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► **To cite this version:**

Frederic O Escosa, Rémi Leprêtre, Vincenzo Spina, Oriol Gimeno-Vives, Ch. Kergaravat, et al.. Polyphased mesozoic rifting from the Atlas to the north-west Africa paleomargin. *Earth-Science Reviews*, 2021, 220, pp.103732. 10.1016/j.earscirev.2021.103732 . hal-03518687

HAL Id: hal-03518687

<https://hal.science/hal-03518687v1>

Submitted on 10 Jan 2022

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Published in Earth-Science Review, in 2021

doi : <https://doi.org/10.1016/j.earscirev.2021.103732>

Polyphased mesozoic rifting from the Atlas to the north-west Africa paleomargin

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Abstract

Based on the interpretation of geological maps, seismic reflection and well data complemented with a bibliographic compilation and field work in the Rif, we investigate the factors that control the rift initiation, its development and the formation of oceanic crust in NW Africa. From SE to NW, we examine the Western Sahara Atlas, the Tindrara, the Guercif, and the Rif basins, to establish their geodynamic evolution in relation with the Mesozoic formation of the Central Atlantic and Maghrebien Tethys oceans, respectively. The Triassic extension was diffuse and developed over Lower Carboniferous horst-and-graben structures formed in the NW passive margin of Gondwana and involved in the subsequent late Carboniferous – early Permian Variscan orogenic system; suggesting that, at the onset of the Triassic rifting, the lithosphere was thermally re-equilibrated and replaced by more fertile lithospheric mantle. Afterwards, extension resumed in the Atlas system during middle to late Pliensbachian and finished during Toarcian. In the Rif and Guercif basins, the extension began later, mainly during the Toarcian, climaxing during Middle Jurassic times with the exhumation along low-angle extensional faults of CAMP gabbroic bodies and the final mantle exhumation during Upper Jurassic in the Rif. The study evidences the prominent role of the Variscan structural and thermal inheritance on the subsequent deformation events. Accordingly, the Paleozoic inverted basins and

horsts localized the Triassic extension. From that, the opening of the Central Atlantic and Maghrebian Tethys oceans activated, respectively, the SE (Atlas) and NW (Tethys) rifted segments of the weakened continental crust where the Jurassic extension was gradually distributed. This led to the final formation of an oceanic domain in the NW paleomargin of Africa.

1. Introduction

The opening of an oceanic domain can be the consequence of a single episode of rifting as observed along the southern margin of the Paleo-Tethys giving birth to the Neo-Tethys Ocean during the Permian (Stampfli et al., 1991; Ricou, 1994; Korsch et al., 2009). It can also result from the superimposition of several rifting events separated by quiescent periods finally leading to fully rifted continental margins. This, for instance, is the case of the failed late Carboniferous–Permian rift over the Variscan Orogen (upper Carboniferous to early Permian) and the further development, in this weakened zone, of the Triassic–Lower Jurassic Central Atlantic rift system (Burg et al., 1994; Oftedal et al., 2005; Redfern et al., 2010).

The specific features and the tectonic evolution of a rift system are mainly controlled by the rate of extension, the presence of inherited structures, and the rheology of the stretched lithosphere that varies according to its composition and thermal state (Steckler and ten Brink, 1986; Kuszniir et al., 1987; Buck, 1991; Stampfli et al., 1991; Hopper and Buck, 1996; Brun, 1999; van Wijk and Cloetingh, 2002; Gueydan and Précigout, 2014). In this regard, the weakening of the lithosphere and ultimate breakup is linked to two main elements: (1) the horizontal tectonic stress acting at the boundaries of the initial rift system, which are mostly related to the so-called “slab pull” mechanism (McKenzie, 1978; Lithgow-Bertelloni and Richards, 1998); and (2) the heat source related to the presence of a mantle plume responsible of thermal weakening at the base of the lithosphere (Burke and Dewey, 1973). These two elements are not self-excluding, so that they may occur in combination through the rift history (Ziegler and Cloetingh, 2004; Huismans et al., 2001; Merle, 2011). In addition to these constrains, the initial set-up of the lithosphere (i.e. the structural inheritance represented, for example, by collapsed orogens, older rift systems or lithospheric-scale faults) also controls the localization and subsequent development of a rift system (Frizon de Lamotte et al., 2015; Duretz et al., 2016; Petri et al., 2019).

Extension leading to the opening of an oceanic domain can affect a wide zone of the lithosphere forming a rift system (Şengör and Burke, 1978; Buck, 1991; Khain, 1992; Ziegler, 1992; Brun, 1999), with the subsequent development of “rim basins” bordering the spreading center. A key controlling factor on the width of a rift system, besides the ones described above, is also the modalities of rheological coupling between the layered crust and subcontinental mantle (Gueydan et al., 2008). At

the beginning of rifting, the deformation is decoupled and distributed over a large region, and as rifting evolves in time and space, it becomes progressively coupled and localized on the site of the future oceanic domain. This evolution leads to a final configuration in which the rift basins are older at the periphery of the system (e.g. “rim basins”) and younger towards the spreading axis. In this regard, numerical and physical scaled models show that the development of a wide rift zone is favored by the presence of a thick and hot crust (e.g. a young mountain belt), deformed over a long period of time at low strain rates (Buck, 1991; Brun, 1999; van Wijk and Cloetingh, 2002).

In cases of polyrifted margins (i.e. when successive rifting episodes separated by sag “quiet” periods occur in the same region, like for example along the Atlantic margin), the locus of rifting may change, leading to the formation of extensional basins distributed in time and space along the rift system. It is, for instance, the case of the Atlantic- Tethys triple junction where several episodes of rifting occurred before the accretion of the oceanic crust and the drifting stage (Favre and Stampfli, 1992; Frizon de Lamotte et al., 2011; Fernandez, 2019; Gimeno-Vives et al., 2019). Because of this polyphased history of rifting, the Tethys in the Maghreb region in NW Africa was fringed to the south by a wide strip of intra-continental rim basins forming the Atlas system separated by less-deformed paleo-highs (e.g. Frizon de Lamotte et al., 2000, 2008, and Frizon de Lamotte et al., 2011).

This paper analyzes a series of poorly inverted rim basins in NW Africa (Fig. 1a) in order to establish their tectonic and geodynamic evolution in relation with the Mesozoic development of the Central Atlantic and Maghrebian Tethys, respectively. From southeast to northwest, the rim basins incorporated into the analysis are (Fig. 1b): the Western Sahara Atlas, the Tendrara Basin, the Guercif Basin/Middle Atlas, and finally the Rif Basin in which the former margin of the Maghrebian Tethys was recognized (i.e., Favre, 1992; Gimeno-Vives et al., 2019; Gimeno-Vives et al., 2020a; Michard et al., 2020). Based on geological maps, vintage subsurface data, well data and the available literature, this study aims to evaluate the structural styles and the polyphased rifting history of extensional rim basins in NW Africa at the Central Atlantic/Maghrebian Tethys junction and to describe their evolution in the regional frame of the NW African paleomargin formation. Then, we will discuss the role of inheritance on the nucleation of the successive rift systems and finally factors that control a differently distributed inversion affecting the Mesozoic rim basins during the Alpine cycle.

2. Geological setting: the successive rifting episodes shaping NW Africa

The investigated areas are in the NW corner of Africa (Fig. 1b) where two main E to NE-trending orogenic systems developed by the Late Cretaceous as a response to the convergence of the African and Eurasian plates (Durand-Delga and Fontboté, 1980). The resulting orogenic belts are the Rif-Tell to the north and the Atlas system to the south (Fig. 1b).

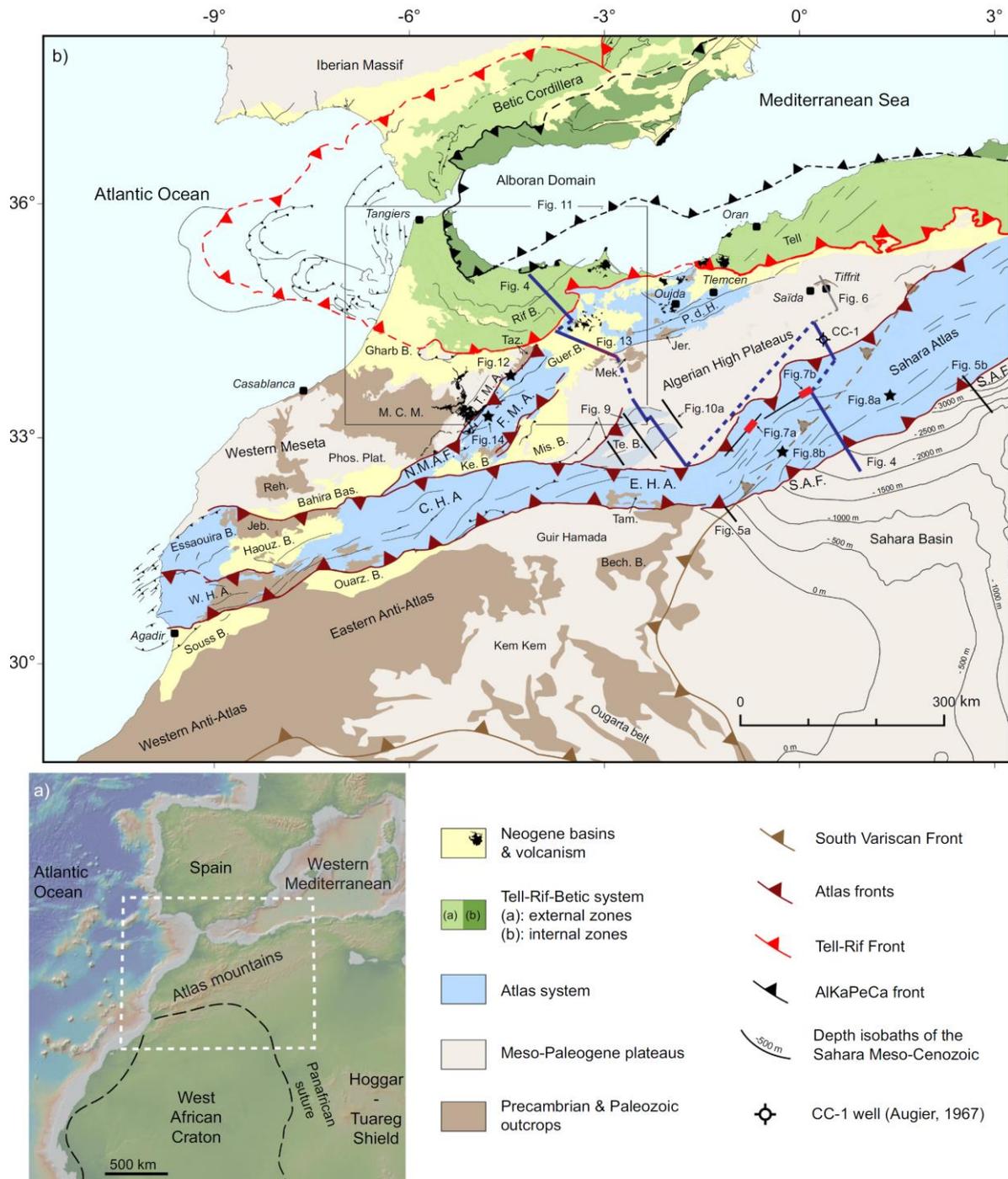
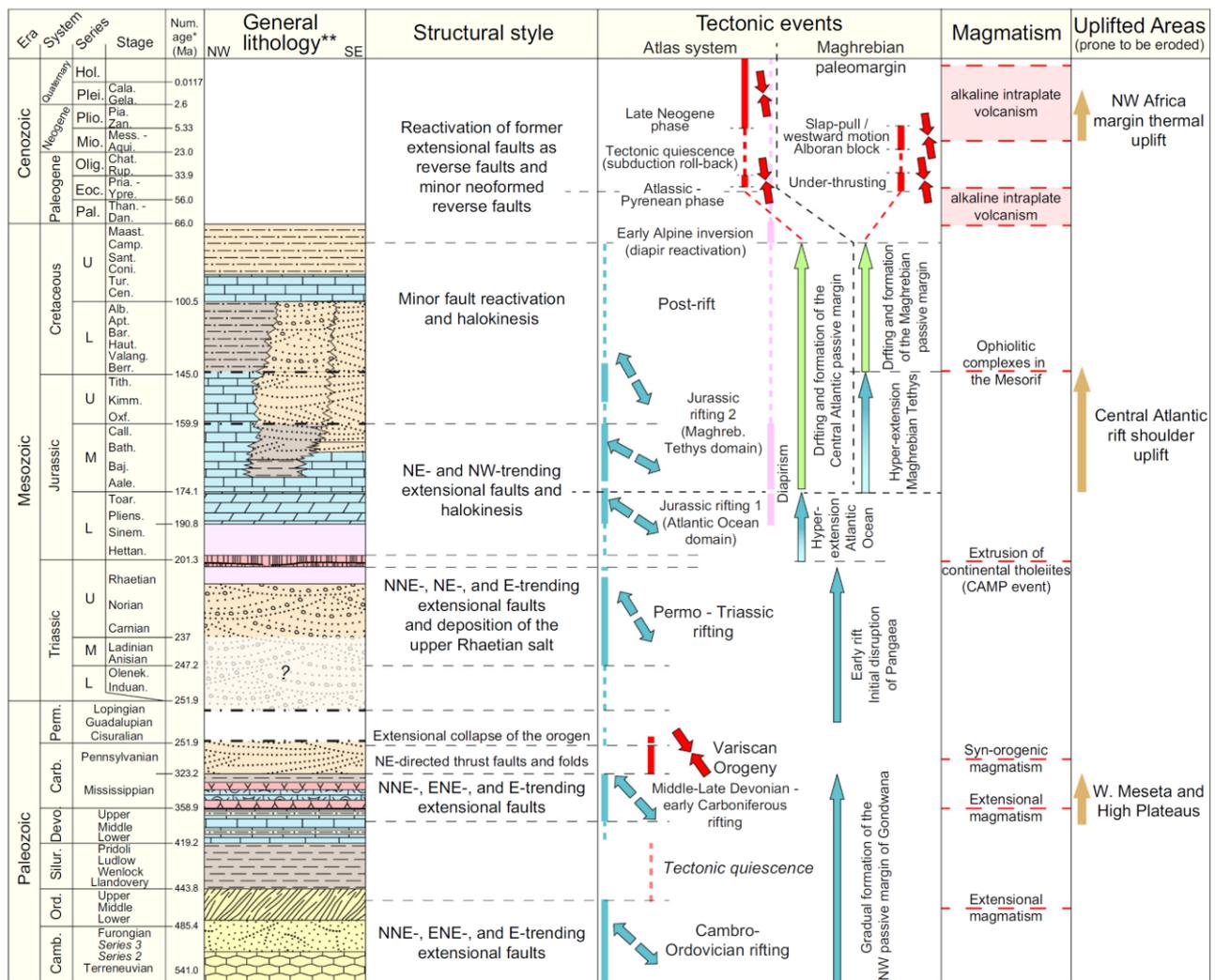


Fig. 1. a) Relief map of the western Mediterranean and NW Africa; b) Structural map of NW Africa depicting the main orogenic systems. Note the location and names of the areas and structures cited along the text. Modified from Frizon de Lamotte et al. (2011). Coordinates are in degrees in World Geodetic System 84. Abbreviations: P. d. H: "Pays des Horsts"; Jer.: Jerada Massif; Taz.: Tazzeka Massif; Mek.: Mekkam Massif; T. M. A.: "tabular" Middle Atlas; F. M. A.: "folded" Middle Atlas; Mis. B.: Missouri Basin; Te. B.: Tendirra Basin; N. M. A. F.: North Middle Atlas Fault; M. C. M.: Moroccan Central Massif; Ke. B.: Kerrouchene Basin; Phos. Plat. Phosphates Plateaus; Reh.: Rehamna Massif; Jeb.: Jebilet Massif; W. H. A.: Western High Altas; C. H. A.: Central High Atlas; E. H. A.: Eastern High Altas; S. A. F.: South Atlas Front; Tam.: Tamlelt Massif; Bech. B.: Bechar Basin.

The Rif in Morocco and Tell in Algeria and Tunisia, formed in response to the closure of the Maghrebian Tethys and the inversion of the associated paleomargins (see review in Leprêtre et al., 2018a). The Atlas system comprises a set of intracontinental doubly-verging fold-and-thrust belts (i.e. High Atlas, Middle Atlas, Sahara Atlas, Aurès and Tunisian Atlas) derived from the inversion of several rim basins associated with the development of the Atlantic and Tethys oceanic domains during the Mesozoic (e.g. Frizon de Lamotte et al., 2000; Frizon de Lamotte et al., 2009; Khomsi et al., 2006; Khomsi et al., 2009; Teixell et al., 2003). South of the High Atlas, the Anti-Atlas and Ougarta constitute older fold-and-thrust belts developed during the Late Paleozoic (Variscan Orogeny) and partially covered by a thin Mesozoic cover (Burkhard et al., 2006; Michard et al., 2008; Fig. 1b). East of the studied area, the Sahara Atlas is bounded to the southeast by the Sahara Basin (Fig. 1b), a flat region consisting of a pile of Paleozoic and Mesozoic sedimentary rocks affected by low magnitude deformation during the Phanerozoic times (Boote et al., 1998; Craig et al., 2008; Frizon de Lamotte et al., 2013).

Rifting in NW Africa developed during successive stages of extension since the Late Paleozoic through Mesozoic times (Fig. 2). The first episode of rifting in the Moroccan Meseta, Atlas and the Anti-Atlas domains was responsible for the development of Paleozoic basins in two phases: a first Cambrian–Ordovician extensional phase accompanied by continental magmatism (Hollard et al., 1982; El Hassani, 1994; Soulimani et al., 2004); and a second one, Middle/Late Devonian to early Carboniferous time interval, responsible for the development of a “platform and basin topography” (sensu Wendt, 1985). This second extensional phase was mainly controlled by NNE-SSW and transverse E-W to ENE-WSW oriented basement faults that contributed to the gradual foundering and disintegration of the NW passive margin of Gondwana (Wendt, 1985; El Kamel and El Hassani, 2006; Wendt et al., 2006; Simancas et al., 2009; Eschard et al., 2010; Frizon de Lamotte et al., 2013). The subsequent late Carboniferous–early Permian Variscan Orogeny developed over this area by the re-activation of the major structural trends inherited from the previous deformation events (Burkhard et al., 2006; Wernert et al., 2016; Delchini et al., 2018). This resulted in the formation of the Moroccan Variscides, comprising the Moroccan Mesetas (i.e., both the Western Meseta and the Algerian High Plateaus on Fig. 1b) and the Anti-Atlas (Hoepffner et al., 2006; Michard et al., 2008; Michard et al., 2010). A third episode of rifting recorded in the region is related to an intermediate phase between the two main contractional stages of the Variscan orogeny during the uppermost Carboniferous–early Permian (Chopin et al., 2014; Delchini et al., 2018). This is possibly linked to partial orogenic collapse and/or globally, by thermal insulation of the Pangea continental aggregation and subsequent large-scale melting localized along preexisting lithospheric weaknesses in different geological contexts (Frizon de Lamotte et al., 2015). Nevertheless, in NW Africa they constituted aborted rift basins scattered over the dismantling orogenic belt (Dercourt et al., 2000).



* Numerical ages according to the International Commission on Stratigraphy (Cohen et al., 2013, updated).
 ** Lithological variations in the NW Africa margin from the Rift Basin (NW) to the western Sahara Atlas (SE).

Color code lithology

- | | | |
|-------------------|-------------------|--------------------------------------|
| Metamorphic rocks | Evaporites | Fluvio-deltaic and eolian sandstones |
| Magmatic rocks | Marine carbonates | Marine sandstones and shales |

Lithology

- | | | | |
|-------------|--|------------------------------|--------------------------|
| Quartzite | Carbonates | Bioclastic carbonates | Tholeiitic flood basalts |
| Metapelites | Sandy shales | Sandstones | Dolomites |
| Marbles | Andesitic and rhyolitic tuffas and ashes | Conglomerates and sandstones | |
| Shales | Evaporites | | |

Fig. 2. Tectonostratigraphic chart of NW Africa depicting a general lithology, the main tectonic events and related structural styles and associated magmatism from the NW Africa margin to the western Sahara Atlas.

The progressive opening of the Central Atlantic Ocean (Stampfli, 2000; Ellouz et al., 2003; Davison, 2005; Leleu et al., 2016), led to the formation of a new phase of rifting affecting the Pangea continent in the Middle to Late Triassic (Fig. 3). During this stage, rifting propagated north-eastward from the Central Atlantic towards the Tethys domain along the northwestern African paleomargin. This rifting phase was controlled by NNE- to NE-trending extensional faults in the Atlantic margin of Morocco (e.g. Essaouira Basin or Western High Atlas), but to the northeast was controlled by NE- to E-trending faults in the High Atlas, Sahara Atlas, Middle Atlas, “Pays des Horsts” and up to the Tunisian Atlas (Fernandez, 2019).

The end of the Triassic rifting is marked over the Atlantic and Tethys margins of Morocco by the flooding of continental tholeiites, dated at 200 ± 1 Ma (Olsen et al., 2003; Knight et al., 2004; Marzoli et al., 2004; Panfili et al., 2019; Gimeno-Vives et al., 2019; Haissen et al., 2021). These basaltic extrusions, forming together the Central Atlantic Magmatic Province (see Fig. 3 for the distribution and limits of the CAMP event in NW Africa), mark the boundary between the Triassic and Jurassic and predate the drifting in the northern central Atlantic segment (Marzoli et al., 1999; Marzoli et al., 2004; Knight et al., 2004; Verati et al., 2007; Nomade et al., 2007). Thereby, spreading of the oceanic crust began during earliest Jurassic in the southern central Atlantic (i.e., the oceanic segment located south of Iberia; Favre and Stampfli, 1992; Stampfli et al., 2001; Sahabi et al., 2004; Stampfli and Borel, 2004; Labails et al., 2010; Kneller et al., 2012) and propagated northeastward to the Tethys realm (Frizon de Lamotte et al., 2011). The Middle to Late Jurassic phase of extension combined with the coeval sedimentation triggered salt migration and the development of diapiric structures made up of Triassic evaporites over the Central High Atlas, Sahara Atlas, Tunisian Atlas, Rif and Tell margins (Saura et al., 2013; Troudi et al., 2017; Teixell et al., 2017; Gimeno-Vives et al., 2019). This process represented the final fragmentation of Pangaea and led ultimately to the opening of the Maghrebian Tethys during the Middle to Upper Jurassic (Favre and Stampfli, 1992; Bracène and Frizon de Lamotte, 2002; Gimeno-Vives et al., 2019).

South of the Maghrebian Tethys, the Atlas system partly bounds the Western Meseta and surrounds the Algerian High Plateaus. Both areas comprise several Paleozoic massifs that remained as poorly subsiding basement highs during the Mesozoic (e.g. in Western Meseta: Ghorbal et al., 2008; Saddiqi et al., 2009). South of the Western Meseta, the Atlas system can be subdivided into an Atlantic domain to the west and southwest (including the Western High Atlas; Fig. 1b) and an eastern/ northeastern domain showing Tethys affinities (i.e. Central High Atlas, Middle Atlas and northeastern extension into the “Pays des Horsts”, Sahara Atlas, Aurès (Algeria) and Tunisian Atlas; Frizon de Lamotte et al., 2008; Fig. 1b). In the Western High Atlas, extension occurred mostly during late Permian–Triassic times, controlled by NNE-trending extensional faults, and thus dominated by the development of the Central

Atlantic Ocean (Hafid et al., 2006). Towards the eastern/northeastern part of the Atlas system instead (i.e. Middle Atlas, Guercif Basin, Sahara Atlas, Aurès Basin and Tunisian Atlas), major extension occurred later. It is controlled by NE-trending extensional faults and is rather associated to the development of the Maghrebien Tethys during Jurassic times. However, at the center of the Atlas system (i.e. Central High Atlas, Tendirra Basin), the developing basins were influenced by both extensional phases due to the progressive northeastwards opening of the Tethys domain.

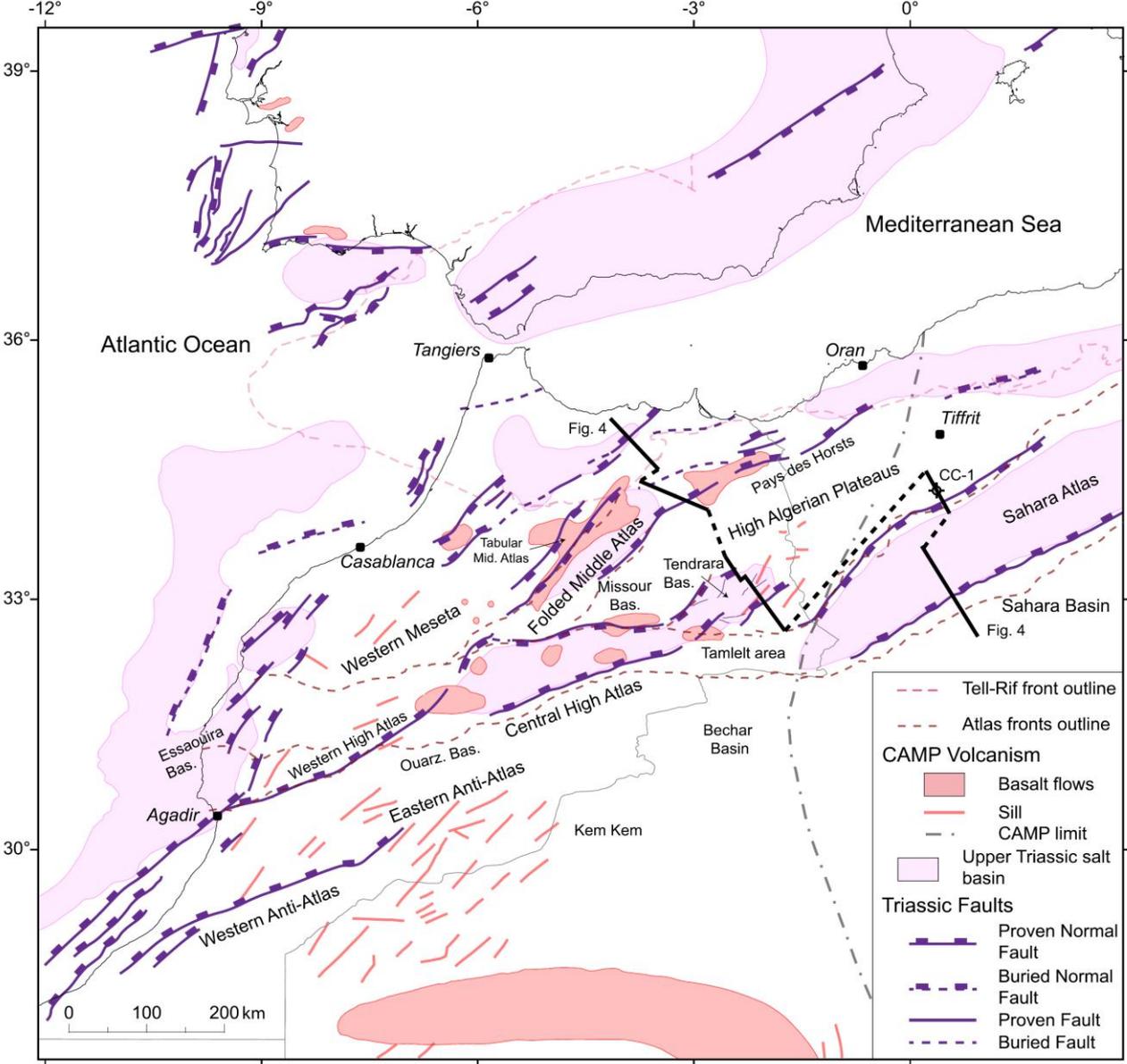


Fig. 3. Upper Triassic tectonic map of northwestern Africa and southern Iberia depicting the locations of the main basins and bounding faults as well as the CAMP volcanism. Note how the CAMP sills are emplaced following the main trends of the extensional faults. Coordinates are in degrees in World Geodetic System 84.

3. General lithostratigraphy of NW Africa

The stratigraphy of the studied basins can be grouped into four main sedimentary groups according to the geodynamic events that affected NW Africa during Paleozoic and Mesozoic times and described in the previous section (Fig. 2).

1. A Paleozoic succession that from bottom to top is composed of: Cambrian–Ordovician marbles, quartzites and metapelites; Silurian black shales; Devonian shallow marine carbonates and sandy shales; early Carboniferous shales and bioclastic marine carbonates interbedded with andesitic and rhyolitic tuffas and ashes; and a late Carboniferous–lowermost Permian assemblage of fluvial and deltaic sediments (Medioni, 1980; Hoepffner, 1981, 1987; Piqué and Michard, 1989). The Cambrian to early Carboniferous succession was deposited along the passive margin of Gondwana (Wendt, 1985; Simancas et al., 2009; Eschard et al., 2010; Frizon de Lamotte et al., 2013). By contrast, the late Carboniferous continental sediments are syn-orogenic and deposited during the inversion of the previous extensional basins during the Variscan Orogeny (Hoepffner et al., 2006; Saber et al., 2007; Graham and Sevastopulo, 2007)

2. An uppermost Permian–Triassic unit lying unconformable over the Paleozoic succession (Saber et al., 2007; Benaouiss et al., 2011). It is represented by: a lower Carnian–Norian coarse-grain fluvial unit (e. g. “TAGI” reservoir in Algeria and eastern Morocco); and an upper Rhaetian finer-grained mudflat to evaporitic unit (e.g. Keuper evaporites; Beauchamp, 1988; Leleu and Hartley, 2010; Ortí et al., 2017). These deposits are syn- to post-rift by reference to the Triassic extensional phase (Fig. 3), which are sealed by the tholeiitic flood basalts of the CAMP (Bertrand, 1991; Youbi et al., 2003; Hafid, 2000). The Triassic evaporitic unit, here and after referred as lower salt, constitutes a regional décollement level activated several times during the subsequent deformation affecting NW Africa.

3. Overlying the previous group, a Lower–Middle Jurassic succession displays major thickness variations over the studied rim basins. It is composed of: a basal evaporitic unit, Hettangian–Sinemurian in age, capped by anhydrites and dolomites (here and after referred as the upper salt unit); a middle dolomitic unit, Pliensbachian–Toarcian in age, and an upper limestone unit Middle Jurassic in age (Du Dresnay, 1987; Zizi, 2002; El Hammichi et al., 2002; Aït Addi, 2006). These are syn-rift deposits by reference to the Lower–Middle Jurassic rifting associated to the development of the Maghrebian Tethys (Favre et al., 1991; Gimeno-Vives et al., 2019; Gimeno-Vives et al., 2020a; Michard et al., 2020). In addition, salt migration triggered by thick-skinned extension was also important on contributing for the thickness variations observed in the upper carbonate unit (Saura et al., 2013; Moragas et al., 2016). It is worth noting that the above described Paleozoic to Norian succession constitutes the subsalt unit in relation to the lower and upper salt, and that the overlying successions constitute the suprasalt

units. So, during the Mesozoic extensional events, the subsalt rocks were mechanically coupled and thus behaved as basement.

Finally, the Upper Jurassic to Lower Cretaceous succession can be attributed to a post-rift phase in relation to the Lower–Middle Jurassic rifting. This period is characterized by a differentiation of the sedimentary environments over NW Africa (Fabre, 2005). There is no deposition on the western sector (Western High Atlas and Moroccan Meseta), clastic sedimentation in the center (Central High Atlas), and carbonate sedimentation in the eastern Atlas system (Vially et al., 1994; Elmi, 1996). By the Upper Cretaceous, the sedimentation becomes homogenous at scale of the whole Northwest African margin (Haddoumi et al., 2008).

4. Structure and evolution description of the Atlas rim basins and the NW-Africa paleomargin

The following subsections are devoted to describe the structure and stratigraphy of the investigated areas in NW Africa from SE to the NW. Fig. 4 shows a SSE-trending composite regional cross-section that illustrates the style of deformation for the Mesozoic basins. The interpretation in depth of the seismic data was constrained by horizon tops from well data. The geological cross-sections were designed perpendicularly to the main structural trends. However, in places where subsurface data were not available, the construction of the cross-sections was complemented using a compilation of geological maps, published scientific works and field observations

4.1. The Sahara Atlas

The Sahara Atlas represents the Algerian eastern continuity of the Eastern High Atlas, with a change of strike from an E-W to a NE-SW orientation (Fig. 1b). The surface geology is known from a set of geological maps (Galmier, 1972a, 1972b) and fieldworks (Bassoulet, 1973; Kazi-Tani, 1986; Mekahli, 1998; Aït Ouali, 1991; Kacemi, 2013). The subsurface data is partly released within few studies but shows a poorly dense network over the area (Kazi-Tani, 1986; Vially et al., 1994; Bracène et al., 1998; Bracène, 2002; Bracène and Frizon de Lamotte, 2002). Only few exploratory wells constraint the geology along the northern and the southern fronts of the Sahara Atlas, where they have generally not reached the basement, nor the Triassic series (see ref. above). Overall, based on the interpretation of the available data, previous authors agree with the control of the sedimentation within NE-SW oriented grabens delimited by thick-skinned extensional faults (Vially et al., 1994; Yelles-Chaouche et al., 2001; Bracène and Frizon de Lamotte, 2002). In terms of structural segmentation, the Sahara Atlas can be sub-divided into the Pre-Atlas to the north and the Atlas Trough to the south (Fig. 4). The latter represents the main depocenter separated from the Pre-Atlas by a structural high developed during

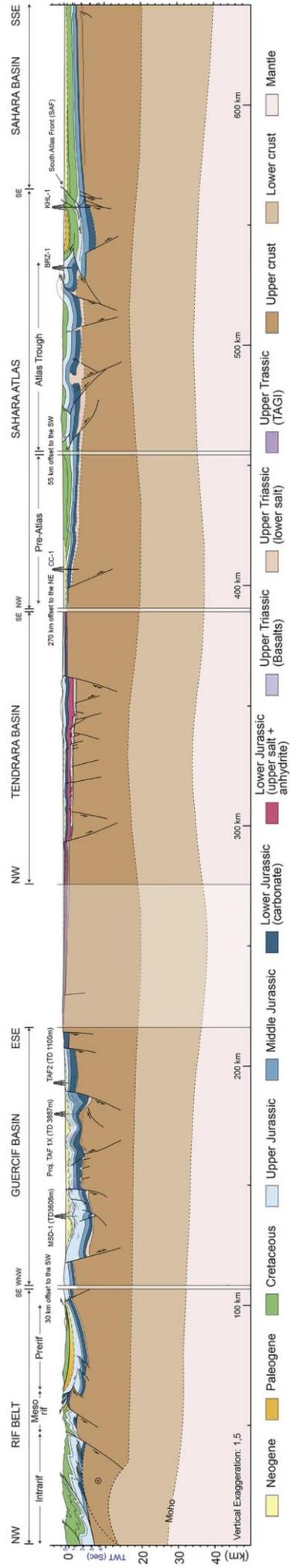


Fig. 4. NW-trending crustal-scale cross-section depicting the main structural features of the studied area in NW Africa (vertical exaggeration of 1,5). See its location in Figs. 1b and 3. The deep crustal structure is extrapolated from geophysical studies conducted by Schwartz et al. (1992), Wigger (1992), Rimi (1999) and Ellouz et al. (2003).

the Early Jurassic (Yelles-Chaouche et al., 2001). In continuity with the Moroccan High Atlas, the Sahara Atlas also constitutes an inverted Mesozoic basin during the Cenozoic (Frizon de Lamotte et al., 2000; Bracène and Frizon de Lamotte, 2002 and references therein).

4.1.1. Paleozoic inheritance

No data are available concerning the Paleozoic basement of the Mesozoic Atlas Trough. However, some indirect information regarding the Paleozoic basement can be derived from the southern fringing Sahara Basin and from the Algerian High Plateaus that lie north of the Pre-Atlas (Fig. 1b). The Sahara Basin is deformed along the so-called South Atlas Front (SAF; Figs. 1b and 4). South of the SAF, along the southern part of the Fig. 4 cross-section, the Paleozoic strata are well-imaged in the seismic data showing a simple “layer cake” geometry with sub-parallel reflectors and low-angle unconformities (Fig. 5). North of the Atlas Trough and below the Triassic series, the poor resolution of the seismic data does not allow to properly interpret the Paleozoic structure (Vially et al., 1994). Nevertheless, indirect evidences can be extracted from well data (e.g. CC-1 well) and the outcrops of the Tifrit- Saïda area (Fig. 6). Here, the Paleozoic strata are deformed by thrust faults developed during the Variscan Orogeny and intruded by syn-orogenic magmatic bodies (circa. 298 Ma, in Remaci-Benaouda, 2005). In between, below the Pre-Atlas and the Atlas through, the Variscan frontal thrust might be located (Fig. 1b) as already suggested in Frizon de Lamotte et al., 2011 (their Fig. 10). Consequently, the Western Sahara Atlas likely samples the Variscan Front re-worked during the subsequent Mesozoic rifting events, and thus localizing the ensuing deformations.

4.1.2. Mesozoic rifting

Seismic imaging and drilling records point out that the Mesozoic rifting started during the Late Triassic (Yelles-Chaouche et al., 2001). Extension was mainly controlled by NE-SW to ENE-WSW oriented thick-skinned extensional faults with maximum throws located along the northern part of the Sahara Atlas (Bassoulet, 1973; Kazi-Tani, 1986; Yelles-Chaouche et al., 2001; Bracène, 2002; Bracène and Frizon de Lamotte, 2002). The major thick-skinned faults were relayed by WNW-trending transverse faults compartmentalizing the area in different sub-basins. These narrow and fault bounded basins were infilled by siliciclastic sediments (Triassic in age) overlain by Late Triassic evaporites (Yelles-Chaouche et al., 2001). On top of the evaporites, a Rhaetian–Hettangian dolomitic level (Bassoulet, 1973) capped the Late Triassic (e.g. see EBD-1 well in Bracène, 2002).

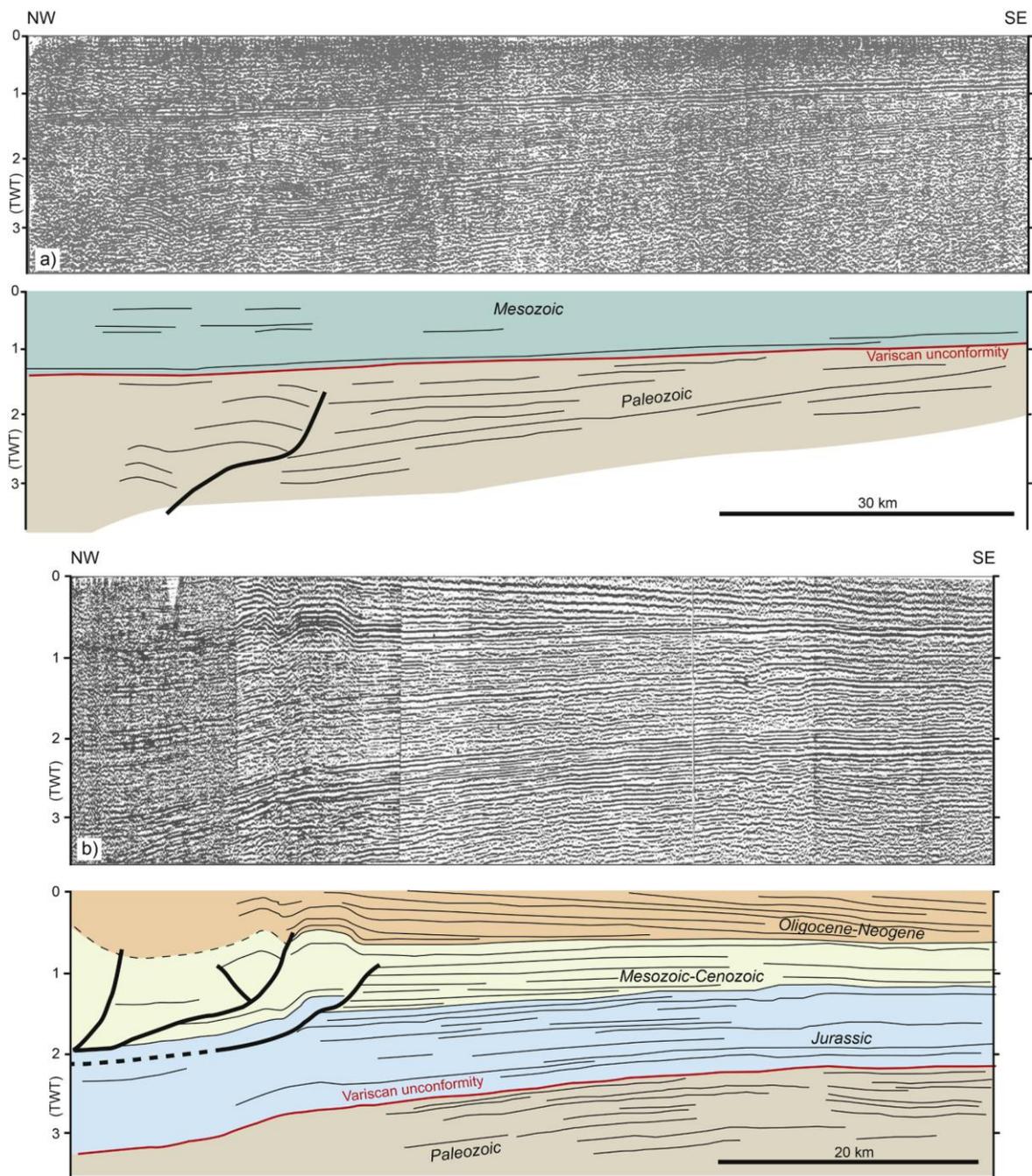


Fig. 5. Interpreted seismic reflection profiles a) and b), located in the Saharan Atlas (see Fig. 1b). Modified from Bracène (2002) and Bracène and Frizon de Lamotte (2002).

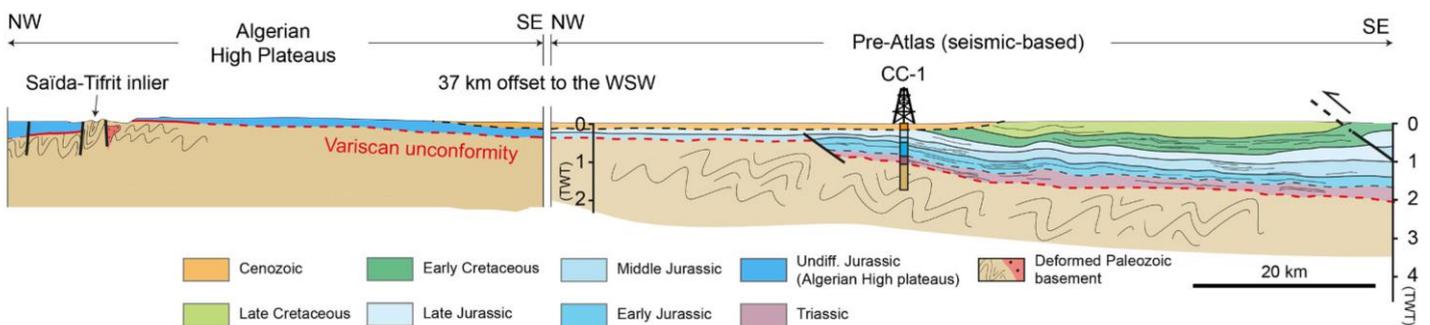


Fig. 6. Seismic-based geological section across the CC-1 well and the Saïda/Tifrit area linking the Pre-Atlas and the High Algerian Plateaus (see location in Fig. 1b).

Along the southern frontal part of the fold-and-thrust belt, the Triassic evaporites crop out within diapirs piercing through the overburden in the core of anticlines (e.g. Azil and Aït Ouali, 2017). Also, subsurface data from the northern frontal part of the belt show diapir growth related structures (Fig. 7a and b). In these diapiric outcrops, there are pieces of Rhaetian–Hettangian dolomites and CAMP basalts acting as stringers folded and embedded with varicolored clays and evaporites (Chabou et al., 2010; Meddah et al., 2007; Meddah et al., 2017). Therefore, the Triassic evaporites were mobilized during the subsequent rifting phase, a process that makes difficult the reconstruction of their original thickness and the exact timing of the Triassic extension.

A second phase of rifting promoted a rapid deepening of the basin and a change in the sedimentation to open marine with the deposition of the Toarcian–early Bathonian limestones on top of the shallow-water Rhaetian–Hettangian dolomites (Aït Ouali, 1991; Mekahli, 1998; Elmi et al., 1998; Yelles-Chaouche et al., 2001). A clear change in the sedimentation from open-marine conditions to a shallow-marine environment occurred at the end of the Bathonian, and the generated space was fed from the west by an important clastic sedimentary system (“grès des Ksour Fm.”; Bassoulet, 1973; Delfaud, 1974 and Delfaud, 1986; Aït Ouali, 1991; Elmi, 1996). This one derived from the erosion of the uplifting footwall blocks of the newly formed Central Atlantic rift shoulder (Leprêtre et al., 2015).

The structural style of the Toarcian–Bajocian/Bathonian rifting phase has been described in few studies (Vially et al., 1994; Yelles-Chaouche et al., 2001). The authors depicted a general deepening of the basin towards the south and the development of a set of NE-SW oriented extensional faults (Vially et al., 1994; Yelles-Chaouche et al., 2001). In the northern part of the basin within the less-subsiding Pre-Atlas, the basement extensional faults offset the Jurassic strata (Fig. 6). On the other hand, towards the Atlas Trough (Fig. 4), there is a structural decoupling between the overburden, affected by suprasalt faults soled into the Triassic salt, and the underlying subsalt strata which are deformed by thick-skinned extensional faults. The movement along these basement faults and the subsequent salt mobilization controlled the variations of the thickness of the Early and Middle Jurassic strata in a similar way than in the Central High Atlas (Saura et al., 2013).

By Bathonian onwards, subsidence was still important with the development of deltaic systems up to the Albian (Bassoulet, 1973; Delfaud, 1986). However, no rifting events have been identified after the Middle Jurassic, but halokinetic-related deformation is likely, even though they are not properly described in the Western Sahara Atlas. Only good evidences of salt mobilization are observed in the Eastern Sahara Atlas (Soua et al., 2009) and the Tunisian Atlas (Zouaghi et al., 2005; Jaillard et al., 2017). Nevertheless, Fig. 8 shows map extracts of commented satellite images of the Western Sahara Atlas where evident halokinetic-related structures can be recognized. They demonstrate the

importance of salt mobilization, although it is not possible to link them either to active tectonics or passive movement due to halokinesis.

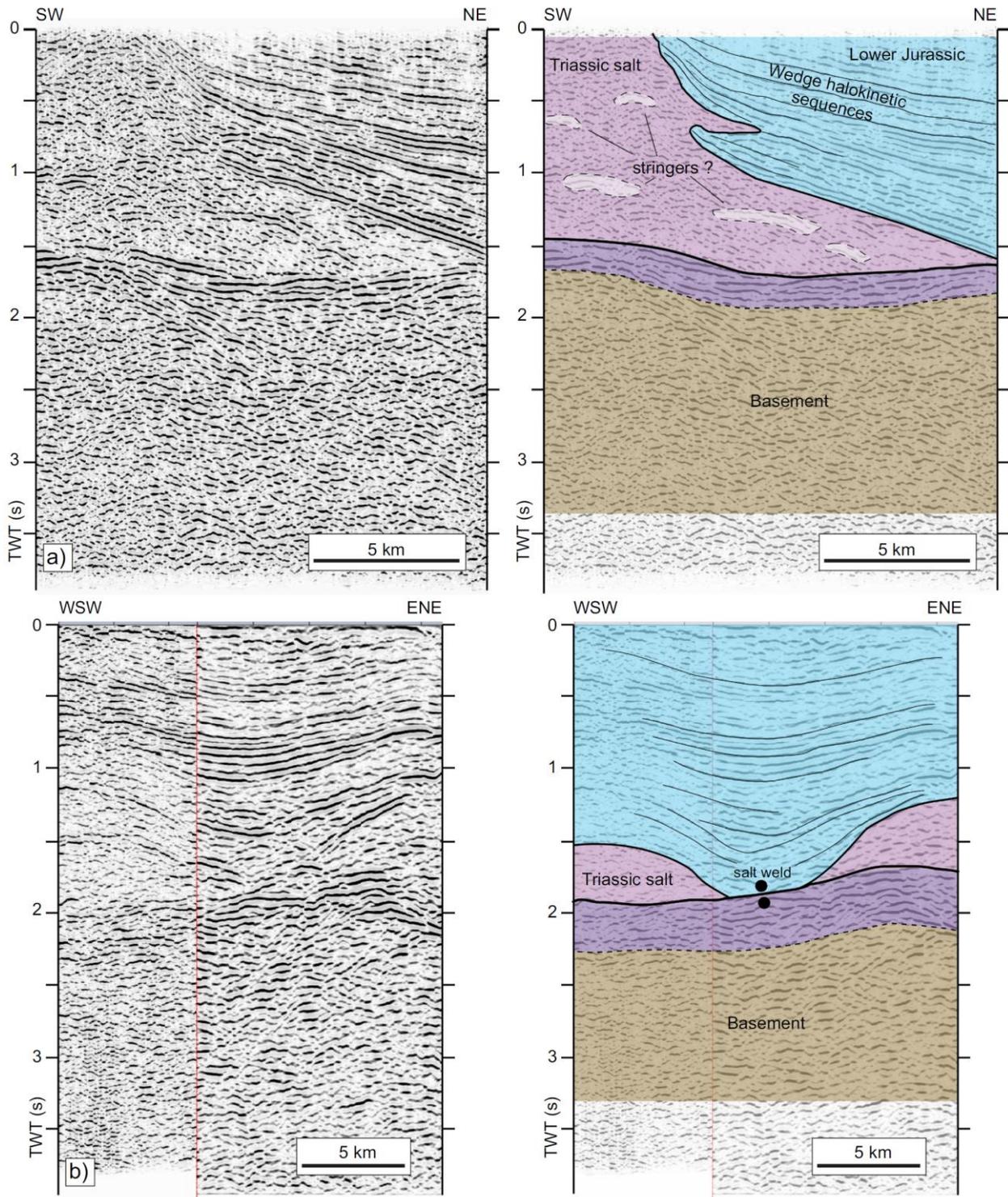


Fig. 7. Interpreted seismic reflection profiles a) and b) across several diapiric structures in the Western Sahara Atlas. See their location in Fig. 1b. Seismic profile in a) modified from Bracène et al. (2003) and in b) modified from Bracène (2002).

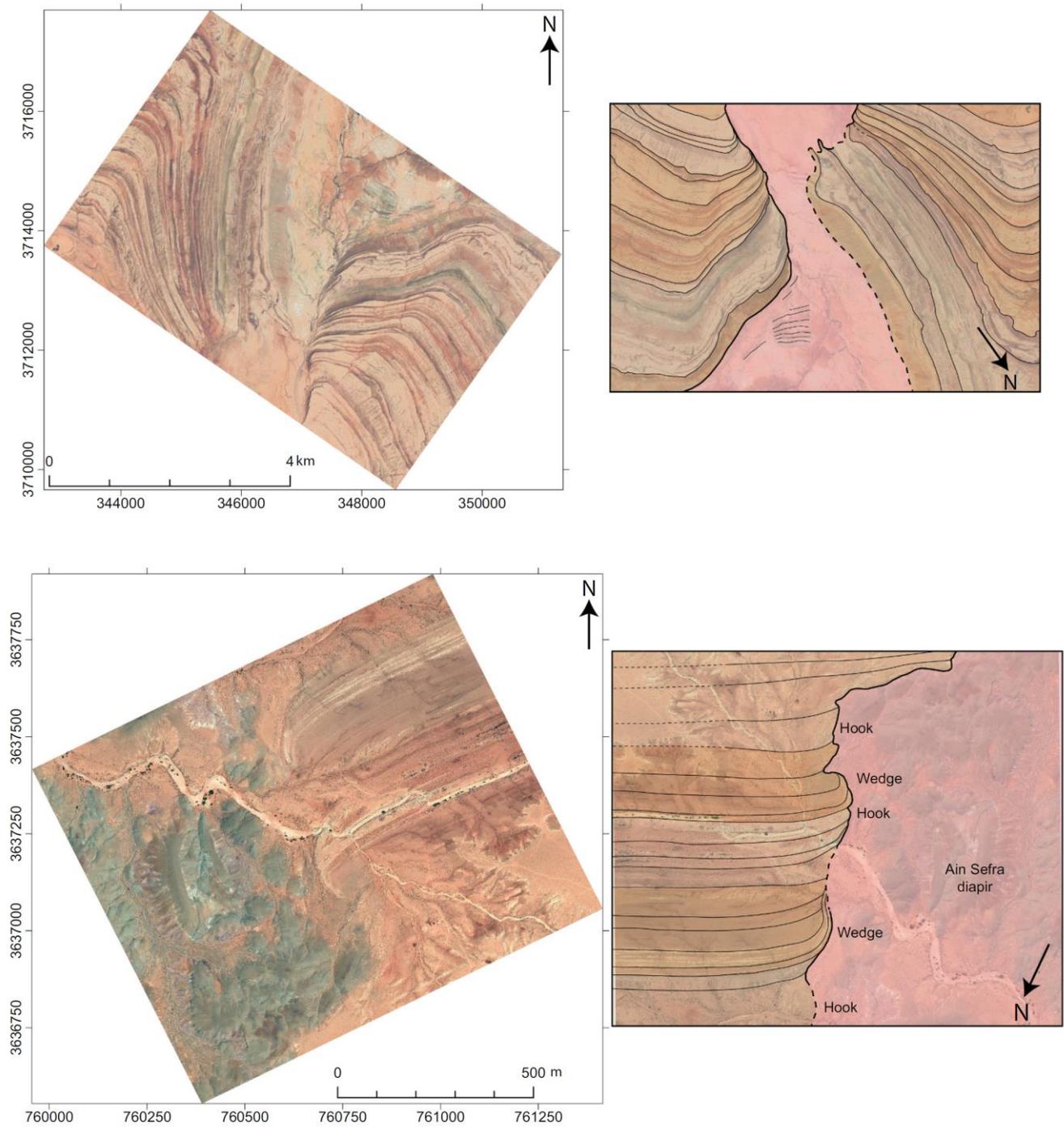


Fig. 8. Interpreted satellite images of diapir-related structures in the central and western Sahara Atlas. See their location in Fig. 1b. Satellite images from the Google Earth repository. Coordinates are in meters in Universal Transverse Mercator, zone 30 northern hemisphere and datum Nord Sahara 1959.

4.2. The Tadrara Basin

The Tadrara Basin is located northwest of the Sahara Atlas, north of the Eastern High Atlas, and southwest of the Algerian High Plateaus (Figs. 1b and 4). Surface and subsurface data show that the Tadrara Basin corresponds to a NE-SW Mesozoic half-graben (Fig. 3), bounded to the south by a set of two main NE-trending basement extensional faults dipping to the NW (Fig. 9a and b). In general, the basin narrows and deepens to the SW, where it connects the Central High Atlas, and shallows to the NE (Fig. 10a), in the transition to the Algerian High Plateaus (Fig. 1b). In contrast to the other studied rim basins, the Tadrara Basin has not been completely incorporated into the Atlas fold-and-thrust belt, so it remains poorly inverted and widely displays the heritage of the extensional architecture.

4.2.1. Paleozoic inheritance

The Paleozoic structure below the Tadrara Basin can be inferred in the central and northeastern part of the basin, where a tabular and thin salt and suprasalt succession (less than 0,75 TWTs thick) allows for a better resolution in depth of the seismic reflection data (Fig. 10a). Even if only few subsurface data are available (Taki, 2012), the deep structure of the basin can be constrained thanks to a series of geological studies conducted in adjacent areas of the Algerian High Plateaus and Eastern High Atlas (e.g. Hoepffner et al., 2005; Hoepffner et al., 2006).

The Paleozoic stratigraphy for the Tadrara Basin can be partially obtained from several wells distributed between the central and northeastern part of the basin (Fig. 9c). The JK-1, located in the NE part of the basin, drilled a > 2300 m thick sequence of lavas, tuff and ashes with intercalations of marine sandstones, siltstones and mudstones Devonian–early Carboniferous in age (Taki, 2012 and references therein), based upon palynological data acquired on the interfingered marine sediments. The TE-1, TE-2 and TE-3 wells, in the central part of the basin (Fig. 9c), encountered 108 m, 103 m, and 1403 m respectively of Carboniferous black silts, shales and carbonates (Taki, 2012). Therefore, two Paleozoic units are distinguished: a Devonian–early Carboniferous volcano-sedimentary unit overlain by a younger Carboniferous siliciclastic unit.

The seismic reflection data also show two main discontinuities in the Paleozoic units (Fig. 10a): an upper horizontal level located at the base of the lower salt unit (see continuous red line at 0,5–0,75 TWTs depth); and a lower SE-dipping level located in an unknown position inside the Paleozoic succession (see dashed red line at 2 TWTs depth). Below the upper detachment, seismic reflection data depicts a series of horizontal reflectors that are onlapping an underlying SE-dipping seismic panel. To the SE, the lower panel is folded and display a growth strata geometry.

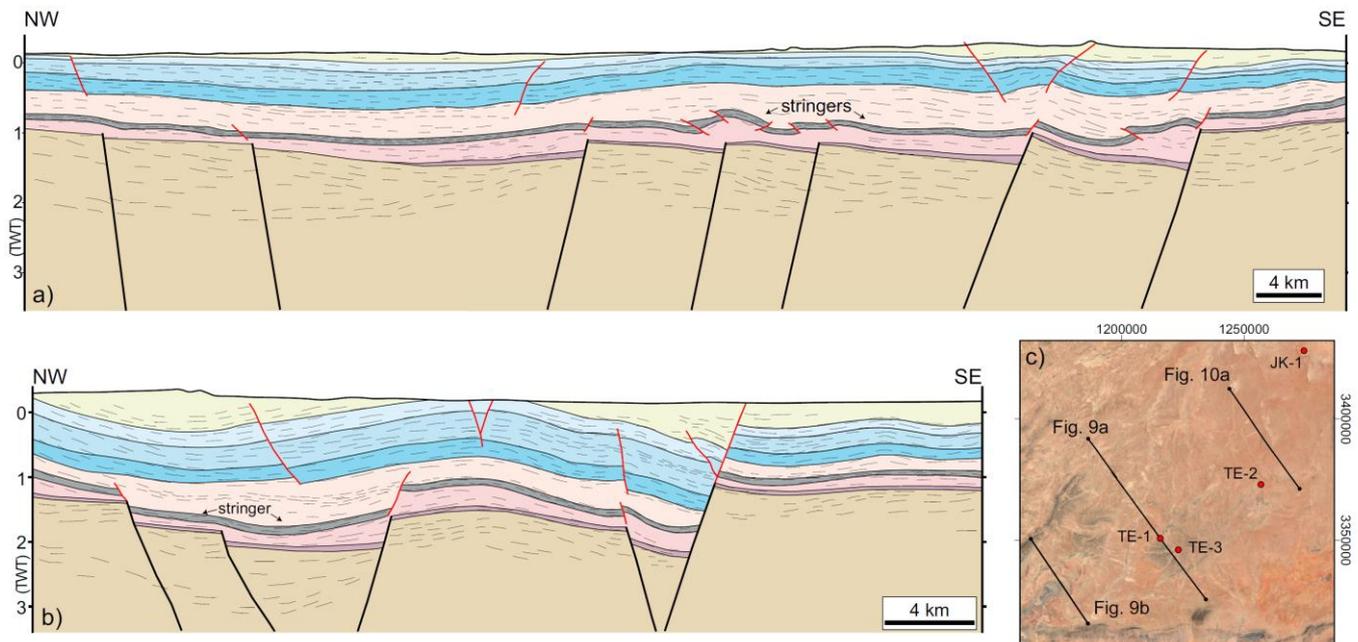


Fig. 9. Interpreted seismic reflection profiles a) and b), depicting the main structural features of the Tendirra Basin. Modified after Taki (2012). See their location in Figs. 1b. Satellite image in c) from the Google Earth repository. Coordinates are in meters in Universal Transverse Mercator, zone 30 northern hemisphere and datum WGS 1984.

Considering these sedimentary and structural features, the SE-dipping panel with growth strata geometries might be interpreted as deposited in half-grabens formed during the Devonian to early Carboniferous extension and controlled by NW-dipping normal faults (Fig. 10b). This is in agreement with geological observations over the Algerian High Plateaus and at the scale of whole Maghreb, where the Middle/Late Devonian to early Carboniferous successions were deposited in an extensional regime during the formation of the NW passive margin of Gondwana (Wendt, 1985; Eschard et al., 2010; Frizon de Lamotte et al., 2013). The sub-horizontal and onlapping unit instead, may be interpreted as deposited synchronously to the uplift of the underlying half-grabens (i.e. syn-contractinal unit during the Variscan Orogeny?); or alternatively, deposited during a later extensional pulse. Considering the first hypothesis, the late Carboniferous Variscan Orogeny that affected the Algerian High Plateaus could promote the inversion of the pre-existing half-grabens and the deposition of the late Carboniferous siliciclastic successions progressively onlapping the uplifted basins (see how depocenters are laterally shifted during inversion; Fig. 10b). In this scenario, shortening reactivated the previous extensional faults as reverse faults but also used the intra-Paleozoic weak rocks (e.g. the Silurian shales, the volcanic ashes or the marine marls) as décollement levels. As a result, an asymmetric pop-up structure could be developed involving the mildly inverted Middle/Late Devonian

to early Carboniferous basins progressively covered by the late Carboniferous syn-orogenic sediments (Fig. 10b).

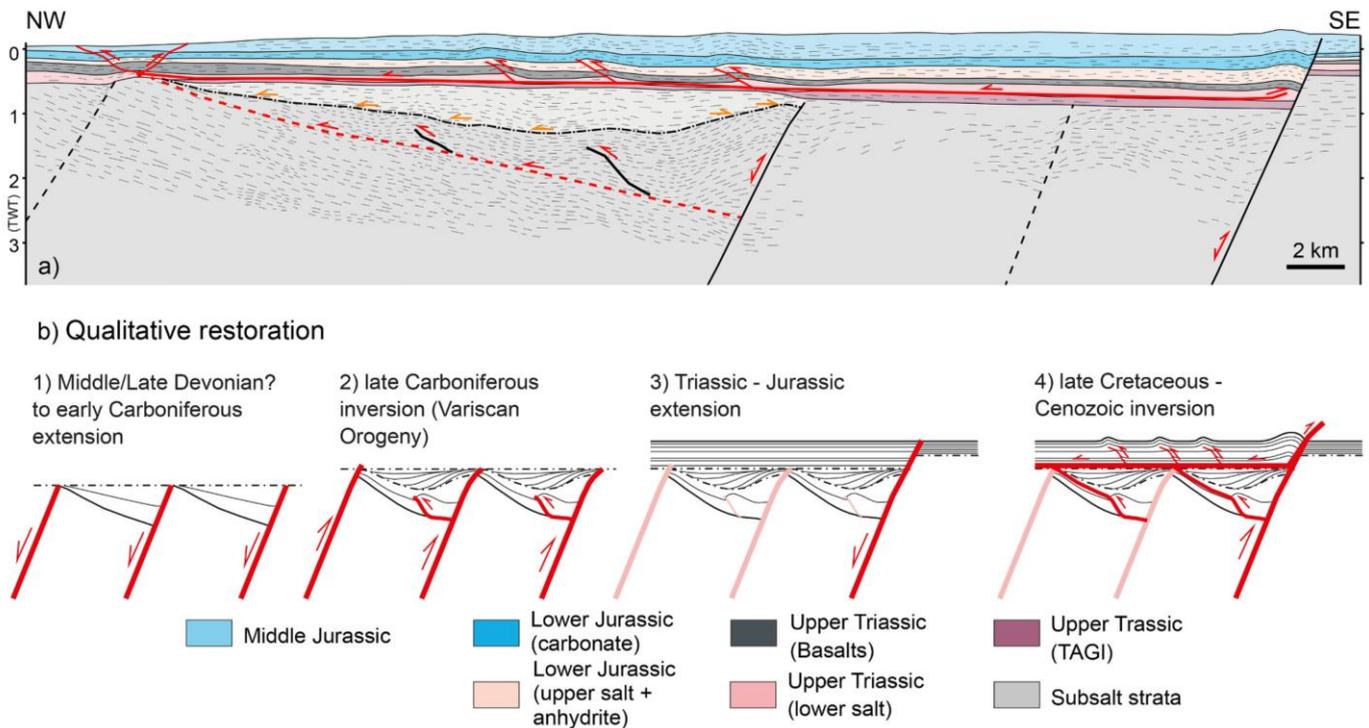


Fig. 10. Interpretation of a seismic reflection profile a) located in the northern part of the Tendirra Basin illustrating the geometry of the Paleozoic basement. Modified after Taki (2012). See its location in Figs. 9c. b) Qualitative sequential restoration depicting the Paleozoic, Mesozoic, and Cenozoic tectonic evolution of the basin. Note the implications of the Paleozoic structural inheritance for the subsequent Mesozoic extension.

4.2.2. Mesozoic rifting

The initial Paleozoic configuration constituted a structural inheritance for the subsequent deformations affecting the Tendirra Basin. The Triassic and Jurassic rifting events reactivated some of the preexisting Paleozoic extensional faults allowing the deposition of thick Triassic and Jurassic sequences showing growth strata (Fig. 9b).

The subsalt basement across the basin is deformed by minor thick-skinned conjugate extensional faults displaying two main orientations: NE-SW and NNE-SSW (Figs. 3 and 9). Because of this structural framework, the basin is compartmentalized in several sub-basins deepening in general towards the SW (compare Fig. 9a and b). The first syn-extensional deposits, the lower Carnian–Norian fine-grain to conglomeratic sandstones, are lying unconformable above the Paleozoic succession. The lower and upper salt units (upper Rhaetian and Hettangian–Sinemurian in age, respectively) show thickness

variations over the faulted blocks: they are thicker in the hanging wall of the bounding thick-skinned faults and thinner in their footwalls. The total salt thickness increases towards the southeast and thins-out towards the NE part of the basin (Fig. 9a, b, and 10a). The maturity of the developed salt structures also varies along the basin. Tabular salt is present in the northern part of the basin (Fig. 10a), long-wavelength salt-cored anticlines are common in the central part (Fig. 9a) and reactive diapirs represent the main building blocks of the basin towards the southwest (Fig. 9b). Additionally, in between both salt units, there is a continuous and stiff intra-salt layer constituted by the tholeiitic CAMP basalts acting as stringers. The deformation of this stiff layer seems to be constrained at some degree by the capacity of salt to move.

The suprasalt (i.e. Pliensbachian to Cretaceous rocks) thickness variations over the Tendirra Basin are controlled by salt migration coeval to extensional deformation characterized by thin-skinned faulting and drape folding (Fig. 9a and b). Above the upper salt unit, the Pliensbachian–Toarcian and Middle Jurassic carbonates are also thicker in the hanging walls of the bounding subsalt faults and thinner in the footwalls (Fig. 9a and b). Finally, the Upper Jurassic to Cenozoic succession does not depict major thickness variations and represents a post-rift stage in the tectonic evolution of the basin. However, in the southern part of the basin, seismic reflection data shows a minor depocenter that could have developed during the Upper Jurassic up to Cretaceous? (Fig. 9b).

4.3. The Middle Atlas and the Guercif Basin

The Middle Atlas represents a Cenozoic fold and thrust belt separating the so-called Western Meseta and the Algerian High Plateaus (Fig. 1b). With a NE-SW orientation, the Middle Atlas branches southward to the Central High Atlas and finishes northward sealed below the Neogene Guercif Basin. This mountain belt can be subdivided into two areas (Fig. 11): a formerly low-subsiding basin to the west, namely the “tabular” Middle Atlas; and the “folded” Middle Atlas to the east (Fig. 12), representing a more subsiding basin during the Mesozoic (Du Dresnay, 1988; Fedan, 1988; Charrière, 1990). This NE-trending branch of the Atlas system has its own rift timing that is somehow shared with its northern extension, presently buried below the Guercif Basin (Fig. 13). The deepest Neogene series in the Guercif Basin are as old as Tortonian up to the Pliocene-Quaternary (Colletta, 1977; Bernini et al., 1992; Zizi, 2002 and references therein).

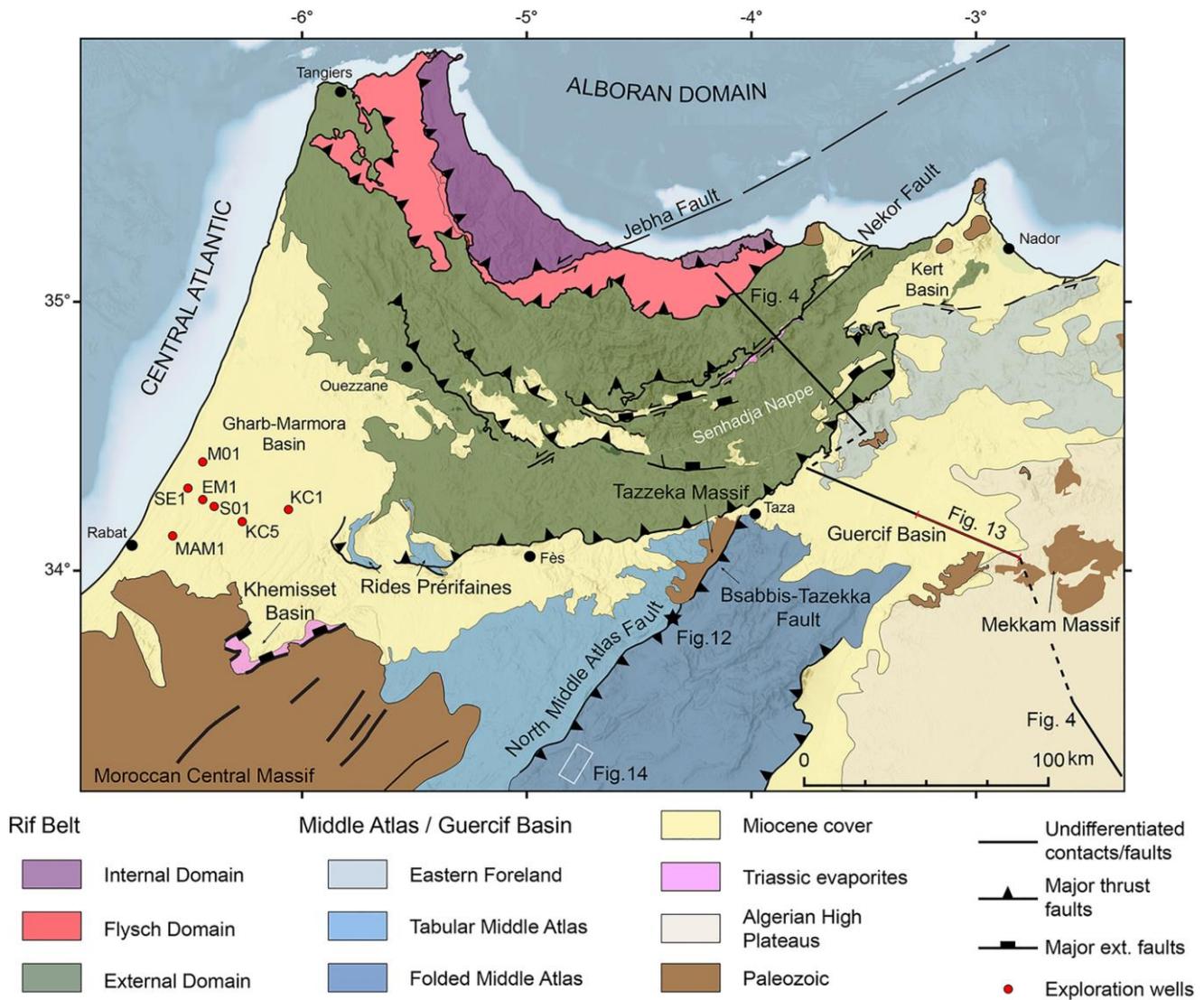


Fig. 11. Structural map of the Rif belt and the Middle Atlas / Guercif Basin. Modified from Gimeno-Vives et al. (2020b). Coordinates are in degrees in World Geodetic System 84.

4.3.1. Paleozoic inheritance

The main Paleozoic outcrop along the Middle Atlas and Guercif areas is the Tazekka massif, forming the NE limit of the tabular Middle Atlas (Fig. 1b and 11). This massif is characterized by folded and metamorphosed, mainly Ordovician in age, schists (Morin, 1973; Rauscher et al., 1982; Hoepffner, 1978, 1987; Charrière, 1990). Locally in the southwest, sediments can be as young as Devonian in age. In terms of deformation, the Tazekka massif shows deep Paleozoic levels with low-grade metamorphism (Hoepffner, 1978). The pre- and syn-orogenic Variscan deformation is guided in-between important NE-SW major faults (e.g. the Bsabbis-Tazekka fault; Fig. 11) suggested to act as major transform faults, either accommodating transpression or transtension during the different deformation events (Hoepffner, 1978; Charrière, 1990). These faults were reactivated during the

Mesozoic rifting, since (1) they control the significant thickness increase when crossing the boundary from the “tabular” to the “folded” Middle Atlas domains (Charrière, 1990; Fig. 12); and (2) the main depocenters of the “folded” Middle Atlas are separated from each other by NE-SW ridges (Fedan, 1988).

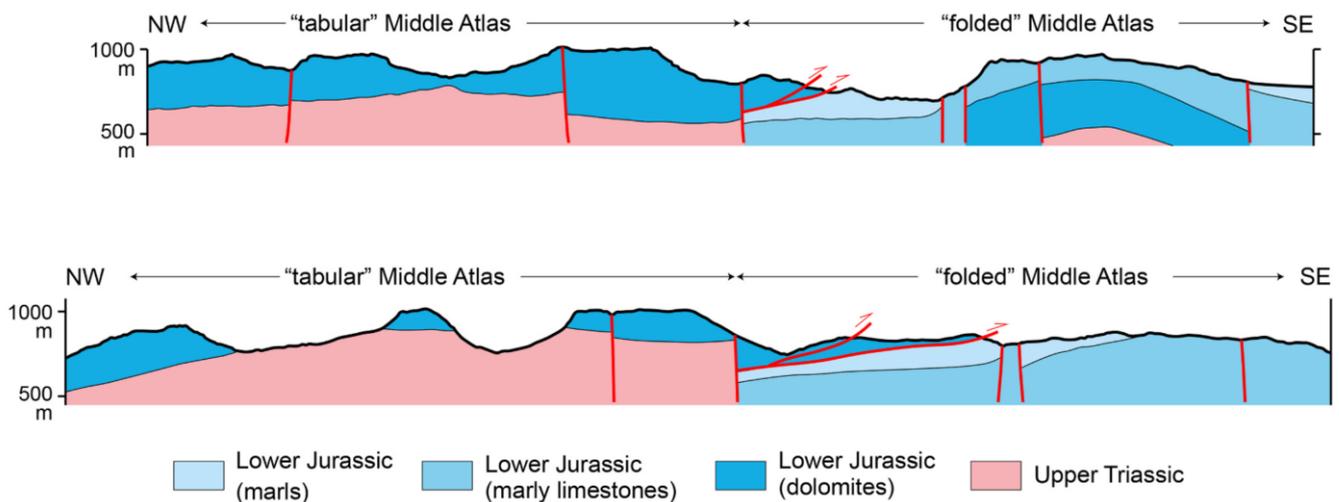


Fig. 12. Geological cross-sections depicting the limit between the folded and tabular Middle Atlas (modified from Charrière, 1990). See their location in Fig. 11.

4.3.2. Mesozoic rifting

The first event of Mesozoic rifting occurred during the Late Triassic and lasted up to the lowermost Early Jurassic (Salvan, 1984; Baudelot et al., 1990; Sabaoui, 1987; Lorenz, 1988; Oujidi et al., 2000; Lachkar et al., 2000; Courel et al., 2003). This extensional phase sets relatively small basins filled with siliciclastic sediments and evaporites -which were subsequently intruded by the basalts of the CAMP. From a structural point of view, a net geographical differentiation of the basin is often stressed in the literature (e.g. Choubert and Faure-Muret, 1960–1962; Oujidi et al., 2000): a clastic basin lies to the SW (i.e. the Kerrouchene Basin, Fig. 1b; Lorenz, 1976; Laville et al., 1995), whereas a salt-floored basin existed in the NE, below the present-day location of the Guercif Basin (Fig. 11), as evidenced by the mobilization of the Triassic salt at the cores of the anticlines deforming the whole stratigraphic pile (Bernini et al., 2000 and references therein). In between, there are relative structural highs without significant subsidence, recording thin clay-sandy sedimentation with some evaporitic levels interbedded with basalts (e.g. Charrière, 1990 for the “tabular” Middle Atlas). Extensional deformation resulted from the activity of NE-trending thick-skinned extensional faults, with N70–90° conjugate faults. For example, more than 1 km of throw has been considered for the major thick-skinned fault bounding the Kerrouchene Basin (Lorenz, 1976; Laville et al., 1995), whereas minor throws occurred on a large fault array inside the basin. The two structural orientations, active during Triassic times, are

important regarding the subsequent evolution of the basin. They separated different blocks along the N70–90° direction that efficiently provided the persistence of a structural high between the Middle and Central High Atlas up to the Middle Jurassic (Choubert and Faure-Muret, 1960–1962; El Arabi et al., 2003).

After the first rifting event, a stable carbonate platform installed on top of the previous lowermost Early Jurassic sediments (Choubert and Faure-Muret, 1960–62; du Dresnay, 1988; Fedan, 1988). A second rifting phase occurred during the Pliensbachian to the early Toarcian times in both “tabular” and “folded” Middle Atlas (Fedan, 1988; Charrière, 1990) but is also seen at the Mekkam/Guercif transition on seismic lines (Zizi, 2002; Fig.13). Thick-skinned extensional faults controlled the sedimentation above tilted fault blocks with a general NE-deepening trend in the “folded” Middle Atlas (e.g. the Aïn Nokra-Skoura-Zloul succession of tilted blocks, lying on the eastern side of the North Middle Atlas Fault (NMAF in Fig. 1b). However, west of the NMAF, the Guigou-Anocer block was poorly subsiding during the same period acting as a relative structural high (Charrière, 1990; El Arabi et al., 1999 and 2001). Overall, the field evidences in the “tabular” and western “folded” Middle Atlas, suggest that the progressive deepening of the sedimentary depocenters was accomplished through SE- and SW-dipping directions of the main thick-skinned faults (Charrière, 1990). Strikingly, the interpretation of the seismic lines and the well correlations in the Guercif Basin, showing the Middle Atlas continuation below the Neogene series (Zizi, 1996; Sani et al., 2000), denotes that extensional faulting mainly occurred within the same Toarcian/Aelian to Bajocian/Bathonian(?) time interval (Fig. 2). It thus advocates for an along-strike consistent record of these Early to Middle Jurassic extensional events.

Accordingly, the fault network controlling the sedimentation is divided into two main directions: (1) NE-SW oriented faults acting as the main separators between the principal depocenters, but also responsible in a minor way for the progressive staircase deepening towards the depocenters; and (2) transverse E-W to NW-SE faults that also promoted the development of a “staircase” of tilted fault blocks towards the main depocenters. More information about the deformations responsible for the onset of Middle Atlas rift on its eastern boundary are available from seismic lines within the Guercif Basin (Zizi, 1996; Sani et al., 2000). Here, a minimum of 2 TWTs of throw is recorded between the Mekkam high to the east and the subsiding Middle Atlas to the west along a single major WNW-dipping fault (Fig. 13). Considering that the Cenozoic inversion minimized the amount of extensional motion along the fault plane, the transition towards the basin really was abrupt along this steep and WNW-dipping major fault. Salt-related structures are not evidenced during this phase of rifting. Only seismic imaging suggests that minor halokinesis might have played a secondary role in early structuring of

some areas in the Guercif Basin (e.g. Sani et al., 2000; their Fig. 3 and some unexplained thicknesses reductions without obvious faulting).

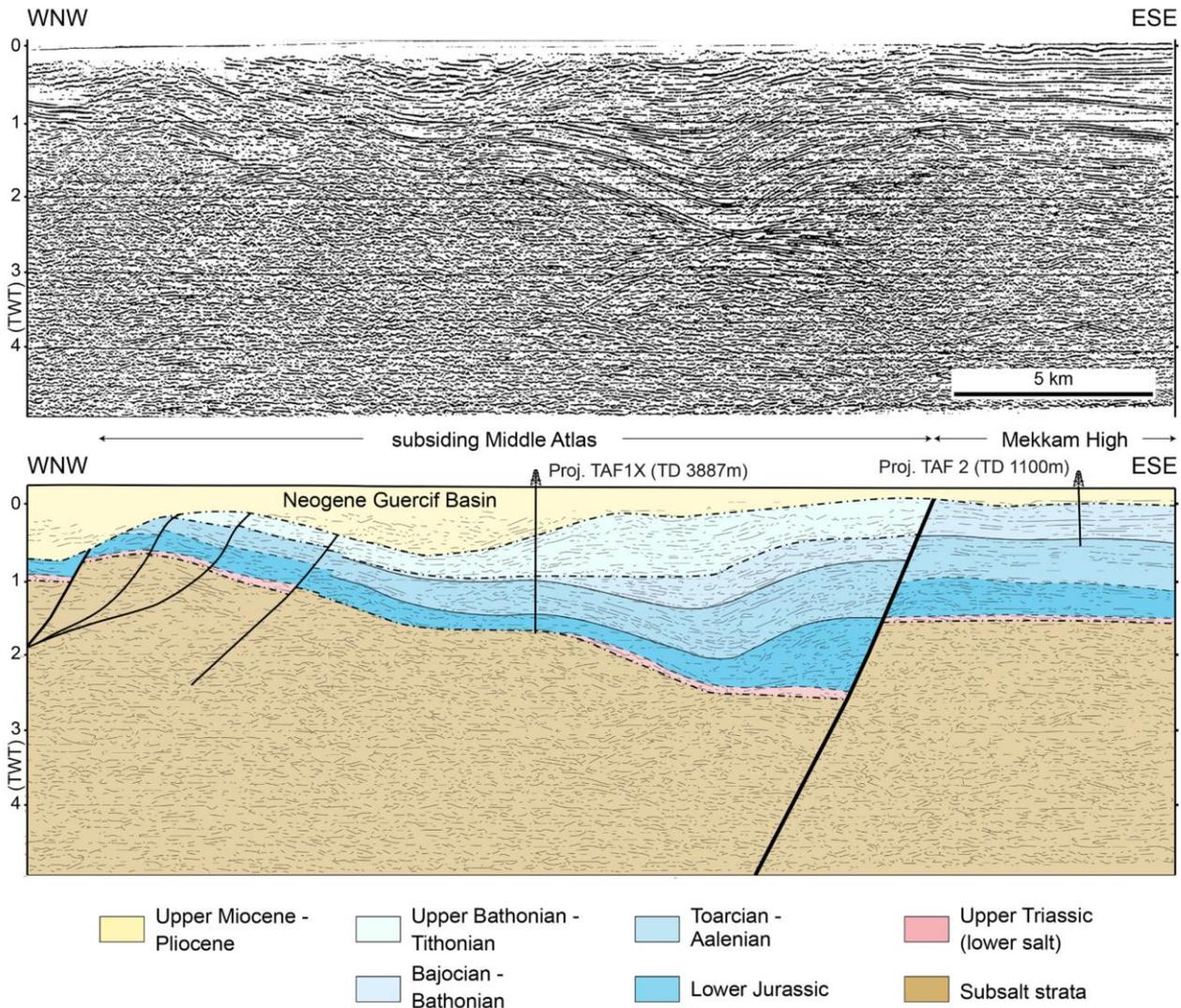


Fig. 13. Interpretation of the seismic reflection profile across the Middle Atlas and the Mekkam High (modified from Zizi, 1996). See location of the seismic reflection profile in Fig. 11.

An important deepening phase was widely recorded in the Middle Atlas during the sedimentation of lower to mid-Bajocian marls, which in some parts of the basin were deposited unconformably over older deposits (du Dresnay, 1988; Charrière, 1990). This second deepening cycle ended with the progressive filling of the basin, the transition from shallow marine to shoreface environments and the final development of red beds from Middle–Late Bathonian to Hauterivian times. These are in turn capped by a widespread unconformity that is Barremian/Aptian(?) in age (Charrière and Haddoumi, 2016). During the Bajocian to Bathonian–Callovian times, a series of elongated ridges were developed separating different depocenters (Fedan, 1988; Charrière, 1990; Charrière and Haddoumi, 2016).

These ridges represented relative structural highs against which the growth strata developed (e.g. see the northeastern Oudiksou syncline border against the Lakhchabt salt wall in Fig. 14; Laville, 1977; Fedan, 1977; Charrière, 1990). Adjacent to these ridges there are growth strata depicting angular unconformities (see cross-sections A, B, C, and D in Fig. 14), with olistolites and occurrences of bipyramid quartz (coming from the Triassic salt) in contemporaneous sediments. These field evidences point out to a strong coupling between the active growth of salt ridges and the rapid subsidence attesting for active halokinesis at the time. The possibility of some additional rifting phases has been suggested by Zizi (2002) based on the interpretation of the seismic reflection profiles in the Guercif Basin. This author suggests that extension had acted from the end of Early Jurassic to Late Jurassic. In fact, as far as it concerns the Middle Atlas, no major extensional phase could be recognized after the Toarcian or only as minor events (between Aalenian and Bajocian; Frizon de Lamotte et al., 2008).

Strikingly, the interpretation of the seismic lines in the Guercif Basin, showing the Middle Atlas continuation below the Neogene series (Zizi, 1996; Sani et al., 2000), denotes that extensional faulting mainly occurred in Callovian to Upper Jurassic times (Fig. 13). Moreover, well correlations in the Guercif Basin also attest for syn-depositional lateral thickness variations Bajocian(?)–Bathonian in whose origin is not clear, but suggested by Sani et al., (2000) to still represent extensional activity (see their Fig. 4). Finally, extension post-dating Hauterivian times is not recorded in the Middle Atlas and the first stages of inversion followed the main High Atlas tectonic schedule (Frizon de Lamotte et al., 2008).

4.4. The Rif Basin

The Rif belt is in northern Morocco at the junction between the Mediterranean Sea and the Central Atlantic Ocean (Fig. 1b). It results from the Cenozoic interaction between the African and Iberian plates with the Alboran Domain in between (Durand-Delga et al., 1960-1962; Suter, 1965; Chalouan et al., 2008; Gimeno-Vives et al., 2020b). The Rif belt is subdivided into three main structural domains (Fig. 11): the internal Rif (Alboran domain), the Flysch and the External Rif. In the frame of this study, we focus exclusively on the external Rif, where remnants of the NW Africa paleomargin are preserved (Favre et al., 1991; Favre, 1992). From SE to NW, the external Rif is subdivided into three structural-paleogeographic domains: the Prerif, the Mesorif and the Intrarif (Suter, 1965; Favre, 1992), defining the proximal, the necking zone and distal paleomargin, respectively (Gimeno-Vives et al., 2019 and Gimeno-Vives et al., 2020a; Michard et al., 2020).

4.4.1. Paleozoic inheritance

Very little information about the structure and Paleozoic tectonic evolution exists in the geological literature dealing with the external Rif. The main reason is that the Upper Triassic evaporites acted as

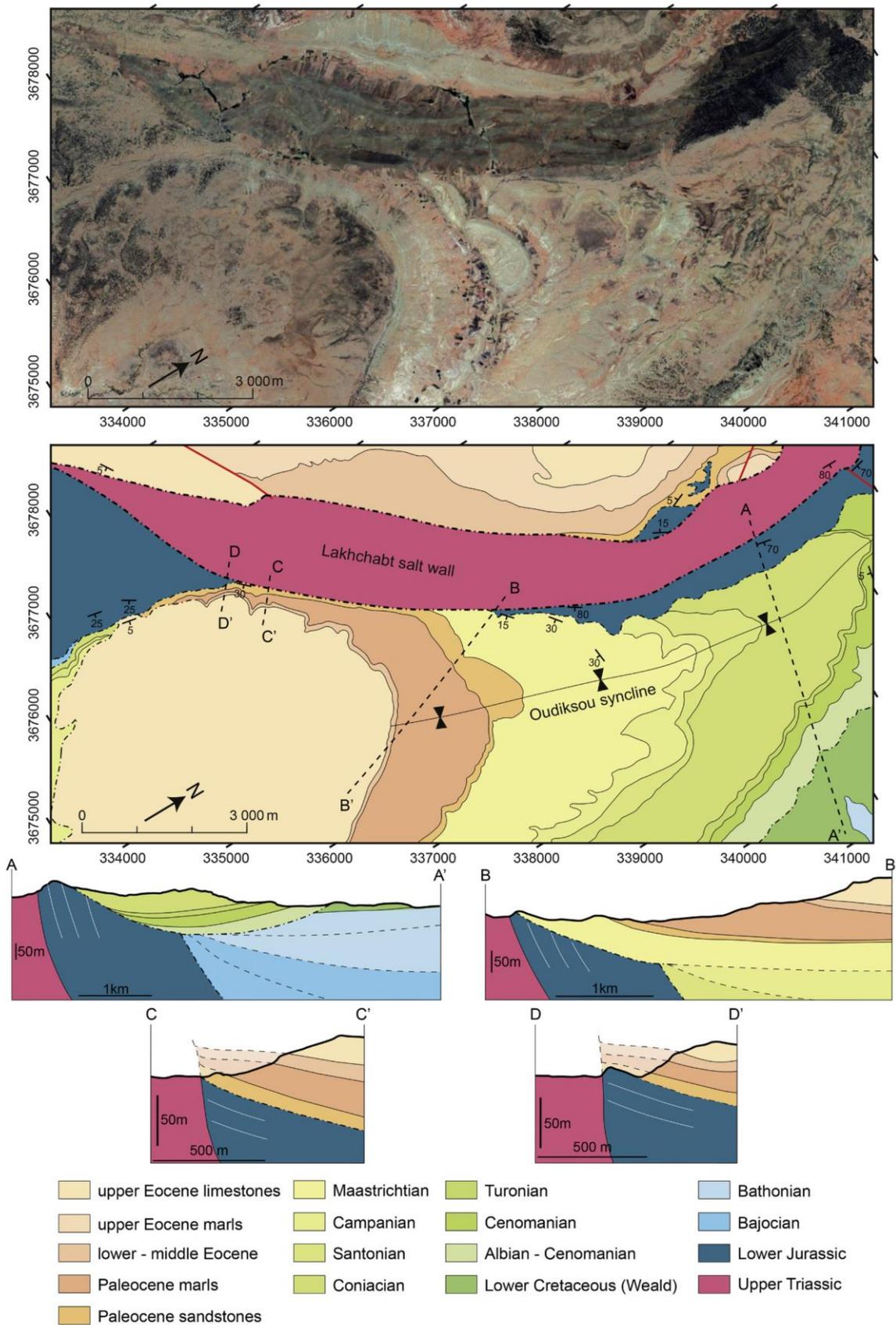


Fig. 14. Satellite image and associated geological map of the Oudiksou syncline adjacent to the Lakhchabt salt wall (“folded” Middle Atlas). Geological map modified from Charrière (1990), and cross-sections reinterpreted from his work. Satellite image from the Google Earth repository. Coordinates are in meters in Universal Transverse Mercator, zone 30 northern hemisphere and datum Nord Sahara 1959. See the location of the geological map in Fig. 11.

a décollement level during both the Mesozoic and Cenozoic tectonic events making, in general, the underlying Paleozoic levels unreachable (Gimeno-Vives et al., 2019 and 2020b). However, basement blocks constituted of gneiss, micaschist, kinzigite, and Paleozoic sedimentary rocks have been recognized embedded in Triassic evaporites (Frizon De Lamotte, 1985; Lacoste, 1934).

In the Gharb-Marmora basin (i.e. representing the foreland of the Rif belt; Fig. 11), a series of wells (MAM1, SE1, M01, EM1, S01, KC5 and KC1) intersect thick upper Miocene strata and thin Mesozoic strata before reaching the Paleozoic succession (Suter, 1980). To the east, in the Tazekka Massif, upper Miocene strata are lying directly over the Paleozoic succession without Mesozoic rocks in between. Therefore, the horst-and-graben geometry observed in the foreland continues below the chain and the major NE-trending faults (i.e. the Nekor Fault Zone) are likely inherited from the Variscan fault array. From the map depicted in Fig. 11 it can be inferred that, beneath the Rif belt, the remnants of the Variscan belt continue but most likely reworked by the stretching developed in the distal part of the margin during the subsequent Triassic and Jurassic extensional phases.

4.4.2. Mesozoic rifting

The NW Africa paleomargin encompassed a complex and polyphased rifting history during the Mesozoic. The first rifting event, occurring during the Late Triassic, remains puzzling since there are very few outcrops presenting pre-Keuper strata. Nevertheless, this rifting event is well defined in adjacent Mesozoic basins that have remained almost unaffected by the subsequent Alpine inversion (e.g. the Khemisset Basin and the Rides Prérifaines see Fig. 11; Et-Touhami, 2000; Zizi, 2002; Sani et al., 2007). Although the Triassic architecture cannot be properly recognized, the existence of a Late Triassic basin is evidenced by the presence of extensive Triassic evaporites cropping out over the external Rif. The importance of this first rifting phase remains difficult to evaluate. However, a coeval thinning of the lithosphere is certain, since large mafic magmatic additions were emplaced within the continental crust at about 200 Ma (Gimeno-Vives et al., 2019; Haissen et al., 2021). Such magmatism is expressed in the Rif under the name of Mesorif Gabbroic Complex "MGC" and has been attributed to the Central Atlantic Magmatic Province (CAMP).

After the emplacement of this widespread magmatism and during the lowermost Jurassic, a large carbonate platform deposited coinciding with a period of relative tectonic quiescence (Papillon, 1989; Favre, 1992). Subsequently, a second rifting phase occurred, characterized by a sudden deepening of the basin, from the Toarcian to Upper Jurassic times. Contrary to the first Late Triassic extension, this phase resulted on successful rifting, thus completing the oceanization and the NW-SE opening of the Maghrebien Tethys. In the Mesorif, gabbroic rocks are overlaid by dismembered and discontinuous

blocks of Lower Jurassic carbonates, and covered by Middle to Upper Jurassic sediments. Such a geometry indicates a decoupling between the suprasalt strata and the basement along the Triassic evaporites. Subsalt deformation allowed the exhumation of the Mesorif Gabbroic Complex up to the surface throughout large detachment fault (see a restored cross-section in Fig. 15).

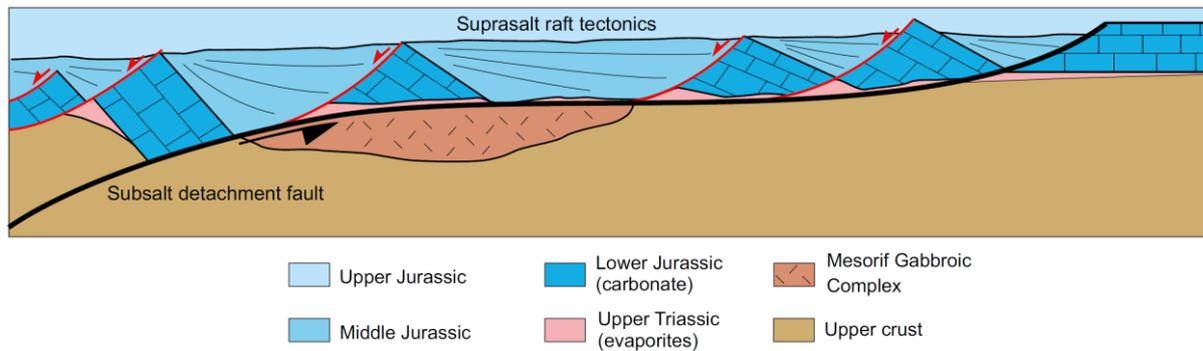


Fig. 15. Restored cross-section in the Rif Basin during Late Jurassic times, depicting the exhumation of the Mesorif Gabbroic Complex along a low-angle extensional fault and associated raft-tectonics extensional system in its hangingwall. Modified from Gimeno-Vives et al. (2020a).

On the other hand, in the detachment hanging wall, this rifting phase is dominated by raft tectonics recording the stretching and dismembering of the Lower Jurassic platform over the Upper Triassic evaporites. Importantly, coeval salt movements are observed throughout the development of salt diapirs in the distal part of the margin (Gimeno-Vives et al., 2019). These features are well documented in the syn-rift sequence Toacian–Bajocian in age, presenting thickness variations and wedge geometries against diapiric structures. The most distal part of the margin, exposed in the Intrarif, exhibits serpentinized peridotites, interpreted as exhumed mantle (Michard et al., 1992; 2007; Gimeno-Vives et al., 2019). The Uppermost Jurassic limestones and calcschists resting directly on top (Fig. 15), suggests a Late Jurassic mantle exhumation age and subsequent drifting, which gave birth to the Maghrebian Tethys.

5. Discussion

Regional observations based on geological and seismic evidences, showed that the pre-orogenic and orogenic structures related to the Variscan belt controlled the extensional deformation during the formation of the NW Africa paleomargin (see Table 1 for a summary of the ages of the tectonic events in NW Africa) and, consequently, the localization and evolution of the Mesozoic rifts. Firstly, it is important to understand how this control works. Then, we will discuss the timing of the Mesozoic rifting events from one basin to the other, in relationship with the geodynamics of the Atlantic and Tethys oceans. And finally, we will consider the timing and modalities of the rift inversion.

Table 1

Comparison of the rifting timings in the different studied basins of NW Africa.

		Sahara Atlas	Tendrarra Basin	Middle Atlas / Guercif Basin	Rif Basin
Paleozoic	Extension	Unknown	Inferred Cambrian to early Carboniferous	Inferred Ordovician to Devonian	Unknown
	Inversion	upper Carboniferous (dated by igneous intrusions of 298 Ma)	Inferred upper Carboniferous	Inferred upper Carboniferous	Inferred upper Carboniferous
Mesozoic	Permo-Triassic rifting	Late Triassic (NE- to ENE-trending faults sealed by the CAMP basalts)	Carnian – Norian and Rhaetian (NE- to NNE-trending faults sealed by the CAMP basalts)	Late Triassic (emplacement of the CAMP magmatic intrusions)	Inferred Late Triassic (emplacement of the Mesorif Gabbroic Complex, CAMP equivalent)
	Jurassic rifting 1 (Atlantic domain)	Toarcian (reactivation of the NE-trending extensional faults)	Pliensbachian – Toarcian (reactivation of the NE- to NNE-trending faults)	uppermost Pliensbachian	
	Jurassic rifting 2 (Maghrebien Tethys domain)	early Bathonian	Middle Jurassic	middle Bajocian to Callovian	Toarcian to Upper Jurassic (successful rifting and oceanization)

5.1. A prominent Paleozoic inheritance

The Atlas Mesozoic rift system developed over a segment of the Variscan belt (i.e. Moroccan Variscides). Given the absence of continental suture in the Moroccan Variscides, this orogeny is considered as an intraplate orogeny entirely deriving from the NW passive margin of Gondwana (Simancas et al., 2009; Michard et al., 2010). Before the Variscan orogeny, this domain was the site of two major periods of rifting, during the Cambrian-Ordovician and Middle/Late Devonian to early Carboniferous times, respectively (Hollard et al., 1982; Wendt, 1985; El Hassani, 1994; Soulaïmani et al., 2004; Wendt et al., 2006; Simancas et al., 2009; Eschard et al., 2010; Frizon de Lamotte et al., 2013; Fig. 2). The later was followed by the Variscan tectonic inversion during the late Carboniferous-early Permian (Chopin et al., 2014; Wernert et al., 2016; Delchini et al., 2018). The NE- to ENE-trending Late Devonian-Early Carboniferous rift basins (Fig. 16a) are of crucial importance for the subsequent tectonic evolution of the Moroccan Variscan belt which directly results from the inversion of these basins (Fig. 16b). A set of arguments shows that the rift basins were particularly deep and hot with high paleogeothermal gradients (Chopin et al., 2014; Wernert et al., 2016; Delchini et al., 2018; Lahfid et al., 2019), similarly to what has been observed in the Aptian-Albian Cretaceous Pyrenean basins (Vacherat et al., 2014; Clerc et al., 2015).

In NW Africa, the general structural array for the Variscan deformation resulted from the inversion of the previously hot Lower Carboniferous basins; for this reason, the extension and the distribution of these basins represents a key factor. During their inversion, the former basin horsts emerged as “pop-up” structures, meanwhile the exhumed deeper parts of the basins reveal a scarce Barrovian metamorphism (e.g. in the Rehamna massif: see Chopin et al., 2014; Fig. 1b). There is not complete agreement on the condition for tectonic inversion in the intraplate setting of the NW Africa Variscides. However, it is largely accepted that the pre-orogenic conditions that allowed accumulating heat likely promoted the tectonic inversion (e.g. for the Rehamna area: Chopin et al., 2014; Wernert et al., 2016;

or for the Jebilet Massif: Delchini, 2018; Delchini et al., 2018; Fig. 1b). The compression produced a general NE-SW to E-W structural trends, accounting for the overall present-day structure in Paleozoic outcrops, showing alternation of Lower Paleozoic topographic highs with Lower Carboniferous basins. The inversion of former extensional basins occurred in two main phases with distinct orientations: a first compression that was oriented NNW-SSE, followed by a second one oriented WNW-ESE (Chopin et al., 2014; Delchini et al., 2018; Fig. 16b).

The trends of Mesozoic rift basins follow almost strictly the structural grain of the Variscan Belt. In particular, it is striking to note how it follows the NE-SW Paleozoic inheritance in the Middle Atlas, Rif and Tendirara basins, whereas it becomes E-W in the Tamlelt area, as the former Paleozoic faults bend (e.g. Houari and Hoepffner, 2003) before trending NE-SW again in the Sahara Atlas (compare Fig. 16b and c). However, the Mesozoic basins indifferently developed over both former basin horsts (“pop-up” structures, e.g. Middle Atlas, on top of Tazekka Paleozoic high) or former Lower Carboniferous basin (e.g. Tendirara Basin). This configuration may suggest that after the Permian emplacement of magmatic intrusions (El Hadi et al., 2006) and at the onset of the Mesozoic rifting period, the lithosphere was thermally re-equilibrated with a flat topography forming a wide levelling surface. So, at the beginning of the Triassic and beneath the whole area, the lithospheric mantle could be homogeneously replaced by more fertile lithospheric mantle (e.g. the consequence of the “mantle plume” effect and related asthenospheric upwelling). So, globally there is a transition from numerous uppermost Carboniferous-early Permian riftings likely triggered by thermal insulation of the Pangea continental aggregation and subsequent large-scale melting and global weakness of the lithosphere, to a scenario of Mesozoic rifting systems that are progressively more confined along inherited structural trends and influenced by a relative increase of the “mantle plume” effect (Frizon de Lamotte et al., 2015). So, this might suppose a “reset” of the Paleozoic thermal inheritance, thus the subsequent Mesozoic rifting events will be progressively more constrained along the main inherited structural (Variscan) trends.

5.2. A generalized rifting episode during the Late Triassic

Overlying a mature (i.e. flattish) topography, the oldest Mesozoic syn-rift sediments are fine grain to conglomeratic sandstones dated lower Carnian–Norian in the Tendirara Basin (Taki, 2012 and references therein). Overlying this initial siliciclastic sequence, syn-rift Rhaetian evaporites (Keuper Fm.) deposited in fault-bounded basins (Fig. 3). The Late Triassic sedimentary sequences were then overlain by the continental flood basalts dated around 200 Ma (Marzoli et al., 1999; Knight et al., 2004; Verati et al., 2007), pertaining to the CAMP (see its distribution in Fig. 16c). In the study area, the CAMP basalts clearly post-date the Triassic rifting. However, the restricted sedimentary environments and evaporitic sedimentation lasted up to the Sinemurian, so being partly post-rift by reference to the

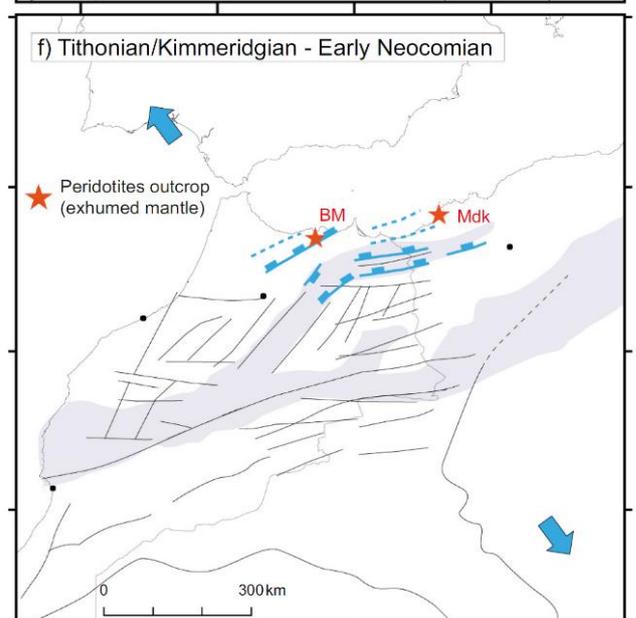
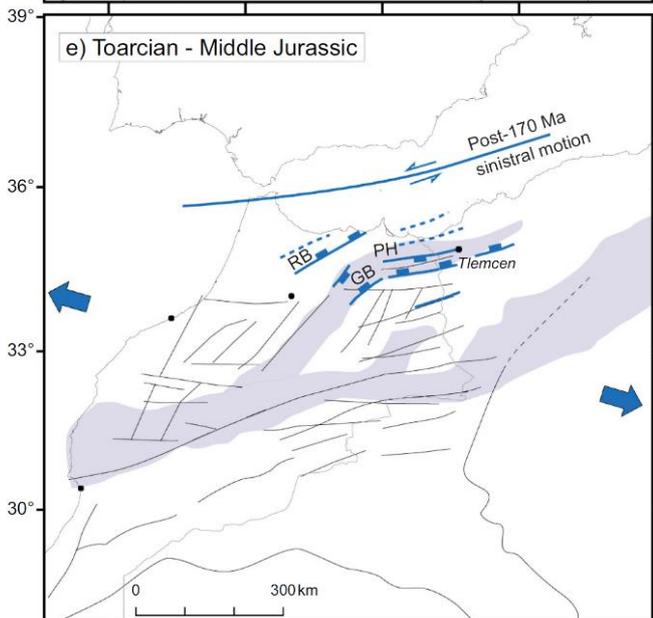
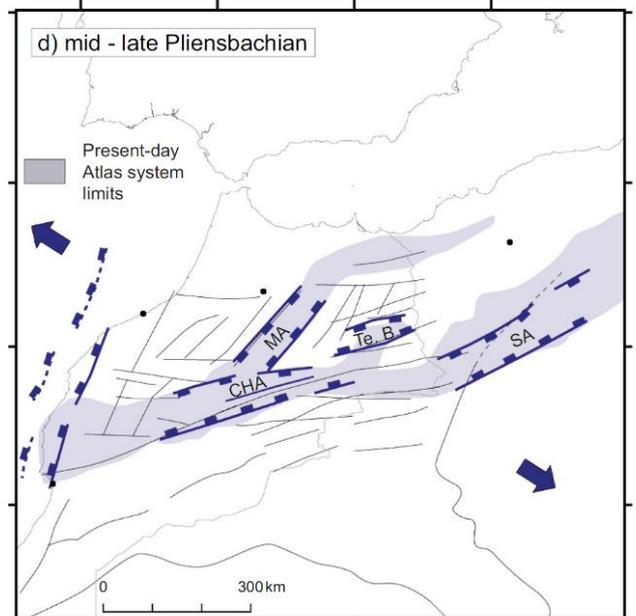
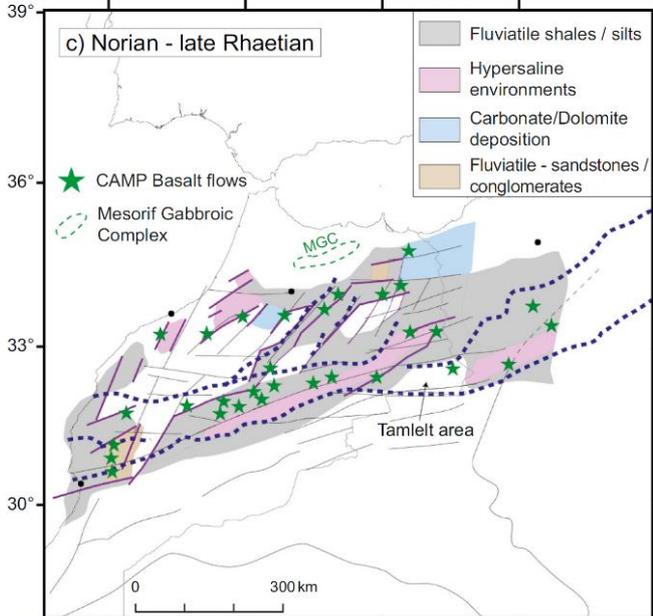
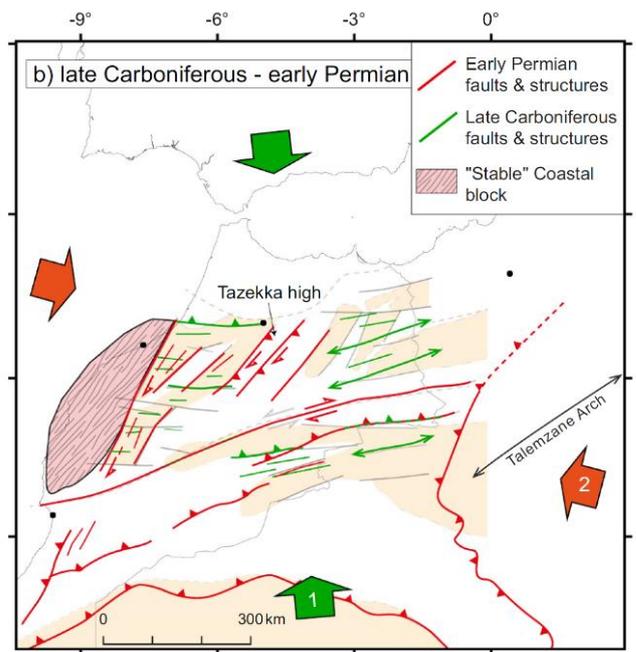
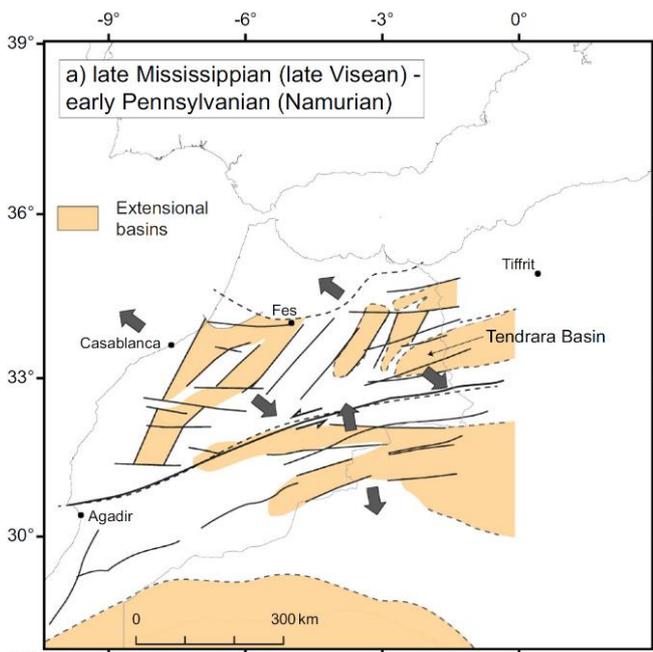


Fig. 16. Time-sequenced tectonic maps of NW Africa depicting: a) the location of the main subsiding areas and the rifting trend during Late Visean to Namurian times (after Hollard (1967), Wendt and Aigner (1985), Piqué and Michard (1989), Bouabdelli (1989), El Hassani (1994), Zahraoui (1994), Fabre (2005), Hoepffner et al. (2005), Wendt et al. (2006); b) the main trends of the Variscan Belt re-using the previous Paleozoic extensional features; c) Upper Triassic basins indifferently developed over both former lower Paleozoic horsts and Lower Carboniferous rift basins (Oujidi et al., 2000; Courel et al., 2003; Meddah et al., 2017; Gimeno-Vives et al., 2019; El Hachimi et al., 2020); d) mid to late Pliensbachian basins associated with the rifting along the Central Atlantic margin (CHA: Central High Atlas; MA: Middle Atlas; Te. B.: Tendirra Basin; SA: Sahara Atlas; e) the north-eastward jump of rifting/spreading of the Central Atlantic domain accommodated by a major E-W transform fault system and activating NE- to ENE-trending extensional faults in the Maghrebian Tethys domain during Toarcian – Middle Jurassic times (RB: Rif basin; GB: Guercif Basin; PH: “Pays des Horst”); f) hyper-extension in the Maghrebian margin triggering mantle exhumation and the start of oceanic accretion in the Maghrebian Tethys (“BM” and “Mdk”: Beni Malek and Medakh peridotite bodies, respectively). All map coordinates are in degrees in World Geodetic System 84.

Triassic rifting. The presence of restricted basins allowing the sedimentation of thick Lower Jurassic evaporite successions could be the consequence of the formation of volcanic ridges combined with the thermal subsidence related to previous rifting in the northern central Atlantic continental margin.

In the Rif, the end of the Upper Triassic rifting episode is marked by the emplacement of the Mesorif Gabbroic Complex within the continental crust (Gimeno-Vives et al., 2019 and Gimeno-Vives et al., 2020a; Haissen et al., 2021; Fig. 16c). For these authors, the Mesozoic Gabbroic Complex pertains to the CAMP underlining the importance of this magmatic event at the scale of the NW African paleomargin. The occurrence of intraplate magmatism might be related to the existence of inherited tectonic/thermal features weakening the continental crust. The origin of the Upper Triassic magmatism and the related emplacement of the Mesorif Gabbroic Complex in the Rif might be favored by the presence of major faults inherited from the Variscan Orogeny or even older tectonic events. These crustal-scale faults separating different accreted Precambrian to Paleozoic terranes (Michard et al., 2010, their Fig. 12) could be reactivated, as the weaker parts of the lithosphere, thus promoting the localization of extensional faulting during the subsequent Mesozoic rift. So, in a similar way than the Paleoproterozoic shear zones onshore Angola reactivated during the opening of the South Atlantic Ocean and localizing syn-rift and syn-exhumation Cretaceous magmatism (e.g. Lucapa and/or Sumbe shear zones: Alberti et al., 1999; Eley et al., 2008; Jelsma et al., 2018), the major fault zones in NW Africa could connect the subcontinental mantle and the crust acting as preferential conduits for the emplacement of magmatic additions into the weakened lithosphere during the Mesozoic rifting.

5.3. Differentiated Jurassic rifting events and opening of the Maghrebian Tethys

After the emplacement of the CAMP basalts everywhere and the Mesozoic Gabbroic Complex in the Rif Basin (Gimeno-Vives et al., 2019), rifting resumed along the Central Atlantic and in the Atlas systems (Figs. 16d). Along the Central Atlantic, the precise timing of this new rifting episode has been debated (Sahabi, 2004; Labails et al., 2010; Biari et al., 2017). However, it is clear that the first oceanic crust occurred no later than 175 Ma in the Atlantic Ocean domain (Klitgord and Schouten, 1986) and that a slow-extension corresponding possibly to hyper-extension of the continental crust and coeval mantle exhumation developed between 195 (late Hettangian) and 175 Ma (late Toarcian) (see discussion in Biari et al., 2017). As suggested by Frizon de Lamotte et al. (2011), in the northern part of Central Atlantic domain, rifting/spreading propagated north-eastward at the Bajocian period, towards the incipient Maghrebian Tethys (Sibuet et al., 2012; Fernandez, 2019). At that time, the relative left-lateral movement between Africa and Newfoundland-Iberia was accommodated along a lithospheric-scale E-W transform fault system (Fig. 16e), likely inherited from the Variscan orogeny (this cannot be argued through field evidences but comes up as a necessary constraint in the Late Variscan stages and the Pangaea assembly; e.g. Edel et al., 2018). This major structure controlled the localization of the incipient Maghrebian Tethys (Sallarès et al., 2011; Leprêtre et al., 2018a; Fernandez, 2019).

In the Atlas System (i.e. the Middle Atlas, the Central High Atlas, Tadrara Basin and the Sahara Atlas), the Jurassic rifting episode resumed by the middle to late Pliensbachian (189 Ma) and likely was over by the Toarcian (Fig. 16d). In the Central High Atlas, this is confirmed by coeval salt mobility as proposed by several authors (e.g. Saura et al., 2013; Jousiaume et al., 2014; Moragas et al., 2017), but a similar timing could be also argued for the Sahara Atlas (Vially et al., 1994; Elmi et al., 1998; Yelles-Chaouche et al., 2001), the Middle Atlas (Fedan, 1988; Charrière, 1990) and our new results in the Tadrara Basin. By the Middle Jurassic, halokinesis stopped along the margins of the Central High Atlas Basin in agreement with the end of the rift-related extension at that time. Salt mobility continued passively in the center of the basin at least up to the Bathonian. In the Middle Atlas, the extension associated to this tectonic phase is known to have stopped during Middle Toarcian (Fedan, 1988; El Arabi et al., 2001), however it seems to have persisted in northernmost parts, in the Guercif area.

In the Rif Basin, the second episode of rifting began later mainly during the Toarcian and after a phase of relative tectonic quiescence (Favre et al., 1991; Elmi et al., 1998; Benest et al., 1999; Chotin et al., 2000; Gimeno-Vives et al., 2019; Fig. 16e). Nevertheless, the deposition of sedimentary breccias during the Early Jurassic (Favre, 1992) suggests a tectonic instability but without significant tectonic subsidence. The main phase of rifting, crowned by the exhumation of the Mesorif Gabbroic Complex along major detachments (Fig. 16f), occurred during Middle Jurassic times. The syn-rift configuration is well-preserved in the Senhadja Nappe (see a review in Gimeno-Vives et al., 2020b; Fig. 11). In the Mesorif, the second episode of Jurassic rifting is sealed by the "Ferrysh Fm" (Wildi, 1983), consisting

of Upper Jurassic turbidites and interpreted as a post-rift stratigraphic succession. In the Intra-rif (i.e. the distal part of the Tethys margin), this rifting was active until the end of the Jurassic, as demonstrated by mantle exhumation via hyper-extension associated to the occurrence of peridotites in the Beni Malek massif (Gimeno-Vives et al., 2019; Michard et al., 1992, 2007; Gimeno-Vives et al., 2020a; Fig. 16f), and followed by drifting during the Lower Cretaceous.

The organization is relatively similar for the “Pays des Horsts” (NE Morocco) and the Tlemcen area (NW Algeria; Fig. 1b). There, scarce tectonic activity is recognized during Early Jurassic. Rifting resumed during the Toarcian, with a relatively quiescent period during the Aalenian, followed by an important Bathonian-Bajocian tectonic activity and related deepening of the sedimentary environments (Fedan, 1988; Charrière, 1990; Benest et al., 1999; Chotin et al., 2000). A Kimmeridgian/Early Cretaceous extensional activity is also demonstrated in the “Pays des Horsts”, with a global northward deepening towards the Maghrebian Tethys (Chotin et al., 2000). In addition, the possible Lower to Middle / Upper Jurassic extensional activity recorded below the Guercif Basin (Zizi, 2002) could be related to the same extension that is recorded in the Rif and the “Pays des Horsts” and Tlemcen area.

Therefore, looking at the scale of the entire Atlas and Rif systems, it appears that the Jurassic rifting developed firstly in a trend oriented NE-to ENE in the Central High Atlas, the Middle Atlas, the Tendirra Basin and the Sahara Atlas and was primarily associated to the opening of the Central Atlantic (Figs. 16d). At the end of the Toarcian, whereas oceanization began in the Central Atlantic and rifting stopped in the High Atlas, the continental rift jumped with a NE- and ENE oriented trend in the Rif Basin, “Pays des Horsts” and Tlemcen area, respectively (Fig. 16e). This led ultimately to mantle exhumation during the uppermost Jurassic in the Rif Basin, and to pulses of extension through Middle and Late Jurassic in the “Pays des Horsts”, Tlemcen area and in the NE Middle Atlas (Fig. 16f). This rifting episode and the subsequent drifting accommodated the relative left-lateral transtensional displacement between Africa and Newfoundland-Iberia and the extensional settings ceased quickly after the Early Cretaceous. Drifting in the Maghrebian Tethys lasted up to the Campanian and the onset of the Africa-Eurasia convergence.

The Jurassic timespan can thus be observed as a nice example of how, due to tectonic inheritance, the far-field stresses and their orientations are being sampled by the different Moroccan and Algerian rifted segments. It is different from the Triassic rifting since the Jurassic geodynamic kinematics (i.e. the openings of the Central Atlantic Ocean and the Maghrebian Tethys) “activated” different segments of the weakened NW African lithosphere over which the extensional deformational was distributed. Instead, the Triassic rifting stage is characterized as a more “diffuse” extensional and thermal event during which the tensile stress due to the Pangaea disruption are widely distributed along the main

inherited structural trends (see Figs. 3 and 16c), causing all the concerned areas to undergo rifting accompanied with important magmatism.

5.4. A contrasted tectonic inversion

In the NW Africa paleomargin, like in the whole Tethys realm, the tectonic inversion began by the Late Cretaceous (e.g. Frizon de Lamotte et al., 2011; Jolivet et al., 2015). However, this first event is poorly expressed in the Rif and Atlas (Froitzheim et al., 1988); as the major inversion occurred later during the Middle–Late Eocene and upper Miocene times (Frizon de Lamotte et al., 2000, 2008, and 2011; Leprêtre et al., 2018b; Gimeno-Vives et al., 2020b; Fig. 2).

In the Atlas system, along our NW-SE cross-section from the Rif Basin to the Sahara Atlas, the Mesozoic rift basins are globally inverted. The peculiar case of the Eastern High Atlas (Tamlelt area, Fig. 1b), an E-W segment joining the ENE-trending Central High Atlas and Sahara Atlas, must be discussed here. It does not correspond to a former basin but represents a Mesozoic structural high bounded by reverse faults since the Variscan orogeny. Therefore, in this case, the Cenozoic inversion directly reactivated Variscan structures, that strike E-W in this area (Houari and Hoepffner, 2003). By contrast, the Tendirra Basin, constituting the continuation of the Central High Atlas, is poorly inverted. The reasons explaining these differences in the structural evolution are not completely understood. However, the spacing imposed by the inversion of the Middle Atlas and the Eastern High Atlas could control the wavelength and the localization of the contractional deformation and the preservation of areas, like for example the Tendirra Basin being poorly affected by compressional deformations. In addition, the tectonic/thermal inheritance of the Mesozoic rifting might also control the locus for the contractional deformation during the inversion of the rim basins (e.g. Gueydan et al., 2019). But in contrast to the Pyrenees, where rifting was directly followed by compression (Rosenbaum et al., 2002; Lagabrielle et al., 2010; Vacherat et al., 2014), in the Atlas system, the time gap between extension (ended by 145 Ma) and subsequent compression (onset at 83 Ma) could be long enough for the thermal re-equilibration and strengthening of the lithosphere.

North of the Atlas system, the geodynamic evolution of the Rif is dominated by the under-thrusting climax of the Maghrebian paleomargin below the Maghrebian Tethys during the Middle–Late Eocene (Leprêtre et al., 2018a; Gimeno-Vives et al., 2020b). During the Oligocene, the convergence between Africa and Europe was being accommodated along a new subduction located parallel to the northern margin of the Maghrebian Tethys. From the Middle/Late Miocene onwards, and by contrast, the deformation in the Rif has been dominated by the westward translation of the Alboran Domain (see a review in Gimeno-Vives et al., 2020b). In the Atlas domain instead, no significant deformation is recorded but the region was the site of a thermal uplift occurring along a stripe crossing Morocco from

the Anti-Atlas to the eastern Rif and causing some dramatic sliding along the borders of the system (see Leprêtre et al., 2018b). The so-called “Morocco Hot Line”, which is underlined by recent volcanism (see review in Missenard and Cadoux, 2012), partly explains the high elevation characterizing Morocco by comparison to Algeria or Tunisia (Missenard et al., 2006). Finally, by the late Miocene, a new contractional event is recorded in the whole area, leading to the reactivation of the Middle–Late Eocene structures.

6. Conclusion

The interpretation of subsurface data combined with field observations and detailed bibliographic revisions of several Mesozoic rim basins in NW Africa have shed new light on the factors that controlled the rift initiation, its development and the final accretion of oceanic crust in the Maghrebian Tethys. From this structural analysis, we suggest the following conclusions:

1. The Late Devonian-Early Carboniferous rift basins are of crucial importance for the tectonic evolution of NW Africa since the subsequent Moroccan Variscan Belt and Meso-Cenozoic deformation events directly resulted from the structural inheritance of these basins. Specifically, the pre-orogenic conditions, that allowed the accumulation of heat, facilitated the inversion of the former Paleozoic basins during the late Variscan Orogeny. Therefore, in addition to the structural control of the Paleozoic extensional faults, the thermal inheritance differently distributed along the Paleozoic basins and horsts influenced the mode in which the Variscan inversion took place in NW Africa.
2. The trends of the subsequent Mesozoic Atlas rift were strictly controlled by the structural inheritance of the Moroccan Variscan Belt. However, the rim basins analyzed in this study were indifferently developed over both former lower Paleozoic horsts and Lower Carboniferous rift basins. This may suggest that at the onset of the Mesozoic rifting period, the lithosphere was thermally re-equilibrated, and the system was rejuvenated with more fertile lithospheric mantle (i.e. “mantle plume” effect). A possible pre-requisite for the subsequent development of the Triassic rifting and the future CAMP event.
3. At the scale of the entire NW Africa paleomargin, the Triassic rifting was widely distributed and controlled by the Variscan structures. The emplacement of large magmatic additions (related to the CAMP) within the continental crust at about 200 Ma (and specifically the Mesorif Gabbroic complex in the Rif Basin) denotes an important thinning and weakening of the lithosphere. This process could be enhanced by crustal-scale faults inherited from previous Paleozoic and/or older deformation events acting as preferential conduits for the emplacement of magmatic bodies into the already weakened continental crust.

4. After the CAMP event and until Bajocian times, rifting/spreading propagated from the northern part of the Central Atlantic Ocean domain north-eastward towards the incipient Maghrebian Tethys. Note again the role played by the tectonic/thermal inheritance on the localization of the new Tethyan oceanic domain along a pre-existent E-W transform fault accommodating the relative left-lateral transtensional displacement between Africa and Newfoundland-Iberia. Following the north-eastward propagation of the Jurassic rifting, extension in the Atlas system (Middle Atlas, the Central High Atlas, Tadrara Basin and the Sahara Atlas) mainly occurred during middle to late Pliensbachian and was over by the Toarcian. However, in the Rif but also in the Pays des Horst, Tlemcen area and Guercif Basin, the extension climaxed during the Middle Jurassic with the exhumation of the Mesorif Grabbroic Complex along major low-angle extensional faults. By the Upper Jurassic, hyper-extension led to mantle exhumation in the distal part of the margin marking the onset of drifting in the Maghrebian Tethys. According to the Jurassic tectonic agenda of NW Africa, the deformation was sampled by the different Moroccan and Algerian rifted segments developed during Triassic times. This rifting period is different from the Triassic since the successive openings of the Central Atlantic Ocean and Maghrebian Tethys activated different segments of the weakened NW African continental crust where the extensional deformation was distributed.

5. Finally, the Alpine inversion of the rim basins in the NW African margin mainly affected the areas that underwent major stretching during the Mesozoic rifting phase. In addition, the spacing imposed by the inversion of the Middle Atlas and the Eastern High Atlas could control the wavelength and the localization of the contractional deformation and the preservation of areas, like for example the Tadrara Basin being poorly affected by compressional deformations.

Declaration of Competing Interest

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

Acknowledgments

This work was funded by the Total Group and comes from the results of a M2 internship conducted by the first author at the same company. The authors wish to acknowledge an anonymous reviewer and the editor for their constructive and fruitful reviews. The views and conclusions contained in this article are solely those of the authors and should not be interpreted as necessarily representing the views or the official policies, either expressed or implied, of the Total Group.

Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.earscirev.2021.103732>

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