Bathonian to the Early Cretaceous, the Atlas domain was characterized by major magmatic activity, represented by many basic rock intrusions now exposed at outcrop within the Triassic heart of the ridges, and by numerous sills and magma flows, interstratified within red continental strata (Hailwood *et al.*, 1971; Monbaron, 1981; Laville and Piqué, 1992; Rahimi *et al.*, 1997; Armando, 1999; Lhachmi *et al.*, 2001; Zayane *et al.*, 2002; Charrière *et al.*, 2005; Haddoumi *et al.*, 2010; Bensalah *et al.*, 2013). From Barremian-Aptian period until Middle Eocene, the Atlas domain was an epicontinental basin, in which depositional sequences were mostly controlled by relative sea level variations, especially recording the significant transgressions of the Aptian, Cenomanian–Turonian, Senonian and Eocene (Charrière *et al.*, 2005; Haddoumi *et al.*, 2010).

The compressive tectonic inversion of the Atlas basin is mostly recorded within mid-Eocene to Pliocene syn-tectonic deposits, remarkably well preserved in the sub-Atlas domain (Fraissinet *et al.*, 1988; Frizon de Lamotte *et al.*, 2000; El Harfi *et al.*, 2001; Tesón and Teixell, 2006; Tesón and Teixell, 2008; Tesón *et al.*, 2010). The shortening amount across the Atlas Mountains is low and estimated between 10 and 25% depending on authors (Froitzheim *et al.*, 1988; Brede, 1992; Laville and Piqué, 1992; Teixell *et al.*, 2003, 2024). Most of the shortening seems to be concentrated in the southern and northern Atlas front thrusts.

2.2 Structural setting of the Imilchil area

2.2.1 Structures description

The studied domain is located in the Imilchil area and is structured in three prominent ENE- WSW minibasins-synclines, 5 to 10 km wide, which from north to south have been named: Ikkassene, Lake Plateau and Ikkou (Saura *et al.*, 2014) (Fig. 3). Four subparallel NE-SW salt walls-ridges bound them: the Tasraft, Tassent, Ikkou and Amagmag. With a width between a few hundreds of meters to 3 kilometers, the cores of the salt-related ridges are mostly composed of Triassic shales and large magmatic intrusions. The geometries proposed in the geological section (Fig. 3b) are based on: 1) measured (fieldwork) or estimated thickness of Jurassic series within the synclines; and 2) the dip of the contact between the ridges and the Jurassic series. For a more precise representation of the salt walls this cross-section needs to be completed by that of Saura *et al.* (2014) (Fig. 3c).

The core of the minibasins is characterized at outcrop by thick Lower and Middle Jurassic depositional sequences (Toarcian to lower Callovian) with a measured thickness close to 3,000 m in the Tilmi and Ikkassene synclines-minibasins and an estimated thickness of over 5,000 meters in the Lakes syncline-minibasin. These depositional sequences are dramatically thinning on the flanks of adjoining ridges where halokinetic patterns are very well preserved. These halokinetic patterns are always accompanied by the progressive steepening of strata reaching a subvertical attitude close to the Triassic ridges. In addition, the thickness of lithostratigraphic units highly changes from a minibasin to another, as shown for example by the Imilchil Formation.

Beside these halokinetic patterns, compressive deformations also affect these ridges. As an example, a southeast dipping thrust is at the origin of the late squeezing of the Ikkou ridge and of its subdivision into two distinct parts. On the northern flank of the ridge, this thrust is combined with small-wavelength folds and a series of curved reverse faults oriented N45° to N90°, taking root in the Ikkou ridge. These deformations do not display any evidence of Jurassic halokinetic movements and are therefore linked to the Cenozoic compression as proved by analog modelling simulating the interaction of halokinetic and compressive processes (Moragas *et al.*, 2016; Vergés *et al.*, 2017).

2.2.2 Structures interpretation

The structural organization of the Imilchil region is thus characterized by wide synclines that acted as Jurassic depocenters, separated by narrow synsedimentary growing ridges (Laville and Harmand, 1982; Laville, 1988; Frizon de Lamotte *et al.*, 2008; Michard *et al.*, 2011; Bouchouata *et al.*, 1995; Ettaki *et al.*, 2007). At first, their origin was debated over different interpretations: 1) Early compressive deformations during Early and Middle Jurassic (Duée *et al.*, 1978; Laville, 1978; Stüder, 1980; Jenny, 1984; Ibouh *et al.*, 1994). These deformations would have been the result of ESE-WNW compressive constraints in a regional transpressive context (Jenny, 1984); 2) Synsedimentary magmatic intrusions localized in relay zones of ENE-WSW major faults arranged “en echelon”, within a left-lateral transtensive context from Aalenian to Bathonian and finally becoming transpressive until the Barremian (Laville and Harmand, 1982; Fedan, 1988; Laville and Fedan, 1989; Laville and Piqué, 1992); and 3) Triassic rift structures inversion during Cenozoic compression (Poisson *et al.*, 1998; Teixell *et al.*, 2003; Beauchamp, 2004).

It is now generally accepted that the Triassic ridges of the Imilchil area are elongated synsedimentary diapiric structures (*e.g.*, salt wall) that are the result of Triassic formations remobilization. The intricate salt-related Jurassic tectono-sedimentary evolution was analyzed in Saura *et al.* (2014) for the first time. The synclines are hence considered as minibasins, in which the high subsidence rate was linked to Triassic material expulsion along adjacent diapiric ridges. It also seems that these Jurassic structures were further deformed during the Neogene compression. However, these deformations are limited and only slightly modify the original structure of the Jurassic basin, which thus remains particularly well preserved.

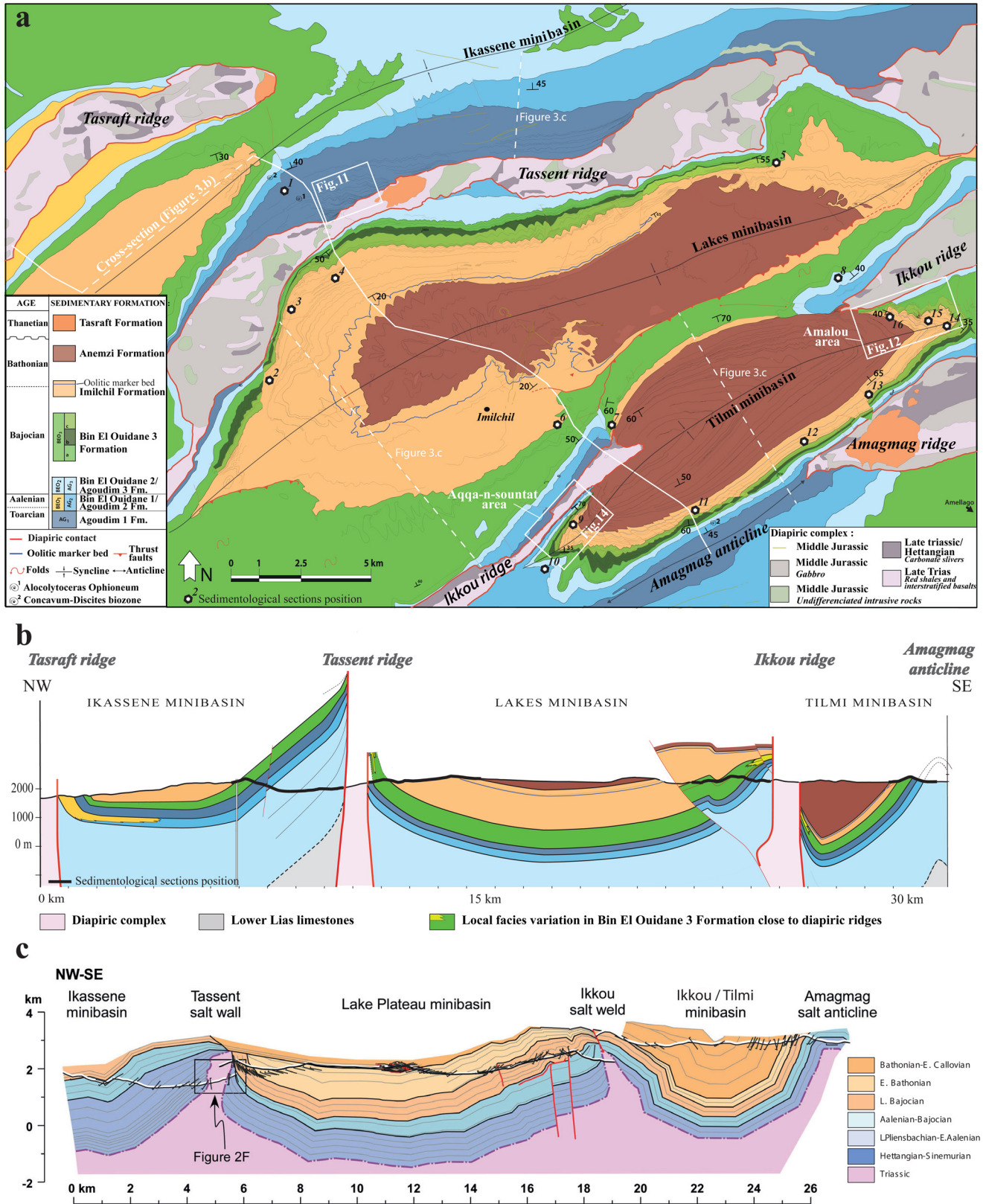


Fig. 3. a) Geological map of the Imilchil area showing the distribution of ridge core rocks and Jurassic sediments. Boxes correspond to three areas studied in detail: the northern flank of the Tassent ridge and the eastern (Amalou area) and western (Aqqa-n-sountat area) tips of the Tilmi mini-basin. b) Geologic cross-section across the western part of the Imilchil area (location in a). In the cross-section are depicted lithostratigraphic units and local facies variations in Bin El Ouidane 3 Formation. c) Cross-section of the Imilchil area from [Saura et al. \(2014\)](#).

This diapiric interpretation joins the one commonly used for analogous structures in the Moroccan Central High Atlas by Bouchouata *et al.* (1995) in the Tazoult ridge, Ettaki *et al.* (2007) in the Izerki ridge, Michard *et al.* (2011) in the Imilchil area, Malaval (2016) in the Zaouiat Ahançal area, Teixell *et al.* (2017) in the Azag minibasin, and Martin-Martin *et al.* (2017) in the Tazoult ridge. Indeed, this interpretation is now considered as the result of a limited Mesozoic basin inversion, already largely influenced by very active syn-rift and post-rift diapirism, responsible for the individualization of several minibasins (Saura *et al.*, 2014). Others similar basins were recently re-interpreted as diapiric basins (e.g., Parizot *et al.*, 2023) in the Corbières area (NE Pyrenees) or Zaagane *et al.* (2025) in Ouarsenis (Western Tell in Northern Algeria).

3 Methods

This study is based on extensive fieldwork that included detailed geological mapping, sixteen sedimentological logs measurement over 15,000 m of cumulative thickness (Annex 1), and sampling for petrographic and biostratigraphic analysis.

The geological map was performed by remote sensing mapping (RMS) and fieldwork using geological maps of Morocco (Fadile, 2003; Saura *et al.*, 2014) as a base. RMS was performed on high resolution Geoeye satellite images (0.5 m resolution). Field mapping focuses on characterizing the envelope of major facies associations, meaning main boundaries of formations but also isochronous surfaces (“time lines”) corresponding to maximum flooding surfaces (MFS) and maximum regression surfaces (MRS), or marked by regional marker beds (hard-grounds, fossils rich layers, etc.). Minor stratification surfaces were also mapped in order to precise the geometry of specific layers around the diapiric structures.

Sixteen sedimentological logs were measured and studied around the three analyzed minibasins in this study. This facies analysis was completed by a petrographic study on thin sections. The evolution of the facies associations over time and space could then be interpreted, followed by the definition of depositional sequences. Using sequence stratigraphy concepts (e.g., Catuneanu *et al.*, 2011) and the physical mapping of certain marker beds, stratigraphic correlations were made between the sedimentological sections. Five regional transects of several tens of kilometers were indeed realized (Annex 1). Two of them, oriented NW-SE, are perpendicular to the diapiric ridges. The first one is located in the western part of the studied area, between Tasraft and Amagmag ridges (NS1), whereas the second one is located in the eastern part between Tassent and Amagmag ridges (NS2). Three transects are parallel to the diapiric ridges with the first one located on the southern flank of the Tassent ridge (EW1), the second one on the northern flank of the Ikkou ridge (EW2), and the third one on the northern flank of the Amagmag ridge (EW3). Finally, the last transect is a composite transect displaying the distribution of sedimentary deposits around the Tilmil syncline, between Ikkou and Amagmag ridges (Fig. 4).

The chronostratigraphic setting of our study is mainly provided by existing biostratigraphic data, especially from the Imilchil geological map at 1/100,000 (Fadile, 2003), completed

by the analysis of ten ammonite species, sampled on five sites located within distal outer platform-basin deposits at the base of the studied series (Toarcian to Lower Bajocian). Paleontological determinations and biostratigraphic attributions were realized by Pascal Neige from the University of Burgundy, based on the zonation established by the *Groupe Français d'Étude du Jurassique* (1997) and on studies carried in the Amellago area (Pierre, 2006; Bourillot *et al.*, 2008).

4 Lithostratigraphy

This section describes 1) the rocks cropping out in the salt walls, and 2) the Jurassic succession cropping out both in the flank and depocenter of the minibasins.

4.1 Triassic series in the core of salt walls

The oldest omnipresent rocks in the core of the ridges and at the origin of diapir rises are a mix of deformed motley clay, siltstone, fine sandstone and occasionally gypsum of Triassic and Early Liassic age. Magmatic intrusive and volcanic rocks, emplaced during the Late Triassic magmatic event associated to the “Central Atlantic Magmatic Province” (CAMP), are associated to this unit (Youbi *et al.*, 2003; Knight *et al.*, 2004). Multi-decametric remnants of carbonate rocks attributed to Upper Triassic and Hettangian (Fadile, 2003) are wrapped in this diapiric unit. They are made of a dark limestone and dolomite facies association, with bird-eyes, stromatolites and dissolution breccia, deposited in a confined inner platform environment, probably associated to the clayey-evaporitic systems.

Jurassic basic magmatic rocks, dated between 175 and 155 ± 5 Ma (Aalenian–Kimmeridgian) according to Hailwood and Mitchell (1971), were emplaced within the salt walls. On some salt walls (Tasraft, Tassent and Amagmag), this diapiric complex unit is overlain by Paleocene fluvio-lacustrine continental deposits, defined as the Tasraft Formation (Charrière *et al.*, 2009).

4.2 Jurassic series in the minibasins

The Jurassic sedimentary succession preserved in the minibasins can be divided into 4 successive lithological units (Fig. 5): (1) a Toarcian to Bajocian marly-limestone unit, the Agoudim Group (Stüder, 1980), which thickness can reach more than 2000 m, (2) a 200 to 500 meters- thick Late Bajocian limestone unit, the Bin El Ouidane Group (Monbaron, 1985; Fadile, 2003), (3) a mixed unit of marly-limestone and clayey-sandstone which thickness changes dramatically (50 to 1300 m) attributed to the latest Bajocian-earliest Bathonian, the Imilchil Formation (Fadile, 2003), and (4) a Bathonian to Early Callovian red clayey-sandstone unit, which thickness can locally exceed 2000 m, the Anemzi Formation (Stüder, 1980). The lithostratigraphic division used in this study is based on the one proposed on the geological map of Morocco (Imilchil geological sheet; Fadile, 2003). Some adaptations are however suggested in order to better consider the sequential organization of the deposits and to clarify the chronostratigraphic correlations at the scale of the Central High Atlas.

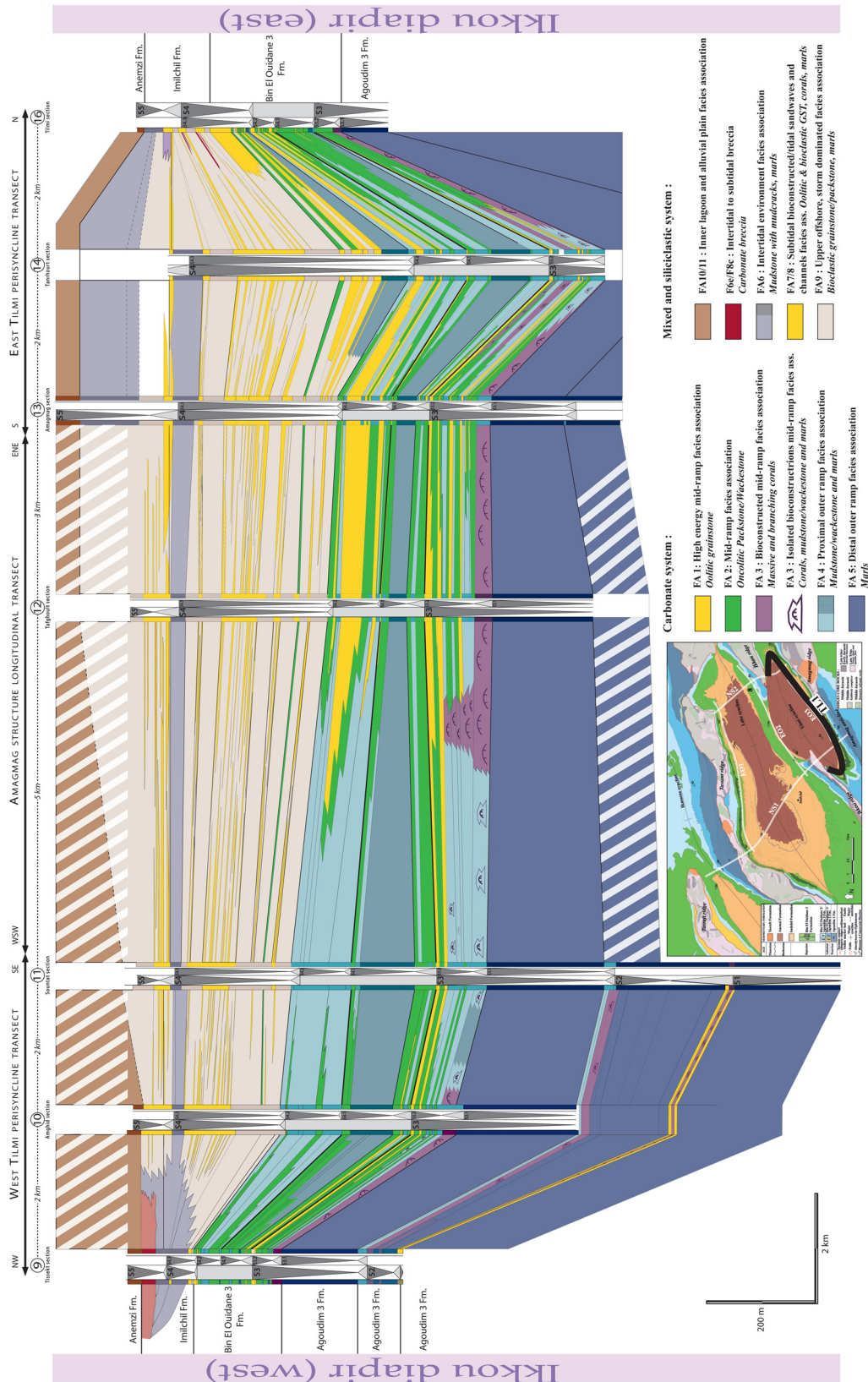


Fig. 4. Correlation transect illustrating sequence geometry and facies distribution in the Tilmi mini-basin between the Ikkou ridge (east and west sections) and the Amagmag ridge. This transect is made of seven sedimentological sections (logs) and is divided in three distinct segments, two are perpendicular to the diapiric ridges and cross the eastern tips (logs n°13, n°14 et n°15) and western tips (logs n°9, n°10 et n°11) of the Tilmi mini-basin, and one is parallel to the Amagmag ridge and corresponds to the southern flank of the Tilmi mini-basin (logs n°11, n°12 and n°13). The other correlation transects (NS1, NS2, EO1, EO2 and EO3) are presented in the annexes (Annexes 2 to 6).

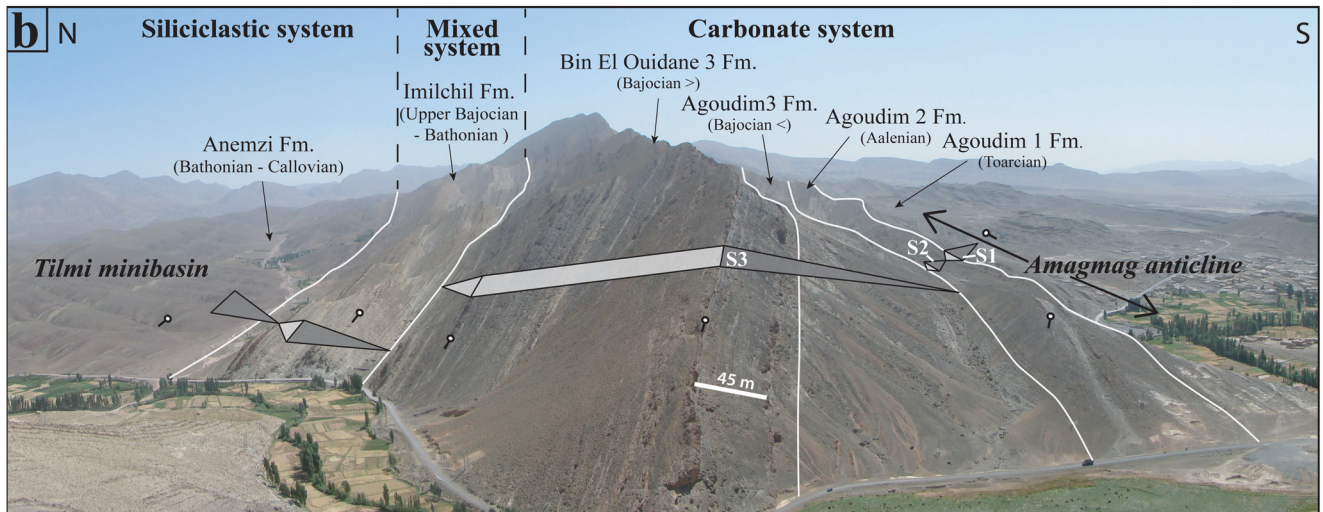
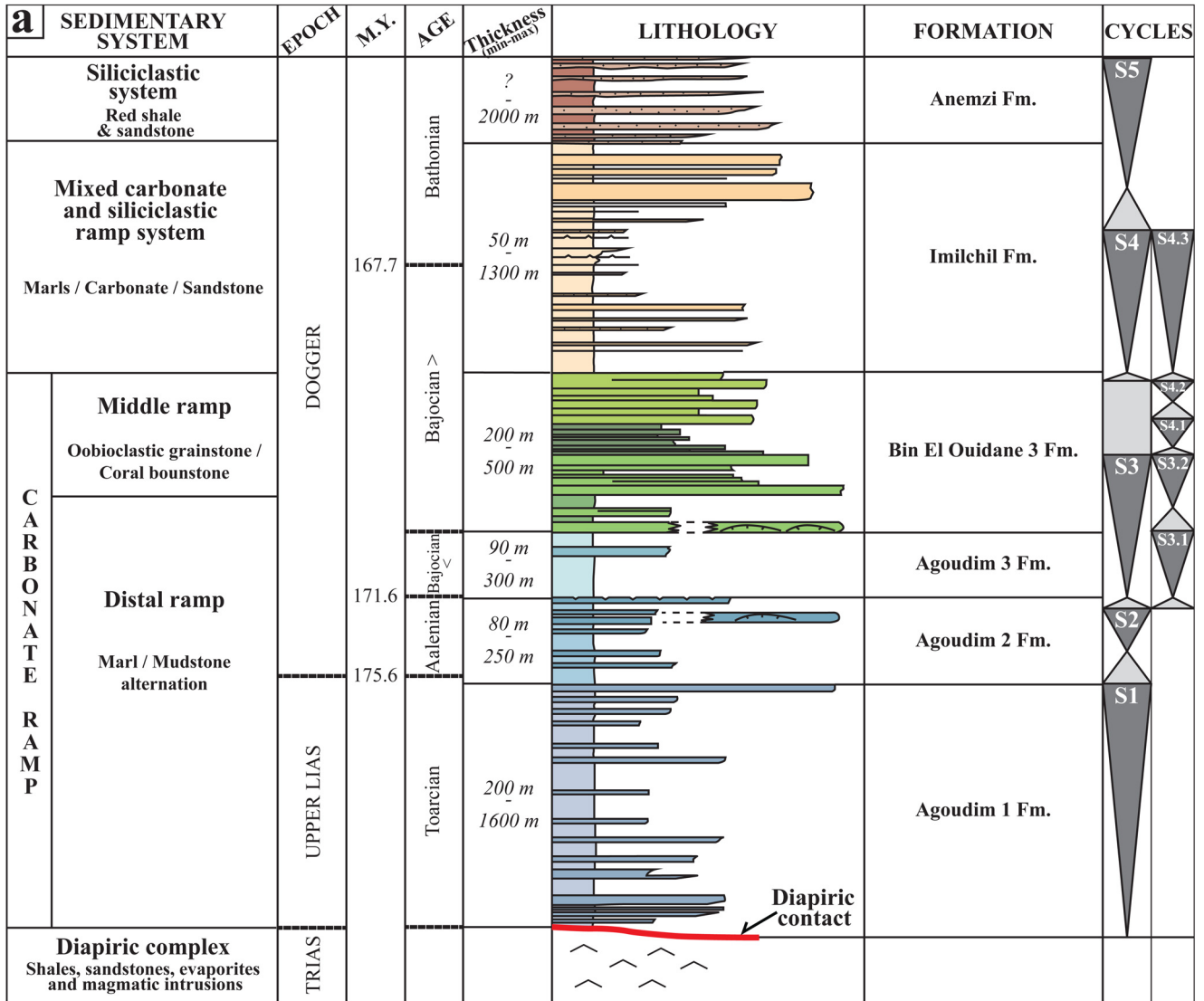


Fig. 5. a) Early-Middle Jurassic synthetic stratigraphic section of the central High Atlas in the Imilchil area. The thicknesses correspond to the maximum and minimum values measured in the field for each formation. b) South of Tilmi mini-basin interpreted field view showing all the lithostratigraphic units and the Transgressive/Regressive sequences.

The marly-limestone unit dated as Toarcian to Bajocian is subdivided in three formations. The Agoudim 1 Formation is mainly made of interbedded grey marls and decimetric mudstone to packstone carbonate beds. These alternations are organized in a succession of thickening-up sequences of several tens-of-meters to several hundred-of-meters thick. It ends with a carbonate layer, several meters thick, made of bioconstructed reefs which is a regional isochronous marker bed. This formation is dated as Toarcian to Aalenian (Fadile, 2003).

The Agoudim 2 Formation is also made of interbedded grey marls and mudstone to packstone carbonate beds. The top of this formation is marked by a condensed layer of regional extent, rich in ammonites that span the Concavum and Discites biozones of the Aalenian-Bajocian limit (association of *Dodicoceras* sp, *Fontannesia* sp, *Eudmetoceras* sp, *Graphoceras* sp and *Graphoceras concavum*; det. Pascal Neige). This condensed layer is another regional isochrone surface.

The Agoudim 3 Formation (included in the Agoudim 2 Formation in Imilchil geological map) is mainly made of grey marls grading upward to a marl-limestone alternation characterized by decimeter-thick beds. The base of this unit is dated as early Bajocian. It is overlain more or less gradually by more massive limestones of the Bin El Ouidane Group. The limit between these two units is diachronous due to the prograding trend of the Bajocian carbonate system.

The Bin El Ouidane Group defined at regional scale (Monbaron, 1985; Jossen, 1990) is here represented by the upper Bajocian Bin El Ouidane 3 Formation (Fadile, 2003). The lower part of this unit usually includes bioconstructed limestones overlain by a facies association made of oolitic, oncolitic and bioconstructed limestones. The middle part of the formation is made of a well-stratified alternation of oncolitic limestones, calcareous mudstones and marls. The upper part is characterized by a massive, relatively homogeneous, unit/bed made of oolitic and oncolitic limestone alternating with marly-limestone deposits. The top of the formation is often marked by the development of bioconstructed reefs. The transition to the Imilchil Formation is gradual but relatively quick and can be considered as a regional isochronous surface. However, this limit can locally be diachronous, due to facies variations located on the flanks of the diapir ridges.

The Imilchil Formation is mostly made of light-colored marly deposits, within which are intercalated carbonate and sandstone beds several centimeters to several meters thick. Red clays and mudstone layers displaying desiccation cracks are intercalated within the clay-rich middle part of the formation. Several meters thick oolitic-bioclastic limestone layers with large-scale cross-stratifications and reefs characterize the upper part of the formation at regional scale. The top of the formation comprises more and more intercalated red clays and sandstone, which marks the transition with the Anemzi Formation. This fast transition is observed at regional scale (Monbaron, 1985; Jossen, 1990). At the scale of the studied area, this cartographic limit can be considered almost isochronous.

The Anemzi Formation is made of a thick and monotonous alternation of red clays and sandstone layers, usually several tens of centimeters thick. These beds appear relatively tabular or slightly channelized. No younger deposits overlay this formation, which upper part is regionally attributed to the

Early Callovian (Fadile, 2003). West of the studied area, in the Beni Mellal region, the age-equivalent Formation of Guettoua is overlain unconformably by Late Jurassic continental deposits (Hadoumi *et al.*, 2010).

5 Facies description

The four lithostratigraphic units described above record the development of three successive depositional systems (Fig. 6). The Agoudim and Bin El Ouidane Groups were deposited on a carbonate ramp system between the Toarcian and the Late Bajocian. The Imilchil Formation witnesses the emplacement of a mixed ramp during the Late Bajocian. Finally, during the Bathonian, the Atlas basin gradually became a site of continental sedimentation dominated by fluvial deposits, as recorded by the Anemzi Formation. Ten sedimentary facies associations characterize juxtaposed or stacked depositional environments within these three depositional systems. The main characteristics of these facies associations as well as the facies description can be found in Table 1 and in the description below.

5.1 The carbonate ramp systems

The Late Toarcian – Late Bajocian carbonate ramp system corresponds to the Agoudim and Bin El Ouidane Groups and gathers five facies associations.

FA1–Cross-bedded oolitic-bioclastic grainstone – high-energy middle ramp environment

These high-energy facies association is made of two mega-ripple cross-bedded grainstone facies, which only differentiate themselves by their composition: oolitic (F1a; Fig. 7a) and bioclastic grainstone (F1b). The mega-ripple cross-beddings usually have a sigmoid shape and show evidence of bidirectional currents. No sedimentary structure proves a significant wave influence. These facies indicate a high-energy hydrodynamic regime dominated by tidal currents (Allen, 1982; Allen and Homewood, 1984). They are interpreted as an oolitic shoal environment on a middle ramp (Tucker and Wright, 1990).

FA2–Oncolitic Packstone-Wackestone – middle ramp environment

This facies association is represented by packstone (F2a; Fig. 7b) and wackestone (F2b) facies with oncolites and bioclasts, including brachiopods and coral debris. Oncolites, that dominate in these facies, form in low to moderately agitated environments and have been described in various settings from inner lagoon to outer platform (Flügel, 2004). Debris of organisms (brachiopods and coral) tend to indicate a middle to outer ramp environment, which is confirmed by the distal position of these deposits compared to the oolitic-bioclastic facies (see correlation transects; Annexes 2 to 6). They are therefore interpreted as middle ramp deposits (Read, 1985).

FA3–Bioconstructed limestone – middle ramp environment

Reef bioconstructions located in Agoudim and Bin El Ouidane formations were the topic of a detailed analysis allowing a specific determination of bioconstructing organisms (B. Lathuillière, pers. com.).

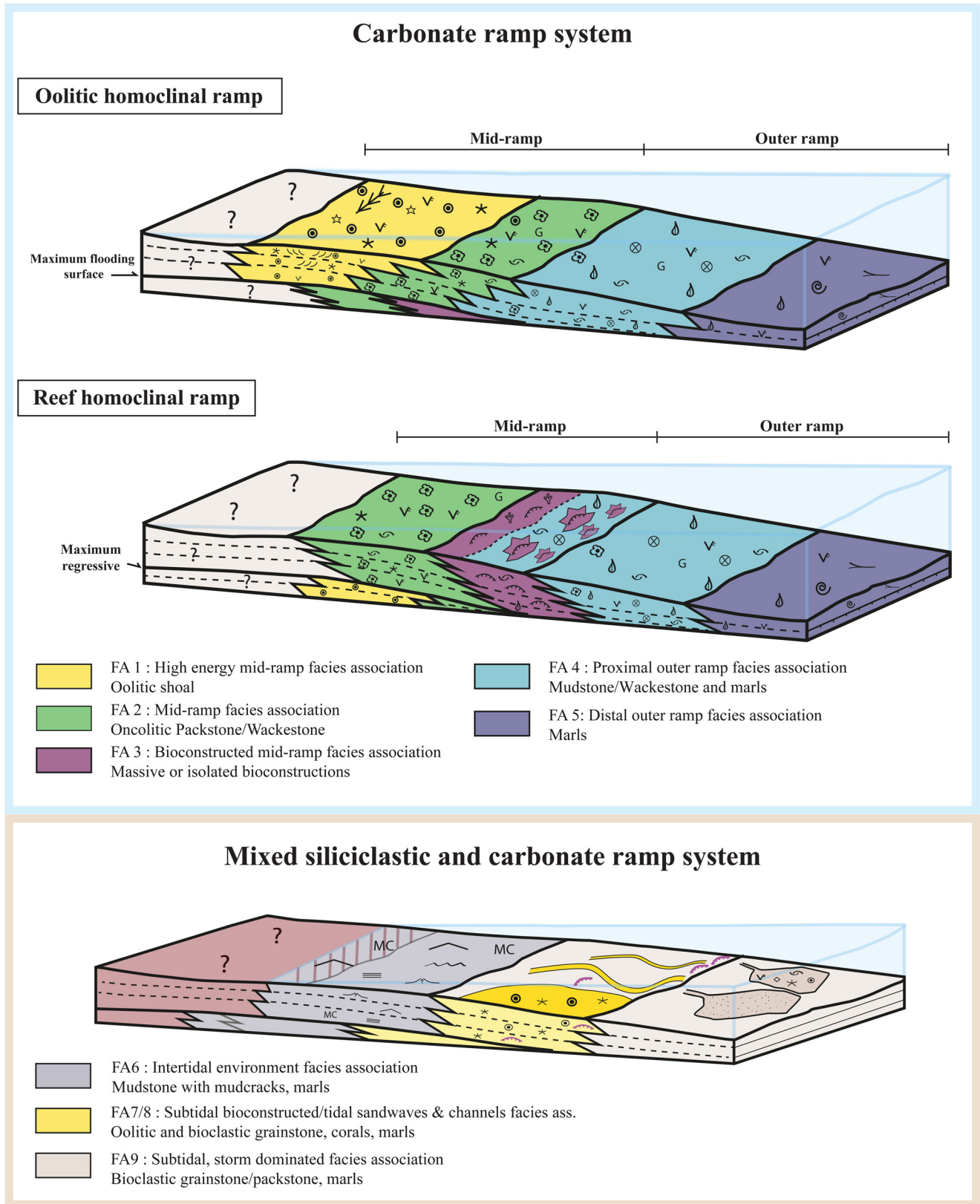


Fig. 6. Depositional models for the carbonate system and for the mixed system. For the carbonate system, two models are defined: the oolitic homoclinal ramp corresponding to the prograding system and the reef homoclinal ramp corresponding to the retrograding system.

Table 1. Facies classification.

	Facies association	Facies code		Facies name	Bed thickness	Stratonomy	Main content	Structure/geometry	Energy	Depositional environment
CARBONATE SYSTEM	FA1	F1a	GST-ON	Oolitic grainstone with megaripple crossbedding	dm-m	Massive	Oolites Bioclasts (echinoderms, bivalves), oolites	Composite megaripple crossbedding (2D and 3D), herringbone	High	High energy mid-ramp
		F1b	GST-B	Bioclastic grainstone with megaripple crossbedding						
	FA2	F2a	PST-ON	Oncolitic and bioclastic packstone	dm-m	Massive to well-stratified	Oncolites (size: 2 cm max.), bioclasts (echinoderms, gastropods, bivalves, lamellibranch), ooids		Moderate	Mid-ramp
		F2b	WST-ON	Oncolitic and bioclastic wackestone	cm-m		Oncolites (size: 1 cm max.), bioclasts (brachiopods, corals, gastropods, bivalves)			
		F3a	BR-TS	Upper Toarcian reef bioherms	m		Massive and branching corals, bioclastic rudstone			
	FA3	F3b	BR-BM	Upper Bajocian A reef bioherms	m-dam	Bioherms	Massive and branching corals (high diversity), bioclastic rudstone		Moderate	Bioconstructed mid-ramp
		F3c	BR-BS	Upper Bajocian B reef bioherms	m-dam		Branching corals			
		F4a	PST-RP	Bioclastic wackestone-packstone	cm-m		Brachiopods (<i>Rhynchonella</i> , <i>Terebratuloid</i>), corals, gastropods, bivalves			
	FA4	F4b	MST-RP	Mudstone	cm-m	Well stratified	Brachiopods (<i>Rhynchonella</i> , <i>Terebratuloid</i>), gastropods, bivalves, corals, pectens, ammonites print		Moderate	Proximal outer ramp
		F4c	MR-RP	Grey marls	dm-m		Brachiopods, bivalves			
		F5a	WST-RD	Bioclastic wackestone-packstone	cm-m		Echinoderms, gastropods, corals debris, brachiopods, ammonites, Belemnites guards	Hard-ground	Moderate to low	
MIXED SYSTEM	FA5	F5b	MST-RD	Mudstone	cm-m	Well stratified	Ammonites	Hard-ground	Low	Distal outer ramp
		F5c	MR-RD	Grey to green marls	m-lm		Ammonites, brachiopods, Belemnites guards		Low	
		F6a	MRN	Beige marls	dm-dam		Brachiopods, gastropods, bivalves		Low	
	FA6	F6b	MST-L	Thinly laminated mudstone with teepee structures	dm-m	Well stratified		Teepee structures, mudcracks, horizontal to wavy lamination, bird-eyes	Low	Intertidal
		F6c	BRE	Breccia	cm				Low	
		F6d	ARG-R	Red shales	dm-m				Low	
		F6e	BRC	Carbonate breccia	m		Roots trace		Low	
	FA7					Well stratified, channelized	Lithoclasts (angular, cm to dm size), mud clasts, carbonate mud matrix	Channelised bedding	High	Subtidal bioconstructions
						Massive to well-stratified	Massive and branching corals, bioclasts	Composite megaripple crossbedding (2D and 3D)	High to moderate	
		F6a	MRN	Beige marls	dm-dam		Brachiopods, gastropods, bivalves		Low	
	FA8	F8a	GR-OB	Oolitic and bioclastic grainstone	dm-dam	Well stratified, channelized sandwaves	Oolites, echinoderms debris	Crossbedding, composite megaripple crossbedding (2D and 3D), herringbone	High	Subtidal sandwaves and channels
		F8b	PST-B	Bioclastic packstone	cm-m		Bioclasts (bivalves, echinoderms, corals, brachiopods), oolites	Bidirectional megaripple crossbedding	High	
		F8c	CGL	Carbonate conglomerate	m		Lithoclasts (sub-angular, cm to dm size), bioclastic packstone matrix		High to moderate	
SLICICLASTIC SYSTEM	FA9	F6a	MRN	Beige marls	dm-dam	Well stratified	Brachiopods, gastropods, bivalves		Low	Storm influenced upper offshore
		F9a	GFM	Fine/medium sandstone with wave ripples	cm-dm			Planar to wavy bedding (HCS), current and wave ripples, graded bedding, erosive base	High to moderate	
		F9b	PF-B	Bioclastic packstone/floatstone	cm-dm		Bioclasts (bivalves and echinoderms), oolites	Graded bedding	High to moderate	
	FA10			Fine to medium sandstone and red shale	dm-m	Well stratified, channel (rare)		Multidirectional current ripples, planar bedding	Low to moderate	Alluvial plain

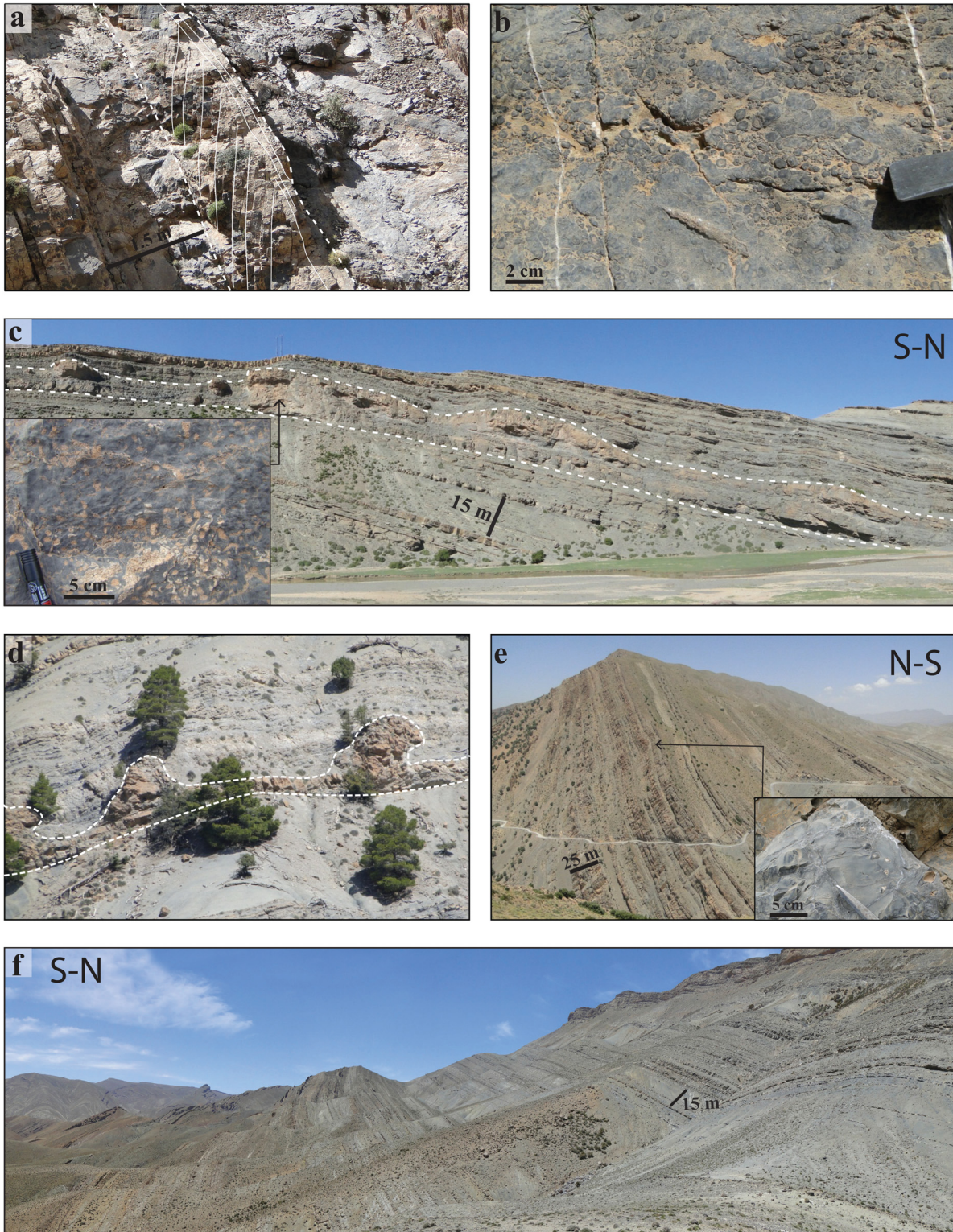


Fig. 7. a) Oolitic grainstone with mega-ripple crossbedding (F1a) characterizing a high- energy middle ramp environment (FA1) (Bin El Ouidane 3 Formation, sequence S4.1) ; b) Packstone with oncolites and bioclasts (bivalves) (F2a), middle ramp environment (FA2) (Bin El Ouidane 3 Formation, sequence S4.1); c) Reef bioconstructions within middle ramp setting (FA3) of Upper Bajocian (F3c) made of branchial corals of *dendrea* genre (box) (upper part of Bin El Ouidane 3 Formation, sequence S4.1); d) Upper Toarcian scattered reefs (F3a) spaced out approximately ten meters apart by type F4a and F4b facies (base of Agoudim 2 Formation, sequence S2); e) Well-bedded limestone from proximal outer ramp environment (FA4) made of an alternation of marls, mudstone and bioclastic packstone/wackestone (Bin El Ouidane 3 Formation south of Tassent ridge); f) Marls and limestone alternation characteristic of distal outer ramp deposits (FA5) (Agoudim 1/2/3 Formation to the north of the Ikkou ridge, sequences S1/ S2/S3.1)

- *Facies 3a – Upper Toarcian reef bioconstructions (Fig. 7d)*: the top of Agoudim 1 Formation is marked by a group of scattered reef structures, several tens of meters long and several meters high, separated from each other by about ten meters. These bioconstructions have a pyramid like shape and grade laterally to proximal outer ramp facies (F4a and F4b). The organisms present in these bioconstructions are diverse: branchial corals *phacéloïdes*, including *Cladophyllia* and *Lochmaeosmia*, massive corals, of which some belong to *Periseris* genre. These corals are associated with many bioclasts, bivalves (including *Chlamis*), echinoderm spines, sponge debris (*Neuropora*), annelids (*Serpula Socialis*) and stromatolites (leiolites and thrombolites).
- *Facies 3b – Upper Bajocian reef bioconstructions – lower part of Bin El Ouidane 3 Formation*: These bioconstructions mark the emplacement of shallow carbonate ramp facies (Bin El Ouidane 3 Formation) prograding over the outer ramp facies (Agoudim 3 Formation). In the proximal zone, they form a ten meters thick massive layer, whereas in the more distal zone they are made of scattered reefs separated by proximal outer ramp deposits (facies F4a and F4b). They are made of various corals: massive forms of *Periseris* genre, *Microsolena*, (massive *cérïoïde*) and *Isastrea*, branchial forms of *Dendrea* and *Stephanastrea* genre, branchial phaceloïde forms, end of *Chladophyllia* and *Thecosmia* genre, secondary genres with *plocoïdes* (*Pseudosoria* genre) and many thrombolytic encrusting. Corals are mostly in living position. These bioconstructions also include many fossils and bioclastic debris: oysters, brachiopods (*Rhynchonella*), echinoderms, etc. The bioclastic elements are found within these bioconstructions but also in bioclastic rudstone levels, just around them.
- *Facies 3c – Upper Bajocian reef bioconstructions – upper part of Bin El Ouidane 3 Formation (Fig. 7c)*: this bioconstruction type appears as reef structures of metric to decametric width, in the shape of several meters-high domes, spaced of about ten meters and separated by facies F4a and F4b. These bioconstructions are almost made only of a monospecific association of branchial corals of *Dendrea* genre, usually recrystallized and floating within a black muddy carbonate matrix. They can be found with rare massive corals of *Isastrea* and *Thecosmia* genre as well as branchial phaceloids debris of *Chladophyllia* genre associated with brachiopods, bivalves and gastropods in the surrounding facies. These three types of reef bioconstructions (FA3a/b/c) developed in a relatively calm depositional environment, below fair-weather wave base. Around some of these, coarse facies witness episodic high-energy conditions, probably associated to storms. These bioconstructions are always located either more proximally than or laterally to proximal outer ramp facies, as shown by the correlation transects (Annexes 2 to 6). They therefore developed in the transition zone between middle and outer ramp. Bathymetry of these bioconstructions was estimated between 20 and 30 meters (B. Lathuillière, pers.com.) for F3a and above 20 meters for F3b and F3c. Around Imilchil village, the optimum development of corals occurred during the Bajocian for bioconstructions of F3b type, at the bottom of the Bin El Ouidane Formation. The drastic

decrease of organisms' diversity within bioconstructions of F3c type at the top of the Bin El Ouidane Formation is due to an increasing input of detrital material in the basin at the transition with the Imilchil Formation.

FA4–Alternating packstone to mudstone and grey marls (Fig. 7e) – proximal outer ramp environment

This facies association is made of an alternation of bioclastic wackestone-packstone (F4a), mudstone (F4b) and grey marls (F4c), the first two being predominant. It indicates a low (F4c, F4b) to moderate (F4a) energy environment below the fair-weather wave base or the main currents. These deposits are found directly more distally or laterally to the reef bioconstructions. They correspond therefore to an environment located in the proximal part of an outer ramp.

FA5–Grey marls and WST-PST with Ammonites (Fig. 7f) – distal outer ramp environment

This facies association comprises an alternation of large intervals of grey to green marls (F5c) with rare decimetric layers of bioclastic wackestone-packstone (F5a) and mudstone (F5b) with ammonites and belemnites. Mostly marly, this facies association represents the most distal depositional environment of the studied sedimentary system. However, the lack of gravity deposits shows the lack of a pronounced slope. This facies association is interpreted as a distal outer ramp environment.

Depositional model of the carbonate system (Fig. 6) represents a gently and uniformly dipping ramp with a slope angle of less than 1 degree. This interpretation is supported by the tabular and regular geometry of deposits, gradual facies transitions, and the absence of instability features or gravity-driven deposits (Read, 1985; Tucker and Wright, 1990). In the studied area, various depositional environments are juxtaposed on this carbonate ramp: high-energy oolitic-bioclastic middle ramp (FA1), middle ramp with muddy oncolitic facies (FA2), proximal outer ramp with carbonates (FA4) and then distal outer ramp with marls and carbonates (FA5). Reef bioconstructions (FA3) developed on the middle ramp during transgressive phases, at the scale of both third-order and higher-order frequencies. The dynamic of this carbonate system is mainly dominated by tidal currents. The very low export rate of oolites on the outer platform and the type of sedimentary structures indicate a very limited influence of waves and storms. These characteristics could be the result of the narrow physiography of the Atlas basin, limiting swell penetration and enhancing tidal currents. The proposed depositional model presents the facies organization of the Upper Toarcian to Bajocian series in the south of the studied area. It therefore, concerns the carbonate ramp system of SE-NW polarity in place at that time in the southern part of the Atlas basin (*cf.* §6). During this time, the axis of the basin was approximately located at the emplacement of the Ikassene minibasin. Another carbonate ramp system of NW-SE polarity extended north of this axis and is represented by the series exposed at outcrop on the Tasraft ridge and within the Ikassene minibasin in the northern part of the studied area. This northern carbonate ramp displays a fairly different facies organization than the southern one. Indeed, the facies associations of oncolitic middle ramp (FA2) and reef bioconstructions (FA3) are missing. The high-energy oolitic middle ramp facies (FA1)

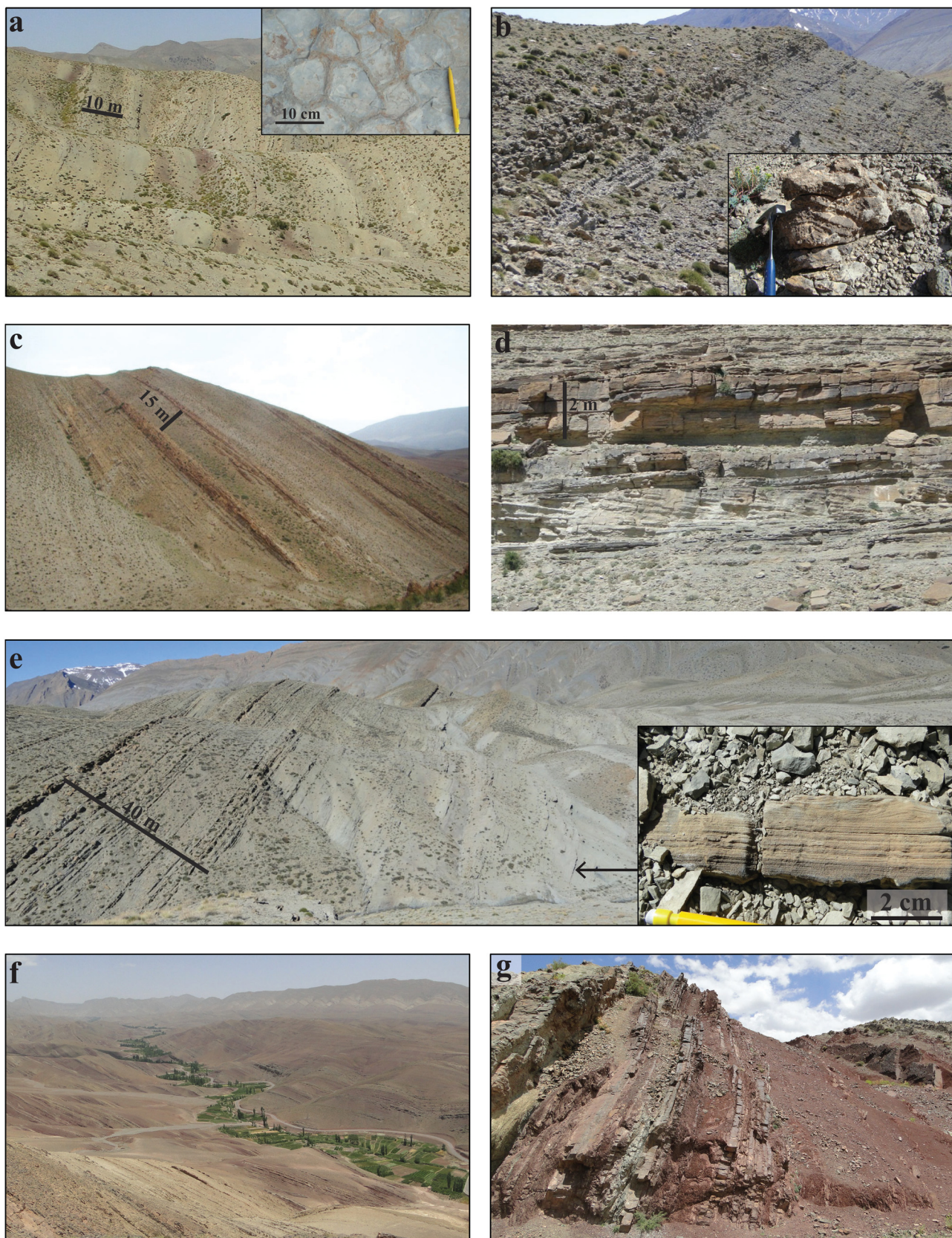


Fig. 8. a) Alternation of beige marls (F6a), red clays (F6d), and finely laminated mudstone with teepee structures and desiccation cracks (F6b) (box) characteristic of intertidal depositional environment (FA6) (Imilchil Formation, sequence S4.2). b) Biostromes interbedded with bioclastic grainstone and beige marls, subtidal reef environment (FA7) (Imilchil Formation, transgressive systems tract of sequence S5). Biostromes are made of branchial and massive corals. The massive corals sometimes present « skullcaps » morphologies (box), typical of a depositional environment subject to an environmental stress. c) Subtidal environment with tidal channels and sandwaves (FA8) characterized by an alternation of marly deposits and oolitic/bioclastic channel bodies, one hundred to several hundred meters long (Imilchil Formation, transgressive systems tract of sequence S5). d) Several meters thick set of oblique crossbedding in oolitic grainstone deposits (F8a) typical of tidal sandwaves (Reyneck, 1963) (Imilchil Formation, transgressive systems tract of sequence S5). e) Alternation of beige marls (F6a), fine- to medium-grained sandstone with wave ripples (F9a) (box) and bioclastic packstone/floatstone (F9b) characteristic of upper offshore environment (FA9) (Imilchil Formation, to the south of the Tassent ridge, sequence S4.2). f) Alternation of meter-thick to several-tens-of-meters thick intervals of red clays and tabular sandstone beds ten centimeters to one meter thick, interpreted as deposited in a distal alluvial environment (FA10) (Anemzi Formation in the mini-basin of Tilmi, sequence S5). g) Fine- to medium-grained sandstone, red to grey clays (Anemzi Formation).

grades directly to the muddy proximal outer ramp facies (FA4). This difference between the northern and southern ramps could be the result of the system orientation regarding main winds and currents and/or the result of the depositional slope. The proximal part of these carbonate ramps is not visible in the Imilchil area. It is however exposed at outcrop in the Amellago (Pierre *et al.*, 2010) and Bin El Ouidane (Monbaron, 1985; Rebouillat, 1983) regions, respectively for the southern and northern ramps. They are mostly dominated by aggrading thick tabular mudstone and wackestone layers with stromatolites and oncolites, separated by marly interbeds and several subaerial exposure surfaces.

5.2 The mixed ramp system

The mixed ramp system, represented by the Imilchil Formation (Late Bajocian–Early Bathonian), gathers four facies associations.

FA6–Alternating marls and limestone with stromatolites and desiccation cracks – Intertidal environment (Fig. 8a).

This facies association shows an alternation of beige marls (F6a), mudstone with microbial laminations, mud-cracks, teepee structures (F6b; zoom Fig. 8a), intraformational breccia with evaporite pseudomorphs (F6c), red clays (F6d) and channelized conglomerates with carbonate elements (F6e). These facies record a very shallow and confined depositional environment, of very low energy, subject to frequent subaerial exposures. It therefore corresponds to an intertidal environment in the inner part of a clayey-carbonate system.

FA7–Coral boundstone – Reef subtidal environment

This facies association appears as biostromes of metric to plurimetric thickness made of corals in living position interrupted by biotrititic grainstone and beige marl levels (Fig. 8b). The bioconstructions are mostly made of branchial corals (phaceloid forms of *Chadophylia* genre) and some massive corals (individuals of *Isastrea* and *Tamnasteria* genre). The massive corals can form boulder colonies with skullcap morphologies (zoom Fig. 8b) (Hofling, 1989) suggesting an environment where bioconstructions are perturbed by a strong sediment flux. The biotrititic grainstone hold many *Nerinees* fragments. The composition of the bioconstructions, their association with *Nerinees* and their alternations with biotrititic grainstone tend to express a shallow subtidal environment under episodic influence of an

active hydrodynamic event. It is interpreted as the result of tidal currents according to other facies associations in this depositional system. The tabular geometry of these bioconstructions, the dominance of resisting corals (branchial and *Isastrea*), and the presence of characteristic skullcap morphologies show evidence of a bio-sedimentary system under stress, probably due to a high clastic sediment input and to the basin confinement (B. Lathuillière, Pers. Com.).

FA8–Cross-bedded oolitic-bioclastic grainstone/packstone – Tidal channels and bars (Fig. 8c).

This facies association is composed of alternating grainstone with oolites and bioclasts (F8a; Fig. 8d), bioclastic packstone (F8b) and beige marls (F6a). The dominant granular facies have a lens shape, sometimes channelized and often made of metric to plurimetric sets of sigmoid cross-bedding, indicating bidirectional currents. Right next to the diapiric ridges, conglomerates with carbonate elements (F8c) are locally interbedded within this facies association. This facies association is dominated by high-energy marine facies, characteristic of a carbonate subtidal environment. The sedimentary structures and the geometry of the granular bodies suggest that they can be interpreted as a complex of tidal channels and bars. The amplification of tidal currents recorded by these deposits could be the result of an increased segmentation of the basin, linked to the syndimentary growth and subaerial exposure of the diapiric ridges, as demonstrated by the conglomeratic deposits flanking the ridges.

FA9–Alternation of marls and graded sandstone – Upper offshore environment under storm influence (Fig. 8e).

This facies association is made of alternating beige marls (F6a), graded centimetric layers of fine- to medium-grained sandstone with sharp base (F9a; zoom Fig. 8e) and bioclastic floatstone/packstone (F9b). These layers often display planar or undulated bedding and/or wave ripples. This facies association is mostly marly and testifies of a low energy environment. However, it records repeated rapid flow events with an oscillatory component, responsible for an accumulation by traction-decantation of thin graded layers. These deposits are interpreted as the result of storm currents in an offshore environment. The lack of real sandy shoreface deposits and the thinness of these stormy layers show however that the storm energy was limited in a shallow offshore environment.

Depositional model of the mixed system (Fig. 6):

No significant lateral facies variation was seen within the Imilchil Formation at the scale of the studied area. However, the organization of the facies associations is clearly expressed

vertically in the stacking pattern, forming third-order transgressive-regressive cycles that can be regionally correlated (*c.f.* §5). Their analysis allows to propose a synthetic depositional model showing the proximal-distal succession of depositional environments, from intertidal (FA6), to subtidal with reef bioconstructions or tidal channels and bars (FA7/8) and to subtidal under storm influence (FA9). This depositional model is characterized by: 1) a low-energy mixed sedimentation dominated by marls and limestone, 2) a very slightly dipping depositional profile underlined by a great extent of facies belts that goes beyond the studied area, and 3) an evolution of the dominant hydrodynamic processes over time, with an initially well-expressed storm energy being replaced by tidal-current energy. This evolution could be the result of an increased segmentation of the basin due to the emersion of diapiric ridges during the deposition of Imilchil Formation *c.f.* §6). Furthermore, the characterization of the mixed system also leans on observations made in the Ahançal valley, located 70 km southwest of the Imilchil area (Malaval, 2016).

5.3 The continental siliciclastic system

Continental siliciclastic deposits of the Anemzi Formation (Bathonian to Early Callovian) are very monotonous and only made of one facies association. It is an alternation of decimetric to metric layers of fine to medium sandstone and silty red clays (FA10) (Fig. 8f). The silty clays represent 80% of the deposits and form intervals of a few meters to tens of meters thick (Fig. 8g). They often display parallel laminations. The sandstone layers are mostly tabular, of hectometric to kilometric lateral extent, and occasionally slightly channelized. They display parallel lamination and multi-directional current ripple cross-lamination. Fossils are absent and bioturbations are rare. This red, azoic facies association is interpreted as deposited in a continental environment. The fine granulometry and the lack of channelized structures are typical of a distal alluvial environment, weakly erosive, that could correspond to the distal part of a “distributive fluvial system” (Nichols and Fisher, 2007; Weissmann *et al.*, 2013). The coarser facies of the age-equivalent Guettoua Formation, exposed at outcrop west of the studied area (Monbaron, 1985), should therefore be the more proximal part of this SW-NE alluvial system. The lack of strictly lacustrine facies association suggests that the minibasins were not endoreic at this time.

6 Architecture of depositional sequences and paleogeographic reconstructions: regional organization and local perturbations

The correlation of depositional sequences between the different minibasins of the Imilchil area provides a precise chronostratigraphic framework allowing to determine the depositional architecture and the paleogeographic evolution of this domain from the Toarcian to the Bathonian. This reconstruction allows the distinction between local perturbations linked to salt walls movements and regional trends linked to the general evolution of the Atlas basin. These reconstructions are illustrated by five stratigraphic correlation transects (Annex 1), either transverse (NS1, NS2) or parallel (EW1,

EW2, EW3) to the diapiric structures (Annexes 2 to 6), and by one correlation transect around the Tilmi minibasin (Fig. 4).

Analysis of the vertical facies succession and inferred deposit geometries in the Imilchil area has allowed for the identification of six transgressive-regressive (T/R) sequences within the Toarcian to Lower Bathonian succession, spanning approximately 15 million years (Fig. 5). The lower portion of the Anemzi Formation (Bathonian-Callovian) was excluded from this analysis due to the homogenous nature of its continental facies, which precludes precise determination of depositional sequences. With the exception of Sequence S1, each sequence comprises a transgressive systems tract culminating in a maximum flooding interval or surface (MFS), succeeded by a regressive systems tract that concludes with a maximum regressive interval or surface (MRS). The MRS is typically marked by the shallowest facies of each cycle (*e.g.*, bioconstructions, oolitic limestone, intertidal facies) rather than a prominent subaerial exposure surface, suggesting these regressive systems tracts can be considered “highstand systems tracts” (Vail *et al.*, 1991). No “falling stage systems tract” or “lowstand systems tract” (Catuneanu, 2011) were observed. Each sequence has a duration of 2 to 4 million years, classifying them as third-order cycles.

6.1 Sequence S1 (Upper Toarcian)

Sequence S1, corresponding to the middle–upper Toarcian Agoudim 1 Formation, is solely defined by its regressive systems tract. It consists of proximal (FA4) and distal (FA5) outer ramp facies, characterized by thickening-up and prograding high-frequency sequences. The maximum progradation of this Toarcian sequence is indicated by middle-ramp reef bioconstructions (FA3) and, less commonly, oolitic bodies (FA1), a level correlated across the entire Imilchil area and marking the top of the Agoudim 1 Formation. This sequence consists of quite homogeneous marls and limestones deposited in an outer ramp environment (FA5) throughout the study area (Fig. 9a and NS1). The depositional environments appear slightly more proximal in the southern Tilmi and Lake minibasins than in the northern, and marlier, Ikassene minibasin. The effects of diapiric movements are particularly well-recorded on the northern flank of the Tassent salt wall. Here, the layers dramatically thin, displaying spectacular wedge patterns that correspond to composite halokinetic sequences over approximately one kilometer (Fig. 10). This structure is associated with an evolution towards more carbonate-rich mudstone facies closer to the Tassent salt wall. Within these deposits, we find lithoclasts derived from the remobilization of perforated hardgrounds. These lateral facies variation is interpreted as a result of a decreasing sedimentation rate around the rising salt wall. This decrease led to lower preservation of marls and the development of hardgrounds, caused by a lower subsidence rate compared to the center of the minibasins. However, the absence of instability features or gravity flow sedimentation means we cannot confirm the existence of sedimentary slopes close to the diapiric salt wall. Outcrop exposure conditions on the flanks of other salt walls in the Imilchil area do not allow for further information on their impact on Sequence S1 deposition.

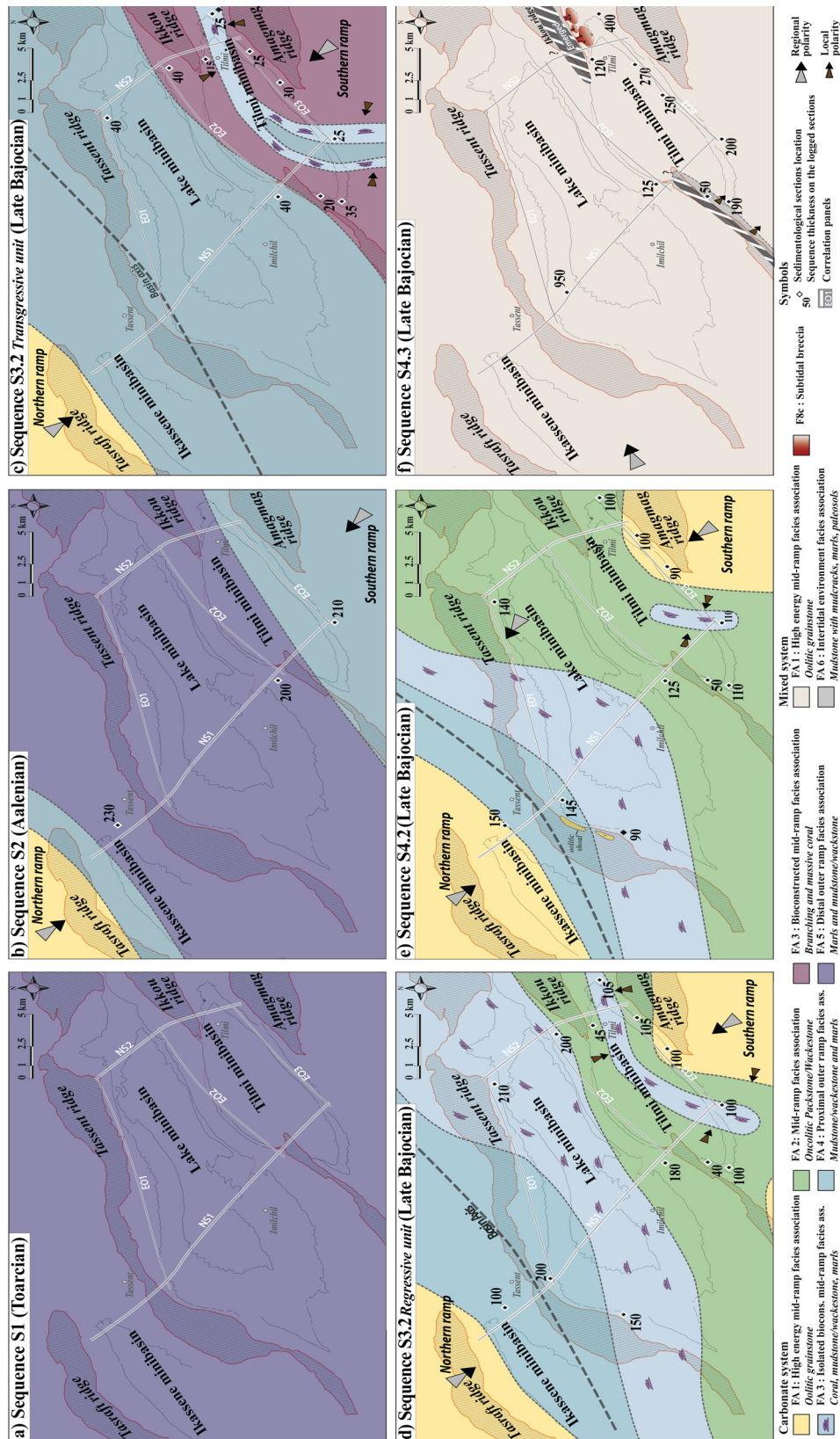


Fig. 9. Paleographic sketch maps showing the main depositional environments during sequence S1 (a), sequence S2 (b), the transgressive unit of sequence S3.2 (c), the regressive unit of sequence S3.2 (d), sequence S4.2 (e), sequence S4.3 (f). The available data for sequence S4.3 give information about the facies only in the south of the Ikkou ridge, they do not provide any evidence concerning the emersion of the other diapiric ridges (Tassent and Amagmag) at this time. Note that the correlation transects used to construct these maps are represented in white.

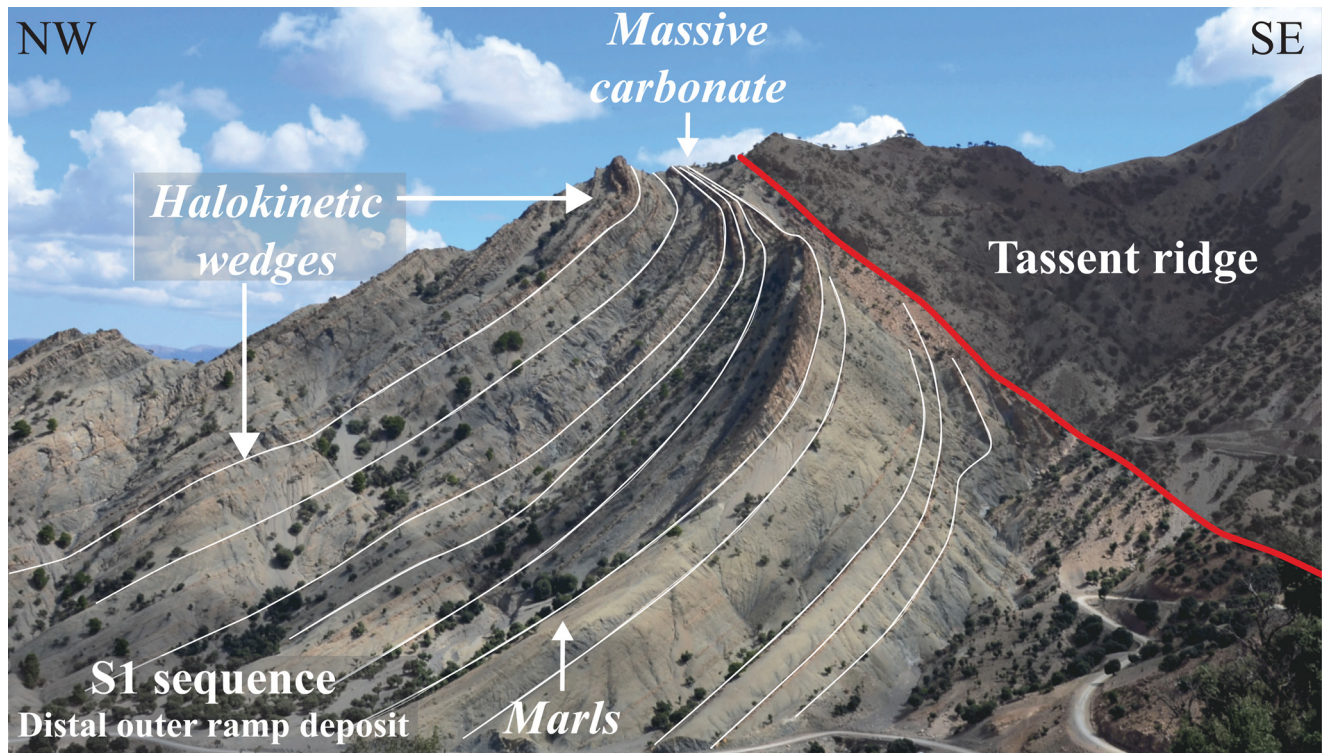


Fig. 10. Field view of the northern flank of Tassent ridge, showing progressive unconformity of the distal outer ramp deposits during sequence S1. See location in [Figure 3a](#).

6.2 Sequence S2 (Aalenian)

Sequence S2, encompassing the Agoudim 2 Formation (latest Toarcian to Late Aalenian), begins with a transgressive systems tract of thinning-up marl-limestone with increasing clay content, indicative of proximal and distal outer ramp environments (FA4, FA5). The subsequent regressive systems tract is composed of an overall thickening-up marl-limestone sequence, culminating in reef bioconstructions in the most proximal domains. This regressive maximum, situated within the upper part of the Agoudim 2 Formation, correlates with the maximum progradation of the Amellago oolitic ramp system on the southern Atlas basin platform (Pierre, 2006) and the maximum progradation of the Bin El Ouidane 1 Limestone on the northern platform (Fadile, 2003). The organization of the deposits within this sequence S2 shows the centripetal progradation of the northern and southern platforms towards an axial depocenter on the Lake and Ikassene minibasins, where a distal ramp sedimentation of marls and limestone persisted ([Fig. 9b](#)). The southern carbonate platform progradation is locally marked by the development of reef bioconstructions at the top of the sequence (FA3), whereas the progradation of the northern platform is recorded by oolitic ramp facies (FA1) near the Tasraft salt wall. Regionally, the thickness of this sequence S2 slightly increases from south to north. From 200 m in the Tilmi minibasin (log 11) to 230 m in the Lake (log 6) and Ikassene (log 1) minibasins, located in a more distal position (NS1). However, the sequence thins significantly on the flanks of the salt walls, as in the case of the

southeastern flank of the Ikkou salt wall, sequence S2 thins to only 60 m, but without any facies variation.

6.3 Sequence S3 (Upper Aalenian- Upper Bajocian)

Sequence S3 initiates with a thin retrograding carbonate interval, overlain by a condensed layer containing *Concavum* and *Discites* biozone ammonites (Late Aalenian to Early Bajocian). This regionally extensive condensed layer represents the top of the Agoudim 2 Formation and is also found atop the Amellago Limestone 60 km to the south ([Pierre, 2006](#)). The sequence then transitions into an outer-ramp marly unit, gradually evolving into a thickening-up alternation of marls and limestone (FA5, FA4), which forms the Lower Bajocian Agoudim 3 Formation. This facies succession indicates a new progradation of the carbonate system, leading to the development of the more massive middle-ramp carbonates of the Bin El Ouidane 3 Formation during the Late Bajocian. A thick, regionally consistent oolitic grainstone unit/bed signifies the maximum progradation. Although precise biostratigraphic dating for the top of this sequence is unavailable, it may correspond to the first progradation prism of the Assoul Limestone, dated as early Late Bajocian in the Amellago area ([Poisson *et al.*, 1998](#)). Sequence S3 is further subdivided into two lower-order sequences. **Sequence S3.1** includes the transgressive systems tract of S3 and an initial prograding prism, ending at the base of the Bin El Ouidane 3 Formation with the first middle-ramp deposits (FA3). The transgressive base of **Sequence S3.2** records the

aggradation and subsequent retrogradation of reef bioconstructions, well-developed in the lower Bin El Ouidane 3 Formation. A subsequent regressive phase led to the progradation of distal middle-ramp oncolitic facies (FA2) and high-energy oolitic ramp (FA1) across the entire study area.

6.3.1 Sequence S3.1 (Upper Aalenian – Lower Bajocian)

The regional drowning recorded at the beginning of Sequence S3 led to the homogenization of the distal ramp facies (FA5) everywhere in the Imilchil area (NS1, NS2 and EW3). The sequence ends with the progradation of oolitic (FA1) and bioconstructed (FA3) carbonate ramps, on the margins of the deeper axial zone, centered on the Lake and Ikassene minibasins (*c.f.* sequence S3.2). Progradation directions of the carbonate ramps are orthogonal to the main NE-SW direction of the diapiric ridges. However, the ridges do not seem to significantly affect these regional progradation. Moreover, in detail, facies belts can be oblique to the diapiric ridges, a fortiori in the deepest parts of the basin (*c.f.* sequence S3.2 on the Tassent ridge). Progradation directions and therefore facies belts are mainly controlled by regional subsidence and thus by the regional structuration of the basin. The thickness of this sequence, around 230 m, increases of a few tens of meters in the Lake minibasin that was, at this time, the axis of the Atlas basin (NS1). However, the sequence thins gradually but quickly on the flanks of the diapiric ridges. On the Ikkou ridge's flanks, its thickness decreases of about 100 m over 4 km (NS1).

6.3.2 Sequence S3.2 (Upper Bajocian)

This sequence records 1) to the south, the development of reef bioconstructions (FA3) in the transgressive systems tract (Fig. 9c) followed by the northwestward progradation of middle-ramp oolitic and oncolitic deposits (FA1 and FA2) within the regressive systems tract (Fig. 9d), and 2) to the north, the aggradation followed by the southeastward progradation of middle-ramp oolitic facies (FA1). These two carbonate systems of opposite polarity displaying different facies successions were separated by a deeper and SSE-NNW trending domain, globally centered on the Tassent area. This sequence presents major thickness variations in the different minibasins. The Lake minibasin was apparently subsiding twice more than the two others (NS1). In the very stable tectonic context characterizing this period, these local subsidence rate variations are interpreted as the result of salt withdrawal associated to diapir-rise. Diapiric activity also influenced facies distribution, especially on the southern ramp. A slightly deeper trough was located along the axis of the Tilmi minibasin and ends southward around the western tip of the Amagmag ridge. This trough is marked by the disappearance of bioconstructions in the transgressive systems tract and of oncolitic facies in the regressive systems tract, for the benefit of slightly deeper deposits. The flanks of the ridges are also characterized by an important thinning of the layers and lateral facies variations. The sequence is indeed thinning from 90 to 50 meters on the northwestern flank of the Amagmag ridge. Bioconstructed deposits (FA3) followed by oncolitic and oolitic facies (FA1, FA2) located close to the ridge grade to outer-ramp marls and limestone (FA4) in the center of the

Tilmi minibasin. The overall prograding sequence S3.2 displays clearer evidence of the influence of diapiric ridges on facies distribution compared to previous sequences. This is partly because the depositional environments of middle ramp become shallower and therefore facies variations become much more sensitive to bathymetric variations compared to deeper outer-ramp environments.

6.4 Sequence S4 (Upper Bajocian)

Sequence S4 begins with an aggrading-retrograding systems tract, represented by the carbonate-dominated succession within the upper half of the Bin El Ouidane 3 Formation. Overlying this unit are distal-ramp or offshore marly deposits, typically marking the commencement of the mixed Imilchil Formation. These facies signify the drowning of the carbonate system and are interpreted as the maximum flooding of Sequence S4. This carbonate system drowning and replacement by a mixed system occurred almost synchronously across the High-Atlas, dated as Late Bajocian (Monbaron, 1981, 1985; Jossen, 1990). The precise reasons for this Late Bajocian decline in carbonate production and its potential link to siliciclastic influx are beyond the scope of this paper. Within the lower Imilchil Formation, Sequence S4 continues with a regressive systems tract, evolving from upper offshore storm-influenced deposits (FA9) to intertidal deposits with mud-cracks, red clays, and evaporite pseudomorphs (FA6). Dating of adjacent deposits (Monbaron, 1985) indicates this regressive event occurred around the Bajocian-Bathonian transition.

6.4.1 Sequence S4.1 (Upper Bajocian)

Sequence S4 is further divided into three lower-order sequences. **Sequence S4.1** starts with a transgressive systems tract characterized by a thinning-up succession of distal middle-ramp oncolitic limestone (FA2), followed by outer-ramp marls and limestone, and finally marls (FA4/FA5). The re-emergence of middle-ramp oncolitic or oolitic facies marks a renewed progradation of the carbonate system during this sequence's regressive phase. This T/R sequence can be correlated across the study area and, based on its stacking pattern, may correspond to the unit dated as the middle part of the Late Bajocian within the Assoul Limestone of the Amellago area (Durllet in Pierre, 2006). Similar to the preceding sequence, **Sequence S4.2** consists of middle and outer ramp deposits, signifying another phase of carbonate system retrogradation and progradation. Sequences S4.1 and S4.2 are characterized by the persistence of the Bin El Ouidane carbonate system and belong to the overall aggrading lower part of Sequence S4. The base of **Sequence S4.3** records a rapid drowning of the carbonate ramp and its replacement by the mixed system, which rapidly exhibits a regressive evolution characteristic of the top of Sequence S4. Variations in the organization of sequences S4.2 and S4.3 at the transition between carbonate and mixed systems may be linked to a faster diapir-rise rate during this period, causing localized diachronism between the Bin El Ouidane and Imilchil formations (see §8).

The retrogradation of the carbonate systems at the beginning of sequence S4.1 led to the homogenization of outer-ramp marls and limestone facies (FA4) in the studied

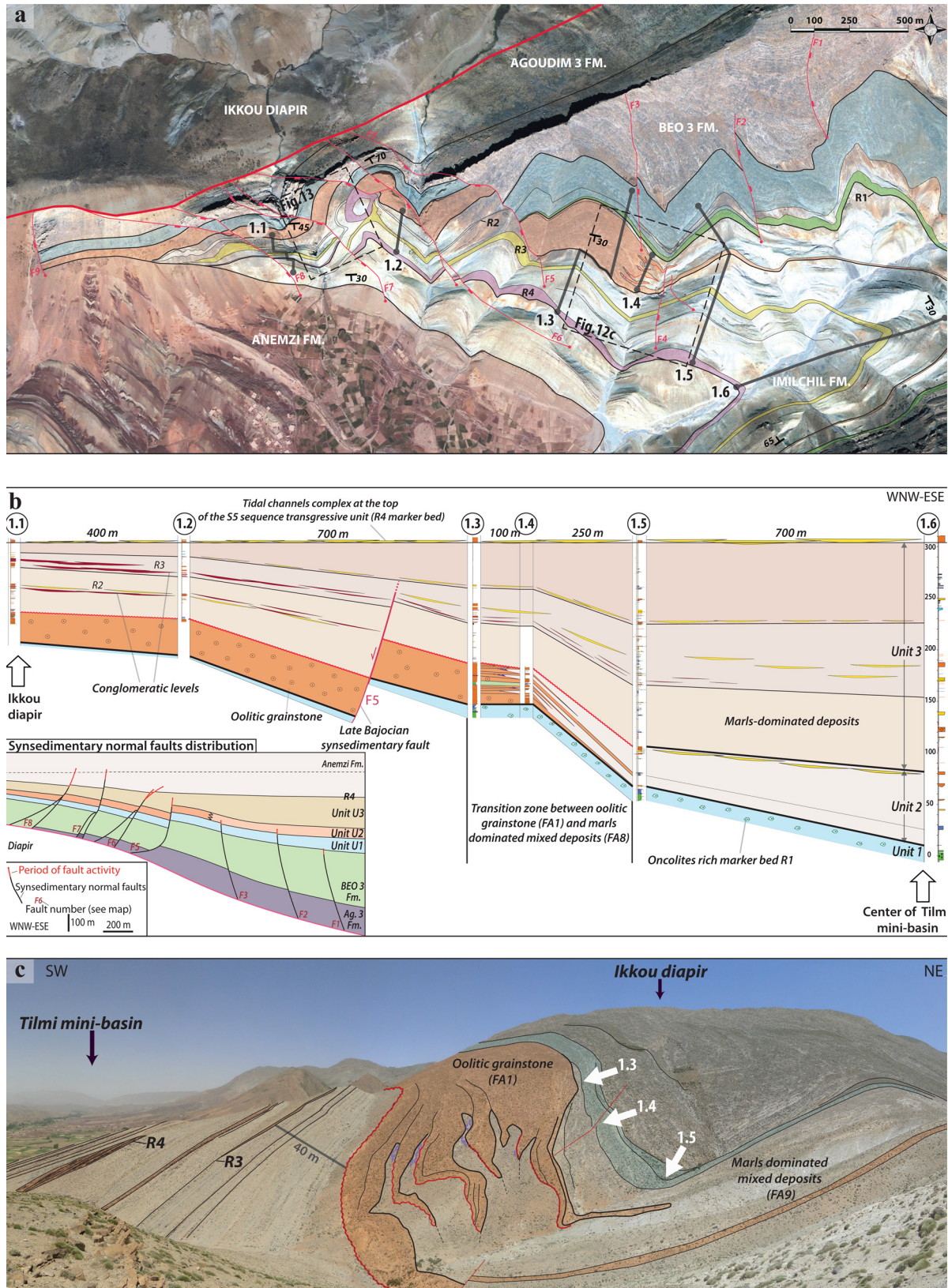


Fig. 11. A) Detailed geological map of the Amalou area. B) Detailed correlation panel (location in A)) of sequence S4.3 (base of Imilchil Formation) on the south flank of Ikkou ridge. C) Field view showing the facies variations in the Amalou area (location in A)) showing oolitic grainstone, close to the ridge, grading to marly dominated mixed deposits in the center of Tilmi mini-basin. See location in Figure 3a.

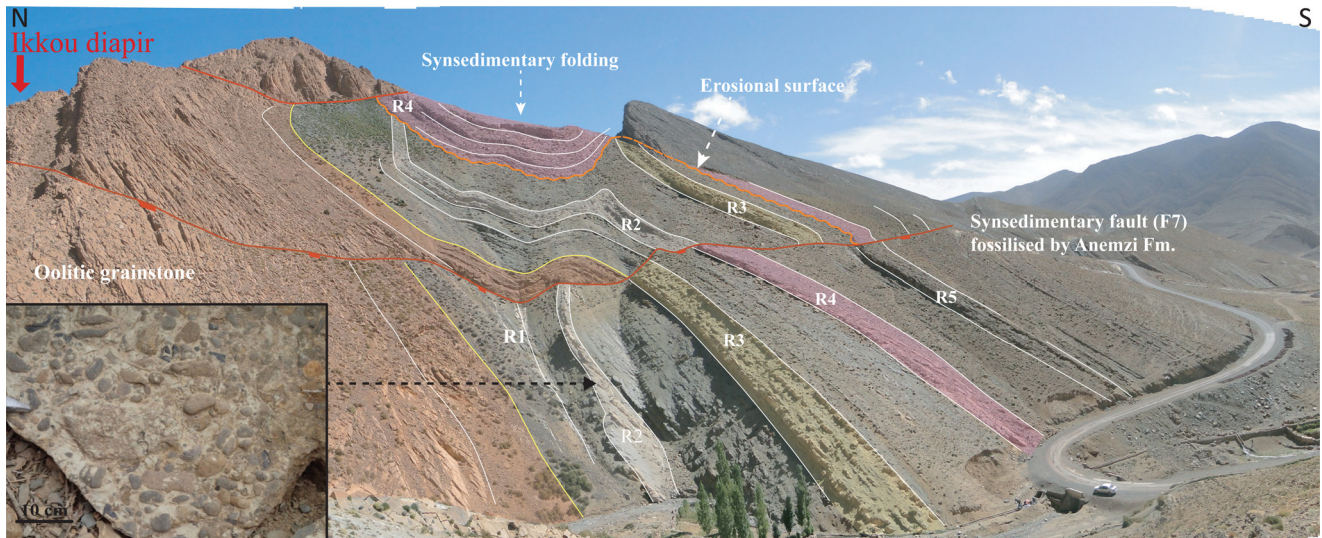


Fig. 12. Field view of the south flank of the Ikkou diapir showing complex halokinetic geometries in sequence S4.3. The layers mostly made of conglomeratic deposits (marker beds R1, R2 and R3) show synsedimentary folding and internal erosions forming angular unconformities that can reach 90° (e.g., marker bed R2). See location in [Figure 12a](#).

area. The upper part of sequence S4.1 records a new progradation of the carbonate ramps. In the Ikassene minibasin to the north, oolitic wedges are prograding southeastward. To the south, the northwestward progradation of the southern ramp is marked by oolitic, oncolitic and reef facies, though these do not extend as far as those in the previous sequence (S3.2). In the Tilmi minibasin, the heterogeneity of facies distribution due to diapiric movements first faded during the regional deepening of the system. This heterogeneity then reappeared in the shallower deposits associated with the progradation of the system. The thickness of sequence S4.1 strongly varies between the different minibasins: 290 m in the Lake minibasin, and only 150 m and 220 m in the Tilmi and Ikassene minibasins respectively (NS1). Thicknesses also strongly vary within the same minibasin: 290 m in the western part and only 70 m in the eastern part of the Lake minibasin (NS2). This high variability of subsidence rates is considered as the result of diapir-rise, responsible for a regionally heterogeneous downbuilding phenomenon.

This sequence displays larger thickness variations than the previous one on the flanks of all diapiric ridges. As an example, this can be seen in the Tilmi minibasin where thicknesses vary from 160 to 50 m on the southwestern flank of the Ikkou ridge. These thinning are locally associated with lateral facies variations. As an example, on the southeastern flank of Ikkou ridge, oncolitic middle-ramp deposits (FA2) laterally change to deeper marls and limestone (FA4) over 1 to 3 km towards the center of the Tilmi minibasin (NS1). On the other hand, no significant facies variation was recorded on the flanks of the Tassent and Amagmag ridges (NS1 and NS2). In that case, the variations of accommodation rates induced by diapiric-ridge growth are probably balanced by sedimentation rates and have therefore no impact on the depositional profiles. The preferential location of facies variations on the flanks of the Ikkou ridge could be due to a quicker growth of this ridge, at the origin of a more distinct topographic high. This topographic high could have induced an earlier development of oncolitic facies.

6.4.2 Sequence S4.2 (Upper Bajocian)

Facies distribution in this mostly aggrading sequence is very similar to the upper part of sequence S3. However, it shows a noticeable progression of the oolitic middle-ramp facies in the north (FA1) and of oncolitic facies (FA2) and reef bioconstructions (FA3) in the south, therefore reducing the width of the deeper domain, still centered on the Tassent region ([Fig. 9e](#)). At this time, the paleogeographic complexity of the Tilmi minibasin was also reduced. This sequence gets thinner on the flanks of all diapiric ridges. These thickness variations are however less important than those of the previous sequence (S4.1). They are systematically associated with very localized facies variations, proving the influence of diapiric ridges on paleo-bathymetry. For example, on the southern flank of the Tassent ridge, an oolitic body (FA1), about twenty meters thick, prograded southeastward over several hundreds of meters and then laterally grades to outer-ramp marls and limestone (FA4) in the Lake minibasin ([Fig. 9e](#)).

As for the previous sequence, sequence S4.2 records a relatively higher activity of the Ikkou ridge, expressed by more marked thickness and facies variations than around other ridges. Thus, on the southern flank of the Ikkou ridge, a 50 m-thick oolitic and oncolitic carbonate unit (FA1, FA2) grades laterally to a twice-thicker outer-ramp marls and limestone unit (FA4) in the center of the Tilmi minibasin. In comparison, the variations of thickness (<10%) and facies are much less marked in the middle ramp deposits on the northern flank of the Amagmag ridge.

6.4.3 Sequence S4.3 (Upper Bajocian)

At the base of this sequence, the marls and limestone sedimentation including offshore storm deposits marks the drowning and end of the Bin El Ouidane carbonate system and the development of the Imilchil mixed system. The overlying intertidal marls and limestone record a quick decrease of bathymetry. This sequence presents a homogeneous facies distribution over the whole studied area ([Fig. 9f](#)). The

thickness of the deposits shows a significant increase of the clayey-carbonate sedimentation rate during the latest Bajocian. However, subsidence rates highly vary from one minibasin to another. The thickness of this sequence is indeed 200 m in the Tilmi minibasin and 900 m in the Lake minibasin along the western transect (NS1). Locally, substantial thickness variations on the flanks of the ridges show an intensification of diapiric movements. The thickness of this sequence varies from 125 m on the northern flank of the Ikkou ridge to 900 m in the center of the Lake minibasin (NS1). Similar order of thinning occurs on the other side of this minibasin, close to the Tassent ridge. Significant thickness variations are recorded in the Tilmi minibasin on the flanks of the Ikkou ridge and at a smaller scale on the flanks of the Amagmag ridge. In the eastern part of the minibasin, the thickness of this sequence varies from north to south from 120 m to 400 m and then 270 m between the Ikkou and Amagmag ridges. Facies variations are very well expressed in the Amalou area on the southern flank of the Ikkou ridge (Fig. 11). At the bottom of the sequence, a 50 m-thick carbonate wedge made of oolitic grainstone (FA1), interbedded with a few transgressive oncolitic layers and bioconstructions (FA2), remains on the flank of the Ikkou ridge and laterally grades, over 400 m and without transition, to deeper marls and limestone deposits (FA9) in the axis of the Tilmi minibasin. This permanence of carbonate systems around diapiric structures, within mainly siliciclastic systems is also described in La Popa Basin (Giles *et al.*, 2008). Above this oolitic body, a shallow-marine sedimentation made of marls and limestone extends homogeneously between the flank of the ridge and the center of the minibasin. This homogeneity is however interrupted by the intercalation of several few-meters-thick conglomeratic layers near the diapiric ridge (Figs. 11 and 12). These conglomerates have an erosive base and are slightly channelized (FA8c). They are made of well-rounded carbonated gravels and pebbles coming from the erosion of Bin El Ouidane Formation. Their extent is limited to 500 m around the ridge for the thickest layers (few-meters-thick) and to 1 km for the thinner layers. The composition of the conglomerates evidenced that Ikkou ridge and its overburden (*e.g.*, Bin El Ouidane) were exposed for the first time, but exposure and erosion of the Ikassene salt wall core cannot be proved.

This Upper Bajocian sequence (S4.3) is affected by syndimentary deformations on the flank of the Ikkou ridge. Disharmonic folds of hectometric wavelength affect the whole 200-m-thick succession at the contact with the ridge. They are associated with several intraformational unconformities (Fig. 12). They affect a well-stratified alternation of clays and limestone whereas the deformation of underlying carbonate formations, more competent, is of larger wavelength. Curved syndimentary normal faults of mean direction N170°E near the center of the minibasin and grading to N120°E before taking root in the Triassic are gradually fossilized from east to west by the upper layers of the Imilchil Formation and the lower layers of the Anemzi Formation. These steep faults (60° to 80°) are dipping eastward in the eastern part and dipping westward in the western part of the studied outcrop. The tectonic regime associated to these faults are not clearly understood but they could associate to dextral strike-slip motion along the ridge. These normal faults could also be associated to a radial fault system (Stewart, 2006;

Davison *et al.*, 2000) linked to the rise of the diapir (“drag zone” *sensu* Alsop *et al.*, 2000). The westward migration of the fault system activity could be linked to the westward dipping of the ridge, involving a higher diapir-rise to the east.

6.5 Sequence S5 (Bathonian)

The base of Sequence S5 indicates a minor transgressive trend, fostering an open marine depositional environment conducive to the development of reef bioconstructions (FA7) and a system of oolitic-bioclastic tidal channels and bars. This Early Bathonian interval (Fadile, 2003) is interpreted as the maximum flooding of Sequence S5. It is succeeded by regressive facies, including clays and carbonates deposited in intertidal to supratidal environments (FA6) at the top of the Imilchil Formation, followed by continental deposits (FA10) of the Anemzi Formation. The lower part of sequence S5 records a regional transgression expressed by the evolution from an intertidal environment (FA6) to a subtidal environment characterized by the emplacement of a complex of oolitic-bioclastic tidal bars and tidal channels (FA8) followed by reef bioconstructions (FA7). The sequence ends with a regressive phase leading to the stacking up of very thick and monotonous continental sediments, forming the Anemzi Formation. The thickness of this sequence is higher than 2000 m in the Tilmi minibasin. Its preserved thickness is much thinner in the Lake minibasin. Since no younger formation is preserved, there is no evidence to interpret that, unlike previous sequences, subsidence was higher in the Tilmi minibasin than in the Lake minibasin.

On the flanks of all diapiric ridges, layers of sequence S5 are involved in very well-marked progressive unconformities attesting intense diapiric activity during the Bathonian. These geometries are combined with many unconformities and inner erosional surfaces as can be seen in the western part of the Tilmi minibasin (Figs. 8, 13a, 13b, 13c). In this area, two angular unconformities are observed on the southern flank of the Ikkou ridge. The first one makes an angle of 20° to 30° and corresponds to the base of sequence S5. It separates intertidal facies (FA6) and intraformational breccia deposits (F6e) (Fig. 13d) alternating with paleosols (Fig. 13e). The composition and angular shape of the lithoclasts suggest a local and sub-aerial erosion of the underlying intertidal deposits (FA6) on this ridge flank. A second angular unconformity defines the base of the Anemzi Formation. Tracing this unconformity northward, it truncates the Imilchil, Bin El Ouidane 3, and Agoudim 3 and 2 Formations (part of sequences S2–S5), ultimately juxtaposing the Anemzi Formation against the diapiric core of the Ikkou ridge. This erosional surface provides evidence for the emergence and erosion of the sedimentary cover close to the diapiric ridge, culminating in the aerial exposure of the Triassic core of the Ikkou ridge during the deposition of sequence S5. These geometries are combined with facies variations within the marine deposits at the base of the sequence. Intertidal to supratidal facies, as well as reef bioconstructions on the flanks of the ridges, grade laterally to subtidal facies dominated by marls towards the axis of the minibasins (*e.g.*, SE flank of Ikkou ridge). These facies variations are not visible within the continental facies of Anemzi Formation, which remain

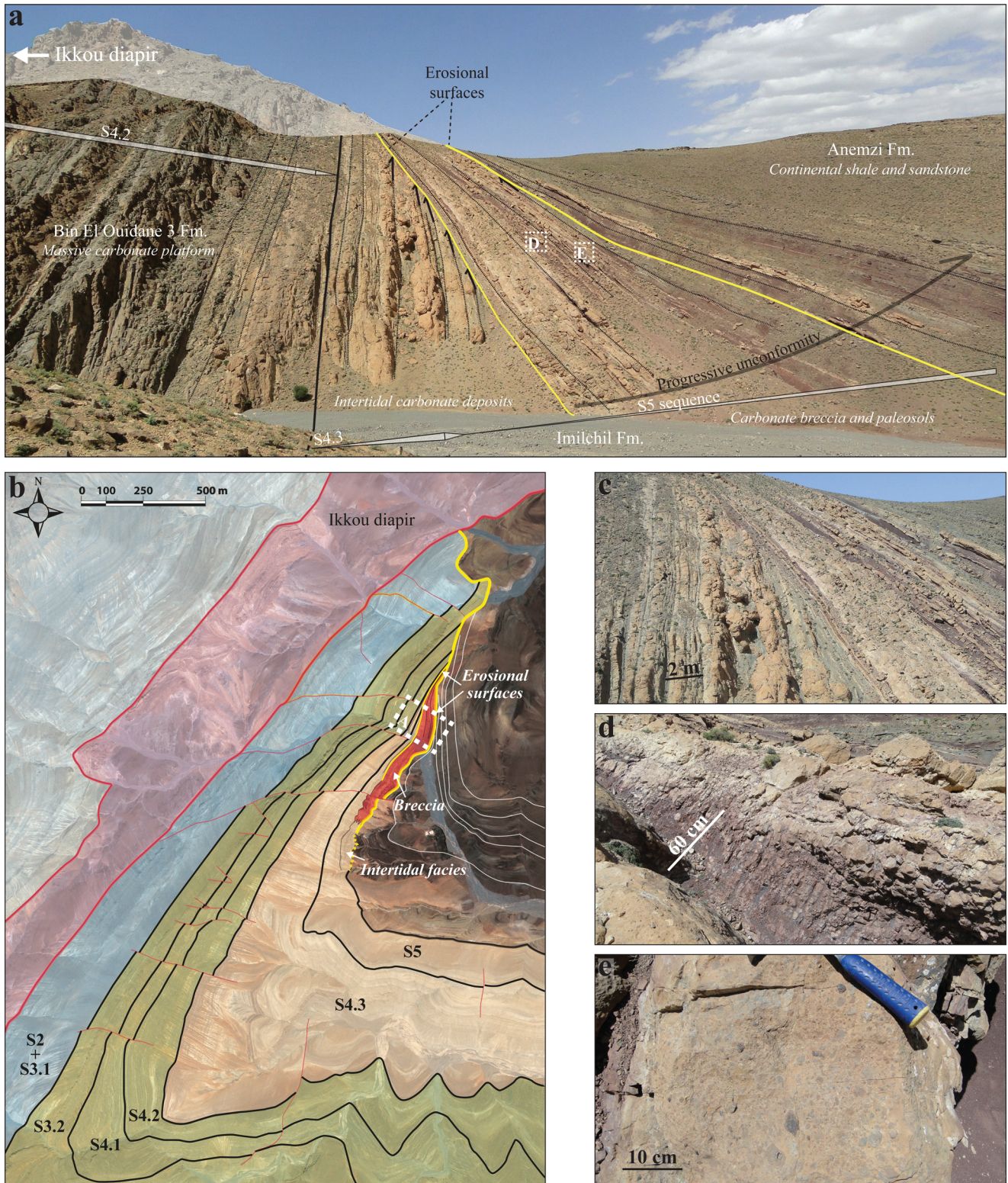


Fig. 13. a) Detailed geological map of the Aqqa-n-Sountat area. b) Field view of the south flank of Ikkou diapir (location in a)) showing the angular unconformities in the lower part of sequence S5. c) Zoom on the 20° to 30° angular unconformity in the lower part of sequence S5 (Imilchil Formation). d) Paleosol. e) Carbonate breccia. See location in [Figure 3a](#).

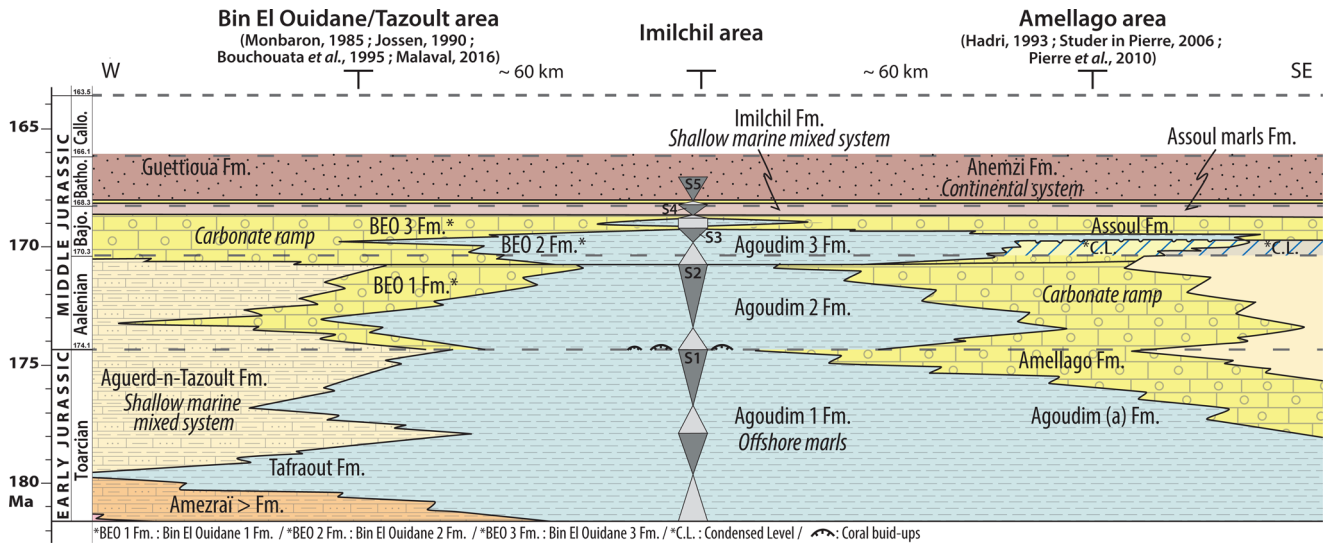


Fig. 14. Stratigraphic framework and sequential organization of the Toarcian to Callovian deposits in the central High Atlas. The southern margin of the Atlas basin is represented by the Amallago region, the central part by the Imilchil region and the northern and western margin by the Zaouiat Ahançal/Tazoult region.

homogeneous on the flanks of the diapirs as well as toward the center of the minibasins.

7 Discussion

7.1 Regional correlations and paleogeographic implications

Sequences defined in the Imilchil region can be correlated with those recognized in the Amellago area on the southern margin of the Atlas basin, 60 kilometers south-east of Imilchil (Hadri, 1993; Pierre *et al.*, 2010) and with those described in the western part of the Atlas basin in the Bin El Ouidane – Tazoult area (Monbaron, 1985; Jossen, 1990; Malaval, 2016) (Fig. 14).

In the Amellago area, sequence S1 can be correlated with the first prograding sequence of the Amellago Limestones (Am1) over the Agoudim marls (a) during the latest Toarcian (cycle 1 of Pierre *et al.*, 2010). After a brief retrogradation of the sedimentary system, sequence S2 corresponds to most of the prograding wedge of the Amellago Limestones (Am2 to Am5) dated as Aalenian (cycles C2 to C5 of Pierre *et al.*, 2010). The great transgressive phase that regionally marks the base of sequence S3 is first recorded in the Amellago area by the retrogradation of the carbonate system at the end of the Aalenian (cycles 6 to 7). It ends with a major flooding of the carbonate ramp, marked here as well as in Imilchil by a condensed layer of regional extent spanning *Discites*, *Laeviuscula* and *Propinquans* biozones. This condensed layer, which marks sequence S3 maximum flooding, is overlain by a regressive systems tract. It starts with the distal-ramp grey marls of the Agoudim member (c) (lateral equivalent of Agoudim 3 Formation) and ends with the onset of oolitic carbonate ramp deposits and reef bioconstructions represented by the Bajocian Assoul Limestones, lateral equivalent to the Bin El Ouidane 3 Formation in the Imilchil area. In Amellago as well as in Imilchil, this carbonate ramp sedimentation ends during the latest Bajocian with the development of an

argillaceous clastic influx, followed consecutively by the drowning of the carbonate system. This drowning event is recorded by the Assoul Marls that are of the same age as the Imilchil Formation (sequence S4). In the western part of the Atlas basin (Bin El Ouidane-Tazoult area), the Aalenian-Bajocian deposits present overall more proximal features and a higher proportion of clay. The sequence organization, even though it is not as clearly expressed, is similar to the one of age-equivalent series in Imilchil.

In the Zaouiat Ahançal area, sequences S1 and S2 as well as the transgressive systems tract of sequence S3 correspond to the aggradation of a thick subtidal to intertidal mixed succession (Taфраout and Aguerd-n-Tazoult Formations), followed by the progressive development of the Lower Bajocian transgressive carbonate ramp of the Bin El Ouidane Group (Jossen, 1990; Bouchouata *et al.*, 1995; Malaval, 2016). After a maximum flooding marked by middle-ramp oolitic deposits (Bin El Ouidane 1 Limestones of Monbaron, 1985), the Bin El Ouidane Formation records a regressive phase marked by inner-ramp marls and limestone deposits (Bin El Ouidane 2 Limestones). This regressive stage corresponds to the one recorded by the progradation of the middle-ramp facies during sequence S3 in Imilchil. In the Zaouiat Ahançal area, this regressive maximum is progressively overlain by middle-ramp oolitic deposits (Bin El Ouidane 3 Limestones) forming a transgressive carbonate unit at the top of the Bin El Ouidane Group. The uppermost Bajocian drowning of the carbonate ramp is first marked by the growth of reef bioconstructions (distal to middle ramp) and then by the deposition of green marls with bioclastic stormy layers at the base of the Tillouguit Formation, which is the same age as the Imilchil Formation. The Tillouguit Formation then displays the same sequence organization as the Imilchil Formation with a regressive trend (S4) followed by a T/R sequence (S5) ending with the red continental series of the Guettoua Formation, more proximal but of the same age as the Anemzi Formation (Monbaron, 1985; Jossen, 1990).

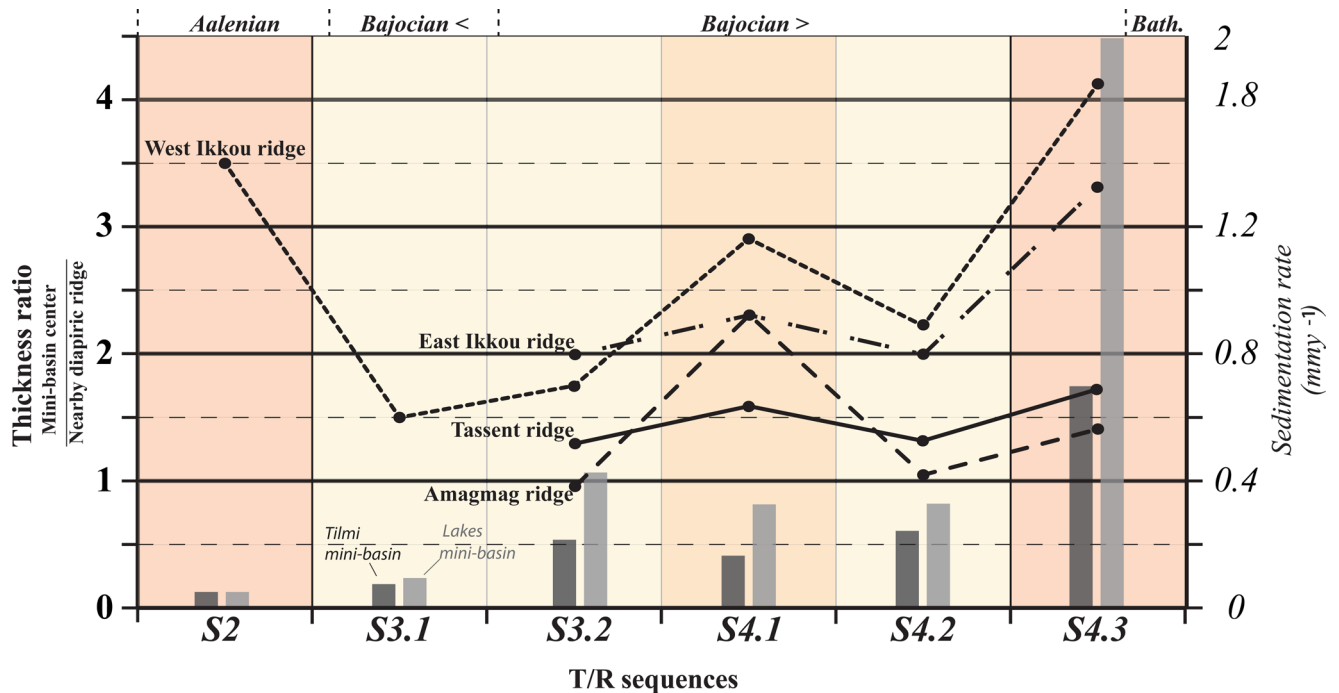


Fig. 15. Table comparing sedimentation rates (sticks) and relative intensity of diapiric movements (curves) for each transgressive-regressive sequence. The relative intensity of diapiric movements is inferred from bedding geometry on the flanks of the diapiric ridges, and more precisely from the ratio of sequence thickness between the center of the mini-basins and the flanks of the ridges. Sedimentation rates are calculated from non-decompacted thicknesses, therefore only their relative value is meaningful. The estimation of the intensity of diapiric movements should also be considered in a relative manner.

These regional correlations show that the Toarcian to Bathonian of the Imilchil region display more distal facies than in adjacent areas and therefore that the studied area was in the axial part of the Atlas basin. They also highlight the main tectono-sedimentary evolution stages of the whole Atlas basin. From Late Toarcian to Late Bajocian, a period of relative tectonic stability favored the development of a carbonate platform over the whole central High-Atlas domain, recording several transgression-regression cycles. During the latest Bajocian, almost simultaneously over the domain, a clayey siliciclastic influx and the drowning of the carbonate platform occurred. It was then replaced by a mixed system of SW-NE polarity. During the Early Bathonian, the increasing terrigenous input led to a rapid progradation of clay and sandstone continental deposits, ensuring the final fill of the subsiding domains of the Atlas basin. The general polarity of the latest Bajocian to Bathonian mixed and then siliciclastic systems tends to show that the origin of the terrigenous sediments was linked to the uplift and erosion of the West Moroccan Arch on the western part of the Atlas domain. Finally, these correlations show that the global paleogeographic evolution of the Atlas basin during Early and Middle Jurassic was controlled by regional to global parameters such as eustasy, regional subsidence and vertical movements along its boundaries but not by diapiric movements that impacted sedimentation at minibasin and salt-wall scales.

7.2 Timing of diapiric growth

In the Imilchil region, diapir movements have been identified in the oldest exposed strata dating back to the

Toarcian. However, studies of the Tazoult and Jebel Azourki salt walls, located to the southwest, indicate that diapirism was active since at least the Pliensbachian (Bouchouata *et al.*, 1995; Saura *et al.*, 2014; Malaval, 2016; Martin-Martin *et al.*, 2017). In the Imilchil region, the growth of diapiric ridges persists until at least the Late Bathonian to Callovian. In contrast, the Tazoult and Jebel Azourki salt walls are locally fossilized by the limestones of the Aït Abdi Formation (Malaval, 2016; Martin-Martin *et al.*, 2017), which transgressive base is dated as Late Aalenian (Jossen, 1990) and corresponds to the transgressive base of sequence S3 (Top of Agoudim 2 Formation). Other salt walls in this western region, such as the Abbadine salt wall, continue to rise until the Callovian, possibly extending into the Cretaceous (ongoing research). Although the growth of diapiric salt walls in the Imilchil area was continuous from the Toarcian to the Callovian, it was not linear. The stratal geometry of the deposits reveals two major active phases identified in all analyzed Tassent, Ikkou, Amagmag salt walls. The first phase occurred during sequences S1 and S2 (Toarcian–Aalenian), while the second phase took place during sequences S4.3 and S5 (latest Bajocian–Early Callovian). Additionally, a phase of reduced diapiric activity is observed during sequences S3 to S4.2 (latest Aalenian–Late Bajocian). Notably, sequence S4.1 indicates a slight intensification in salt wall rise compared to the surrounding sequences.

These phases of diapir growth, characterized by varying intensities can be partially observed in other areas of the Atlas basin. In the Tounfite region, located about ten kilometers northeast of Imilchil, two phases of diapir growth deformation,

previously interpreted as being of compressive origin, have been recognized and separated by a quiet phase (Stüder, 1980): the first one during the Toarcian–Aalenian and the second one during the Bajocian–Bathonian. In the Rich region, more than a hundred kilometers east of Imilchil, an intensification of diapiric movements was identified during the Late Bajocian–Early Bathonian, through the analysis of the Azag minibasin, (Teixell *et al.*, 2017), in accordance with the regional eastwards migration of subsidence indicated in Moragas *et al.* (2018). In the westernmost part of the Atlas basin, the Jebel Azourki, Tazoult and Abbadine salt walls were very active from Middle Liassic to Aalenian (Malaval, 2016; Moragas *et al.*, 2016; Martin-Martin *et al.*, 2017). The Jebel Azourki and Tazoult salt walls were fossilized in the Bajocian, while the Abbadine salt wall resumed during the Bathonian, as indicated by the halokinetic depositional sequences of this age preserved on its flanks (Monbaron, 1985). These similarities across the Atlas basin demonstrate that regional tectonic and sedimentary factors significantly influenced diapir movements.

In the Atlas basin, the initiation of diapir growth could result from an extensive phase of deformation identified during the Early Pliensbachian (Malaval, 2016). The decrease in the diapir rise rate observed during the deposition of sequences S3 to S4.2 (Bajocian), as well as the fossilization of some ridges further west, took place during a very stable tectonic phase characterized by the development of wide carbonate platforms throughout the basin. Given the diapir-related deformations affecting the Middle Liassic carbonates exposed at outcrop west of the studied area (Bouchouata *et al.*, 1995; Malaval, 2016; Martin-Martin *et al.*, 2017), we do not consider that the accumulation of carbonate deposits and their early diagenesis significantly influenced the decrease in diapir rise rate. The resumption of particularly active diapirism at the end of the Bajocian and during the Bathonian occurred during a specific tectono-sedimentary context in the evolution of the Atlas basin. Local fault activity during the growth of the Tilmi minibasin is observed in the Imilchil Formation suggesting strike-slip movements along the diapiric salt walls and the deeper structures they are associated with. Additionally, basic magmatic intrusions dated as Bathonian indicate an active geodynamic context (Laville and Harmand, 1982; Laville and Piqué, 1992). Finally, a significant eastward terrigenous flux occurred from the Late Bajocian and primarily during the Bathonian. This suggests an uplift and erosion of the western domain (West Moroccan Arch), which may be thermally related to the magmatism recognized during this period. This terrigenous flux contributed to the rapid disappearance of the carbonate systems, which were replaced by a mixed carbonate-siliciclastic system. This transition was associated with a significant increase of the sedimentation rate in the minibasins and a notable rise in the diapir growth rate.

At the end of Bajocian (sequence S4.3), a correlation can be established between the increasing diapir growth rate and significant terrigenous input, which elevates the sedimentation rate (Fig. 15). This observation fits with the model suggesting that evaporites or clay flows, responsible for diapir rise and the subsidence of minibasins, are significantly influenced by sediment load (Banham and Mountney, 2013; Peel *et al.*, 2020). However, this relationship does not hold true for the Aalenian sequence S2, during which the sedimentation rate in the Imilchil area remains low, in contrast to the western, more

proximal part of the basin (Malaval, 2016). In addition to these temporal fluctuations, the intensity of diapiric movements also varies between Imilchil structures. Stratal geometries along the rims of Tilmi minibasin indicate a more rapid growth of the Ikkou ridge compared to the Amagmag ridge. Unlike the Ikkou ridge, there is no evidence of subaerial exposure on the Amagmag ridge in the upper part of the Imilchil Formation (uppermost Bajocian). The emersion of this ridge probably occurred later, during the deposition of the Anemzi Formation. This possible north-to-south migration of the ridge growth seems to confirm the correlated migration of subsidence in the minibasins. During sequence S4.3, subsidence in the Lake minibasin is significantly greater than in the Tilmi minibasin (Imilchil Formation). The subsequent evolution of these depocenters cannot be determined due to the absence of marker beds at the top of the Anemzi Formation. The segmentation of the basin by these longitudinal diapir ridges likely played a crucial role in influencing the accumulation of fluvial deposits and the differential subsidence of the minibasins (Banham and Mountney, 2013). Ultimately, basin segmentation was limited by the complete expulsion of the Triassic clays and evaporites initially present beneath the minibasin, leading to a “roofing” phenomenon. We could hypothesize that the Lake minibasin underwent an earlier final filling, located closer to the northwestern margin of the basin, which was more subsiding during previous stages. This may have allowed for a migration of the depocenter toward the Tilmi minibasin, where subsidence potential was still preserved (Saura *et al.*, 2014).

7.3 Jurassic halokinetic depositional sequences

The Toarcian to Callovian period is characterized by three successive depositional systems in the Imilchil area: i) a system of carbonate ramps prograding from all directions toward the Atlas basin axis, located to the north of Imilchil; ii) a shallow mixed carbonate-siliciclastic system resulting from a significant eastward clastic flux from the latest Bajocian; and iii) an alluvial siliciclastic system from the Early Bathonian. The influence of diapir movements on these depositional systems is expressed differently at the scale of the minibasins (multi-kilometer scale) and at the scale of the close flanks of the diapir ridges (hectometer to kilometer scale).

Regionally, the primary effect of diapir growth relates to variations in subsidence rates, which influence the thickness of sequences not only between minibasins but also within each individual minibasin. These variations were first very limited until the Early Bajocian (Agoudim Formation, Sequence 1 to Sequence 3.1), even though a slight thickening is observed in the Ikassene minibasin, the axis of the Atlas basin. These variations are slightly more pronounced within the Upper Bajocian middle ramp carbonates (Bin El Ouidane Group, S3.2 to S4.2), with the Lake minibasin being slightly more subsiding than adjacent minibasins. From the latest Bajocian onward, the differences in subsidence rates between minibasins become quite significant. The mixed deposits of the Imilchil Formation (S4.3) are particularly thick in the Lake minibasin, while the continental deposits of the Anemzi Formation (S5) are predominantly preserved in Tilmi minibasin. At this stage, the emergence of the ridges leads to a clear compartmentalization of the Atlas basin, resulting in the formation of distinct erosive wedges and pinch-outs.

The second effect of diapirism at a multi-kilometer scale concerns facies distribution. Initially, the diapirs exert little influence on facies distribution within the relatively deep marls and limestones of the Agoudim Group. Then, the progradation of the Bin El Ouidane oolitic/bioconstructed ramp is slightly orthogonal to the general NE-SW orientation of the diapir ridges. These ridges do not appear to significantly affect regional progradation since, looking closely, facies belts can be seen to run obliquely to the diapir ridges, particularly in the deepest part of the basin (*cf.*, sequences S3.2 to S4.2 on the Tassent ridge). Progradation directions and associated facies belts are primarily controlled by regional subsidence and basin structuration, which also significantly influenced the orientation of the diapir ridges. From the latest Bajocian onward, sediments deposited in mixed systems (Imilchil Formation) and later in continental systems (Anemzi Formation) were quite homogeneous, showing no differentiation between the various minibasins, despite highly variable subsidence rates. At this stage, sedimentary flux was very high, sufficient to compensate for variations in subsidence rates and maintain a flat depositional profile across the studied area. Stratigraphic correlations with laterally equivalent series in the western part of the Atlas basin support this conclusion at a broader scale. One consequence of the high growth rate that led to the emersion of the diapir ridges was the compartmentalization of the basin, which intensified the effects of tidal currents. This process resulted in the development of a complex of tidal channels and sandwaves (FA8) within the Imilchil Formation.

Finally, the impact of diapirism on sedimentation is most evident in close proximity of the Central High Atlas ridges. This effect is observed over distances ranging from a few hundred meters to 2 km, characterized by the systematic thinning of the layers, which may be associated with facies variations. The evolution of the sedimentary systems (in terms of composition and bathymetry) and the diapir ridges (immersed or emerged) resulted in four main types of geometries and facies organization for the sedimentary wedges on the flanks of the ridges.

Type A sedimentary wedges are made of outer-ramp marls and limestone that are relatively homogeneous (FA4 and FA5), but exhibit progressive unconformities within a distance of 2 km from the ridge (Fig. 16a). These wedges are found in the relatively deep deposits of sequences S1, S2 and S3.1. The absence of significant facies variations indicates that the submarine topography resulting from diapir movements is insufficient for the deposition profile to reach shallower facies zones. This persistence of a subdued underwater topography is further supported by the lack of gravity failure features and gravity-related deposits in these wedges.

Type B sedimentary wedges are characterized by progressive unconformities and facies variations between oolitic-bioclastic middle ramp deposits (FA1), with oncolites (FA2) or bioconstructions (FA3), and outer ramp marls and limestone (FA4 and FA5; Fig. 16b). These facies variations occur over distances of 1 to 3 km from the flanks of the ridges. These wedges are found within the overall prograding and shallow sediments of sequences S3.2, S4.1, S4.2 and S4.3. In a shallow marine environment, the growth of diapir ridges tends to modify facies zonation, which is strongly influenced by paleo-bathymetric conditions. However, the submarine paleo-

topographies resulting from diapirism remain quite subtle. Notably, there are no gravity failure features affecting carbonate sedimentation, unlike other examples of diapir structures (Rezák *et al.*, 1985; Giles *et al.*, 2008; Bosence, 1998; Malaval, 2016). The transition from type A to type B wedges is more related to a decrease in water depth resulting from progradation of carbonate ramp systems than to changes in the dynamics of the diapir ridges. No evidence suggests the emersion of the diapir ridges at this stage. Additionally, there is no asymmetry in deposits observed on the opposite flanks of the same ridge, in contrast to observations in the La Popa basin, where wedges differ on the diapirs' flanks due to prevailing winds (Giles *et al.*, 2008). In the Imilchil area, a differentiation in facies zonation is observed at the basin scale between the northern and southern ramps, independent of diapir control.

Type C sedimentary wedges are also characterized by progressive unconformities and facies variations but they differ from the previous types by the presence of angular unconformities, inner erosion surfaces and interbedded conglomeratic layers (Fig. 16c). These wedges are found within the shallow-marine mixed system of the Imilchil Formation (top of S4.3, bottom of S5). Angular unconformities are associated with the bending of layers on the flanks of the ridges, creating "hooks" structures typical of diapiric settings (Giles and Rowan, 2012). These features can be further enhanced by short-wavelength disharmonic syndimentary folds. The conglomeratic layers, which extend up to 1 km from the diapir, consist of carbonate pebbles sourced from underlying strata, providing evidence of subaerial exposure at the top of the ridge during this stage. This period coincides with a time of very active diapirism at the end of the Bajocian. However, there is currently no evidence to confirm that the Triassic core of the ridges was exposed at the surface. The emersion of the ridges flanks (salt-walls) led to the segmentation of the basin, which contributed to the amplification of tidal processes during this period.

Type D sedimentary wedges are characterized by the onlap termination of strata against the Triassic core of the diapir ridges (Fig. 16d). These wedges are found exclusively within the continental layers of the Anemzi Formation, which record the most pronounced growth of the diapiric ridges. They indicate the piercing and aerial exposure of the ridge's Triassic core. Locally, complete erosion of the series flanking the ridges results in the intercalation of polygenic conglomerates within the continental deposits of the Anemzi Formation. Although no Triassic elements have been identified within these conglomerates, they have been noted further east around the Jbel Taoudelt ridge (Stüder, 1980).

In summary, the different sedimentary wedges, A to D identified in this study, were deposited sequentially, recording both the increasing activity of the diapiric ridges and the gradual filling of the basin. This infill evolved from shallowing-upward marine deposits to thick continental series. As a result, wedge geometries became more pronounced, and facies variations became more distinct in the shallow-marine carbonate deposits, which are particularly sensitive to changes in water depth. The initial subaerial exposure of the ridges is indicated by interbedded conglomeratic layers, while their ultimate emergence is shown by the onlap terminations of the series onto the flanks of the uplifted ridges.

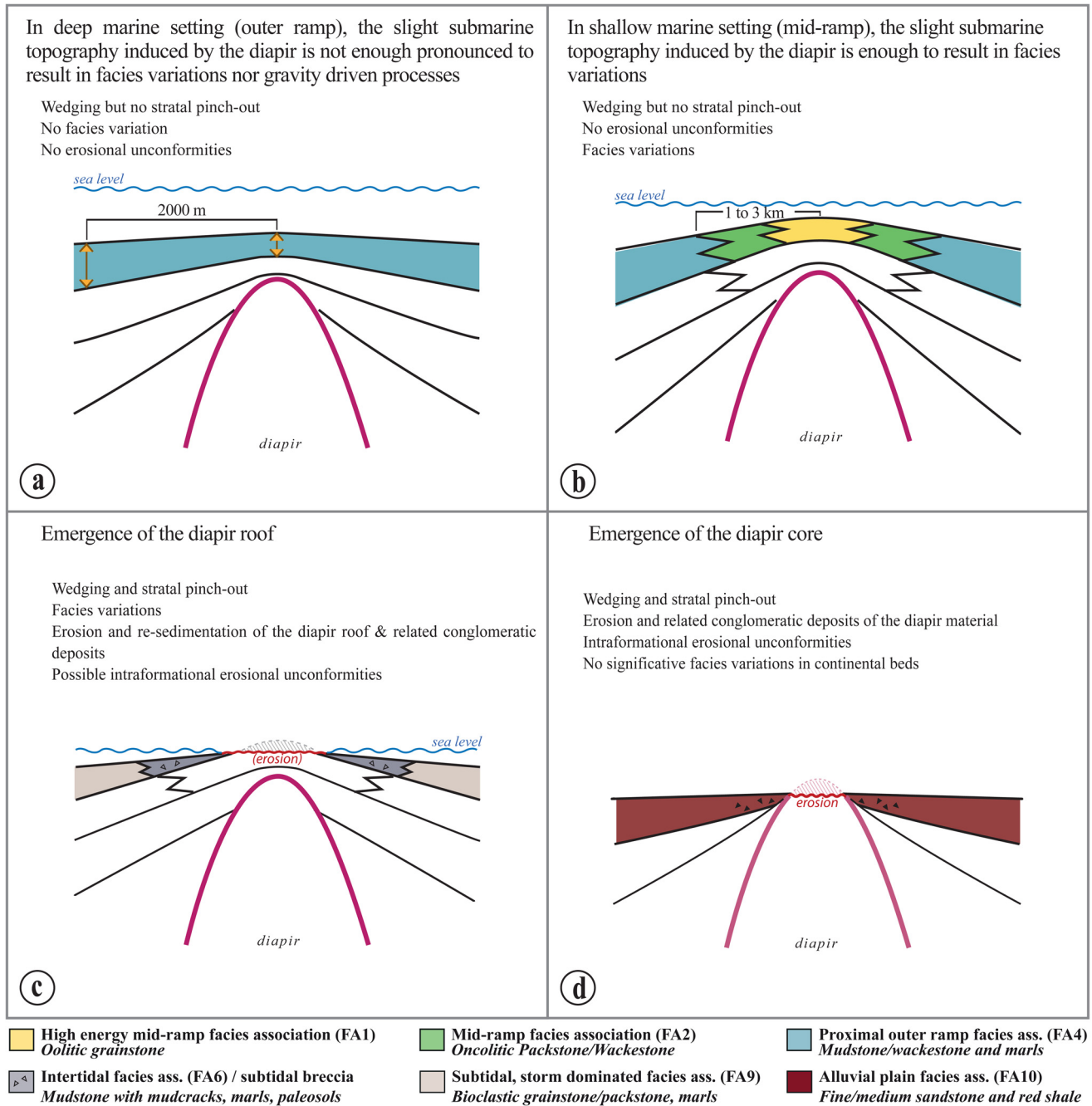


Fig. 16. Simplified sketches showing the diapiiric growth effects on stratigraphic geometries and facies distribution.

7.4 Comparison with the model of halokinetic sequences

The halokinetic-sequence concept, introduced by Giles and Lawton (2002), describes a succession of strata bounded by angular unconformities on the flanks of a diapiir. These unconformities become conformable surfaces with increasing distance from the diapiir, reflecting the growth of the diapiir itself. The ratio between the rate of diapiir rise and the sedimentation rate controls the formation of two distinct geometries (Giles and Rowan, 2012). “Hook halokinetic sequences” feature pronounced folding of the layers in the

immediate vicinity of the diapiir, sharp angular unconformities (up to 90°) and very rapid facies variations. These occur when the sedimentation rate is lower than the diapiir-rise rate. Conversely, “wedge halokinetic sequences” display a broader growth “fan like” structure, gentler angular unconformities (less than 30°), and more gradual facies variations. They occur when the sedimentation rate is higher than the diapiir-rise rate. In the context of the Imilchil Jurassic strata, the halokinetic sequence concept has limited applicability. In fact, angular unconformities are rarely observed on the flanks of the diapiirs, noted only locally and exclusively in the latest sequences (S4.3 and S5) within very shallow-marine (type C wedge) and

continental (type D wedge) deposits at the top of the Imilchil Formation and in the Anemzi Formation. No angular unconformities were identified in the marine deposits of earlier sequences, even on outcrops located in the close vicinity (< 1 km) of diapiric ridges. Instead, the sedimentary wedges on the flanks of these ridges exhibit very progressive growth structures over distances of 1 to 3 kilometers. Hook geometries with sharp angular unconformities (90°) were observed only in select locations, often combined with disharmonic folding and synsedimentary faults, within growth (fan-like) rather than tabular structure. Therefore, the geometry of these deposits is more comparable to *Tapered Composite Halokinetic Sequences*, which correspond to a stack of wedge halokinetic sequences (Giles and Rowan, 2012). However, this comparison cannot be strictly established, because (1) inner unconformities bounding halokinetic sequences appear only in Bathonian strata, and (2) no cyclic facies organization controlled by the growth of ridges was observed. The identified depositional sequences can be correlated across the entire studied area, indicating they do not result from local diapiric activity. Furthermore, there is no evident relationship between transgressive or regressive systems tracts and the geometry of these wedges. This contrasts with observations in the La Popa basin, where the generally more condensed transgressive systems tracts display hook geometries, while more developed regressive systems tracts show wedge geometries (Giles and Rowan, 2012; Rowan and Giles, 2023).

The predominance of “wedge” geometries, coupled with the scarcity of “hook” structures, the absence of inner unconformities and the lack of sequential motif that would indicate true halokinetic sequences in the marine Toarcian to Bajocian deposits, suggest that the rate of diapir growth was lower than the sedimentation rate during this period. It is only from the latest Bajocian onward that an acceleration in diapiric movements, along with the exposure of ridges, facilitated the formation of halokinetic unconformities. The notably high sedimentation rate during this time hindered the development of “hook” geometries and the sequential organization of facies associated with each ridge. Consequently, the growth of ridges only modulated the regional depositional framework at the basin scale rather than establishing distinct sequences tied to individual ridges.

8 Conclusions

This study investigated the detailed geometry, ages, and accumulation rates of several thousand-meter-thick halokinetic depositional sequences within the Imilchil diapiric area within the Central High Atlas in Morocco. Our research focused on the well-preserved Ikassene, Lakes, and Tilmi minibasins, whose evolution was directly controlled by the confining Tasraft, Tassent, and Ikkou salt walls.

Overall, the Jurassic sedimentation in the Imilchil minibasins began with a thick deposit of Toarcian–middle Bajocian ramp carbonates (more than 2,000 m). This was succeeded by over 1,000 m of middle Bajocian–middle Bathonian mixed carbonate-siliciclastic deposits, which finally graded upwards into a siliciclastic system.

Salt diapirism initiated in the Pliensbachian (middle Early Jurassic) and persisted until the Callovian (latest Middle Jurassic). While salt wall growth rates slowed slightly during the late Bajocian, coinciding with peak carbonate progradation, they accelerated significantly from the latest Bajocian onward. This later acceleration was coeval and related to the major increase in siliciclastic deposition and its eastward regional migration. Importantly, the spatial variability of diapiric growth created both submarine and subaerial paleotopographies over the ridges, leading to differential subsidence among the minibasins.

The most evident impact of the Central High Atlas diapiric activity is found on the salt wall flanks, where halokinetic depositional patterns directly reflect shallow-marine facies variations. Initially, subtle submarine paleotopographies from salt ridge growth did not significantly influence outer ramp carbonate sedimentation until the Late Bajocian. However, as the basin shallowed, these diapiric paleotopographies began to localize facies changes, promoting grainy bioclastic deposits or reefal bioconstructions atop the ridges. The eventual emersion and erosion of these ridges, which began in the Bathonian, is evidenced by restricted conglomeratic deposits at the minibasin margins.

This work provides compelling evidence for the intricate tectono-halokinetic-sedimentary links that governed Jurassic sedimentation across the High Atlas region. Beyond this regional focus, our findings offer invaluable insights into the fundamental interplay and dynamic feedback mechanisms between salt tectonics and basin evolution at diverse geological scales.

Acknowledgments

The core of the structural analysis was performed at Univ. Bordeaux, CNRS, Bordeaux INP (EPOC, UMR 5805) and was part of a collaborative project with Geosciences Barcelona (Geo3Bcn–CSIC, Spain) funded by Equinor ASA (Norway).

Supplementary material

Annex 1: Base map.

Annex 2: East-West transect 1.

Annex 3: East-West transect 2.

Annex 4: East-West transect 3.

Annex 5: North-South transect 1.

Annex 6: North-South transect 2.

The Supplementary Material is available at <https://www.bsgf.fr/10.1051/bsgf/2025023/olm>.

References

- Aït Addi A, Chellai EH, Ben Ismail MH. 1998. Les paléoenvironnements des faciès du Lias supérieur-Dogger du Haut Atlas d'Errachidia (Maroc). *Afr Geosci Rev* 5: 39–48.
- Aït Addi A, Chafiki D. 2013. Sedimentary evolution and palaeogeography of mid-Jurassic deposits of the Central High Atlas, Morocco. *J Afr Earth Sci* 84: 54–69.

- Allen JRL. 1982. Sedimentary structures, their character and physical basis, Vol. 2: Developments in sedimentology 30B. Amsterdam-New York: Elsevier Scientific Pub. Co., 663 p.
- Allen PA, Homewood P. 1984. Evolution and mechanics of a Miocene tidal sandwave. *Sedimentology* 31 (1): 63–81.
- Alsop GI, Brown JP, Davison I, Gibbling MR. 2000. The geometry of drag zones adjacent to salt diapirs. *J Geol Soc* 157 (5): 1019–1029.
- Armando G. 1999. Intracontinental alkaline magmatism: geology, petrography, mineralogy and geochemistry of the Jbel Hayim Massif (Central High Atlas – Morocco). Mémoires de Géologie, Université de Lausanne, Vol. 31, 106 p.
- Banham SG, Mountney NP. 2013. Evolution of fluvial systems in salt-walled minibasins: a review and new insights. *Sediment Geol* 296: 142–166.
- Beauchamp W, Barazangi M, Demnati A, El Alji M. 1996. Intracontinental rifting and inversion: Missour Basin and Atlas Mountains, Morocco. *Am Assoc Petrol Geol Bull* 80 (9): 1459–1482.
- Beauchamp W. 2004. Superposed folding resulting from inversion of a synrift accommodation zone, Atlas Mountains, Morocco. In: McClay KR, ed. Thrust tectonics and hydrocarbon systems. AAPG Memoir, Vol. 82, pp. 635–646.
- Benaouiss N., Courel L., Beauchamp J. (1996) Rift-controlled fluvial/tidal transitional series in the Oukai'meden Sandstones, High Atlas of Marrakesh (Morocco), *Sedimentary Geology*, Volume 107, Issues 1–2, P. 21–36 ISSN 0037-0738, [https://doi.org/10.1016/S0037-0738\(96\)00013-9](https://doi.org/10.1016/S0037-0738(96)00013-9).
- Bensalah MK, Youbi N, Mata J, Madeira J, Martins L, El Hachimi H, *et al.* 2013. The Jurassic-Cretaceous basaltic magmatism of the Oued El-Abid syncline (High Atlas, Morocco): physical volcanology, geochemistry and geodynamic implications. *J Afr Earth Sci* 81: 60–81.
- Bosence DWJ, Al-Aawah MH, Davison I, Rosen BR, Vita-Finzi C, Whitaker E. 1998. Salt domes and their control on basin margin sedimentation: a case study from the Tihama Plain, Yemen. In Purser BH, Bosence DWJ, eds. *Sedimentation and Tectonics in Rift Basins Red Sea: Gulf of Aden*. Londres: Chapman & Hall, p. 448–454.
- Bosence DWJ. 2005. A genetic classification of carbonate platforms based on their basinal and tectonic settings in the Cenozoic. *Sediment Geol* 175 (1-4): 49–72.
- Bouchouata A. 1994. La ride de Talmest-Tazoult (Haut-Atlas Central, Maroc). Lithostratigraphie, biostratigraphie et relations tectonique – sédimentation au cours du Jurassique. Thèse de Doctorat, Université de Toulouse III, Strata, Toulouse, (serie 2 : mémoires), Vol. 25, 223 p.
- Bouchouata A, Canérot J, Souhel A, Alméras Y. 1995. Stratigraphie séquentielle et évolution géodynamique du Jurassique dans la région de Talmest-Tazoult (Haut Atlas central, Maroc). *Comptes Rendus de l'Académie des Sciences*, t. 320, (série IIA), p. 749–756.
- Bourillot R, Neige P, Pierre A, Durllet C. 2008. Early-Middle Jurassic lycoceratid ammonites with constrictions from Morocco: palaeobiogeographical and evolutionary implications. *Palaeontology* 51: 597–609.
- Bracene R, Patriat M, Ellouz N, Gaulier JM. 2003. Subsidence history in basins of northern Algeria. *Sediment Geol* 156 (1-4): 213–239.
- Brede R, Hauptmann M, Herbig HG. 1992. Plate tectonics and intracratonic mountain ranges in Morocco – The mesozoic-cenozoic development of the Central High Atlas and the Middle Atlas. *Geol Rundschau* 81 (1): 127–141.
- Brede R. 1992. Structural aspects of the Middle and the High Atlas (Morocco) phenomena and causalities. *Geol Rundschau* 81 (1): 171–184.
- Brun JP, Fort X. 2004. Compressional salt tectonics (Angolan margin). *Tectonophysics* 382 (3-4): 129–150.
- Brun JP, Fort X. 2008. Entre sel et terre. Structures et mécanismes de la tectonique salifère. Paris, Vuibert, Coll. «Interactions», 154 p.
- Catuneanu O, Galloway WE, Kendall CGStC, Miall AD, Posamentier LHW, Strasser A, *et al.* 2011. Sequence stratigraphy: methodology and nomenclature. *Newslett Stratigr* 44 (3): 175–245.
- Charrière A, Haddoumi H, Mojon PO. 2005. Découverte du Jurassique supérieur et d'un niveau marin du Barrémien dans les “couches rouges” continentales du Haut Atlas central marocain : implications paléogéographiques et structurales. *CR Palevol* 4 (5): 385–394.
- Charrière A, Haddoumi H, Mojon PO, Ferrière J, Cuche D, Zili, L. 2009. Mise en évidence par charophytes et ostracodes de l'âge Paléocène des dépôts discordants sur les rides anticlinales de la région d'Imilchil (Haut Atlas, Maroc) : conséquences paléogéographiques et structurales. *CR Palevol* 8 (1): 9–19.
- Cochet E, Duffaud F, Guy M, Issenmann O, Taussac R. 1970. Carte géologique du Maroc au 1/100 000, Tamanar. *Notes et Mémoires du Service Géologique du Maroc*, (201).
- Courel L, Aït Salem H, Benaouiss N, Et-Touhami M, Fekirine B, Oujidi M, Soussi M, Tourani A. (2003) Mid-Triassic to Early Liassic clastic/evaporitic deposits over the Maghreb Platform, Palaeogeography, Palaeoclimatology, Palaeoecology, Volume 196, Issues 1–2, Pages 157–176, ISSN 0031-0182, [https://doi.org/10.1016/S0031-0182\(03\)00317-1](https://doi.org/10.1016/S0031-0182(03)00317-1).
- Davison I, Alsop I, Birch P, Elders C, Evans N, Nicholson H, *et al.* 2000. Geometry and late-stage structural evolution of Central Graben salt diapirs, North sea. *Mar Petrol Geol* 17 (4): 499–522.
- Dewey JF, Helman ML, Knott SD, Turco E, Hutton DHW. 1989. Kinematics of the western Mediterranean. *Geol Soc London Spec Publ* 45 (1): 265–283.
- Domènech M, Teixell A, Babault J, Arboleya ML. 2015. The inverted Triassic rift of the Marrakech High Atlas : a reappraisal of basin geometries and faulting histories. *Tectonophysics* 666: 177–191.
- Domènech M, Teixell A, Stockli DF. 2016. Magnitude of rift-related burial and orogenic contraction in the Marrakech High Atlas revealed by zircon (U-Th)/He thermochronology and thermal modeling. *Tectonics* 35 (11): 2609–2635. .
- Du Dresnay R. (1971). Extension et développement des phénomènes récifaux jurassiques dans le domaine atlasique marocain, particulièrement au Lias moyen Bull. Soc. géol. Fr. Delate in text Du Dresnay 1972.
- Duée G, Hervouët Y, Laville E, Luca de P, Robillard D. 1978. L'accident nord moyen-atlasique dans la région de Boulemane (Maroc) : une zone de coulissement synsédimentaire. *Ann Soc Géol Nord* 98: 145–162.
- El Arabi EH. 2007. La série permienne et triasique du rift haut-atlasique: nouvelles datations; évolution tectono-sédimentaire, Thèse de Doctorat non publiée, Université Hassan II Casablanca, Fac. Sc. Ain Chok, 225 p.
- El Harfi A, Lang J, Salomon J, Chellai EH. 2001. Cenozoic sedimentary dynamics of the Ouarzazate foreland basin (central High Atlas Mountains, Morocco). *Int J Earth Sci* 90: 393–411.
- El Hariri K, Dommergues JL, Meister C, Souhel A, Chafiki D. 1996. Les ammonites du Lias inférieur et moyen du Haut-Atlas central de Béni Mellal (Maroc) : taxinomie et biostratigraphie à haute résolution. *Geobios* 29 (5): 537–576.
- Ellouz N, Patriat M, Gaulier JM, Bouatmani R, Sabounji S. 2003. From rifting to Alpine inversion: Mesozoic and Cenozoic

- subsidence history of some Moroccan basins. *Sediment Geol* 156 (1-4): 185–212.
- Elmi S, Amhoud H, Boutakiout M, Benshili K. 1999. Cadre biostratigraphique et environnemental de l'évolution du paléorelief du Jebel Bou Dahar (Haut Atlas oriental, Maroc) au cours du Jurassique inférieur et moyen. *Bull Soc Géol France* 170 (5): 619–628.
- Embry AF, Johannessen EP. 1992. T-R sequence stratigraphy, facies analysis and reservoir distribution in the uppermost Triassic-Lower Jurassic succession, western Sverdrup Basin, Arctic Canada. In Vorren TO, Bergsager E, Dahl-Stamnes OA, Holter E, Johansen B, Lie E, Lund TB, eds. Arctic geology and petroleum potential vol. 2 (Special Publication), Norwegian Petroleum Society (NPF), p. 121–146.
- Frederic OE, Lepretre R, Spina V, Gimeno-Vives O, Kergaravat C, Mohn G. 2021. Dominique Frizon de Lamotte – Polyphased mesozoic rifting from the Atlas to the north-west Africa paleomargin. *Earth-Sci Rev* 220 (2021) 103732.
- Ettaki M, Chellai EH, Milhi A, Sadki D, Boudchiche L. 2000. Le passage Lias moyen – Lias supérieur dans la région de Todra-Dadès : événements bio- sédimentaires et géodynamiques (Haut Atlas central, Maroc). *CR Acad Sci* 331 (10): 667–674.
- Ettaki M, Ibouh H, Chellai EH, Milhi A. 2007. Les structures “diapiriques” liasiques du Haut-Atlas central, Maroc: exemple de la ride d'Ikerzi. *Afr Geosci Rev* 14 (1-2): 79–94.
- Fadile A. 2003. Carte géologique du Maroc au 1/100 000, Imilchil. Notes et Mémoires du Service Géologique du Maroc, Rabat, n° 397.
- Faugères E, Brun JP. 1984. Modélisation expérimentale de la distension continentale. *CR Acad Sci* 299(Série II): 365–370.
- Fedan B. 1988. Evolution géodynamique d'un bassin intraplaque sur décrochements: Le Moyen Atlas (Maroc) durant le Méso-cénozoïque. Thèse Sciences, Université Mohammed V de Rabat, 388 p.
- Flügel E. 2004. Microfacies of Carbonate Rocks, Analysis, Interpretation and Application. Springer-Verlag Berlin Heidelberg, 976 p.
- Fort X, Brun J-P., Chauvel F. 2004. Salt tectonics on the Angolan margin, syndensimentary deformation processes. *Bull Am Assoc Petrol Geol* 88 (11): 1523–1544.
- Frank J. Peel, Michael R. Hudec, Ruud Weijermars, (2020) Salt diapir downbuilding: Fast analytical models based on rates of salt supply and sedimentation, Journal of Structural Geology, Volume 141, 104202, ISSN 0191-8141, <https://doi.org/10.1016/j.jsg.2020.104202>
- Fraissinet C, Zouine EM, Morel JL, Poisson A, Andrieux J, Faure-Muret A. 1988. Structural evolution of the southern and northern central High Atlas in Paleogene and MioPliocene times) In: Jacobshagen V, ed. *The Atlas system of Morocco*. Berlin Heidelberg New York: Springer, pp 273–291.
- Frizon de Lamotte D, Saint Bezar B, Bracene R. 2000. The two main steps of the atlas building and geodynamics of the Western Mediterranean. *Tectonics* 19: 740–761.
- Frizon De Lamotte D, Zizi M, Missenard Y, Hafid M, El Azzouzi M, Maury RC, *et al.* 2008. The Atlas system. In Michard A, Saddiqi O, Chalouan A, Frizon de Lamotte D, eds. *Continental evolution: the geology of Morocco*. Springer-Verlag Berlin Heidelberg: Lecture Notes in Earth Sciences 116, p. 133–202.
- Frizon de Lamotte D, Raulin C, Mouchot N, Wrobel-Daveau J-C., Blanpied C, Ringenbach JC. 2011. The southernmost margin of the Tethys realm during the Mesozoic and Cenozoic: initial geometry and timing of the inversion processes. *Tectonics* 30 (3): 1–22.
- Froitzheim N, Stets J, Wurster P. 1988. Aspects of Western High Atlas tectonics. In: dans Jacobshagen V, ed. *The Atlas system of Morocco, Lectures notes Earth Sciences*, p. 219–244.
- Giles KA, Lawton TF. 2002. Halokinetic sequence stratigraphy adjacent to the El Papalote Diapir, Northeastern Mexico. *Am Assoc Petrol Geol Bull* 86 (5): 823–840.
- Giles KA, Druke DC, Mercer DW, Hunnicutt-Mack L. 2008. Controls on Upper Cretaceous (Maastrichtian) heterozoan carbonate platforms developed on salt diapirs, La Popa Basin, NE Mexico. In Lukasik J, Simo JA, eds. Controls on carbonate platform development. Society for Sedimentary Geology (SEPM), Tulsa: Special Publication Vol. 89, p. 107–124.
- Giles KA, Rowan MG. 2012. Concepts in halokinetic-sequence deformation and stratigraphy. *Geol Soc London Spec Publ* 363 (1): 7–31.
- Haddoumi H, Charrière A, Mojon PO. 2010. Stratigraphie et sédimentologie des « Couches rouges » continentales du Jurassique-Crétacé du Haut Atlas central (Maroc): implications paléogéographiques et géodynamiques. *Geobios* 43 (4): 433–451.
- Hadri M. 1993. Un modèle de plate-forme carbonatée au Lias-Dogger dans le Haut Atlas central au nord-ouest de Goulmina, Maroc. Thèse de Doctorat, Université Paris- Sud, Orsay, 285 p. (inédit).
- Hafid M, Zizi M, Bally AW, Ait Salem A. 2006. Structural styles of the western onshore and offshore termination of the High Atlas, Morocco. *CR Geosci* 338 (1-2): 50–64.
- Hailwood EA, Mitchell JG. 1971. Palaeomagnetic and radiometric dating results from Jurassic intrusions in South Morocco. *Geophys J Int* 24 (4): 351–364.
- Hlaïem A. 1999. Halokinesis and structural evolution of the major features in eastern and southern Tunisian Atlas. *Tectonophysics* 306 (1): 79–95.
- Hofling R. 1989. Substrate-induced morphotypes and intra-specific variability in Upper Cretaceous scleractinians of the eastern Alps (West Germany, Austria). *Mem Assoc Australas Palaeontol* 8: 51–60.
- Hudec MR, Jackson MPA. 2007. Terra infirma: understanding salt tectonics. *Earth Sci Rev* 82 (1-2): 1–28.
- Hudec MR, Jackson MPA. 2009. Interaction between spreading salt canopies and their peripheral thrust systems. *J Struct Geol* 31 (10): 1114–1129.
- Ibouh H, Bouabdelli M, Zargouni F. 1994. Indices de tectonique synsédimentaire dans les dépôts aaléno-bajociens de la région d'Imilchil (Haut Atlas Central, Maroc). *Miscellanea del Servizio Geologico Nazionale (Roma)* 5: 305–310.
- Jabour H, Dakki M, Nahim M, Charrat F, El Alji M, Hssain M, *et al.* 2004. The Jurassic depositional system of Morocco, geology and play concepts. *MAPG Memory* (1): 5–39.
- Jackson MPA, Talbot CJ. 1991. A glossary of salt tectonics. The University of Texas at Austin, Bureau of Economic Geology, Geological Circular Vol. 91-4, 44 p.
- Jackson MPA, Vendeville BC, Schultz-Ela DD. 1994. Structural dynamics of salt systems. *Ann Rev Earth Planet Sci* 22: 93–117.
- Jenny J, Le Marrec A, Monbaron M. 1981. Les Couches rouges du Jurassique moyen du Haut Atlas central (Maroc) : corrélations lithostratigraphiques, éléments de datations et cadre tectono-sédimentaire. *Bull Soc Géol France* 23 (6): 627–639.
- Jenny J. 1984. Dynamique de la phase tectonique synsédimentaire du Jurassique moyen dans le Haut-Atlas central (Maroc). *Eclogae geologicae Helvetiae*, 77 (1): 143–152.

- Jenny J. 1988. Mémoire explicatif de la carte géologique du Maroc, 1904 feuille d'Azilal au 1/100 000. Notes et Mémoires du Service 1905 Géologique du Maroc, n° 339 bis.
- Jossen JA. 1990. Carte géologique du Maroc au 1/100 000, Zawyat Ahançal. Notes et Mémoires du Service Géologique du Maroc, n° 355.
- Joussiaume Remi. (2016). Les relations entre diapirisme et sédimentation : Exemple du Jurassique moyen de la région d'Imilchil, Haut-Atlas central, Maroc.
- Kenter J.A.M., and Campbell A.E. 1991. Sedimentation on a Lower Jurassic carbonate platform flank: geometry, sediment fabric and related depositional structures (Djebel Bou Dahar, High Atlas, Morocco): *Sedimentary Geology*, 72, p. 1–34.
- Knight KB, Nomade S, Renne PR, Marzoli A, Bertrand H, Youbi N. 2004. The Central Atlantic Magmatic Province at the Triassic-Jurassic boundary: paleomagnetic and 40Ar/39Ar evidence from Morocco for brief, episodic volcanism. *Earth Planet Sci Lett* 228: 143–160.
- Khoms S, Roure F, Vergés J. 2022. Hinterland and foreland structures of the eastern Maghreb Tell and Atlas thrust belts: tectonic controlling factors, pending questions, and oil/gas exploration potential of the Pre-Triassic traps. *Arab J Geosci* 15 (6): 462. <https://doi.org/10.1007/s12517-022-09707-x>.
- Lachkar N. 2000. Dynamique sédimentaire d'un bassin extensif sur la marge sud-téthysienne: le Lias du Haut Atlas de Rich (Maroc). Thèse de Doctorat, Université de Bourgogne, Dijon, 275 p. (unpublished).
- Lachkar N, Guiraud M, Harfi A, Dommergues J-L., Dera G, Durlot C. 2009. Early Jurassic normal faulting in a carbonate extensional basin: Characterization of tectonically driven platform drowning (High Atlas rift, Morocco). *J Geol Soc* 413–430.
- Laville E. 1978. Incidence des jeux successifs d'un accident synsédimentaire sur les structures plicatives du versant nord du Haut Atlas central (Maroc). *Bull Soc Géol France* (7): 20/3, 329–337.
- Laville E, Harmand CL. 1982. Évolution magmatique et tectonique du bassin intracontinental mésozoïque du Haut-Atlas (Maroc); un modèle de mise en place synsédimentaire de massifs "anorogéniques" liés à des décrochements. *Bull Soc Géol France* 24 (2): 213–227.
- Laville E, Petit JP. 1984. Role of synsedimentary strike-slip faults in the formation of Moroccan Triassic basins. *Geology* 12 (7): 424–427.
- Laville E. 1988. A multiple releasing and restraining stepover model for the Jurassic strikeslip basin of the Central High Atlas (Morocco). In: Manspeizer W, ed. *Triassic- Jurassic Rifting, Continental Breakup and the Origin of the Atlantic Ocean and Passive Margins, Part A. –Develop. –Geotect.*, Vol. 22, p. 499–523.
- Laville E, Fedan B. 1989. Le système atlasique marocain au Jurassique. Évolution structurale et cadre géodynamique. *Sci Géol Mém Strasbourg* (84): 3–28.
- Laville E, Piqué A. 1992. Jurassic penetrative deformation and Cenozoic uplift in the central High Atlas (Morocco): a tectonic model. Structural and orogenic inversions. *Geol Rundschau* 81 (1): 157–170.
- Laville E, Piqué A, Amrhar M, Charroud M. 2004. A restatement of the Mesozoic Atlasic rifting (Morocco). *J Afr Earth Sci* 38 (2): 145–153.
- Lhachmi A, Lorand JP, Fabries J. 2001. Petrogenesis of the Anemzi Mesozoic alkaline intrusion, central High Atlas, Morocco. *J Afr Earth Sci* 32 (4): 741–764.
- Malaval M. 2016. Enregistrement sédimentaire de l'activité diapirique associée à la ride du Jbel Azourki. Haut-Atlas Central, Maroc. Thèse de doctorat, Université de Bordeaux. pp. 1–437.
- Martín-Martín JD, Vergés J, Saura E, Moragas M, Messenger G, Razin P, *et al.* 2017. Diapiric growth within an Early Jurassic rift basin: The Tazoult salt wall (Central High Atlas, Morocco). *Tectonics* 25: 35. <https://doi.wiley.com/10.1002/2016TC004300>.
- Marzoli A, Bertrand H, Knight KB, Cirilli S, Buratti N, Vérati C, *et al.* 2004. Synchrony of the Central Atlantic magmatic province and the Triassic-Jurassic boundary climatic and biotic crisis. *Geology* 32 (11): 973–976.
- Marzoli A, Jourdan F, Puffer JH, Cuppone T, Tanner LH, Weems RE, *et al.* 2011. Timing and duration of the Central Atlantic magmatic province in the Newark and Culpeper basins, eastern U.S.A. *Lithos* 122 (3–4): 175–188.
- Mattauer M, Tapponnier P, Proust F. 1977. Sur les mécanismes de formation des chaînes intracontinentales ; l'exemple des chaînes atlasiques du Maroc. *Bull Soc Géol France* 7 (3): 521–526.
- Merino-Tomé O, Della Porta G, Kenter JAM, Verwer K, Harris PM, Adams E, Playton T, Corrochano D. 2012. Sequence development in an isolated carbonate platform (Lower Jurassic, Djebel Bou Dahar, High Atlas, Morocco): influence of tectonics, eustasy and carbonate production. *Sedimentology* 59:118–155.
- Michard A. 1976. Éléments de géologie marocaine, Notes et Mémoires du Service Géologique du Maroc, n° 252, 422 p.
- Michard A, Ibouh H, Charrière A. 2011. Syncline-topped anticlinal ridges from the High Atlas: a Moroccan conundrum, and inspiring structures from the Syrian Arc, Israel. *Terra Nova* 23 (5): 314–323.
- Monbaron M. 1981. Sédimentation, tectonique synsédimentaire et magmatisme basique: l'évolution paléogéographique et structurale de l'Atlas de Béni Mellal (Maroc) au cours du mésozoïque; ses indices sur la tectonique tertiaire. *Eclogae Geol Helvetiae* 74 (3): 625–638.
- Monbaron M. 1985. Carte géologique du Maroc au 1/100.000, Beni-Mellal. Notes et Mémoires du Service Géologique du Maroc, n° 341.
- Moragas M, *et al.* (2016), Jurassic rifting to post-rift subsidence evolution in the central High Atlas during coeval salt diapirism, *Basin Res.*, <https://doi.org/10.1111/bre.12223>.
- Moragas M, Vergés J, Saura E, Martín-Martín J-D., Messenger G, Merino-Tomé O, *et al.* 2018. Jurassic rifting to post-rift subsidence analysis in the Central High Atlas and its relation to salt diapirism. *Basin Res.* 29. <https://onlinelibrary.wiley.com/doi/10.1111/bre.12223>.
- Moragas M, Baqués V, Travé A, Martín-Martín JD, Saura E, Messenger G, *et al.* 2020. Diagenetic evolution of lower Jurassic platform carbonates flanking the Tazoult salt wall (Central High Atlas, Morocco). *Basin Res* 32 (3): 546–566. <https://doi.org/10.1111/bre.12382>.
- Nichols GJ, Fisher JA. 2007. Processes, facies and architecture of fluvial distributary system deposits. *Sediment Geol* 195 (1): 75–90.
- Orszag-Sperber F, Harwood G, Kendall AC, Purser BH. 1998. A review of the evaporites of the Red Sea-Gulf of Suez rift. In Purser BH, Bosence DWJ, eds. *Sedimentation and Tectonics in Rift Basins Red Sea: – Gulf of Aden*. Londres: Chapman & Hall, p. 409–426.
- Parizot O, Frizon de Lamotte D, Missenard Y. 2023. A new look at old debates about the Corbières (NE-Pyrenees) geology: salt tectonics and gravity gliding. *BSGF - Earth Sciences Bulletin* 194: 6.
- Perthuisot V, Bouzenoune A, Hatira N, Henry B, Laatar E, Mansouri A, *et al.* 1999. Les diapirs du Maghreb oriental ; part des déformations alpines et des structures initiales crétacées et éocènes dans les formes actuelles. *Bull Soc Géol France* 170 (1): 57–65.
- Pierre A. 2006. Un exemple de référence pour les systèmes de rampes oolitiques. Un affleurement continu de 37 km de long (falaises jurassiques d'Amellago, Haut Atlas, Maroc). Thèse de Doctorat, Université de Bourgogne, Dijon, 242p.

- Pierre A, Durllet C, Razin P, Chellai EH. 2010. Spatial and temporal distribution of ooids along a Jurassic carbonate ramp: Amellago transect, High Atlas, Morocco. *Geol Soc London Spec Publ* 329 (1): 65–88.
- Piqué A, Charroud M, Laville E, Aït Brahim L, Amrhar M. 2000. The Tethys southern margin in Morocco; Mesozoic and Cenozoic evolution of the Atlas domain. In : Crasquin-Soleau S, Barrier É, eds. *Peri-Tethys Memoir 5: new data Peri Tethyan sedimentary basins*, Mémoire du Muséum national d'Histoire naturelle, vol. 182, p. 93–106.
- Poisson A, Hadri M, Milhi A, Julien M, Andrieux J. 1998. The central High- Atlas (Morocco). Litho- and chrono-stratigraphic correlations during Jurassic times between Tinjdad and Tounfite. Origin of subsidence. *Mém Mus Natl Hist Nat* 179: 237–256.
- Poprawski Y, Basile C, Agirrezabala LM, Jaillard E, Gaudin M, Jacquin T. 2014. Sedimentary and structural record of the Albian growth of the Bakio salt diapir (the Basque Country, northern Spain). *Basin Res* 26 (6): 746–766.
- Poprawski Y, Basile C, Agirrezabala LM, Jaillard E, Gaudin, Lopez M. 2016. Halokinetic sequences in carbonate systems: an example from the middle Albian bakio breccias formation (Basque Country, Spain). *Sediment Geol* 334: 34–52. <https://doi.org/10.1016/j.sedgeo.2016.01.013>.
- Purkis SJ, Harris PMM, Ellis J. 2012. Patterns of sedimentation in the contemporary Red Sea as an analog for ancient carbonates in rift settings. *J Sediment Res* 82 (11): 859–870.
- Rahimi A, Saidi A, Bouabdelli M, Beraaouz EH, Rocci G. 1997. Crystallization and fractionation of the post-Liassic intrusive series of Tasraft (central High Atlas, Morocco). *CR Acad Sci (série Ila)* 324 (3): 197–203.
- Read JF. 1985. Carbonate platform facies models. *Am Assoc Petrol Geol Bull* 69 (1): 1–21.
- Rebouillat J-P. 1983. Les milieux de sédimentation et les étapes de la transgression du Dogger dans la région de Demnat, Haut Atlas central (Maroc) ». – Thèse 3e cycle Univ. Dijon.
- Reineck HE. 1963. Sedimentgefüge im Bereich der südlichen Nordsee. *Abhandlungen der Senckenbergische Naturforschende Gesellschaft* 505: 1–138.
- Rezak R, Bright TJ, McGrail DW. 1985. Reefs and banks of the northwestern Gulf of Mexico: their geolocial, biological and physical dynamics. New York: Wiley, 323 p.
- Ribes C, Kergaravat C, Bonnel C, Crumeyrolle P, Callot JP, Poisson A, *et al.* 2015. Fluvial Sedimentation in a salt-controlled minibasin: stratal patterns and facies assemblages, Sivas Basin, Turkey. *Sedimentology* 62 (6): 1513–1545.
- Rowan MG, Lawton TF, Giles KA, Ratliff RA. 2003. Near-salt deformation in La Popa basin, Mexico, and the northern Gulf of Mexico: A general model for passive diapirism. *Am Assoc Petrol Geol Bull* 87 (5): 733–756.
- Rowan, Mark & Giles, Katherine. (2023). Different scales of salt–sediment interaction during passive diapirism. *AAPG Bulletin*. 107. 7-22. [10.1306/0104202221069](https://doi.org/10.1306/0104202221069).
- Rowlands G, Purkis S, Bruckner A. 2014. Diversity in the geomorphology of shallow-water carbonate depositional systems in the Saudi Arabian Red Sea. *Geomorphology* 222: 3–13.
- Sadki D. 1992. Les variations de faciès et les discontinuités de sédimentation dans le Lias-Dogger du Haut-Atlas Central (Maroc) ; chronologie, caractérisation, corrélations. *Bull Soc Géol France* 163 (2): 179–186.
- Saura E, Vergés J, Martín-Martín JD, Messenger G, Moragas M, Razin P, *et al.* 2014. Syn- to post-rift diapirism and minibasins of the Central High Atlas (Morocco): the changing face of a mountain belt. *J Geol Soc* 171 (1): 97–105. <https://www.lyellcollection.org/doi/10.1144/jgs2013-079>.
- Shelley DC, Lawton TF. 2005. Sequence stratigraphy of tidally influenced deposits in a salt-withdrawal minibasin: Upper sandstone member of the Potrerillos Formation (Paleocene), La Popa Basin, Mexico. *Am Assoc Petrol Geol Bull* 89 (9): 1157–1179.
- Souhel A. 1996. Le Mésozoïque dans le Haut-Atlas de Beni Mellal au Maroc. Stratigraphie, sédimentologie et evolution géodynamique. *Strata, Toulouse*, (serie 2 : mémoires), 27: 265.
- Stewart SA. 2006. Implications of passive salt diapir kinematics for reservoir segmentation by radial and concentric faults. *Mar Petrol Geol* 23 (8): 843–853.
- Stüder MA. 1980. Tectonique et pétrographie des roches sédimentaires, éruptives et métamorphiques de la région de Tounfit-Tirhist (Haut-Atlas central, Maroc). Thèse de Doctorat, Université de Neuchâtel, 102 p. (inédit).
- Teixell A, Arboleya ML, Julivert M, Charroud M. 2003. Tectonic shortening and topography in the central High Atlas (Morocco). *Tectonics* 22 (5).
- Teixell A, Barnolas A, Rosales I, Arboleya ML. 2017. Structural and facies architecture of a diapir-related carbonate minibasin (Lower and Middle Jurassic, High Atlas, Morocco). *Mar Petrol Geol*. <https://doi.org/10.1016/j.marpetgeo.2017.01.003>.
- Teixell A, Hudec RH, Arboleya ML, Fernandez N. 2024. 3D variation of shortened salt walls from the Moroccan Atlas: Influence of salt inclusions and suprasalt sedimentary wedges. *J Struct Geol* 183. <https://doi.org/10.1016/j.jsg.2024.105125>.
- Tesón E, Teixell A, Ayarza P, Arboleya M, Alvarez-Lobato F, Garcia-Castellanos D, *et al.* (2006). Geometry and evolution of the Ouarzazate basin in the foreland of the High Atlas Mountains (Morocco).
- Tesón E, Teixell A. 2008. Sequence of thrusting and syntectonic sedimentation in the eastern Sub-Atlas thrust belt (Dadès and Mgoun valleys, Morocco). *Int J Earth Sci* 97 (1): 103–113. <https://doi.org/10.1007/s00531-006-0151-1>.
- Tesón E, Pueyo EL, Teixell A, Barnolas A, Agustí J, Furió M. 2010. Magnetostratigraphy of the Ouarzazate Basin: Implications for the timing of deformation and mountain building in the High Atlas Mountains of Morocco. *Geodin Acta* 23 (4): 151–165. <https://doi.org/10.3166/ga.23.151-165>.
- Tucker ME, Wright VP. 1990. Carbonate Sedimentology. Blackwell Scientific Publications, Oxford London Edinburgh Boston Melbourne Berlin Paris Vienna, 482 p.
- Vail PR, Audemard E, Bowman SA, Eisner PN, Perez-Cruz C. 1991. The stratigraphic signatures of tectonics, eustasy and sedimentology-an overview. In Einsele G, Ricken W, Seilacher A, eds. *Cycles and events in stratigraphy*. Berlin: Springer-Verlag, p. 617–659.
- Vendeville BC, Jackson MPA. 1992a. The rise of diapirs during thin-skinned extension. *Mar Petrol Geol* 9 (4): 331–354.
- Vendeville BC, Jackson MPA. 1992b. The fall of diapirs during thin-skinned extension. *Mar Petrol Geol* 9 (4): 354–371.
- Vergés J, Moragas M, Martín-Martín JD, Saura E, Casciello E, Razin P, *et al.* 2017. Salt Tectonics in the Atlas Mountains of Morocco. In Soto G, Flinch JJ, Tari JF, eds. *Permo-Triassic Salt Provinces of Europe, North Africa and the Atlantic Margins*. Elsevier, pp. 563–579. <https://doi.org/10.1016/B978-0-12-809417-4.00027-6>.
- Vergés J, Moragas M, Ruh J. 2019. Multidisciplinary study of the Central High Atlas Diapiric Province in Morocco: results from analogue and numerical models. In Rossetti F, *et al.* eds. *The*

- structural geology contribution to the Africa-Eurasia geology: basement and reservoir structure, ore Mineralisation and tectonic modelling. *Advances in Science, Technology & Innovation*, pp. 225–228. https://doi.org/10.1007/978-3-030-01455-1_48.
- Weissmann GS, Hartley AJ, Scuderi LA, Nichols GJ, Davidson SK, Owen A, *et al.* 2013. Prograding distributive fluvial systems- Geomorphic models and ancient examples. In Driese SG, Nordt LC, eds. *New frontiers in paleopedology and terrestrial paleoclimatology*. SEPM Special Publication, Vol. 104, p. 131–147.
- Wilmsen M, Neuweiler F. 2008. Biosedimentology of the Early Jurassic post extinction carbonate depositional system, central High Atlas rift basin, Morocco. *Sedimentology* 55 (4): 773–807.
- Youbi N, Martins LT, Munha JM, Ibouh H, Madeira J, Ait Chayeb EM, *et al.* 2003. The Late Triassic-Early Jurassic volcanism of Morocco and Portugal in the geodynamic framework of the opening of the central Atlantic Ocean. In Hames WE, McHone JG, Renne PR, Ruppel C, eds. *The Central Atlantic Province; insights from fragments of Pangea*. Amer. Geophys. Union, Geophys. Monograph Vol. 136, pp. 179–207.
- Zouaghi T, Bédir M, Ayed-Khaled A, Lazzez M, Soua M, Amri A, *et al.* 2013. Autochthonous *versus* allochthonous Upper Triassic evaporites in the Sbiba graben, central Tunisia. *J Struct Geol* 52: 163–168.
- Zaagane M, Leprêtre R, Refas S, Bendella M, Mouassa S, Hachemi S, *et al.* 2025. The Ouarsenis “Grand Pic” : an exceptional example of a large-scale preserved halokinetic feature in the Western Tell (N. Algeria), BSGF, Forthcoming article.
- Zayane R, Essaifi A, Maury RC, Pique A, Laville E, Bouabdelli M. 2002. Cristallisation fractionnée et contamination crustale dans la série magmatique jurassique transitionnelle du Haut Atlas central (Maroc). *CR Géosci* 334 (2): 97–104.

Cite this article as: Joussiaume R, Malaval M, Razin P, Grélaud C, Martín-Martín JD, Saura E, Moragas M, Vergés J, Messenger G, Hunt D. 2026. Diapiric ridges and minibasins in the Central High Atlas: impact on geometries and facies distribution (Lower-Middle Jurassic, Morocco), *BSGF - Earth Sciences Bulletin* 197: 1. <https://doi.org/10.1051/bsgf/2025023>