Influence of basement heterogeneity on the architecture of low subsidence rate Paleozoic intracratonic basins (Ahnet and Mouydir basins, Central Sahara)

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Abstract

The Paleozoic intracratonic North African Platform is characterized by an association of arches (ridges, domes, swells or paleo-highs) and low subsidence rate syncline basins of different wavelengths (75–620 km). The structural framework of the platform results from the accretion of Archean and Proterozoic terranes during the Pan-African orogeny (750–580 Ma). The Ahnet and Mouydir basins are successively delimited from east to west by the Amguid El Biod, Arak-Foum Belrem, and Azzel Matti arches, bounded by inherited Precambrian subvertical fault systems which were repeatedly reactivated or inverted during the Paleozoic.
Major unconformities are related to several tectonic events such as the Cambrian–Ordovician extension, Ordovician–Silurian glacial rebound, Silurian–Devonian “Caledonian” extension/compression, late Devonian extension/compression, and “Hercynian” compression. The deposits associated with these arches and syncline basins exhibit thickness variations and facies changes ranging from continental to marine environments. The arches are characterized by thin amalgamated deposits with condensed and erosional surfaces, whereas the syncline basins exhibit thicker and well-preserved successions. In addition, the vertical facies succession evolves from thin Silurian to Givetian deposits into thick Upper Devonian sediments. Synsedimentary deformations are evidenced by wedges, truncations, and divergent onlaps. Locally, deformation is characterized by near-vertical planar normal faults responsible for horst and graben structuring associated with folding during the Cambrian–Ordovician–Silurian period. These structures may have been inverted or activated during the Devonian compression and the Carboniferous. The sedimentary infilling pattern and the nature of deformation result from the slow Paleozoic reactivation of Precambrian terranes bounded by vertical lithospheric fault zones. Alternating periods of tectonic quiescence and low-rate subsidence acceleration associated with extension and local inversion tectonics correspond to a succession of Paleozoic geodynamic events (i.e. far-field orogenic belt, glaciation).

Keywords: intracratonic basin, Paleozoic, arches, low-rate subsidence, tectonic heritage, terranes, Central Sahara

1 Introduction

Paleozoic deposits fill numerous intracratonic basins, which may also be referred to as “cratonic basins”, “interior cratonic basins”, or “intracontinental sags”. Intracratonic basins are widespread around the world (see Fig. 6 from Heine et al., 2008) and exploration for non-conventional petroleum has revived interest in them. They are located in “stable” lithospheric
areas and share several common features (see Allen and Armitage, 2011 and references therein) such as their geometries (i.e. large circular, elliptical, saucer-shaped to oval), their stratigraphy (i.e. filled with continental to shallow-water sediments), their low rate of sedimentation (an average of 7 m/Myr), their long-term subsidence (sometimes more than 540 Myr), and their structural framework (reactivation of structures and emergence of arches also referred to in the literature as “ridges”, “paleo-highs”, “domes”, and “swells”). Multiple hypotheses and models have been proposed to explain how these slowly subsiding, long-lived intracratonic basins formed and evolved (see Allen and Armitage, 2011 and references therein or Hartley and Allen, 1994). However, their tectonic and sedimentary architectures are often poorly constrained.

The main specificities of intracratonic basins are found on the Paleozoic North Saharan Platform. The sedimentary infilling during c. 250 Myr is relatively thin (i.e. around a few hundred to a few thousand meters), of great lateral extent (i.e. 16 million km²), and is separated by major regional unconformities (Beuf et al., 1971; Carr, 2002; Eschard et al., 2005, 2010; Fabre, 1988, 2005; Fekirine and Abdallah, 1998; Guiraud et al., 2005; Legrand, 2003). Depositional environments were mainly continental to shallow-marine and homogeneous. Very slow and subtle lateral variations occurred over time (Beuf et al., 1971; Carr, 2002; Fabre, 1988; Guiraud et al., 2005; Legrand, 2003). The Paleozoic North Saharan Platform is arranged (Fig. 1) into an association of long-lived broad synclines (i.e. basins) and anticlines (i.e. arches swells, domes, highs, or ridges) of different wavelengths (λ: 75–620 km). Burov and Cloetingh, (2009) report deformation wavelengths of the order of 200–600 km when the whole lithosphere is involved and of 50–100 km when the crust is decoupled from the lithospheric mantle. This insight suggests that the inherited basement fabric influences basin architecture at a large scale. Intracratonic basins are affected by basement involved faults which are often reactivated in response to tectonic pulses (Beuf et al., 1971;
In this study of the Ahnet and Mouydir basins, a multidisciplinary workflow involving various tools (e.g. seismic profiles, satellite images) and techniques (e.g. photo-geology, seismic interpretation, well correlation, geophysics, geochronology) has enabled us to (1) make a tectono-sedimentary analysis, (2) determine the spatial arrangement of depositional environments calibrated by biostratigraphic zonation, (3) characterize basin geometry, and (4) ascertain the inherited architecture of the basement and its tectonic evolution. We propose a conceptual coupled model explaining the architecture of the intracratonic basins of the North Saharan Platform. This model highlights the role of basement heritage heterogeneities in an accreted mobile belt and their influence on the structure and evolution of intracratonic basins. It is a first step towards a better understanding of the factors and mechanisms that drive intracratonic basins.

2 Geological setting: The Paleozoic North Saharan Platform and the Ahnet and Mouydir basins

The Ahnet and Mouydir basins (Figs 1 and 3) are located in south-western Algeria, north-west of the Hoggar massif (Ahaggar). They are N–S oval depressions filled by Paleozoic deposits. The basins are bounded to the south by the Hoggar massif (Tuareg Shield), to the west by the Azzel Matti arch, to the east by the Amguid El Biod arch and they are separated by the Arak-Foum Belrem arch.

Figure 2 synthesizes the lithostratigraphy, the large-scale sequence stratigraphic framework delimited by five main regional unconformities (A to E), and the tectonic events proposed in the literature (cf. references under Fig. 2) affecting the Paleozoic North Saharan Platform.
During the Palaeozoic, the Ahnet and Mouydir basins were part of a set of the super-continent Gondwana (Fig. 1). This super-continent resulted from the collision of the West African Craton (WAC) and the East Saharan Craton (ESC), sandwiching the Tuareg Shield (TS) mobile belt during the Pan-African orogeny (Craig et al., 2006; Guiraud et al., 2005; Trompette, 2000). This orogenic cycle followed by the chain's collapse (c. 1000–525 Ma) was also marked by phases of oceanization and continentalization (c. 900–600 Ma) giving rise to the heterogeneous terranes in the accreted mobile belt (e.g. Trompette, 2000). Then, there is evidence of a complex and polyphased history throughout the Palaeozoic (Fig. 2), with alternating periods of quiescence and tectonic activity, individualizing and rejuvenating ancient NS, NE–SW, or NW–SE structures in arch and basin configurations (Badalini et al., 2002; Bennacef et al., 1971; Beuf et al., 1968a, 1971; Boote et al., 1998; Boudjema, 1987; Chavand and Claracq, 1960; Coward and Ries, 2003; Craig et al., 2006; Eschard et al., 2005, 2010, Fabre, 1988, 2005; Frizon de Lamotte et al., 2013; Guiraud et al., 2005; Logan and Duddy, 1998; Lüning, 2005; Wendt et al., 2006). The Palaeozoic successions of the North Saharan Platform are predominantly composed of siliciclastic detrital sediments (Beuf et al., 1971; Eschard et al., 2005). They form the largest area of detrital sediments ever found on continental crust (Burke et al., 2003), dipping gently NNW (Beuf et al., 1969, 1971; Fabre, 1988, 2005; Fröhlich et al., 2010; Gariel et al., 1968; Le Heron et al., 2009). Carbonate deposits are observed from the Mid–Late Devonian to the Carboniferous (Wendt, 1995, 1985; Wendt et al., 2006, 1997, 1993; Wendt and Kaufmann, 1998). From south to north, the facies progressively evolve from continental fluviatile to shallow marine (i.e. upper to lower shoreface) and then to offshore facies (Beuf et al., 1971; Carr, 2002; Eschard et al., 2005, 2010, Fabre, 1988, 2005; Fekirine and Abdallah, 1998; Guiraud et al., 2005; Legrand, 1967a).

3 Data and methods
A multidisciplinary approach has been used in this study integrating new data in particular from the Ahnet and Mouydir basins (see supplementary data 1):

- Geographic Information System analysis (GIS);
- The basins and the main geological structures were identified from Landsat satellite images;
- Seismic section interpretation;
- Sedimentological and well-log analysis;
- Biostratigraphy and sequence stratigraphy;
- Geochronology and geophysical data.

The Paleozoic series of the Ahnet and Mouydir basins are well-exposed over an area of approximately 170,000 km² and are well observed in satellite images (Google Earth and Landsat from USGS). Furthermore, a significant geological database (i.e. wells, seismic records, field trips, geological reports) has been compiled in the course of petroleum exploration since the 1950s. The sedimentological dataset is based on the integration and analysis of cores, outcrops, well-logs, and of lithological and biostratigraphic data. Facies described from cores and outcrops of these studies were grouped into facies associations corresponding to the main depositional environments observed on the Saharan Platform (table 1). Characteristic gamma-ray patterns (electrofacies) are proposed to illustrate the different facies associations. The gamma-ray (GR) peaks are commonly interpreted as the maximum flooding surfaces (MFS) (e.g. Catuneanu et al., 2009; Galloway, 1989; Milton et al., 1990; Serra, 2009). Time calibration of well-logs and outcrops is based on palynomorphs (essentially Chitinozoans and spores), conodonts, goniatites, and brachiopods (Wendt et al., 2006).
Synsedimentary extensional and compressional markers are characterized in this structural framework based on the analyses of satellite images (Figs 4 and 5), seismic profiles (Fig. 6), 21 wells (W1 to W21), and 12 outcrop cross-sections (O1 to O12). Wells and outcrop sections are arranged into three E–W sections (Figs 9, 10 and supplementary data 2) and one N–S section (supplementary data 3). Satellite images (Figs 4 and 5) and seismic profiles (Fig. 6) are located at key areas (i.e. near arches) illustrating the relevant structures (Fig. 3). The calibration of the key stratigraphic horizon on seismic profiles (Figs 9 and 10) was settled by sonic well-log data using PETREL and OPENDTECT software. Nine key horizons easily extendable at the regional scale are identified and correspond to major unconformities: top Infra-Cambrian, top Ordovician, top Silurian, top Pragian, top Givetian, top mid-Frasnian, top Famennian, top Quaternary and top Hercynian unconformities (Figs 9 and 10). The stratigraphic layers are identified by the integration of satellite images (Google Earth and Landsat USGS: https://earthexplorer.usgs.gov/) and the 1:20,000 geological map of Algeria (Bennacef et al., 1974; Bensalah et al., 1971).

Subsidence analysis characterizes the vertical displacements of a given sedimentary depositional surface by tracking its subsidence and uplift history (Van Hinte, 1978). The resulting curve details the total subsidence history for a given stratigraphic column (Allen and Allen, 2005; Van Hinte, 1978). Backstripping is also used to restore the initial thicknesses of a sedimentary column (Allen and Allen, 2005; Angevine et al., 1990). Lithologies and paleobathymetries have been defined using facies analysis or literature data. Porosity and the compaction proxy are based on experimental data from Sclater and Christie, (1980). In this study, subsidence analyses were performed on sections using OSXBackstrip software performing 1D Airy backstripping (after Allen and Allen, 2005; Watts, 2001; available at: http://www.ux.uis.no/nestor/work/programs.html).

4 Structural framework and tectono-sedimentary structure analyses
The structural architecture of the North Saharan Platform (Fig. 1) is characterized by an association of syncline basins and anticlines (i.e. arches, domes, etc.). The basins (or sub-basins) are mostly circular to oval. They are bounded by arches which correspond to the mainly N–S Azzel-Matti, Arak-Foum Belrem, Amguid El Biod, and Tihemboka arches, the NE–SW Bou Bernous, Ahara, and Gargaf arches, and the NW–SE Saoura and Azzene arches (Fig. 1). The basins are structured by major faults frequently associated with broad asymmetrical folds displayed by three main trends (Fig. 1): (1) near-N–S, varying from N0° to N10° or N160°, (2) from N40° to 60°, and (3) N100° to N140° directions (Figs 1, 3A, and 4). These fault zones are about 100 km (e.g. faults F1 and F2, Fig. 4) to tens of kilometers lengths (e.g. faults F3 to F8, Fig. 4).

4.1 Synsedimentary extensional markers

Extensional markers are characterized by the settlement of steeply west- or eastward-dipping basement normal faults associated with colinear syndepositional folds of several kilometers in length (e.g. Fig. 5A–A’, 5B–B’, 5C–C’, 5E–E’ and 6A), represented by footwall anticline and hanging wall syncline-shaped forced folds. They are located in the vicinity of different arches (Fig. 3) such as the Tihemboka arch (Figs 4BB’, 5A–A’ and 5B–B’), Arak-Foum Belrem arch (Figs 4A–A’, 5C–C’ to 5F–F’ and 6A, 6C), Azzel Matti arch (Fig. 6B), and Bahar El Hamar area intra-basin arch (Fig. 6D). These tectonic structures can be featured by basement blind faults (e.g. fault F5 in Fig. 5C–C’, fault F1 in Figs 5D–D’, 5F–F’, and 6A). The deformation pattern is mainly characterized by brittle faulting in Cambrian–Ordovician series down to the basement and fault-damping in Silurian series (e.g. fault F2 in Fig. 5AA’, faults F1 to F6 in Fig. 6B). The other terms of the series (i.e. Silurian to Carboniferous) are usually affected by folding except (see F1 faults in Figs 5F–F’, 6B, 6D and 6C) where the brittle deformation can be propagated to the Upper Devonian (due to reactivation and/or inversion as suggested in the next paragraph).
In association with the extensional markers, thickness variations and tilted divergent onlaps of the sedimentary series (i.e. wedge-shaped units, progressive unconformities) in the hanging wall syncline of the fault escarpments are observed (Figs 5 and 6). These are attested using photogeological analysis of satellite images (Fig. 5) and are marked by a gentler dip angle of the stratification planes away from the fault plane (i.e. fault core zone). The markers of syndepositional deformation structures are visible in the hanging-wall synclines of Precambrian to Upper Devonian series (Figs 5 and 6).

The footwall anticline and hanging-wall syncline-shaped forced folds recognized in this study are very similar to those described in the literature by Schlische (1995), Withjack et al. (1990, 2002), Withjack and Callaway (2000), Khalil and McClay (2002), and Grasemann et al. (2005). The wedge-shaped units (DO0 to DO3; Figs 4, 5 and 6) associated with the hanging-wall synclines are interpreted as synsedimentary normal fault-related folding. The whole tectonic framework forms broad extensional horsts and graben related to synsedimentary forced folds controlling basin shape and sedimentation (Figs 4, 5 and 6).

Following Khalil and McClay (2002), Lewis et al. (2015), Shaw et al. (2005), and Withjack et al. (1990), we use the ages of the growth strata (i.e. wedge-shaped units) to determine the timing of the deformation. The main four wedge-shaped units identified (DO0 to DO3) are indicative of the activation and/or reactivation of the normal faults (extensional settings) during Neoproterozoic, Cambrian–Ordovician, Early to Mid Silurian and Mid to Late Devonian times.

In planar view, straight (F1 in Fig. 4A-A’) and sinuous faults (F2, F3, F3’, F4, F4’, and F5 in Fig. 4AA’) can be identified. The sinuous faults are arranged “en echelon” into several segments with relay ramps. These faults are 10 to several tens of kilometers long with vertical throws of hundreds of meters that fade rapidly toward the fault tips. The sinuous geometry of
normal undulated faults as well as the rapid lateral variation in fault throw are controlled by
the propagation and the linkage of growing parent and tip synsedimentary normal faults
(Marchal et al., 2003, 1998; Fig. 4A-A’).

According to Holbrook and Schumm (1999), river patterns are extremely sensitive to tectonic
structure activity. Here we find that the synsedimentary activity of the extensional structures
is also evidenced by the influence of the fault scarp on the distribution and orientation of
sinuous channelized sandstone body systems (dotted red lines in Fig. 4B-B’).

### 4.2 Synsedimentary compressional markers (inversion tectonics)

After the development of the extensional tectonism described previously, evidence of
synsedimentary compressional markers can be identified. These markers are located and
preferentially observable near the Arak-Foum Belrem arch (Fig. 5F-F’, F2 in Fig. 6C), the
Azzel Matti arch (2 in Figs 6B), and the Bahar El Hamar area intra-basin arch (2 in Fig. 6D).
The tectonic structures take the form of inverse faulting reactivating former basement faults
(F1’ in Fig. 5F-F’, F1 in Fig. 6C, F1’ in Fig. 6D, F1 in Fig. 6B). The synsedimentary inverse
faulting is demonstrated by the characterization of asymmetric anticlines especially
observable in satellite images and restricted to the fault footwalls (Figs 4A-A’ along F1-F2).

Landsat image analysis combined with the line drawing of certain seismic lines reveals
several thickness variations reflecting divergent onlaps (i.e. wedge-shaped units) which are
restricted to the hanging-wall asymmetric anticlines (2 in Figs 5F-F’, 6B, 6C and 6D). The
compressional synsedimentary markers clearly post-date extensional divergent onlaps at
hanging-wall syncline-shaped forced folds (1 in Figs 6B, 6C and 6D). This architecture is
very similar to classical positive inversion structures of former inherited normal faults
(Bellahsen and Daniel, 2005; Bonini et al., 2012; Buchanan and McClay, 1991; Ustaszewski
et al., 2005). Tectonic transport from the paleo-graben hanging-wall toward the paleo-horst
footwall (F1, F2-F2’, F4-F4’ in Fig. 6B; F1-F1’ in Fig. 6B) is evidenced. Further positive
tectonic inversion architecture is identified by tectonic transport from the paleo-horst footwall
to the paleo-graben hanging wall (F1-F1’ in Fig. 5F-F’; F1, F5, and F6 in Fig. 6C). This
second type of tectonic inversion is very similar to the transported fault models defined by
Butler (1989) and Madritsch et al. (2008). The local positive inversions of inherited normal
faults occurred during Silurian–Devonian (F4’ Fig. 6B) and Mid to Late Devonian times (Figs
6B, 6C and 6D). A late significant compression event between the end of the Carboniferous
and the Early Mesozoic was responsible for the exhumation and erosion of the tilted
Paleozoic series. This series is related to the Hercynian angular unconformity surface (Fig.
6B).

5 Stratigraphy and sedimentology

The whole sedimentary series described in the literature is composed of fluvial Cambrian
(Beuf et al., 1968a, 1968b, 1971; Eschard et al., 2005, 2010), glacial Ordovician (Beuf et al.,
1968a, 1968b, 1971; Eschard et al., 2005, 2010), argillaceous deep marine Silurian (Eschard
et al., 2005, 2010; Legrand, 1986, 1996; Lüning et al., 2000) and offshore to embayment
Carboniferous (Wendt et al., 2009) deposits. In this complete sedimentary succession, we
have focused on the Devonian deposits as they are very sensitive to and representative of
basin dynamics. The architecture of the Devonian deposits allows us to approximate the main
forcing factors controlling the sedimentary infilling of the basin and its synsedimentary
deformation. Nine facies associations organized into four depositional environments are
defined to reconstruct the architecture and the lateral and vertical sedimentary evolution of the
basins (Figs 9, 10, supplementary data 2 and 3).

5.1 Facies association, depositional environments, and erosional unconformities
Based on the compilation and synthesis of internal studies (Eschard et al., 1999), published papers on the Algerian platform (Beuf et al., 1971; Eschard et al., 2010, 2005; Henniche, 2002) and on the Ahnet and Mouydir basins (Biju-Duval et al., 1968; Wendt et al., 2006) plus the present study, eleven main facies associations (AF1 to AF5) and four depositional environments are proposed for the Devonian succession (Table 1). They are associated with their gamma-ray responses (Figs 7 and 8). They are organized into two continental/fluvial (AF1 to AF2), four transitional/coastal plain (AF3a to AF3d), three shoreface (AF4a to AF4c), and two offshore (AF5a to AF5b) sedimentary environments.

5.1.1 Continental fluvial environments

This depositional environment features the AF1 (fluvial) and the AF2 (flood plain) facies association (Table 1). Facies association AF1 is mainly characterized by a thinning-up sequence with a basal erosional surface and trough cross-bedded intraformational conglomerates with mud clast lag deposits, quartz pebbles, and imbricated grains (Table 1). It passes into medium to coarse trough cross-bedded sandstones, planar cross-bedded siltstones, and laminated shales. These deposits are associated with rare bioturbations (except at the surface of the sets), ironstones, phosphorites, corroded quartz grains, and phosphatized pebbles. Laterally, facies association AF2 is characterized by horizontally laminated and very poorly sorted silt to argillaceous fine sandstones. They contain frequent root traces, plant debris, well-developed paleosols, bioturbations, nodules, and ferruginous horizons. Current ripples and climbing ripples are associated in prograding thin sandy layers.

In AF1, the basal erosional reworking and high energy processes are characteristic of channel-filling of fluvial systems (Allen, 1983; Owen, 1995). Eschard et al. (1999) identify three fluvial systems (see A, B, and C in Fig. 8) in the Tassili-N-Ajjers outcrops: braided dominant (AF1a), meandering dominant (AF1b), and straight dominant (AF1c). They differentiate them...
by their different sinuosity, directions of accretion (lateral or frontal), the presence of mud drapes, bioturbations, and giant epsilon cross-bedding. Gamma-ray signatures of these facies associations (A, B, and C in Fig. 8) are cylindrical with an average value of 20 gAPI. The gamma ray shapes are largely representative of fluvial environments (Rider, 1996; Serra, 2009; Wagoner et al., 1990). The bottom is sharp with high value peaks and the tops are frequently fining-up, which may be associated with high values caused by argillaceous flood plain deposits and roots (Eschard et al., 1999). AF2 is interpreted as humid floodplain deposits (Allen, 1983; Owen, 1995) with crevasse splay s or preserved levees of fluvial channels (Eschard et al., 1999). Gamma-ray curves of AF2 (D, Fig. 8) show a rapid succession of low to very high peak values, ranging from 50 to 200 gAPI. AF1 and AF2 are typical of the Pragian “Oued Samene” Formation (Wendt et al., 2006). In the Illizi basin, these facies are mainly recorded in the Cambrian Ajiers Formation and the Lochkovian to Pragian “Middle Barre” and “Upper Barre” Formations (Beuf et al., 1971; Eschard et al., 2005).

5.1.2 Transitional coastal plain environments

This depositional environment comprises facies associations AF3a (delta/estuarine), AF3b (fluvial/tidal distributary channels), AF3c (tidal sand flat), AF3d (lagoon/mudflat) (table 1). AF3a is mainly dominated by sigmoidal cross-bedded heterolithic rocks with mud drapes. It is also characterized by fine to coarse, poorly sorted sandstones and siltstones often structured by combined flow ripples, flaser bedding, wavy bedding, and some rare planar bedding. Mud clasts, root traces, desiccation cracks, water escape features, and shale pebbles are common. The presence of epsilon bedding is attested, which is formed by lateral accretion of a river point bar (Allen, 1983). The bed surface sets are intensively bioturbated (Skolithos and Planolites) indicating a shallow marine subtidal setting (Pemberton and Frey, 1982). Faunas such as brachiopods, trilobites, tentaculites, and graptolites are present. AF3b exhibits a
fining-up sequence featured by a sharp erosional surface, trough cross-bedded, very coarse-grained, poorly sorted sandstone at the base and sigmoidal cross-bedding at the top (Figs 7 and 8). AF3c is formed by fine-grained to very coarse-grained sigmoidal cross-bedded heterolithic sandstones with multidirectional tidal bundles. They are also structured by lenticular, flaser bedding and occasional current and oscillation ripples with mud cracks. They reveal intense bioturbation composed of *Skolithos* (Sk), *Thalassinoides* (Th), and *Planolites* (Pl) ichnofacies indicating a shallow marine subtidal setting (Frey et al., 1990; Pemberton and Frey, 1982). AF4d is characterized by horizontally laminated mudstones associated with varicolored shales and fine-grained sandstones. They exhibit mud cracks, occasional wave ripples, and rare multidirectional current ripples. These sedimentary structures are poorly preserved because of intense bioturbation composed of *Skolithos* (Sk), *Thalassinoides* (Th), and *Planolites* (Pl). Fauna includes ammonoids (rare), goniatites, calymenids, pelecypod molds, and brachiopod coquinas.

In AF3a, both tidal and fluvial systems in the same facies association can be interpreted as an estuarine system (Dalrymple et al., 1992; Dalrymple and Choi, 2007). The gamma-ray signature is characterized by a convex bell shape with rapidly alternating low to high values (30 to 60 gAPI) due to the mud draping of the sets (see E Fig. 8). These forms of gamma ray are typical of fluvial-tidal influenced environments with upward-fining parasequences (Rider, 1996; Serra, 2009; Wagoner et al., 1990). AF3a is identified at the top of the Pragian “Oued Samene” Formation and in Famennian “Khenig” Formation (Wendt et al., 2006) in the Ahnet and Mounydir basins. In the Illizi basin, AF3a is mostly recorded at the top Cambrian of the Ajiers Formation, in the Lochkovian “Middle bar”, and at the top Pragian of the “Upper bar” Formation (Beuf et al., 1971; Eschard et al., 2005). The AF3b association can be characterized by a mixed fluvial and tidal dynamic based on criteria such as erosional basal contacts, fining-upward trends or heterolythic facies (Dalrymple et al., 1992; Dalrymple and
Choi, 2007). They are associated with abundant mud clasts, mud drapes, and bioturbation indicating tidal influences (Dalrymple et al., 2012, 1992; Dalrymple and Choi, 2007). The major difference with the estuarine facies association (AF3a) is the slight lateral extent of the channels which are only visible in outcrops (Eschard et al., 1999). The gamma-ray pattern is very similar to the estuarine electrofacies (see F Fig. 8). AF3c is interpreted as a tidal sandflat laterally present near a delta (Lessa and Masselink, 1995) and associated with an estuarine environment (Leuven et al., 2016). The gamma-ray signature (see G Fig. 8) is distinguishable by its concave funnel shape with alternating low and high peaks (25 to 60 gAPI) due to the heterogeneity of the deposits and rapid variations in the sand/shale ratio. These facies are observed in the “Tigillites Talus” Formation of the Illizi basin (Eschard et al., 2005). In AF4d, both ichnofacies and facies are indicative of tidal mudflat/lagoonal depositional environments (Dalrymple et al., 1992; Dalrymple and Choi, 2007; Frey et al., 1990). The gamma-ray signature has a distinctively high value (80 to 130 gAPI) and an erratic shape (see H Fig. 8). AF4d is observed in the “Atafaitafa” Formation and in the Emsian prograding shoreface sequence of the Illizi basin (Eschard et al., 2005). It is also recorded in the Lochkovian “Oued Samene” Formation and the Famennian “Khenig” Sandstones (Wendt et al., 2006).

5.1.3 Shoreface environments

This depositional environment is composed of AF4a (subtidal), AF4b (upper shoreface), and AF4c (lower shoreface) facies associations (Table 1). AF4a is characterized by the presence of brachiopods, crinoids, and diversified bioturbations, by the absence of emersion, and by the greater amplitude of the sets in a dominant mud lithology (Eschard et al., 1999). AF4b is heterolithic and composed of fine to medium-grained sandstones (brownish) interbedded with argillaceous siltstones and bioclastic carbonated sandstones. Sedimentary structures include oscillation ripples, swaley cross-bedding, flaser bedding, cross-bedding, convolute bedding, wavy bedding, and low-angle planar cross-stratification. Sediments were affected by
moderate to highly diversified bioturbation by *Skolithos* (Sk), *Cruziana*, *Planolites*, (Pl) *Chondrites* (Ch), *Teichichnus* (Te), *Spirophytons* (Sp) and are composed of ooids, crinoids, bryozoans, stromatoporoids, tabulate and rugose corals, pelagic styliolinids, neritic tentaculitids, and brachiopods. AF4c can be distinguished by a low sand/shale ratio, thick interbeds, abundant HCS, deep groove marks, slumping, and intense bioturbation (Table 1).

AF4a is interpreted as a lagoonal shoreface. The gamma-ray pattern (see I Fig. 8) is characterized by a concave bell shape influenced by a low sand/shale ratio with values fluctuating between 100 and 200 gAPI. AF4a is identified in the “Tigillites Talus” Formation and the Emsian sequence of the Illizi basin (Eschard et al., 2005) and in the Lochkovian “Oued Samene” Formation (Wendt et al., 2006). AF4b is interpreted as a shoreface environment. The presence of swaley cross-bedding produced by the amalgamation of storm beds (Dumas and Arnott, 2006) and other cross-stratified beds is indicative of upper shoreface environments (Loi et al., 2010). The gamma-ray pattern (see J and K Fig. 8) displays concave erratic egg shapes with a very regularly decreasing-upward trend and ranging from offshore shale with high values (80 to 60 gAPI) to clean sandstone with low values at the top (40 to 60 gAPI). AF4b is observed in the “Atafaitafa” Formation corresponding to the “Passage zone” Formation of the Illizi basin (Eschard et al., 2005). AF4c is interpreted as a lower shoreface environment (Dumas and Arnott, 2006; Suter, 2006). The gamma-ray pattern displays the same features as the upper shoreface deposits with lower values (i.e. muddier facies) ranging from 100 to 80 gAPI (see J and K Fig. 8).

5.1.4 Offshore marine environments

This depositional environment is composed of AF5a and AF5b facies associations (table 1). AF5a is mainly defined by wavy to planar-bedded heterolithic silty-shales interlayered with fine-grained sandstones. It also contains bundles of skeletal wackestones and calcareous
mudstones. The main sedimentary structures are lenticular sandstones, rare hummocky cross-bedding, mud mounds, low-angle cross-bedding, tempestite bedding, slumping, and deep groove marks. Sediments can present rare horizontal bioturbation such as Zoophycos (Z), Teichichnus (Te), and Planolites (Pl). AF5b is characterized by an association of black silty shales with occasional bituminous wackestones and packstones. It is composed of graptolites, goniatites, orthoconic nautiloids, pelagic pelecypods, limestone nodules, tentaculitids, ostracods, and rare fish remains. Rare bioturbation such as Zoophycos (Z) is visible.

In AF5a, the occurrence of HCS, the decrease in sand thickness and grain size together with the fossil traces indicate a deep marine environment under the influence of storms (Aigner, 1985; Reading, 2002). The gamma-ray pattern is serrated and erratic with values well grouped around high values from 120 to 140 gAPI (see L Fig. 8). Positive peaks may indicate siltstone to sandstone ripple beds. AF5b is interpreted as lower offshore deposits (Aigner, 1985; Stow and Piper, 1984; Stow et al., 2001). Here again the gamma-ray signature is serrated and erratic with values well grouped around 140 gAPI (see L Fig. 8). Hot shales with anoxic conditions are characterized by gamma-ray peaks (>140 gAPI). These gamma-ray patterns are typical of offshore environments dominated by shales (Rider, 1996; Serra, 2009; Wagoner et al., 1990). AF5a and AF5b are observed in the Silurian “Graptolites shales” Formation and the Emsian “Orsine” Formation of the Illizi basin (Beuf et al., 1971; Eschard et al., 2005; Legrand, 1996, 1986). The “Meden Yahia” and “Termutasset” Shales have the same facies (Wendt et al., 2006).

5.2 Sequential framework and unconformities

The high-resolution facies analysis, depositional environments, stacking patterns, and surface geometries observed in the Paleozoic succession reveal at least two different orders of depositional sequences (large and medium scale, Fig. 7) considered as
transgressive/regressive T/R (e.g. Catuneanu et al., 2009). The sequential framework proposed in Fig. 7B result from the integration of the vertical evolution the main surfaces (Fig. 7A) and the gamma-ray pattern (Fig. 8). The Devonian series under focus exhibits nine medium-scale sequences (D1 to D9, Fig. 7; Figs 9, 10, supplementary data 2 and 3) bounded by 10 major sequence boundaries (HD0 to HD9), and nine major flooding surfaces (MFS1 to MFS9). The correlation of the different sequences at the scale of the different basins and arches is used to build two E–W (Figs 9, 10, supplementary data 2) and one N–S (supplementary data 3) cross-sections.

The result of the analysis of the general pattern displayed by the successive sequences reveal two major patterns (Figs 9 and 10) limited by a major flooding surface MFS5. The first pattern extends from the Oued Samene to Adrar Morrat Formations and is dated from the Lochkovian to Givetian. D1 to D5 medium-scale sequences indicate a general proximal clastic depositional environment (dominated by fluvial to transitional and shoreface facies) with intensive lateral facies evolution. This first pattern is thin (from 500 m in the basin depocenter to 200 m around the basin rim) and with successive amalgamated surfaces on the edge of the arches between the “Passage zone” and “Oued Samene” Formations (Figs 5C–C’, 6A, 6C, 6D, and 9). It is delimited at the bottom by the HD0 surface corresponding to the Silurian/Devonian boundary. D1 to D3 are composed of T–R sequences with a first deepening transgressive trend indicative of a transition from continental to marine deposits bounded by a major MFS and evolving into a second shallowing trend from deep marine to shallow marine depositional environments. D1 to D3 thin progressively toward the edge and the continental deposits, in the central part of the basin, pass laterally into a major unconformity. The amalgamation of the surfaces and rapid lateral variations of facies between the Ahmet basin and Azzel Matti and Arak–Foum Behrem arches demonstrate a tectonic control related to the presence of subsiding basins and paleo-highs (i.e. arches).
D4 and D5 display the same T-R pattern with a reduced continental influence and upward decrease in lateral facies variations and thicknesses. The D5 sequence is mainly composed of shoreface carbonates. Evidence of mud mounds preferentially located along faults are well-documented in the area for that time (Wendt et al., 2006, 1997, 1993; Wendt and Kaufmann, 1998). This change in the general pattern indicates reduced tectonic influence.

MFS5, at the transition between the two main patterns, represents a major flooding surface on the platform and is featured worldwide by deposition of “hot shales” during the early Frasnian (Lüning et al., 2004, 2003; Wendt et al., 2006).

The second pattern extends from the “Meden Yahia”, “Temertasset” to “Khenig” Formations dated Frasnian to Lower Tournaisian. This pattern is composed of part of D5 to D9 medium-scale sequences. It corresponds to homogenous offshore depositional environments with no lateral facies variations. However, local deltaic (fluvio-marine) conditions are observed during the Frasnian at the Arak Foum Belrem arch (Fig. 10). A successive alternation of shoreface and offshore deposits is organized into five medium-scale sequences (part of D5, and D6 to D9; Figs 9 and 10). This pattern corresponds to the general maximum flooding (Lüning et al., 2003, 2004; Wendt et al., 2006) under eustatic control with no tectonic influences.

6 An association of low rate extensional subsidence and positive inversion pulses

The backstripping approach (Fig. 11) was applied to five wells (W1, W5, W7, W17, and W21). The morphology of the backstripped curve and subsidence rates can provide clues as to the nature of the sedimentary basin (Xie and Heller, 2006). In intracratonic basins, reconstructed tectonic subsidence curves are almost linear to gently exponential in shape, similar to those of passive margins and rifts (Xie and Heller, 2006). The compilation of tectonic backstripped curves from several wells in peri-Hoggar basins (Fig. 11A, see Fig. 1
for location) and from wells in the study area (Fig. 11B) display low rates of subsidence (from 5 to 50 m/Myr) organized in subsidence patterns of: Inversion of the Low Rate Subsidence (ILRS type c, red line, Fig. 11C), Deceleration of the Low Rate Subsidence (DLRS type b, black line), and Acceleration of the Low Rate Subsidence (ALRS type a, blue line).

Each period of ILRS, DLRS, and ALRS may be synchronous among the different wells studied (see B1 to J, Fig. 11B) and some wells of published data (see D to J Fig. 11A).

The Saharan Platform is marked by a rejuvenation of basement structures, around arches (Figs 2, 3, and 4), linked to regional geodynamic pulses during Neoproterozoic to Paleozoic times (Fig. 11). A compilation of the literature shows that the main geodynamic events are associated with discriminant association of subsidence patterns:

(A) Late Pan-African compression and collapse (patterns a, b, and c, A Fig. 11A). The Infra-Cambrian (i.e. top Neoproterozoic) is characterized by horst and graben architecture associated with wedge-shaped unit DO0 in the basement (Fig. 9 and 10). This structuring probably related to Pan-African post-orogenic collapse is illustrated by intracratonic basins infilled with volcano-sedimentary molasses series (Ahmed and Moussine-Pouchkine, 1987; Coward and Ries, 2003; Fabre et al., 1988; Oudra et al., 2005).

(B) Cambrian-Ordovician geodynamic pulse (Fig. 11A-B). Highlighted by the wedge-shaped units DO1 (Figs 5A-A’ and 6), the horst-graben system is correlated with deceleration (DLRS pattern a, B1) and with local acceleration of the subsidence (ALRS pattern b, B2). The Cambrian-Ordovician extension is documented on arches (Arak-Foum Belrem, Azzel Matti, Amguid El Biod, Tihemboka, Gargaf, Murizidié, Dor El Gussa, etc.) of the Saharan Platform by synsedimentary normal faults, reduced sedimentary successions (Bennacef et al., 1971; Beuf et al., 1971, 1968a, 1968b; Beuf and Montadert, 1962; Borocco and Nyssen, 1959; Claracq et al., 1958; Echikh, 1998; Eschard et al., 2010; Fabre, 1988; Ghienne et al., 2013,

(C) Late Ordovician geodynamic pulse (i.e. Hirnantian glacial and isostatic rebound; Fig. 11A-B). Late Ordovician incisions mainly situated at the hanging walls of normal faults (Fig. 6C and 6D) are interpreted as Hirnantian glacial valleys (Le Heron, 2010; Smart, 2000) and followed by local inversion of low rate subsidence (ILRS of type c, C in Fig. 11A).

(D) Silurian extensional geodynamic pulse (D, Figs 11A-B). The Silurian post-glaciation period is featured by the reactivation and sealing of the inherited horst and graben fault system (i.e. wedge-shaped unit DO2; Figs 5B-B’, 5C-C’, 6A and 6B). It is linked to an acceleration of the subsidence (ALRS of pattern b in Fig. 11A-B).

(E) Late Silurian geodynamic pulse (Caledonian compression; E Fig. 11A-B). Late Silurian times are marked by reactivation and local positive inversion of the former structures (Figs 5C-C’ and 6B); by truncations located at fold hinges (Figs 5C-C’ and 6); and by a major shift from marine to fluvial/transitional environments (Figs 9, 10 supplementary data 2 and 3). Backstripped curves register an inversion of the subsidence (ILRS of pattern c, in Fig. 11A-B). The Caledonian event is mentioned as related to large-scale folding or uplifted arches (e.g. the Gargaff, Tihemboka, Ahara, and Amguid El Biod arches) and it is associated with breaks in the series and with angular unconformities (Beuf et al., 1971; Biju-Duval et al., 1968; Boote et al., 1998; Boudjema, 1987; Boumendjel et al., 1988; Carruba et al., 2014; Chavand and Claracq, 1960; Coward and Ries, 2003; Dubois and Mazelet, 1964; Echikh, 1998; Eschard et al., 2010; Fekirine and Abdallah, 1998; Follot, 1950; Frizon de Lamotte et al., 2013; Ghienne et al., 2013; Gindre et al., 2012; Legrand, 1967b, 1967a; Magloire, 1967).

(F) Early Devonian tectonic quiescence (F Figs 11A-B). This is characterized by a deceleration of the low rate subsidence (DLRS of pattern a, F in Figs 11A-B).
(G) Middle to late Devonian geodynamic pulse (extension and local inversions, G Fig. 11A-B). The Mid to Late Devonian period is characterized by large wedge hiatuses and truncations associated with the reactivation of horst and graben structures and local positive inversion (OD3 in Figs 5D-D’, 6, 9, 10 supplementary data 2 and 3). This period is characterized by inversion and acceleration of low rate subsidence (patterns c and b: ILRS - ALRS, Fig. 11A-B). Some of the Middle to Late Devonian hiatuses (Early Eifelian) are noticed in the Ahnet basin (Hassan Kermadji et al., 2008, 2003; Kermadji, 2007; Kermadji et al., 2009; Wendt et al., 2006), in the Reggane (Jäger et al., 2009), on the Amguid Ridge (Wendt et al., 2006), and in the Illizi basin (Boudjema, 1987; Chaumeau et al., 1961).

(H to K) Pre-Hercynian to Hercynian geodynamic pulses (Fig. 11A-B). This period is organized in Early Carboniferous pre-Hercynian (H, Fig. 11A-B) to Late Carboniferous–Early Permian Hercynian (K, Fig. 11A-B) compressions limited by Mid Carboniferous tectonic quiescence (J, Fig. 11A-B). The Carboniferous period is characterized by a normal reactivation and local positive inversion of the previous structural patterns involving reverse faults, overturned folds, transpressional flower structures along strike-slip fault zones (Figs 3, 5F F’, 6B, 6C and 6D). The major Carboniferous tectonic event on the Saharan Platform impacted all arches and it is mainly controlled by near-vertical basement faults with a strike-slip component (Boote et al., 1998; Caby, 2003; Liégeois et al., 2003; Haddoum et al., 2001, 2013; Zazoun 2008; J. Wendt et al., 2009 Carruba et al., 2014). Two major hiatuses (i.e. Mid Tournaissian to Mid Visean–Serpukhovian) are recognized (Wendt et al., 2009b).

The geodynamic pulses attest to the reactivation of the terranes and associated lithospheric fault zones. This observation questions the nature of the Precambrian basement and associated structural heritage.
7 Precambrian structural heritage: accreted lithospheric terranes limited by vertical strike-slip mega shear zones

The 800 km² outcrop of basement rocks of the Hoggar shield provides an exceptional case of an exhumed mobile belt composed of accreted terranes of different ages. The Hoggar shield is composed of several accreted, sutured, and amalgamated terranes of various ages and compositions resulting from multiple phases of geodynamic events (Bertrand and Caby, 1978; Black et al., 1994; Caby, 2003; Liégeois et al., 2003).

To reconstruct the nature of the basement, a terrane map (Fig. 12) was put together by integrating geophysical data (aeromagnetic anomaly map: https://www.geomag.us/, Bouguer gravity anomaly map: http://bgi.omp.obs-mip.fr/), satellite images (7ETM+ from Landsat USGS: https://earthexplorer.usgs.gov/) data, geological maps (Berger et al., 2014; Bertrand and Caby, 1978; Black et al., 1994; Caby, 2003; Fezaa et al., 2010; Liégeois et al., 1994, 2003, 2005, 2013), and geochronological data (e.g. U-Pb radiochronology, see supplementary data 5). Geochronological data from published studies were compiled and georeferenced (Fig. 1). Thermo-tectonic ages were grouped into eight main thermo-orogenic events (Archean, Eburnean (i.e. Paleoproterozoic), Kibarian (i.e. Mesosproterozoic), Neoproterozoic oceanization-rifting, Neoproterozoic Pan-African orogeny, Caledonian orogeny, and Hercynian orogeny). Geochronological data show that the different terranes were reworked during several main thermo-orogenic events. Twenty-three well preserved terranes in the Hoggar were identified and grouped into Archean, Paleoproterozoic, and Mesoproterozoic–Neoproterozoic juvenile Pan-African terranes (see legend in Fig. 1). In the West African Craton, the Reguibat shield is composed of Archean terrains in the west and of Paleoproterozoic terranes in the east (Peucat et al., 2005, 2003). The two main events deduced from geochronological data are the Neoproterozoic (i.e. Pan-African) and Paleoproterozoic (i.e. Eburnean) episodes (Bertrand and Caby, 1978). Aeromagnetic anomaly surveys are
commonly used to analyze geological features such as rock types and fault zones (Gibson and Millegan, 1998; Schubert, 2007; Vacquier et al., 1951). In this study, these data highlight the geometries and the extension of the different terranes under the sedimentary cover. Four main domains can be identified from the aeromagnetic anomaly map, delimited by contrasted magnetic signatures and interpreted as suture zones (thick black lines, Fig. 12A). The study area is bounded to the south by the Tuareg Shield (TS), to the north, by the south Atlasic Range, to the west by the West African Craton (WAC) and at the east by the East Saharan Craton (ESC) or Saharan Metacraton (Abdelsalam et al., 2002).

The magnetic disturbance features (Fig. 12A) show three main magnetic trends. A major NS sinuous fabric and two minor sinuous 130–140°E and N45°E trends. The major NS lineaments coincide with terrane boundaries and mega-shear zones (e.g. 4°50’, 4°10’, WOSZ, EOSZ, 8°30’, RSZ shear zones; Fig. 1). Sigmoidal-shaped terranes 200 to 500 km long and 100 km wide are characterized (red lines in Fig. 12A). The whole assemblage forms a typical SC-shaped shear fabric (cf. Choukroune et al., 1987) associated with vertical mega-shear zones and suture zones (e.g. WOSZ, EOSZ, 4°10’, 4°50’ or 8°30’ Hoggar shear zones in Fig. 1). The SC fabrics combined with subvertical lithospheric shear zones are typical features of the Paleoproterozoic accretionary orogens (Cagnard et al., 2011; Chardon et al., 2009). This architecture is concordant with the Neoproterozoic collage of the Tuareg Shield (i.e. mobile belt) between the West African Craton and the East Saharan Craton (i.e. cratonic blocks) described by Coward and Ries, (2003) and Craig et al. (2006).

The gravimetric anomaly map (Fig. 12B) shows a correlation between gravimetric anomalies and tectonic architecture (intracratonic syncline-shaped basin and neighboring arches). Positive anomalies (> 66 mGal) are mainly associated with arches whereas negative anomalies are related to intracratonic basins (< 66 mGal). Nevertheless, negative anomaly disturbance is found in the Hoggar massif probably due to Cenozoic volcanism and the
Hoggar swell (Liégeois et al., 2005) or to Eocene Alpine intraplate lithospheric buckling (Rougier et al., 2013). Arches are linked to Archean to Paleoproterozoic continental terranes in contrast to syncline-shaped basins which are associated with Meso-Neoproterozoic terranes (Figs 1 and 12A-B-C).

8 Low subsidence rate intracratonic Paleozoic basins of the Central Sahara provide a basis for an integrated modeling study

Paleozoic intracratonic basins with similar characteristics (architecture, subsidence rate, stratigraphic partitioning, alternating episodes of intraplate extension and short duration compressions with periods of tectonic quiescence, etc.) have been documented in North America (e.g. Allen and Armitage, 2011; Beaumont et al., 1988; Burgess, 2008; Burgess et al., 1997; Eaton and Darbyshire, 2010; Pinet et al., 2013; Potter, 2006; Sloss, 1963; Xie and Heller, 2006), South America (e.g. Allen and Armitage, 2011; de Brito Neves et al., 1984; de Oliveira and Mohriak, 2003; Milani and Zalan, 1999; Soares et al., 1978; Zalan et al., 1990), Russia (e.g. Allen and Armitage, 2011; Nikishin et al., 1996) and Australia (e.g. Harris, 1994; Lindsay and Leven, 1996; Mory et al., 2017). However, the nature of the potential driving processes (lithospheric folding, far-field stresses, local increase in the geotherm, mechanical anisotropy from lithospheric rheological heterogeneity, etc.) associated with the formation of intracratonic Paleozoic basins remains highly speculative (e.g. Allen and Armitage, 2011; Armitage and Allen, 2010; Braun et al., 2014; Burgess and Gurnis, 1995; Burov and Cloetingh, 2009; Cacace and Scheck-Wenderoth, 2016; Célérier et al., 2005; Gac et al., 2013; Heine et al., 2008; Leeder, 1991; Vauchez et al., 1998).

The multiscale and multidisciplinary analysis performed in this study enable us to document a model of Paleozoic intracratonic Central Saharan basins coupling basin architecture and basement structures (Fig. 13). While we do not provide any quantitative explanations for the dynamics of these basins, our synthesis highlights that their subsidence is not the result of a
single process and we attempt here to make a check-list of the properties that a generic model of formation of such basins must capture:

(A) The association of syncline-shaped wide basins and neighboring arches (i.e. paleo-highs). The structural framework shows a close association of syncline-shaped basins, inter-basin principal to secondary arches, and intra-basin secondary arches (see Fig. 3).

(B) By local horst and graben architecture linked to steep-dipping planar normal faults and associated with normal fault-related fold structures (i.e. forced folds; a, Fig. 13A). Locally, the extensional structures are disrupted by positive inversion structures (b, Fig. 13A) or transported normal faults (c, Fig. 13A).

(C) A low rate of subsidence ranging between 5 to 50 m/Myr (Fig. 11).

(D) Long periods of extension and tectonic quiescence are interrupted by brief periods of compression or glaciation/deglaciation events (Beuf et al., 1971; Denis et al., 2007; Le Heron et al., 2006). These periods of compression are possibly related to intraplate compression linked to distal orogenies (i.e. Late Silurian Caledonian event, Late Carboniferous Hercynian, Frizon de Lamotte et al., 2013, Ziegler et al., 1995) or to intraplate arch uplift related to magmatism (Derder et al., 2016; Fabre, 2005; Frizon de Lamotte et al., 2013; Moreau et al., 1994).

(E) Synsedimentary divergent onlaps and local unconformities are identified from integrated seismic data, satellite images, and borehole data (Figs 4, 5, 6, 9 and 10). The periods of tectonic activity are characterized by normal to reverse reactivation of border faults, emplacement of wedge-shaped units, and erosional unconformities neighboring the arches (Figs 3, 4, 5, 6, 9, 10 and 13).

(F) The stratigraphic architecture displays a lateral facies variation and partitioning between distal marine facies infilling the intracratonic basins (i.e. offshore deposits) and proximal
amalgamated facies (i.e. fluvio-marine, shoreface) associated with prominent stratigraphic hiatus and erosional unconformities in the vicinity of the arches.

(G) A close connection is evidenced between the period of tectonic deformation and the presence of erosional unconformities (i.e. 2, 3, 6, 8, 10 geodynamic events in Fig. 13B). By contrast, the periods of tectonic quiescence and extension are characterized by low lateral facies variations, thin deposits, and the absence of erosional surfaces.

(H) The Precambrian heritage corresponds to Archean to Paleoproterozoic terranes identified in the Hoggar massif and reactivated during the Meso–Neoproterozoic Pan-African cycle (Fig. 1). The Precambrian lithospheric heterogeneity illustrated by the different characteristics of Precambrian terranes (wavelength, age, nature, fault zones) spatially control the emplacement of the syncline-shaped intracratonic basins underlain by Meso–Neoproterozoic oceanic terranes and the arches underlain by Archean to Paleoproterozoic continental terranes (Figs 1, 3 and 13). Many authors suggest control of the basement fabrics is inherited from the Pan-African orogeny in the Saharan basins (Beuf et al., 1968a, 1971; Boote et al., 1998; Carruba et al., 2014; Coward and Ries, 2003; Eschard et al., 2010; Guiraud et al., 2005; Sharata et al., 2015).

9 Conclusion

Our integrated approach using both geophysical (seismic, gravity, aeromagnetic, etc.) and geological (well, seismic, satellite images, etc.) data has enabled us to decrypt the characteristics of the intracratonic Paleozoic Saharan basins and the control of the heterogeneous lithospheric heritage of the horst and graben architecture, low rate subsidence, association of long-lived broad synclines and anticlines (i.e. arches swells, domes, highs or ridges) with very different wavelengths (λ) (tens to hundreds of kilometers). A coupled basin architecture and basement structures model is proposed.
This study highlights a tight control of the heterogeneous lithosphere over the structuring of the intracratonic Central Saharan basin. This particular type of basin is characterized by a low rate of subsidence and fault activation controlling the homogeneity of sedimentary facies and the distribution of the main unconformities. The low rate activation of vertical mega-shear zones bounding the intracratonic basin during Paleozoic times contrasts markedly with classic rift kinematics and architecture. Three different periods of tectonic compressional pulses, extension and quiescence are identified and controlled the sedimentary distribution. An understanding of tectono-sedimentary interaction is key to understanding the distribution of the Paleozoic petroleum reservoirs of this first-order oil province.

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Terrane names: Tassendjanet (Tas), Tassendjanet nappe (Tas n.), Ahnet (Ah), In Ouzzal Granulitic Unit (IOGU), Iforas Granulitic Unit (UGI), Kidal (Ki), Timétrine (Tim), Tilemsi (Til), Tirek (Tir), In Zaouatene (Za), In Teidini (It), Iskel (Isk), Tefedest (Te), Laouni (La), Azrou-n-Fad (Az), Egéré-Aleskod (Eg-Al), Serouenout (Se), Tazat (Ta), Issalane (Is), Assodé (As), Barghot (Ba), Tchilit (Tch), Aouzegueur (Ao), Edemb (Ed), Djanet (Dj); Shear zone and lineament names: Suture Zone East Saharan Craton (SZ ESC), West Ouzzal Shear Zone (WOSZ), East Ouzzal Shear Zone (EOSZ), Raghane Shear Zone (RSZ), Tin Amali Shear Zone (TASZ), 4°10’ Shear Zone, 4°50’ Shear Zone, 8°30’ Shear Zone.
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OTh: In Tahouite Fm (Early to Late Ordovician, Floian to Katian), OTj: Tamadjert Fm (Late Ordovician, Hirnantian), sIm: Imirhou Fm (Early Silurian), sdAs1: Asedjrad Fm 1 (Late Silurian to Early Devonian), dAs2: Asedjrad Fm 2 (Early Devonian, Lochkovian), dSa: Oued Samene Fm (Lower Devonian, Pragian). See Fig. 3 for map and cross-section location.
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<table>
<thead>
<tr>
<th>Code</th>
<th>Formation</th>
<th>Age</th>
<th>Description</th>
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</thead>
<tbody>
<tr>
<td>OTj</td>
<td>Tamadjert Fm</td>
<td>Late Ordovician, Hirnantian</td>
<td></td>
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<tr>
<td>slm</td>
<td>Imirhou Fm</td>
<td>Early to Mid Silurian</td>
<td></td>
</tr>
<tr>
<td>sdAt</td>
<td>Atafaïtafa Fm</td>
<td>Middle Silurian</td>
<td></td>
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<tr>
<td>dTf</td>
<td>Tifernine Fm</td>
<td>Middle Silurian</td>
<td></td>
</tr>
<tr>
<td>sdAs1</td>
<td>Asedjrad Fm 1</td>
<td>Late Silurian to Early Devonian</td>
<td></td>
</tr>
<tr>
<td>dAs2</td>
<td>Asedjrad Fm 2</td>
<td>Early Devonian, Lochkovian</td>
<td></td>
</tr>
<tr>
<td>dSa</td>
<td>Oued Samene Fm</td>
<td>Early Devonian, Pragian</td>
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<td>diag</td>
<td>Oued Samene shaly-sandstones Fm</td>
<td>Early Devonian, Pragian</td>
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<td>d2b</td>
<td>Givetian</td>
<td>Early Devonian, Frasnian</td>
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<td>d3a</td>
<td>Meden Yahia Fm</td>
<td>Late Devonian, Famennian</td>
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<td>d3b</td>
<td>Meden Yahia Fm</td>
<td>Late Devonian, Famennian</td>
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<td>Khenig sandstones</td>
<td>Late Famennian to early Tourmaisan</td>
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<tr>
<td>hTn2</td>
<td>late Tourmaisan</td>
<td>Early Visean</td>
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</tr>
<tr>
<td>hV1</td>
<td>early Visean</td>
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</tr>
</tbody>
</table>

See Figs. 1 and 3 for map and cross-section location.
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fault with a tectonic transport from footwall to hanging wall). See figure 3 for map and cross-
section location.
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<table>
<thead>
<tr>
<th>Facies associations</th>
<th>Textures/Lithology</th>
<th>Sedimentary structures</th>
<th>Biotaxonomic biotic grains</th>
<th>Ichnofacies</th>
<th>Depositional environments</th>
</tr>
</thead>
<tbody>
<tr>
<td>AF1</td>
<td>Conglomerates, mud to coarse sandstones, siltstones, shales</td>
<td>Trough cross-bedding, mud clasts, lag deposits, Bivalve and ammonite structures, interspersed pebbles, lenticular laminations, oblique stratification</td>
<td>Rare solitary intercalations, interbedded pebbles, sandstones, ironstones, phosphorites, corrodent quartz grains, calcareous matrix, brachiopod coquinas, phosphatized pebbles, hematite, anatase, quartz</td>
<td>Rare bioturbation</td>
<td>Fluvial</td>
</tr>
<tr>
<td>AF2</td>
<td>Silt to argillaceous fine sandstone</td>
<td>Current ripples, climbing ripples, enevase spay, root traces, palaeosols, plant debris</td>
<td>Nodules, ferrigenous horizon</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AF3a</td>
<td>Fine to coarse sandstones, argillaceous siltstones, shales (heterolithic)</td>
<td>Trough cross-bedding, some planar bedding, facet bedding, mud clasts, mud drapes, root trace, desiccation cracks, water escape, wavy bedding, shale pebble, sigmoidal cross-bedding</td>
<td>Brachiopods, trilobites, tentaculites graptolites</td>
<td>Bioturbation, Skolithos (Sk), Planolites (Pl)</td>
<td>Delta/Tidal estuarine channels</td>
</tr>
<tr>
<td>AF3b</td>
<td>Very coarse-grained poorly sorted sandstone</td>
<td>Trough cross-bedding, sigmoidal cross-bedding, abundant mud clasts and mud drapes</td>
<td>Increasing upward bioturbation Skolithos (Sk)</td>
<td></td>
<td>FLuvial/Tidal distributary channels</td>
</tr>
<tr>
<td>AF3c</td>
<td>Fine-grained to very coarse-grained heterolithic sandstone</td>
<td>Sigmoidal cross-bedding with multidirectional tidal bundles, wavy, lenticular, facet bedding, occasional current and oscillation ripples, occasional mud cracks</td>
<td>Absence of ammonoids, goniatites, calyptrodirids, pelecypod molds, brachiopods, echinoderm coquinas</td>
<td>Intense bioturbation, Skolithos (Sk), Planolites (Pl), Thalassinoides (Th)</td>
<td>Tidal sand-flat</td>
</tr>
<tr>
<td>AF3d</td>
<td>Mudstones, varicolored shales, thin sandstone layers</td>
<td>Occasional wave ripples, mud cracks, horizontal lamination, rare multi-directional ripples</td>
<td></td>
<td></td>
<td>Lagoon/Mudflat</td>
</tr>
<tr>
<td>AF3e</td>
<td>Silty mudstone associated with coarse to very coarse argillaceous sandstone, poorly sorted, heterolithic, silty mudstone</td>
<td>Sigmoidal cross-bedding, abundant mud clasts, wavy, lenticular cross-bedding and facet bedding, abundant current and oscillation ripples, mud drapes, root trace, desiccation cracks</td>
<td>Shell debris (echinoids, brachiopods)</td>
<td>Strongly bioturbated Skolithos (Sk), Planolites (Pl)</td>
<td>Subtidal</td>
</tr>
<tr>
<td>AF3f</td>
<td>Fine to very coarse sandstones interbedded with argillaceous siltstone and mudstone, bioclastic carbonates sandstones, brownish sandstones and clays, silts</td>
<td>Oscillation ripples, swaley cross-bedding, bidirectional bedding, facet bedding, rare hummocky cross-bedding, mud cracks (syneresis), convolute bedding, wavy, lenticular cross-bedding, combined flow ripples, planar cross low angle stratification, cross-bedding, ripple marks, centimeter bedding, shale pebbles</td>
<td>Ooids, crinoids, bryozoans, coral clasts, fossil debris, stromatoporoids, tabulates, colonial magnesium-calcium corals, spiny pelagic stylolites, nettic tectalitids, brachiopods, iron ostracods, abundant microfauna</td>
<td></td>
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</tr>
<tr>
<td>AF3g</td>
<td>Silty shales in fine sandstones (heterolithic)</td>
<td>Hummocky cross-bedding, planar bedding, combined flow ripples, convolute bedding, dish structures, mud drapes, remnant ripples, flat lenses, slumping</td>
<td>Intense bioturbation, Cruziana, Skolithos (Sk), Planolites (Pl), Thalassinoides (Th), Tentaculinus (T), Spirophyton (Sp), Diplocraterion (Dipl), Thalassinoides (Te), Chondrites (Ch), Regenetites (Ro), Climacichnus (CI)</td>
<td></td>
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<td>AF3h</td>
<td>Grey silty-shales, bundles of skeletal wackestones, silty shales, interlayers fine grained sandstones, calcareous mudstones, black shales, polychrome clays (black, brown, grey, green, red, pink), grey and reddish shales</td>
<td>Lenticular sandstones, rare hummocky cross-bedding, mud mounds, mud buildings, low-angle cross-bedding, temporary bedding, slumping, deep groove marks</td>
<td>Intensive burrowing, fluvial debris, horizontal burrows, skeletal remains (goniatites, orthocnemic nautiloids, pelagic pelecypod Buchia, anoxic conditions: lime nodule nodules, goniatites, Buchia, tenuicrurites, ostroconid and rare fish remains, Tornoceras, Aulacotomoceras, Lobatonoceras, Mantosiceras, Costamanticoceras and Virginoceras, graptolites</td>
<td>Zoophycos (Z), Thalassinoides (Te), Planolites (Pl)</td>
<td>Upper offshore</td>
</tr>
<tr>
<td>AF3i</td>
<td>Black silty-shales (mudstones), bituminous mudstones- wackestones, packstones</td>
<td>Rare structures</td>
<td>Parallel-aligned stylolites, goniatites, orthocnemic nautiloids, pelagic pelecypod Buchia, anoxic conditions: lime nodule nodules, goniatites, Buchia, tenuicrurites, ostroconid and rare fish remains, Tornoceras, Aulacotomoceras, Lobatonoceras, Mantosiceras, Costamanticoceras and Virginoceras, graptolites</td>
<td>Zoophycos (Z)</td>
<td>Lower offshore</td>
</tr>
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Table 1: Synthesis of facies associations (AF1 to AF5), depositional environments, and electrofacies in the Devonian series compiled from internal (Eschard et al., 1999) and published studies (Beuf et al., 1971; Biju-Duval et al., 1968; Wendt et al., 2006).
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Interpretation of the basement is based on Figs. 1, 3 and supplementary data 4. Outcrop location is in Fig. 3.
Figure 10: NE–W cross-section between the Reggane basin, Azzel Matti arch, Ahnet basin, Arak-Foum Belrem arch, Mouydir basin, and Amguid El Biod arch (well locations in fig. 3).

Well W18 biozone calibration is based on Kermandji et al. (2009): biozone (Tm) *tidikeltense microbaculatus* (Lochkovian, Lower Devonian), biozone (Es) *emsiensis spinaeformis* (Lochkovian–Pragian, Lower Devonian), biozone (Ac) *arenorugosa caperatus* (Pragian, Lower Devonian), biozone (Ps) *poligonalis subgranifer* (Pragian–Emsian, Lower Devonian), biozone (As) *annulatus svalbardiae* (Emsian, Lower Devonian), biozone (Mp) *microancyreus protea* (Emsian–Eifelian, Lower to Middle Devonian), biozone (Vl) *velatus langii* (Eifelian, Middle Devonian). Well W19 and W20 biozones calibration from internal reports (Abdesselam-Rouighi, 1991; Khiar, 1974) is based on Magloire’s (1967) classification: biozone H (Pridoli, Upper Silurian), biozone I (Lochkovian, Lower Devonian), biozone J (Pragian, Lower Devonian), biozone K (Emsian, Lower Devonian), biozone L1–5 (Middle Devonian to Upper Devonian). Interpretation of the basement is based on Figs. 1, 3 and supplementary data 4. Outcrop location is in Fig. 3.
Solid Earth Discuss., https://doi.org/10.5194/se-2018-50
Manuscript under review for journal Solid Earth
Discussion started: 27 June 2018
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Figure 11: (A) Tectonic backstripped curves of the Paleozoic North Saharan Platform (peri-Hoggar basins) compiled from literature: 1: HAD-1 well in Ghadamès basin (Makhous and Galushkin, 2003a); 2: Well RPL-101 in Reggane basin (Makhous and Galushkin, 2003a); 3: L1-1 well in Murzuq basin (Galushkin and Eloghbi, 2014); 4: TGE-1 in Illizi basin (Makhous and Galushkin, 2003b); 5: REG-1 in Timimoun basin (Makhous and Galushkin, 2003a); 6: Ghadamès-Berkine basin (Allen and Armitage, 2011; Yahi, 1999); 7: well in Sbœa basin (Tournier, 2010); 8: well B1NC43 in Al Kufrah basin (Holt et al., 2010). (B) Tectonic backstripped curves of wells in the study area: 1: well W17 in Ahnet basin; 2: well W5 in Ahnet basin; 3: well W7 in Ahnet basin; 4: well W21 in Mouydir basin; 5: well W1 in Reggane basin; (C) The data show low rate subsidence with periods of deceleration (Deceleration of Low Rate Subsidence: DLRS), acceleration (Acceleration of Low Rate Subsidence: ALRS), or inversion (Inversion of Low Rate Subsidence: ILRS) synchronous and correlated with regional tectonic pulses (i.e. major geodynamic events). A: Late Pan-African compression and collapse (type a, b, and c subsidence), B: Undifferentiated Cambrian–Ordovician (type a, b, and c subsidence), B1: Cambrian–Ordovician tectonic quiescence (type a subsidence), B2: Cambrian–Ordovician extension (type b subsidence), C: Late Ordovician glacial and isostatic rebound (type c subsidence), D: Silurian extension (type b subsidence), E: Late Silurian Caledonian compression (type c subsidence), F: Early Devonian tectonic quiescence (type a subsidence), G-H: Middle to late Devonian extension with local compression (i.e. inversion structures, type b and c subsidence), I: Early Carboniferous extension with local tectonic pre-Hercynian compression (type c and b subsidence), J: Middle Carboniferous tectonic extension (type b subsidence), K: Late Carboniferous–Early Permian Hercynian compression (type c subsidence).
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