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# Stratigraphy and tectono-sedimentary processes of allochthonous and autochthonous Devonian deposits of the Tisdafine Basin, Eastern Anti-Atlas, Morocco

## *Stratigraphie et processus tectono-sédimentaires des dépôts dévoniens allochtones et autochtones du bassin de Tisdafine, Anti-Atlas oriental, Maroc*

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**Abstract.** The Tisdafine Basin lies in the Eastern Anti-Atlas (southern Morocco), north of the Proterozoic Saghro-Ougnat massifs, in the Gondwana craton-western Variscid Meseta contact zone. Despite this crucial structural position, its allochthonous and autochthonous Devonian successions have been poorly studied. As an interdisciplinary tectono-sedimentary approach, field investigation, high-resolution biostratigraphy, microfacies analysis, clay mineralogy, and investigations of syn- and postsedimentary tectonic movements were used to reconstruct the Devonian basin history. Results from the wider Tinejdad region are used for comparisons with the Devonian south and east of the Proterozoic belt (northern Maïder and Tafilalt). In the western Bou Tisdafine region, the lower Emsian olistolite which is embedded in the upper Tournaisian-Visean Aït Yalla Formation remnants a shallow pelagic carbonate platform developed at the southern basin margin. A second olistolite preserved the transformation from Eifelian neritic trilobite limestone, resembling the northern Maïder region, to a condensed, mostly pelagic Givetian to middle Famennian platform. Breccia marker beds testify to the widespread Givetian/basal Frasnian synsedimentary tectonics from the Meseta to Eastern Anti-Atlas. During the FrasnianFamennian/boundary, the Upper Kellwasser Event interval is typically developed as dark goniatite limestone, followed after a sort gap by the occurence of Famennian griotte facies. The Bou Tisdafine Upper Devonian becames separated from the southern parts of the Eastern Anti-Atlas by the emerged Ougnat massif. In the eastern part of the basin, the Oued Ferkla Devonian is an isolated, large glide block derived from a western continuation of the Tafilalt Platform. Its fluctuating upper Emsian to middle Givetian pelagic facies records platform pro- and retrogradation phases, controlled by sea-level changes known from the Tafilalt, with a fine representation of the global Daleje, Choteč, Kačák, and *Pumilio* Events. Eifelian synsedimentary tectonic pattern correlates with the contemporaneous crustal disintegration of the whole region. The change from upper Givetian to middle Frasnian thick marls suggests a southern basin extension during the widespread tectonism of that interval. The adjacent Koudiat Inegh succession represents the true facies of the eastern Tisdafine Basin. It is characterized by black shales and distal turbiditic limestones, with a very strong diagenetic overprint (common late dolomitization) and a deformation style (with cleavage), which differs strictly from the contemporaneous beds of the glide blocks. The transition from platform to basin occurred close the base of the Eifelian, as part of the overall structural differentiation. Basinal facies persisted in the Upper Devonian. Unfortunately, the early (Emsian-middle Famennian) and main (upper Tournaisian-Visean) phases of the basin and its southern margin are separated by a gap of outcrop and strata.

**Keywords.** Anti-Atlas, Devonian, stratigraphy, facies, synsedimentary tectonics.

**Résumé.** Le bassin de Tisdafine se situe dans l'Anti-Atlas oriental (sud du Maroc), au nord de l'axe protérozoïque Jebel Sarhro-Jebel Ougnat, à la transition entre le craton stable précambrien du Gondwana et la Meseta occidentale hercynienne. En dépit de cette structuration, les successions dévoniennes allochtones et autochtones présentes dans ce bassin, ont été peu étudiées. Ce travail tente de reconstituer l'histoire du bassin au Dévonien dans un cadre d'approche tectono-sédimentaire pluridisciplinaire à partir de levées de terrain, de la biostratigraphie à haute résolution, de l'analyse des microfaciès, de la minéralogie de l'argile et de l'analyse des mouvements tectoniques syn- et post-sédimentaires. Les résultats obtenus au niveau de la région de Tinejdad sont utilisés pour comparer entre le Dévonien au sud et à l'est de la ceinture protérozoïque (au nord du Maïder et du Tafilalt). Dans la partie occidentale, à Bou Tisdafine, une première olistolite de l'Emsien inférieur emballée dans la Formation Aït Yalla, du Tournaisien supérieur-Viséen, constitue le témoin d'une plateforme carbonatée pélagique peu profonde développée à la marge sud du bassin. Une seconde olistolite plus à l'est, montre le caractère d'une plateforme carbonatée de type condensé, comme en témoigne la transformation du calcaire à trilobites néritique de l'Eifélien. Cette plateforme ressemble à celle située au nord de Maïder, essentiellement pélagique, d'âge Givétien à Famennien moyen. Des niveaux bréchiques représentent les témoins d'une tectonique synsédimentaire généralisée à l'échelle de la Meseta et de l'Anti-Atlas oriental au passage Givétien-Frasnien inférieur. L'intervalle de l'Événement Kellwasser supérieur est signalé à la limite Frasnien/Famennien, manifesté sous la forme typique d'un calcaire à Goniatites sombres, suivi par un faciès à griotte du Famennien. Le Dévonien supérieur de Bou Tisdafine est séparé des parties méridionales de l'Anti-Atlas oriental par le Haut Ougnat émergent. Dans la partie orientale du bassin, le Dévonien de l'Oued Ferkla constitue une unité en bloc de glissement isolé, dérivé d'un continuum occidental de la plateforme du Tafilalt. Son faciès pélagique, fluctuant de l'Emsien supérieur au Givétien moyen, enregistre les phases de pro- et de rétrogradation de la plateforme, contrôlées par les variations du niveau de la mer, connues dans le Tafilalt, révélant ainsi de manière fine les Evénements globaux de Daleje, Choteč, Kačák et Pumilio. La tectonique synsédimentaire de l'Eifélien serait à l'origine de la dislocation de la croûte à l'échelle de toute la région. Les variations des épaisseurs des marnes au passage Givétien supérieur-Frasnien moyen, suggèrent une extension du bassin méridional en relation avec les mouvements tectoniques que connaît cet intervalle. La série adjacente de Koudiat Inegh représente le véritable faciès oriental du bassin de Tisdafine. Elle est caractérisée par des schistes noirs et des calcaires turbiditiques distaux, avec une très forte empreinte diagénétique (dolomitisation tardive) et un style de déformation (avec clivage) qui se distingue nettement des bancs issus des blocs basculés. La transition d'un milieu de plateforme vers le bassin s'est produite à la base de l'Eifélien, dans le cadre d'une différenciation structurale globale avec une expression de faciès qui se maintient globalement jusqu'au Dévonien supérieur. Une lacune au niveau des affleurements sépare le bassin de sa marge sud, entre la phase précoce hercynienne (Emsien-Famennien moyen) et la phase majeure (Tournaisien supérieur-Viséen).

**Mots clés :** Anti-Atlas, Dévonien, stratigraphie, faciès, tectonique synsédimentaire.

## INTRODUCTION

The Palaeozoic Tisdafine Basin is situated in southern Morocco, roughly between Tinerhir (= Tinghir) in the west and Tinejdad in the east (Figs. 1, 2), extending from there to the northeast (Touroug region). It belongs to the Palaeozoic cover of the Eastern Anti-Atlas of Morocco. It is bounded to the north by the South Atlas Fault (S.A.F, Fig. 1) and in the south by the South Meseta Fault Zone (S.M.F.Z.). It occupies an important transitional position between the Variscan Moroccan Meseta and the stable cratonic successions of the Anti-Atlas, whose continuous sedimentary history began in the Neoproterozoic. The main questions addressed in this work are improvements of the regional litho-, bio- and event stratigraphy, sedimentology, and the role of synsedimentary tectonic activity that associated the Devonian sedimentation in the Eastern Anti-Atlas region in north of the Proterozoic Saghro- Ougnat massifs (Fig. 2). Although the structural aspects have already been raised on the Moroccan Palaeozoic by several geologists (e.g. Hollard 1960, Michard *et al.* 1982, Montenat *et al.* 1996, Belfoul *et al.* 2002, Hoepffner et al. 2005, Burkhard *et al.* 2006, Robert-Charrue 2006, Baidder *et al.* 2008, 2016, Rytina *et al.* 2013, Raadi 2014, Hejja *et al.* 2020), the sequence and timing of tectonic events and palaeogeographic changes remain to be refined by better stratigraphic data. Therefore, we have initiated new field-based research in the Tisdafine Basin.

The present manuscript provides the first results of studies in the wider Tinejdad region (Fig. 2). This is a poorly known area, where Devonian outcrops, were first reported by Clariond and Termier (1933), are small compared to the main Anti-Atlas. Apart from preliminary data in Ward *et al.* (2013) and the unpublished studie of Hejja (2013), these outcrops have not been studied in detail. A new sampling of several sections enables a better understanding of the eastern Tisdafine Basin and a correlation with the neighboring basins. Biostratigraphic marker fossils, especially conodonts and ammonoids were determined and the age of strata was specified with high resolution. Furthermore, lithostratigraphic and sedimentological analysis, based on section logging and sediment petrography (microfacies), enable us to place the basin in its geodynamic context. The focus is on the evolution of the depositional environments in time and space, from carbonate platform to siliciclastic basins, and on the structural dynamics responsible for the dislocation of this platform and its dismantling during the opening of the Tisdafine Basin.

## REGIONAL GEOLOGICAL BACKGROUND

The Palaeozoic deposits of the eastern Anti-Atlas belt consist essentially of sedimentary rocks that cover the volcanic and volcano-sedimentary outcrops of the Neoproterozoic Saghro and Ouarzazate groups. The Paleozoic series of the Tisdafine Basin are tabular and slightly folded in contact with shear and thrust zones but there are complex nappe structures in the Tinerhir area to the west (Cerrina Feroni *et al.* 2010). The sediments portray mostly an outer shelf environment, but locally there are shallow-water limestones, sometimes with reefal corals (e.g., in Givetian blocks of Taourirt n'Khellil, Rytina *et al.* 2013).

The Palaeozoic series of the Anti-Atlas south of the Jebel Saghro-Ougnat axis (Tafilalt and Maïder) is generally similar along with its extent. Successions of the Cambrian to Lower Carboniferous include only marine facies although there is evidence for structural highs (islands) characterized by erosion and non-deposition. During the Ordovician, the sedimentation is exclusively detrital; it forms an extensive series up to the upper Ashgill (now Hirnantian), which is marked by a significant sea level fall, interpreted as reflecting the glaciation of the end-Ordovician (Clerc *et al.* 2013, Colmenar & Alvaro 2014, Ghienne *et al.* 2014). The marine lower Silurian was deposited regionally on a perfect subhorizontal peneplain. The upper Llandovery (Telychian) is characterized by a generally high amount of siliceous material, marked by phtanite horizons and sandy shales with graptolites (Willefert in Destombes *et al.* 1985). In contrast, the Devonian successions show strong overall subsidence, with diversified lithofacies that testify to variable palaeo-environments located in the middle of the eastern Anti-Atlas Platform. After the still rather uniform Emsian, a subdivision into adjacent shelf basins and platforms (e.g. Wendt *et al.* 1984, Wendt 1989, 2021) arised from dense block faulting (e.g. Raddi *et al.* 2007, Baidder *et al.* 2008). The Devonian platform deposits are broadly condensed carbonate (Hollard 1981), grading along gentle or steep slopes into thick, often anoxic shales and marls. Subordinate reefal limestones and some large mudmounds are known from the northern and southern Maïder and from the eastern and southern Tafilalt. Devonian volcanism is restricted to the famous Hamar Laghdad (Lakhdad) area of the eastern Tafilalt Platform (e.g. Brachert *et al.* 1992, Montenat *et al.* 1996, Oumejjoud Formation of Becker *et al.* 2018). The Carboniferous of the Tisdafine Basin begins with olistostromes (Michard *et al.* 1982, Rytina *et al.* 2013), which precede the upper Tournaisian to Visean flysch of Jbel Tisdafine (Hindermeyer 1954, 1955, Michard *et al.* 1982, Destombes *et al.* 1983, Graham & Sevastopulo 2008, supposed deltaic succession of Soualhine *et al.* 2003). High subsidence rates with dominant fine-grained siliciclastics also characterize the Lower Carboniferous of the Maïder and Tafilalt (e.g., Klug *et al.* 2006, Kaiser *et al.* 2011). However, there are distinctive calcareous mudmound levels in the middle/upper Visean (Wendt *et al.* 2001).

The Devonian outcrops of the central to eastern Tisdafine Basin form a NE-SW striking, discontinuous stripe north/northeast of the Jebel Ougnat (Fig. 3). It involves formations with different lithologies of the Emsian to Famennian age. Each succession represents an individual lithostratigraphic sequence with particular sedimentary deposits and environments. Where the Lower Carboniferous does not cover the Devonian, the landscape reveals, in some places, depending on the degree of erosion, tectonically dismembered Devonian units (Koudiat Inegh, Fig. 3). The studied two western outcrops (Bou Tisdafine-West and Bou Tisdafine) lie isolated as large olistolites within the poorly exposed Lower Carboniferous siliciclastics of the Aït Yalla Formation (El Boukhari *et al.* 2007, Dal Piaz *et al.* 2007). The outcrops northeast of Tinejdad belong to the isolated, folded, larger-sized Devonian units that are not in contact with the Carboniferous of the Jebel Tisdafine or Jebel Azguine (Fig. 3).

## SECTION DESCRIPTIONS

### Section Bou Tisdafine-West

The section Bou Tisdafine-West lies ca. 6 km east of Taourirt n'Khellil (see Dal Piaz *et al.* 2007 and Fig. 3), at GPS: N31° 26' 34.88"; W5° 19' 11.42". It is accessible ca. hundred meters to the south of national route N°10, which links Tinejdad and Tinghir. The section was sampled and logged in 2011 and 2012 by R.T. Becker and Marie-Kristin Rytina (see Rytina 2013) and noted as locality Ol2 in Rytina *et al.* (2013). The outcrop consists of an oval-shaped, ca. 45 m wide lenticular olistolite that is completely surrounded, without fault evidence, by thin-bedded, greenish-grey siltstones and platy, dark brownish weathering fine sandstones of the Carboniferous Ait Yalla Formation (Fig.

4A). There is no evidence for faulting. The exposed Devonian forms a very low hill in the plain, with beds dipping uniformly with ca. 45° to the north. The ca. 7.5 m thick succession is characterized by alternating middle-grey, detrital, partly crinoidal limestones and poorly exposed, greenish-grey marl, often with crinoidal limestone lenses (Fig. 4B). A few adjacent small blocks with a similar lithology were detached from the main block during initial transport or subsequent erosion. There is only poor macrofauna (some gastropod cross-sections) but in thin-section and microfossil samples, limestone beds are fossiliferous with a fauna of brachiopods, crinoids, dacryoconarids, orthocones, and conodonts. Their microfacies reveal several types: mudstones with a more or less homogeneous micritic matrix and early diagenetic pyrite, strongly bioturbated dacryoconarid wacke-packstone with abundant crinoid debris, and dacryoconarid grainstones with bimodal current orientation and erosive bases. In addition, there was some primary porosity corresponding to clearly visible, sparite-filled inter-granular spaces. The fauna and microfacies indicate a subphotic shallow pelagic carbonate platform or ramp with calm deposition interrupted by episodic bottom current events.

The base of the succession (Sample AI OL2 Base from Bed 1) did not yield conodonts. Sample AI Ol2 P2, from slightly higher in the first sequence of crinoidal limestones (beds 3-5), provided *Caudicriocus curvicauda*, dominant *Caudicriocus celtibericus,* and two *Ctenopolygnathus pireneae*. These three species characterize the upper Pragian *Caudicriocus celtibericus* Zone of classical (Bohemian) chronostratigraphy (Slavík 2004, Slavík *et al.* 2007, Aboussalam *et al.* 2015), which roughly correlates with the basal Emsian sensu the Kitab (Uzbekistan) GSSP section. The lower crinoidal limestones at Bou Tisdafine West are equivalent of the "Pragian Limestone" of the Tafilalt and of the Ihandar Formation in the northern Maïder.

A sample from the upper part of the section (AI OL2 P9) contained *Criteriognathus miae, Criteriognathus steinhornensis,* and *Latericriodus bilatericrescens bilatericrescens*. This fauna falls into the *Criteriognathus steinhornensis* Zone in the middle part of the lower Emsian (Aboussalam *et al.* 2015). At the top (Bed 17, Sample AI Ol2 Top), we found a moderately rich but monotypic *Criteriognathus steinhornensis* fauna. Therefore, the upper part can be correlated with the *Anetoceras* and *Mimagoniatites* Limestones (Units I/J sensu Becker *et al.* 2013) of the upper Seheb El Rhassel Formation of the Tafilalt (e.g. Aboussalam *et al.* 2015, Hartenfels *et al.* 2018) and with the Upper Member of the Bou Tiskaouine Formation of the Maïder (Hollard 1981). However, the characteristic early ammonoids of these two regions do not occur at Bou Tisdafine-West.

**Section Bou Tisdafine**

The Bou Tisdafine section is located 22 km west of Tinejdad and 10 km from Taourirt n'Khellil (GPS: N31°27'12,53"; W5°14'13,17"). It can be accessed by a small track to the south from the national road N10 connecting Tinejdad and Tinghir; the distance to the main road is very short (20 m). The outcrop is highly condensed since it spans only about forty meters from the Eifelian to the middle Famennian (Fig. 5). It was first noted by Le Maitre (1947) and Hindermeyer (1955: recognition of upper Emsian, Eifelian, and Frasnian goniatites), and much later mentioned by Dal Piaz *et al.* (2007) and Becker & El Hassani (2020). The outcrop lithofacies reflects distinctive changes in sedimentary environments and phases of the palaeogeographic evolution. Its lower part has been exploited by the local "trilobite industry", creating two trenches in trilobite-rich Eifelian beds. This section is possibly the type-locality of the lichid *Basseiarges mellishae* (Corbacho & López-Soriano 2013) but their given locality information is rather vague. More recently, Eifelian phacopids and scutelluids from the trenches were described by Gottlob (2020) but the full faunal spectrum has not yet been established.

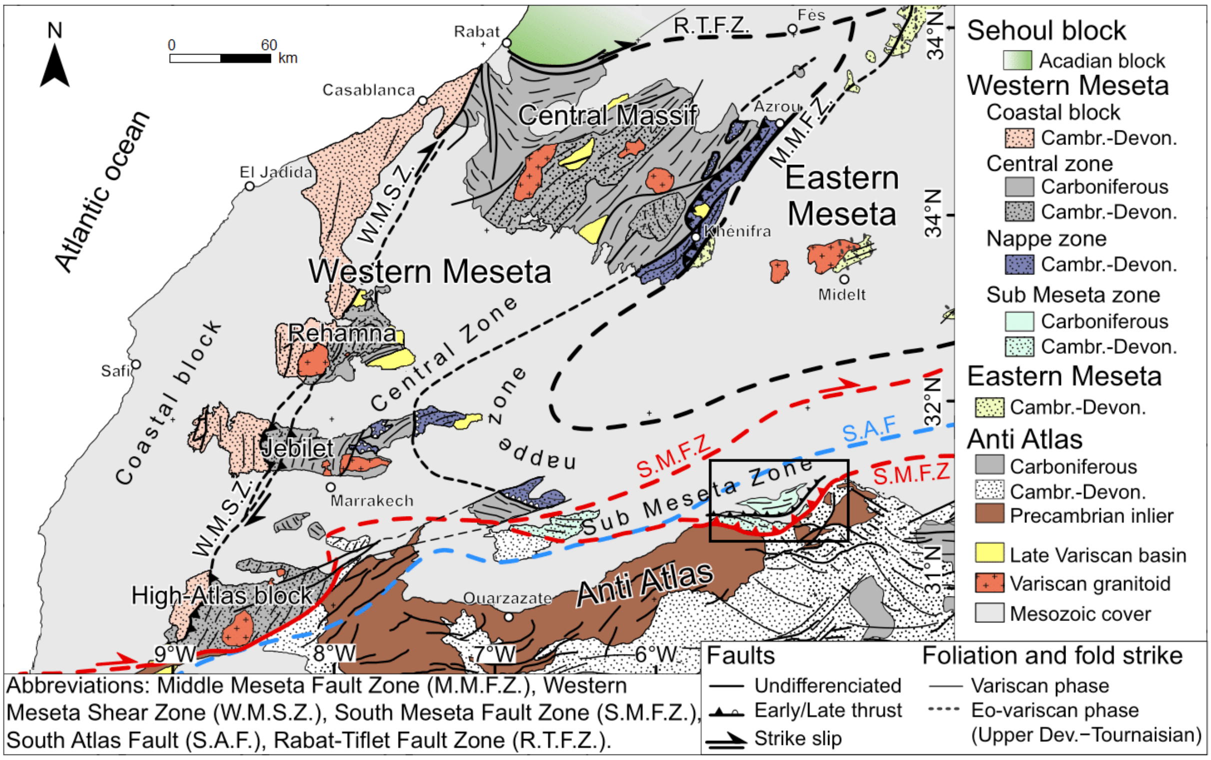


Figure 1. Map of the Moroccan Variscides (after Chopin *et al.* 2014, Hoepffner *et al.* 2005, and Michard *et al.* 2008, 2010), with the location of the Tisdafine Basin (frame = ca. Fig. 3).

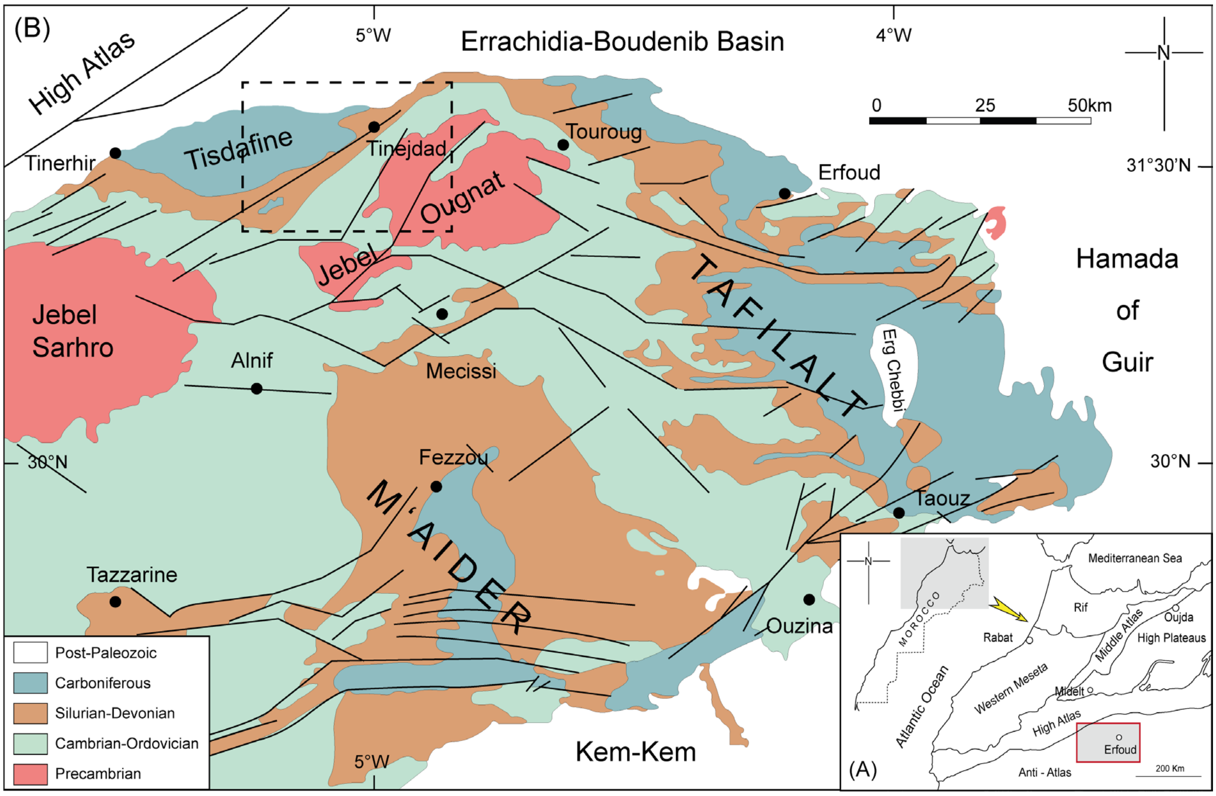


Figure 2. Synthetic geological map of the eastern Anti-Atlas (after Michard *et al.* 2008, Baidder *et al.* 2008, Baidder *et al.* 2016), with the location of Tinejdad region (frame = Fig. 3).

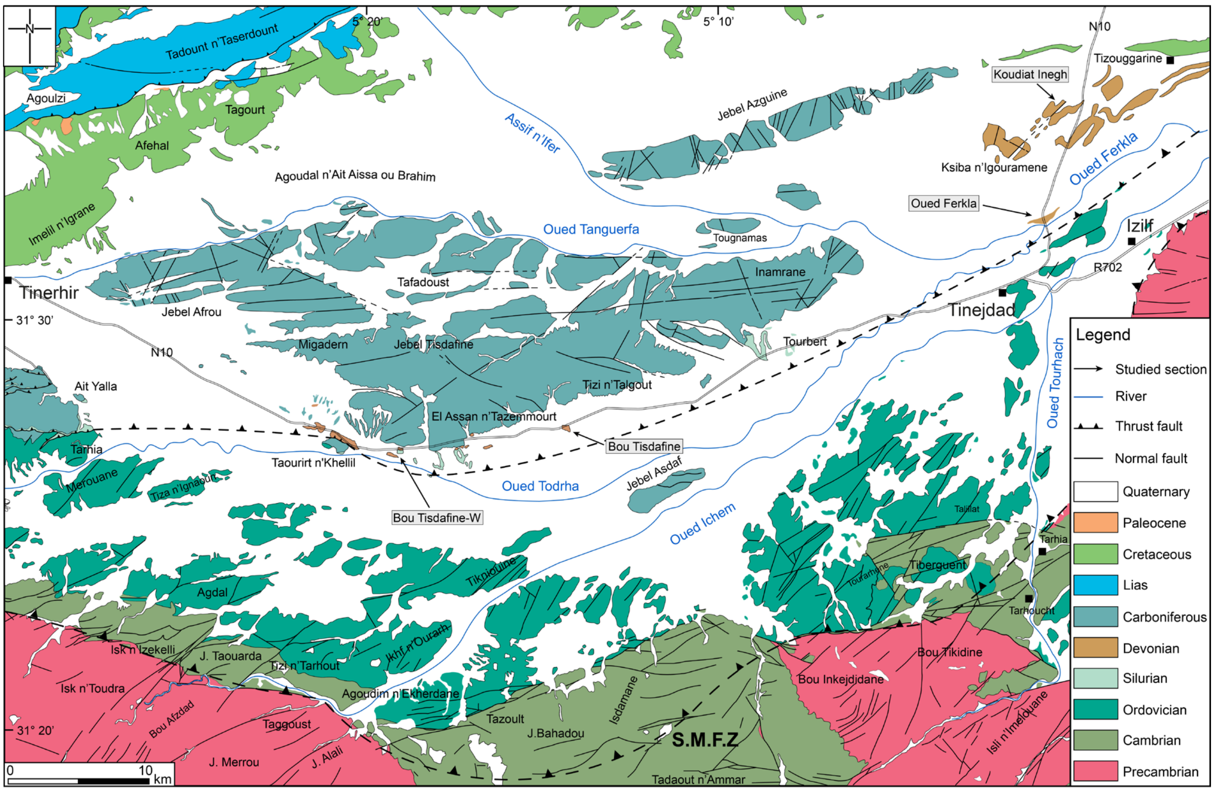


Figure 3. Geological map of the northern border of the eastern Anti-Atlas with the location of the studied sections (from geological map of Morocco, 1/200.000, Todrha-Ma´der, Destombes & Hollard 1988).

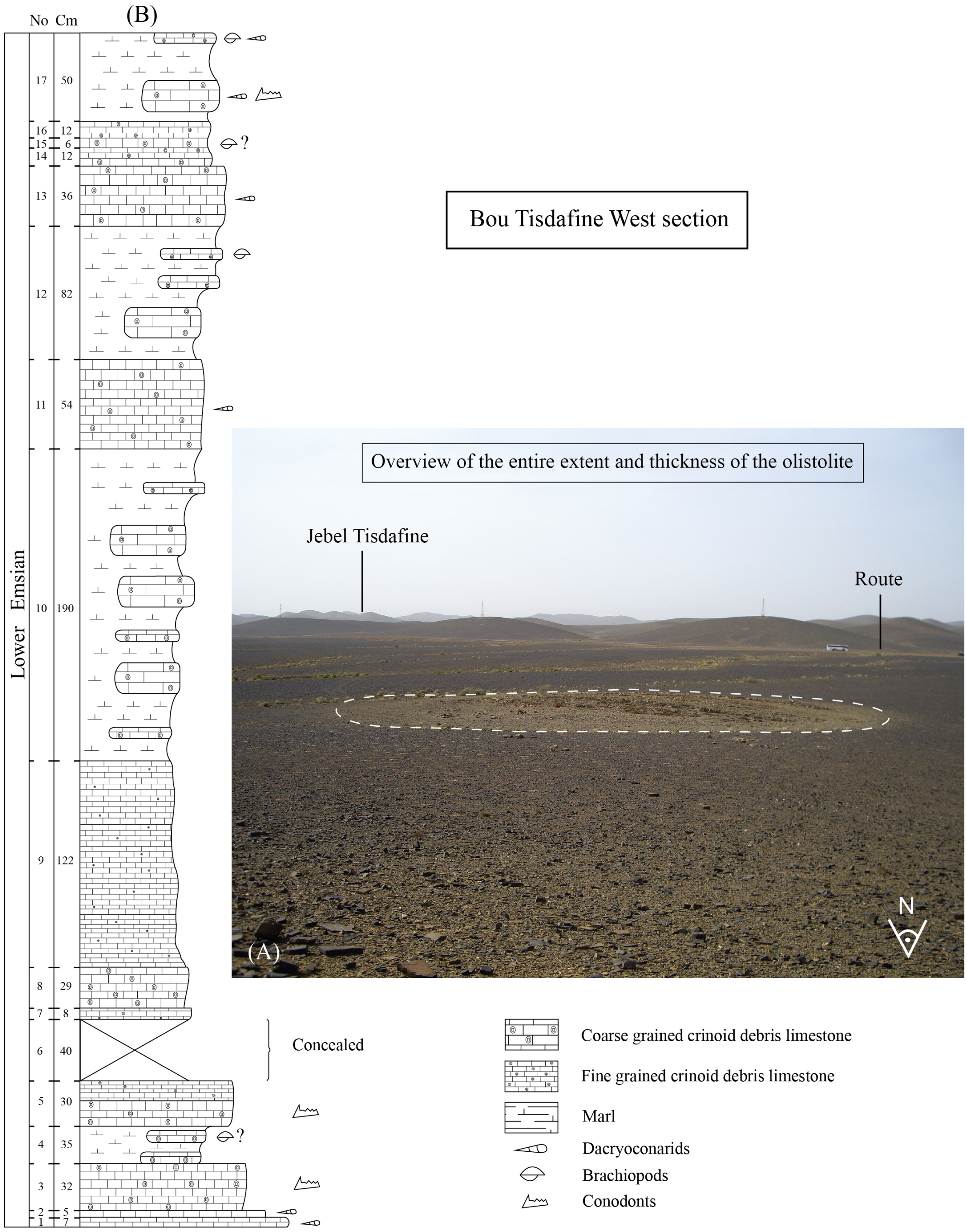


Figure 4. Sketch of outcrop, sample positions, lithostratigraphic details, and faunal records of the Bou Tisdafine-West section (after Rytina 2013).

The outcrop begins with a ca. 10 m thick succession of fine-grained and flaser-bedded to nodular limestones with centimetric marl interbeds (Fig. 5). Because of the trenching, there are covered intervals with arbitrary thicknesses. In thin section, the variable texture is bioturbated mudstone and bioclastic wackestone or packstone, sometimes with sparite cement. The wacke-packstones are rich in fossils including trilobites, bivalve filaments, brachiopods, dacryoconarids, foraminifers, and gastropods (Gottlob 2020). The macrofauna is composed of trilobites, crinoid ossicles, orthocones, very rare goniatites (one *Fidelites* sp.), and bivalves. A conodont sample from the base yielded *Polygnathus costatus, Polygnathus praetrigonicus, Polygnathus robusticostatus, Linguipolygnathus linguiformis, Icriodus retrodepressus*, and *Belodella resima*. This is a typical lower Eifelian assemblage (lower *Polygnathus costatus* Zone; e.g. Belka *et al.* 1997, Becker & Aboussalam 2013). The trilobites encountered were good markers of the Eifelian age, too. Apart from the lichid, there are species of *Gerastos*, *Struveaspis, Drotops, Austerops, Pedinopariops,* and *Thysanopeltis* (Gottlob 2020). The upper levels of this sequence are marked by bioturbation, indicating a moderately shallow and episodically agitated environment. However, there are no organisms in the euphotic zone. In general, there are similarities with the Eifelian of the northern Maïder (El Otfal Formation), which is famous for its trilobite beds.

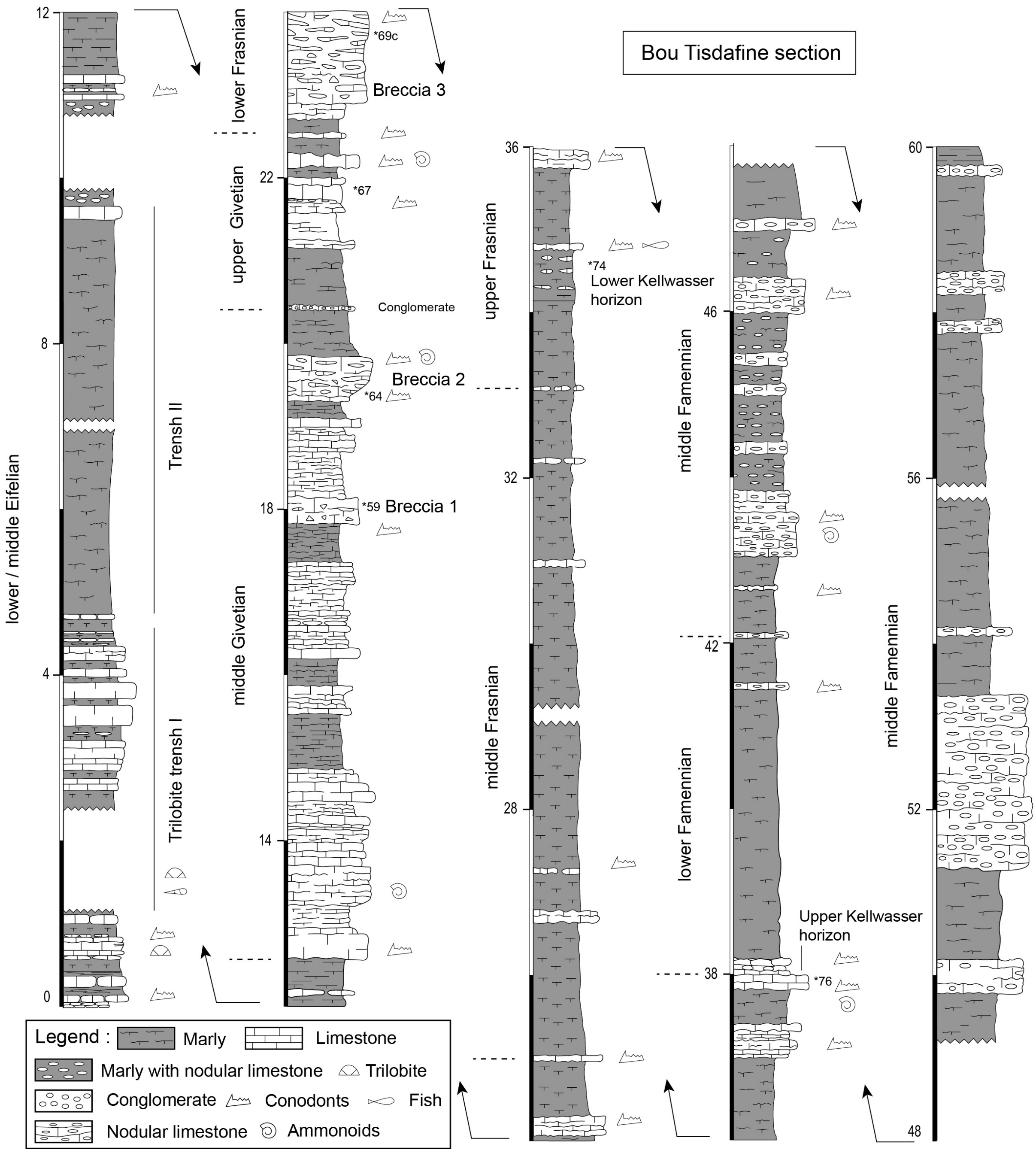


Figure 5. Simplified section log for the Eifelian to middle Famennian of the Bou Tisdafine section.

The precise position of the Eifelian/Givetian boundary is not yet known within a strongly condensed interval. There are no black shales or limestones indicating the global Kačák Events (e.g., Walliser & Bultynck 2001, Becker *et al.* 2018a); the event interval may be missing due to a hiatus. The middle Givetian is represented by marls and gray limestones with intercalations of nodular, black limestones. It is only ca. 4 m thick. The microscopic analysis shows wackestone to packstone. The micropaleontological content includes foraminifers, ostracods, dacryoconarids, crinoids, corals, and brachiopod debris. A conodont association of *Icriodus brevis*, *Polygnathus varcus*, *Polygnathus timorensis*, and *Linguipolygnathus linguiformis* (Figs. 6.1-4) fall in the *Polygnathus rhenanus-varcus* Zone of the lower half of the middle Givetian. The flaser-bedded limestones include occasional tabulate coral accumulations (intraclastic coral rudstones, Fig. 7A), solitary rugose corals, and goniatites (*Sellagoniatites*); the nautiloids are represented by oncoceratids. The lithological succession and fauna suggest a deep neritic to shallow pelagic setting, interrupted by an allochthonous episode of reef-type, nodular coral limestone. It represents either a storm bed or proximal debris flow deposit. There is locally no evidence for the two *Pumilio* Beds, which characterize the middle Givetian of the Tafilalt (Lottmann 1993, Becker *et al.* 2018a, b). Instead, there are three breccia units with resedimented micritic limestone clasts. Based on rich open, deeper-water conodont faunas, including *Polygnathus ansatus* and *Linguipolygnathus mucronatus*, the two lower breccias (Beds 59, 64, Fig. 5) fall in the lower part of the *Polygnathus. ansatus* Zone in the upper half of the middle Givetian. They are followed by an alternation of marls and thin-bedded limestones that display hematite pebbles at the top (Bed 67, Fig. 7B), an indicator of dysoxic sediment starvation surfaces. At this level, there are rich pelagic conodont faunas of the upper Givetian *Polygnathus cristatus ectypus* to *Polygnathus dengleri* zones (e.g. with *Schmidtognathus peracutus, Klapperina disparilis*, and *Polygnathus dengleri dengleri*), resembling those of Tafilalt (Aboussalam & Becker 2007). Above follows the most distinctive (Figs. 5, 7C, Bed 69), coarse, hematite-impregnated breccia (Fig. 7C) with characteristic, large, flat, and angular micritic limestone clasts. Its rich conodont fauna is typical for a deep-water setting and belongs to the basal Frasnian MN 2 or *Ancyrodella rotundiloba* Zone. The Givetian/Frasnian boundary is marked by a disconformity, as on the Tafilalt Platform (Aboussalam & Becker 2007). The three middle/upper Givetian breccias represent debris flows originating (in the original setting of the olistolite) from an adjacent repeatedly active fault scarp. The Givetian of section Bou Tisdafine has only slight similarities with the northern Maïder (Taboumakhlouf Formation), where coral limestones are much more dominant (e.g. Hollard 1974, Bultynck 1985, Kazmierczak & Schröder 1999). The Givetian of the Tafilalt Platform is very differently developed in pelagic ammonoid facies (e.g. Bultynck 1987, Aboussalam & Becker 2007, 2011).

The main part of the Frasnian constitutes a succession of fine-grained limestone beds and nodular limestones with thick marl intercalations (ca. 12 meters thick). At the base, directly above the ferrugineous breccia, *Palmatolepis transitans* indicates the MN 4 Zone sensu Klapper (1989) or *Palmatolepis transitans* Zone, the last zone of the lower Frasnian. Therefore, the lower Frasnian is locally incomplete. The microscopic analysis revealed a bioclastic wackestone to packstone texture with debris from various organisms. The macrofauna is composed of crinoids, orthocones, goniatites (rare gephuroceratids), fragments of brachiopods, and some fish bones (Fig. 7D). At the top, there is a 50 cm thick nodular level, light-grey in color, which is essentially formed by poorly sorted limestone nodules of different sizes. It is a monogenic conglomerate (pseudo-breccia) with limestone debris and ferruginous cement. The middle-upper Frasnian interval consists of alternations of thick marl levels and thin nodular limestone. At the base, *Palmatolepis punctata* and *Ancyrodella. lobata* were found. The precise position of the middle/upper Frasnian boundary is not yet clear. At the top of Bed 74, a thin grey limestone with *Palmatolepis winchelli, Ancyrognathus. asymmetricus, Ancyrognathus amana,* other conodonts, and placoderm remains indicate the Lower Kellwasser level (see, e.g., Schindler 1990, Becker *et al.* 2018a). The upper part of the upper Frasnian interval is marked by a blackish, secondarily (by weathering) bleached bed rich in fossils (Bed 76, Fig. 5), a condensed level that looks like the Kellwasser facies. This interpretation is supported by abundant manticoceratids, including the Upper Kellwasser index genus *Crickites*. Very rich conodont faunas include *Ancyrognathus ubiquitus* and *Palmatolepis ultima*, the index species for the terminal Frasnian (upper part of Upper Kellwasser level) MN Zone 13c of Girard *et al.* (2005).

From a geodynamic point of view, the presence of conglomerates is a strong indicator of vertical movements that occur in tilted blocks, known particularly from the Meseta (Piqué 1979, Fadli 1990, Tahiri 1991, Zahraoui 1991, Becker *et al.* 2015). This type of movement was reported in the Eastern Anti-Atlas by Destombes (1985) and began in the lower Ordovician. It is further characterized by tension slits and asymmetrical folds, from centimeter to metric amplitude, with axis oriented N110 to N130, dipping steeply to the SE, and damping upwards. Upper Frasnian synsedimentary block movements associated with the Kellwasser facies are also known from the Tafilalt (Wendt & Belka 1991). The excavated first Famennian limestone does not belong to the basalmost Famennian *Palmatolepis subperlobata* Zone (Spalletta *et al.* 2017, former Lower *triangularis* Zone). Based on *Palmatolepis praeterita* sensu Schülcke (1995) and *Ancyrognathus sinelaminus*, it falls already in the slightly younger *Palmatolepis delicatula platys* Zone (former Middle *triangularis* Zone). This implies a short sedimentary gap at the stage boundary, as in many Frasnian/Famennian boundary sections.

The regional Devonian succession ends with lower-middle Famennian griotte-type facies. Then, it forms a succession with strongly nodular limestone and dark marl interbeds, sometimes with intercalated lumachelles. The thickness is about 15 m, but this is an estimate due to many covered intervals. Microscopic observations reveal a general mud-wackestone texture, partly rich in crinoids, bivalves, and ostracods. In some samples, well-preserved remains belong to crinoids, orthocones, bivalves, and ammonoids (cheiloceratids). The conodonts *Palmatolepis crepida* (Fig. 6.5) and *Polygnathus* sp. (Fig. 6.6) confirm a lower Famennian age. This corroborates the Famennian presence on the southern flank of Jebel Tisdafine noted by Hindermeyer (1955). Higher beds yielded middle Famennian conodonts (*Scaphignathus velifer* Zone). Possibly younger strata are masked by the Quaternary deposits. In the massive conglomerate/breccia of Taourirt n´Khellil to the west, the youngest known clasts belong to the upper Famennian *Palmatolepis gracilis manca* Zone (Rytina *et al.* 2013, Hartenfels *et al.* 2013).

The Frasnian and Famennian beds of the Bou Tisdafine section do not resemble the successions south of the Jebel Ougnat. The Upper Devonian of the Jebel Gherghiz (= Rheris) of the northern Maïder is very incomplete, consisting of sandstones and shallow, crinoidal facies (Wendt *et al.* 1984, Fröhlich 2004). Towards the SE, in the Bou Dib region, lies a turbiditic to shaly deep pelagic basin (Bou Dib flysch of Hollard 1967). The condensed, pelagic Tafilalt Platform is characterized by a thick interval of black, organic-rich limestones and shales (extended Kellwasser-type, hypoxic facies) spanning all of the upper Frasnian and lower Famennian (e.g., Buggisch & Clausen 1972, Wendt *et al.* 1984, Wendt & Belka 1991, Becker 1993).

**Oued Ferkla**

The Oued Ferkla section is located on the right (northern) slope of the Oued Ferkla, near the bridge, along the road connecting Tinejdad and Goulmima (GPS: N31°32’12.18”; W5°0’37.61”). The first studies in this section were conducted by Clariond & Termier (1933) and Hindermeyer (1955). Ward *et al.* (2013) logged the Emsian to middle Givetian section, provided first conodont, foraminifer, and microfacies data, and specified the local position of global events. In parallel, Hejja (2013: Section TI) studied the succession at a lower resolution. Our work is based on the preliminary data of Ward *et al.* (2013), refining the stratigraphy and sedimentology.

The section shows, from bottom to top, variable sedimentation that reflects environmental changes. Sequences are characterized by a different organization and variable composition that reflect facies shifts with regressions and transgressions (onlap and offlap). According to microfacies, granulometry, calcimetry, and mineralogy (X-ray analysis of bulk rock and of the clay phase), we distinguish eleven lithostratigraphic units, which are as follows (Fig. 8):

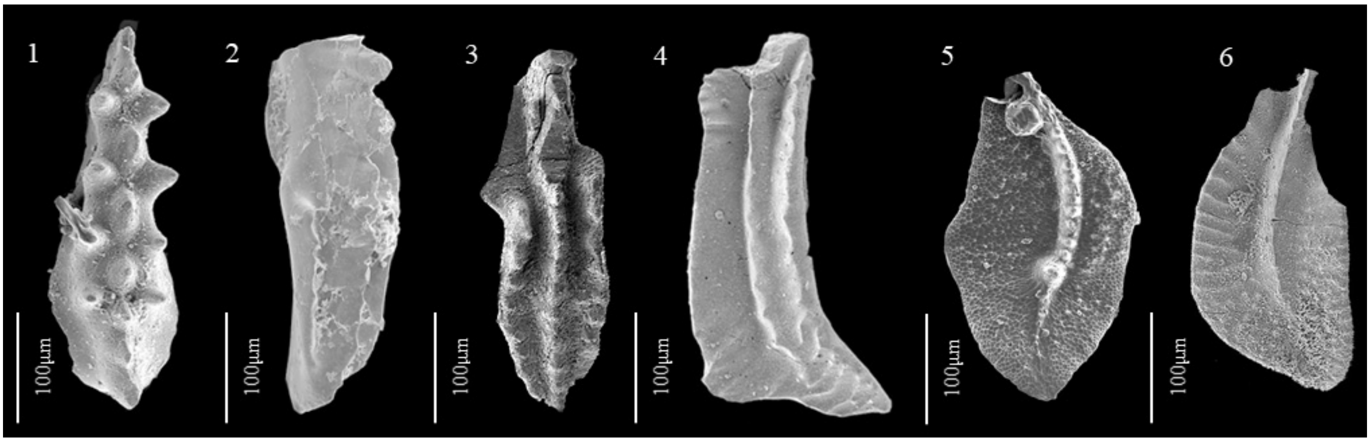


Figure 6. Middle Givetian (1-4) and Famennian (5-6) conodonts from the Bou Tisdafine section. 1. *Icriodus brevis*, 2. *Polygnathus varcus* (free blade broken off), 3. *Polygnathus timorensis*, 4. *Linguipolygnathus linguiformis*, 5. *Palmatolepis crepida*, 6. *Polygnathus* sp. (anterior platform incomplete).

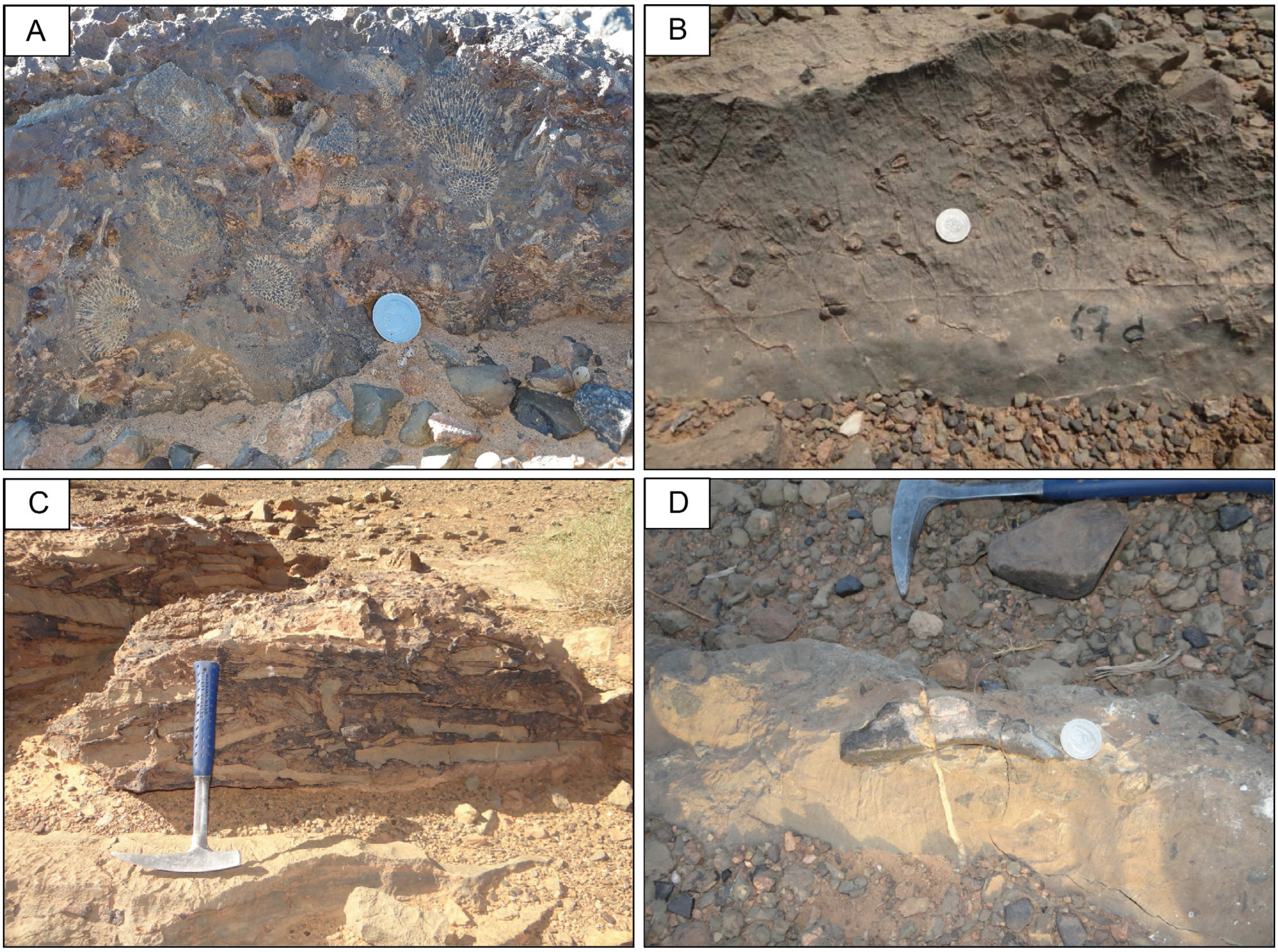


Figure 7. Field photos of the Bou Tisdafine section. A. Givetian bluish coral limestone (rudstone) with unsorted intraclasts, favositids and thamnoporids, deposited by a major storm or debris flow. B. Top of upper Givetian Bed 67 with hematite nodules, representing a discontinuity surface. C. Hematite-impregnated flat pebble breccia with elongated micrite clasts, lower Frasnian (MN 2 or *Ancyrodella rotundiloba* Zone). D. Lower Frasnian Placoderm bone. Coin diameter = 23.5 mm.

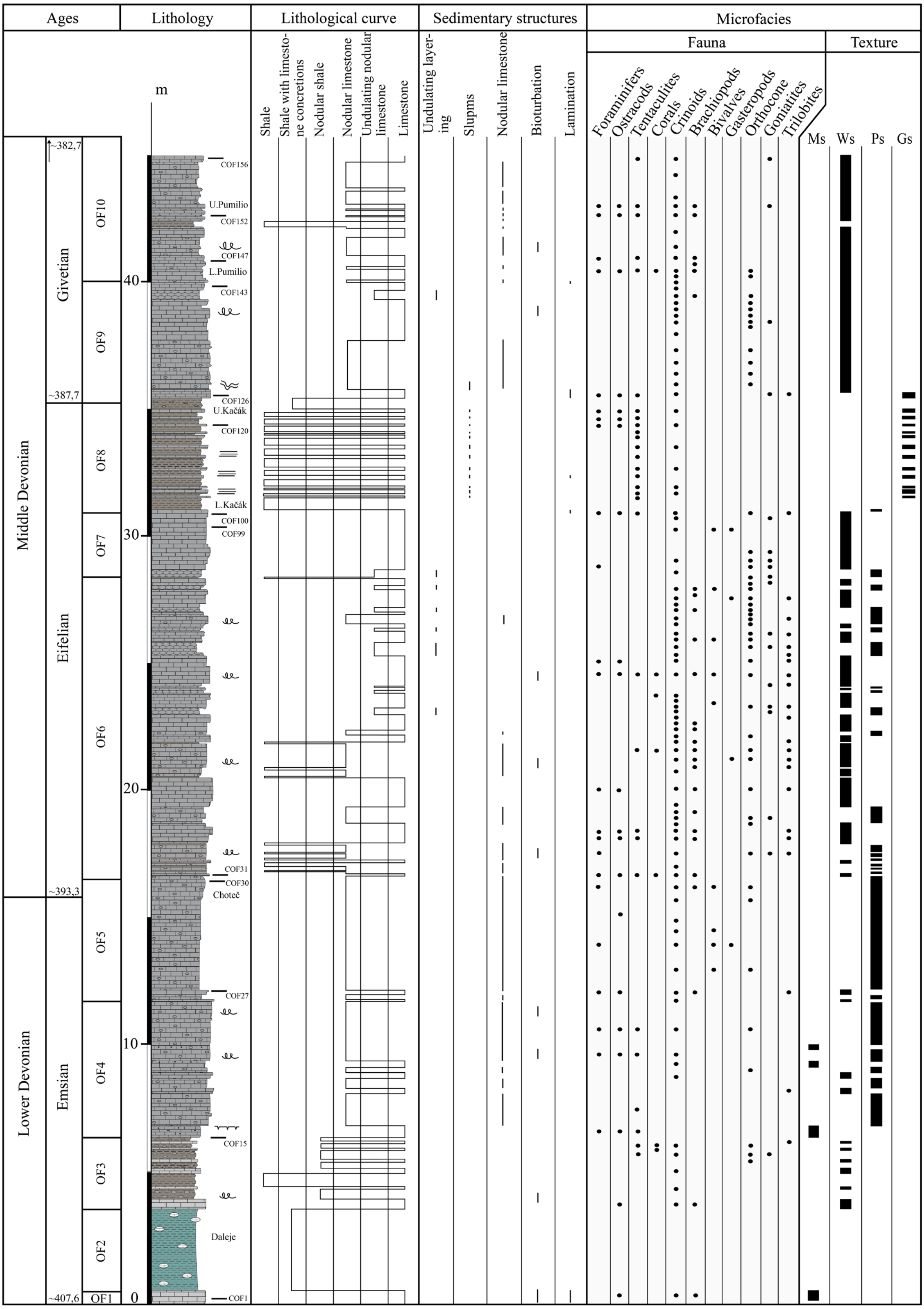


Figure 8. Lower Emsian to middle Givetian litho-, chrono-, conodont and event stratigraphy at Oued Ferkla, showing the macroscopic lithology, position of conodont samples (COF numbers), sedimentary structures, bioclasts, and general texture of thin-sections (Ms = mudstone, Ws = wackestone, Ps = packstone, Gs = Grainstone).

***Unit OF1 (top lower Emsian)***

The base of the succession is represented by a unit up to 1 m thick of dark gray limestone, slightly laminated and bioturbated. In a thin-section, it consists of a dark, organic-rich mudstone to middle-grey wackestone (Fig. 9A). The carbonate matrix contains a silty fraction, predominantly quartz, which is more or less rounded or sub-rhomboedal. A few quartz elements organized into compact aggregates of large, distinctly rhomboedal, polycrystalline grains. The unit contains a varied fauna made up of mollusc filaments, brachiopod fragments, ostracods, recrystallized dacryoconarids, and crinoid debris. The conodont association includes *Eolinguipolygnathus vigieri*, *Linguipolygnathus inversus*, *Eolinguipolygnathus laticostatus*, and *Caudicriodus ultimus*. It enables the attribution to the upper part of the lower Emsian, the *Eolinguipolygnathus laticostatus* Zone (see Aboussalam *et al.* 2015).

***Unit OF2 (basal upper Emsian)***

The second unit consists of 3 m thick greenish, silty shale/marl, comprising limestone concretions without macrofauna. The matrix is homogeneous and contains early diagenetic pyrite. The texture is a recrystallized (micro- to pseudosparitic) mudstone (Fig. 9B) with some bioclasts. The grains consist of lamellar crystals forming loosely packed aggregates, which are stacked on top of each other. The concretions appear to be completely of diagenetic origin, which explains the lack of macrofauna and conodonts. The sudden change from Unit OF1 to OF2 records the transgressive global Daleje Event (Fig. 10A) at the base of the upper Emsian (e.g., House 1985, Tonarova *et al.* 2017), which is widespread in the eastern Anti-Atlas (Aboussalam *et al.* 2015, Becker *et al.* 2018a, b). Unit OF2 correlates with the lower Er-Remlia Formation of the Maïder (Hollard 1974) and with the lower Unit K at the base of the (revised) Amerboh Formation in the Tafilalt region (Aboussalam *et al.* 2015, Hartenfels *et al.* 2018). The Er-Remlia Formation has been assigned to the *Icriodus fusiformis* Zone (Bultynck 1985, Aboussalam *et al.* 2015), which correlates with a gap (*laticostatus-bultyncki* Interregnum) in the southern Moroccan polygnathid zonation.

***Unit OF3 (main upper Emsian)***

Unit OF3 is a 3 m thick succession of limestone and marl and corresponds to Unit IIb of Ward *et al.* (2013). The beds contain relatively few fossils, such as crinoid debris, dacryoconarids, bivalves (*Panenka*), rare brachiopod remains, and orthocones. At the base, a gray-beige limestone bank with a nodular top and iron concretion runs along the outcrop (Bed 3 of Ward *et al.* 2013). It is a benchmark level within the Lower Devonian deposits throughout most of the region. This limestone is a bioturbated bioclastic wackestone, displaying internal nodules (Fig. 9C); it represents an environment favorable to marine benthos. The micrite matrix is partly replaced by diagenetic sparite. Large burrows occur at the top (Fig. 10C). This alternation evolves towards the top into finer facies with a ravinement surface.

Near the top of Unit OC3, some trilobites and loose ammonoids (*Latanarcestes noeggerathi* auct, *Sellanarcestes* sp.) (Fig. 10B) were found. They are typical for the upper Emsian *Linguipolygnathus bultyncki* Zone/Subzone, which is equivalent the lower part of the global *Linguipolygnathus serotinus* Zone (Aboussalam *et al.* 2015). Referring to Bultynck (1985), Unit OF3 is an equivalent of the upper Er-Remlia Formation of the northern Maïder and the upper part of Unit K of the Tafilalt.

***Unit OF4 (higher upper Emsian)***

Unit OF4 is a more than 6 m thick succession of massive, flaser-bedded and nodular limestone, shifting exclusively to nodular towards the top. It is equivalent to the massive lower part of Unit III in Ward *et al.* (2013). The microfacies shows a fairly heterogeneous micritic matrix, sometimes microsparitic, with numerous aggregates of pyrite. The terrigenous silt contributes substantially to the carbonate mud matrix. The general texture is bioclastic packstone (Fig. 9D) with some mudstone/wackestone alternations at the base. The fauna is quite varied. There are crinoid fragments, dacryoconarids, goniatites (*Anarcestes simulans, Sellanarcestes applanatus, Sellananarcestes wenkenbachi*), orthocones, bivalves, rare brachiopods, trilobites, foraminifers, and ostracods, whose valves are sometimes joined. Debris of orthocones and goniatites are sometimes micritized and pyritized. Some bioclasts are partially silicified. The facies is well bioturbated with vortex-shaped burrows and pyrite encrustations. There are microbial, small to bushy ferromanganese tufts, small sulphate-reduction cavities, and calcite cement patches, reminiscent of the microfacies in the Middle Devonian at Jbel Mech Irdane (Casier *et al.* 2010). The fossil assemblage is affected by diagenetic overprint, resulting in a state of preservation, which makes determinations difficult. At the base of the unit, hardground surfaces are well marked showing ferruginous encrustations and traces of bioturbation (Fig. 10D). The goniatite fauna places Unit OF4 in the *Anarcestes* Zone (LD IV-D, Becker & House 1994) in the middle part of the upper Emsian (Becker & House 1994, Ebbighausen *et al.* 2011). The conodont record is very poor. At the base, only the long-ranging *Belodella resima* was obtained from Bed 15. Unit OF4 is an equivalent of the lithologically similar Tazoulait Formation of the northern Maïder (see Bultynck 1985) and the lower *Anarcestes* Limestone (lower Unit L, middle Amerboh Formation) of the Tafilalt (Becker *et al.* 2013, Hartenfels *et al.* 2018).

***Unit OF5 (Emsian-Eifelian transition)***

Unit OF5 (upper part of Unit III in Ward *et al.* 2013) is also characterized by a succession of flaser-bedded and nodular limestone, with a thickness of 4.5 m, but it is morphologically more recessive than the underlying Unit OF4 (Figs. 10E-F). Beds are almost amalgamated and coarsen upwards. The sedimentary sets are organized at the base in calcareous banks with a grey-beige patina of about ten cm, with levels of bedded crypto-algae in filaments. The general microfacies texture is wackestone to packstone. The fossil assemblage is quite varied, showing dominant fragmentary bivalve bioclasts (Figs. 9E-F), orthocones, variably abundant dacryoconarids, and foraminifers. The bioturbation is weak in samples with organic-rich, dark micrite matrix (Fig. 9E). There is no grading or sorting of bioclasts. Diagenetic processes are manifested by calcitic cement joints, whose crystals are equidimensional, non-ferrous, and with light to dark luminescence. Crinoid debris show recrystallization. The increase of marl, goniatites and large orthocones testify along an outer shelf slope the change of the environment towards a hemipelagic basin. Conodonts from Bed 27 at the base (*Belodella resima, Neopanderodus perlineatus, Linguipolygnathus bultyncki*) fall in higher parts of the upper Emsian *Linguipolygnathus bultyncki* Zone (= *Linguipolygnathus serotinus* Zone of the global scale, Aboussalam *et al.* 2015). At the top, a thin, dark, microcrystalline limestone followed by marl indicates the level of the global Choteč Event (House 1985, Koptíková 2011; Fig. 10E), which occurs quite extensively in the Tafilalt (e.g. Becker & House 1994, Klug 2003, Becker & Aboussalam

2013, Hartenfels *et al.* 2018). In Ward *et al.* (2013), its position was correctly shown in the section log (near the top of their Unit III), but erroneously it was noted at the base of Unit III in the caption of Fig. 2. The Choteč Event falls within the lower Eifelian at the top of the *Polygnathus partitus* Zone. As in the northern Maïder (within the El Otfal Formation, Bultynck, 1985) and northern Tafilalt (in the upper part of the nodular *Anarcestes* Limestone, Unit L sensu Becker *et al.* 2013; upper Amerboh Formation), the Lower/Middle Devonian boundary is lithologically and faunistically indistinctive at Oued Ferkla. The deepening at the base of Unit OF5 resembles the trend at the base of the El Otfal Formation of the northern Maïder (Ward *et al.* 2013).

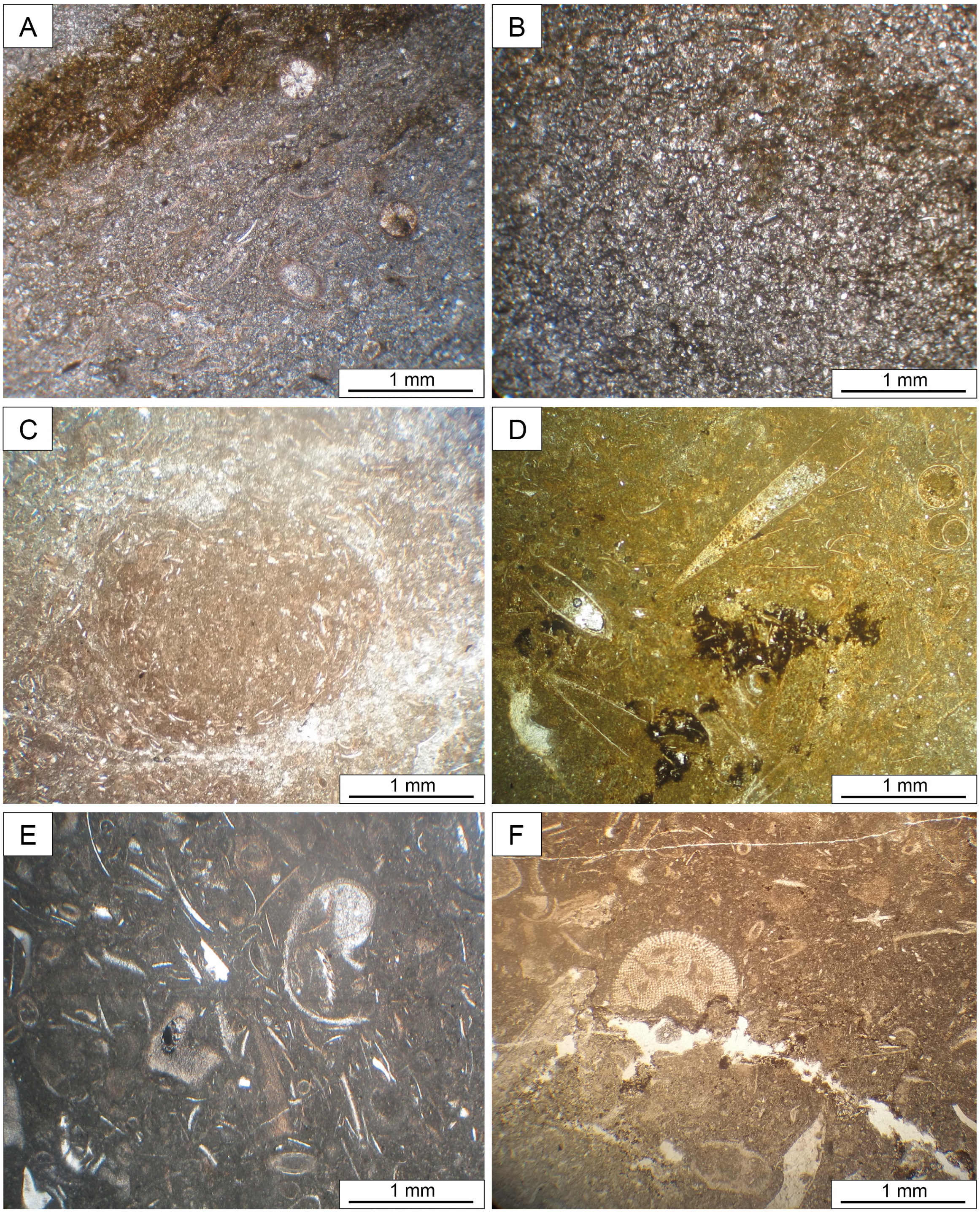


Figure 9. Microfacies of the lower and upper Emsian at Oued Ferkla. A. Silty, mud- to wackestone with fined shell debris and recrystallized cross-sections of dacryoconarids, Unit OF1. B. Strongly recrystallized (pseudosparitic) mudstone, concretion within Unit OF2. C. Bioclastic wackestone with fine shell debris and nodule-forming bioturbation. The micritic matrix is slightly recrystallized and replaced by calcitic micro- and orthosparite, base of Unit OF3. D. Bioclastic wackestone with abundant, ribbed dacryoconarids, ostracods, crinoid and indeterminate shell debris, base of Unit OF4. E. Dark-grey wackestone to packstone with abundant mollusc and crinoid debris and organic-rich, dense micrite matrix. Diagenesis led to calcitic recrystallization of bioclasts, which are partly recognizable as phantoms, Unit OF5. F. Bioturbated, bioclastic wackestone to packstone with recrystallized fine mollusc debris, dacryoconarids, and crinoid ossicles, Unit OF5.

***Unit OF6 (lower Eifelian)***

Unlike the previous unit, Unit OF6 (Unit IV of Ward *et al.* 2013) shows very reduced marly intercalations between fine-grained, flaser-bedded, and nodular limestones, which have a thickness of ca. 11.5 m. There is a network of normal faults oriented N10 to N20, with a slight dip of 5° to 10° towards the NW, the crevices of which are filled with calcite (Fig. 11A). The beds show cyclicity, with couplets of more recessive, nodular limestones with undulating stratification, and more solid and thicker, detrital beds. The microscopic texture is wackestone to packstone (Fig. 12A) with stylolites formed at diagenetic pressure solution contacts, which are well expressed throughout the formation. Skeletal elements consist of small solitary rugose corals (deep-water forms), orthocones, brachiopods, crinoid debris, and fragments of trilobites (phacopids, Fig. 12A). The microfauna includes ostracods and foraminifers. Among the ammonoids, we recognize *Subanarcestes marhoumensis* (Fig. 11B), *Pinacites eminens* (Fig. 11C), and *?Cabrieroceras* sp., which relates the unit to the lower Eifelian (MD I-C/D sensu Becker & House 1994; possibly basal I-E). The age is confirmed by a new association of conodonts from the base (top of Bed 30): *Belodella resima*, *Icriodus* aff. *orri* (Fig. 13.1), *Icriodus struvei*, *Icriodus mariae*, *Icriodus anterodepressus, Linguipolygnathus linguiformis*, *Linguipolygnathus* aff. *alveolus* (Fig. 13.2), *Polygnathus partitus*, and *Polygnathus costatus* (Fig. 13.3). These characterize the *Polygnathus*

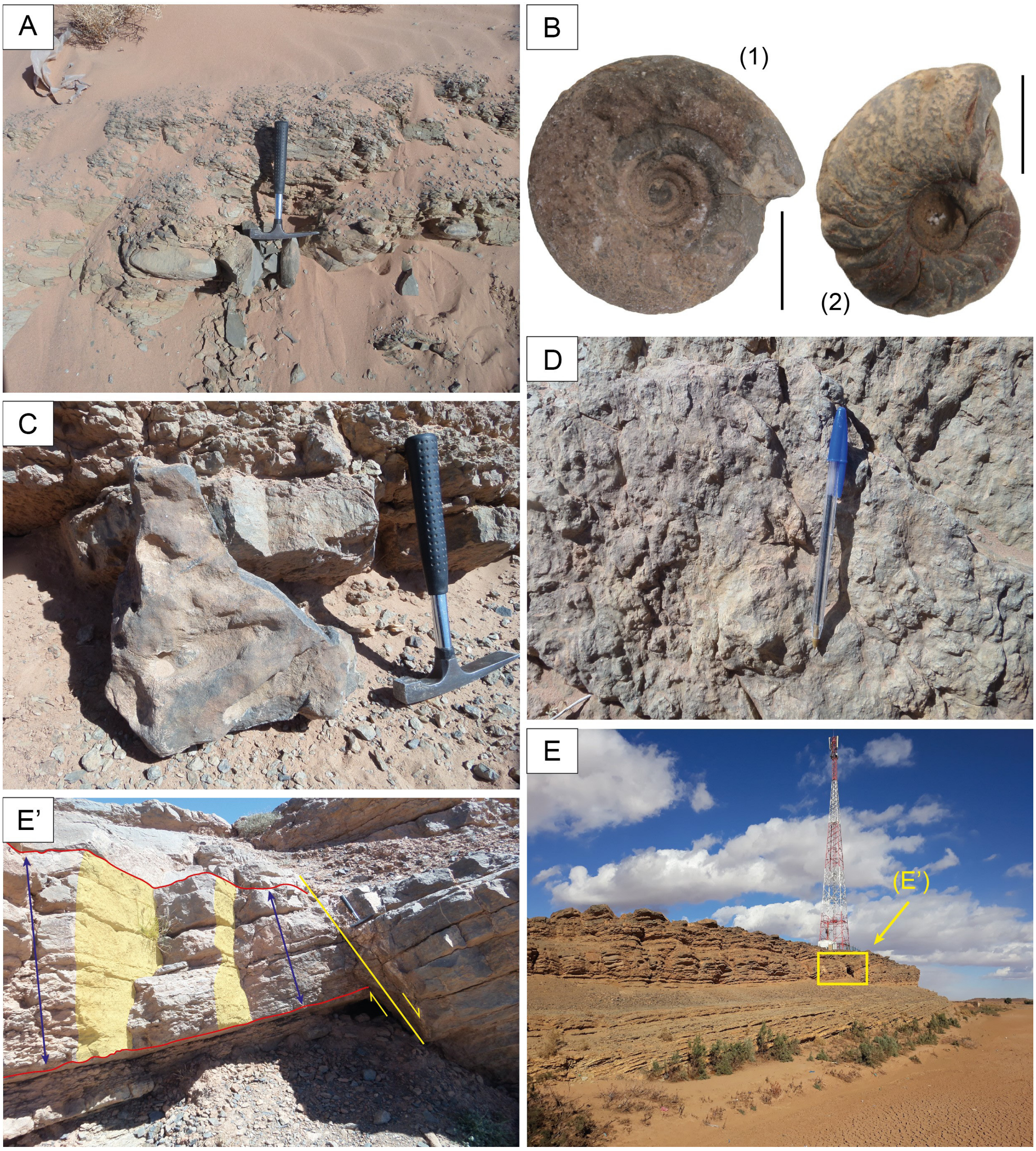


Figure 10. Field photos at Oued Ferkla. A. Unit OF1, greenish, silty shales with some limestone concretions, Daleje Shale equivalents, lower part of upper Emsian. B. Some ammonoids from Unit OF3: (1): *Sellanarcestes* sp., (2) *Latanarcestes noeggerathi* auct., scale bar = 2 cm. C. Large trace fossil (burrows) at the top of the basal bed of Unit OF3. D. Hardground surface in the lower part of Unit OF4. E. Overview of eastern part of the Oued Ferkla section, with Unit OF2 at the base, the nodular Unit OF3, the more solid limestones of Unit OF4, and the Choteč Event level as an incision in the upper cliff. F. Detail of F (yellow rectangle), showing a multi-phase strike-slip fault in Unit OF4, followed by more nodular limestones (Unit OF5) below the incision of the Choteč Event level near the top.

*costatus* Zone/Subzone in the lower Eifelian. Aboussalam in Ward *et al.* (2013) reported from Bed 31 additional taxa, including *Linguipolygnathus bultyncki, Linguipolygnathus pinguis*, and *Icriodus corniger corniger*, which are all known to occur at this level.

The organization of limestones in sequences with regressive trends and the richness of marine fauna suggests a moderately deep ramp environment, with minor sea-level related fluctuations between shallow and deeper pelagic settings, always below the photic zone and below the storm wave base. There is a better correlation with the goniatite-rich lower Eifelian of the Tafilalt (e.g. Becker & House 1994; Unit M, lower Bou Tchrafine Formation, Becker *et al.* 2013, Hartenfels *et al.*, 2018) than with the shallower, more neritic, brachiopod- and trilobite-rich succession of the northern Maïder (Taboumakhloûf Formation, Bultynck 1985) or with the lower Eifelian trilobite beds of our section Bou Tisdafine.

### *Unit OF7 (upper Eifelian)*

Unit OF7 (Unit V of Ward *et al.* 2013) consists of ca. 2 m thick, massive, dark gray limestones without marl interbeds. The microfacies reveals a fine-grained bioturbated

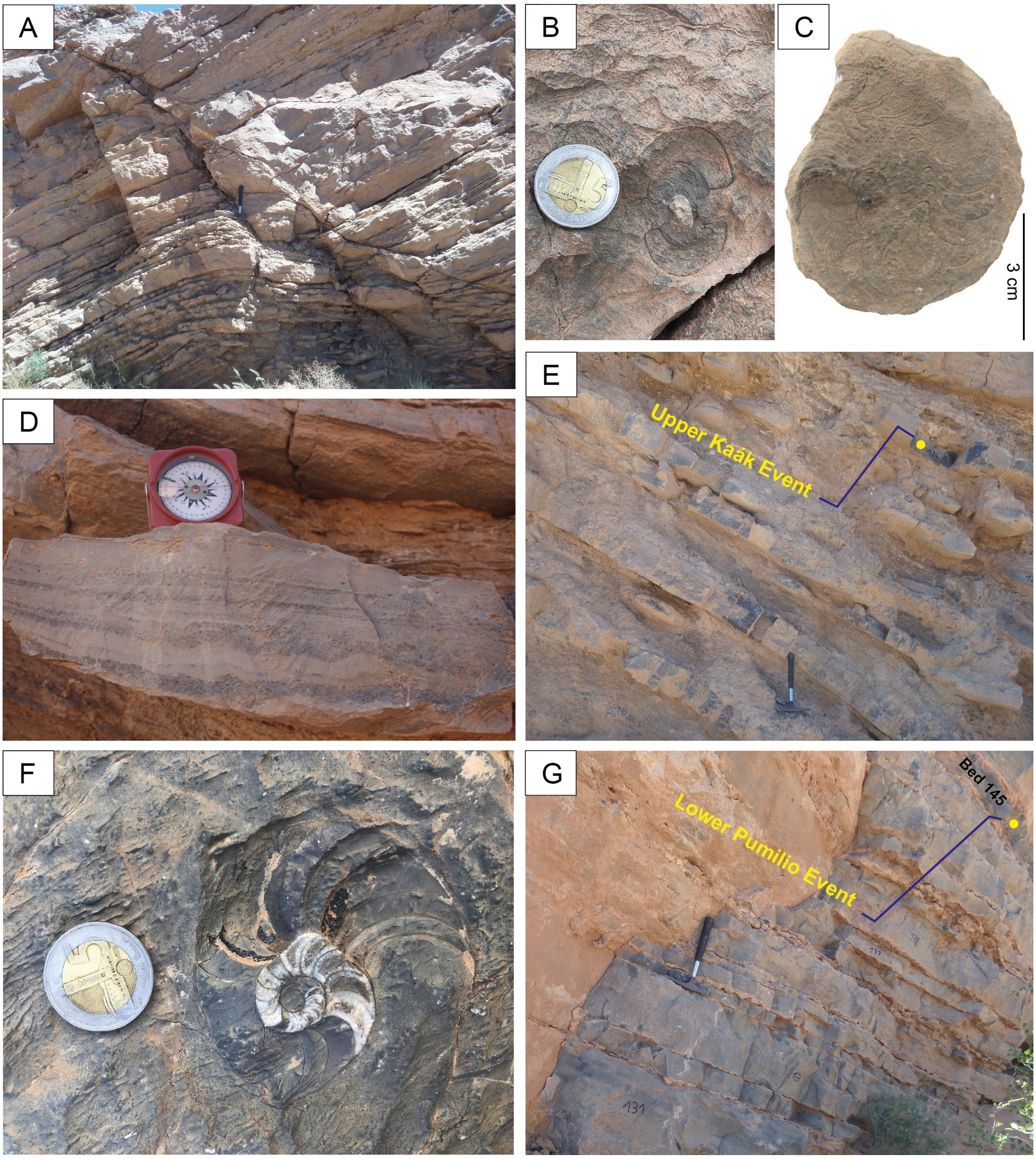


Figure 11. Field photos of the Middle Devonian at Oued Ferkla. A. Undulating normal fault within the Eifelian Unit OF6, eastern side of coad cut. B. Cross-section of a convolute *Subanarcestes marhoumensis*, Unit OF6. C. *Pinacites eminens*, lateral view of a slightly corroded mould, Unit OF6. D. Wavy- to plainly laminated amalgamation of detrital distal debris flow and fine-grained layers in Bed 108, Unit OF8, Kačák Event Interval (*Polygnathus ensensis* Zone). E. Upper part of Unit OF8, with black marl and limestones nodules of Bed 125 near its top, representing possibly the Upper Kačák Event Interval. F. Corroded *Sellagoniatites* sp. embedded in a hardground surface of Unit OF10. G. Upper part of Unit OF9 (Beds 131-144), overlain by the Lower *Pumilio* Event level (Bed 145) defining the base of Unit OF10 within the *Polygnathus rhenanus-varcus* Zone.

packstone texture with micrite matrix and bioclasts composed of abundant small, elongated, straight, or curved mollusc filaments (Fig. 12B), and dacryoconarids. A strong compaction and recrystallization resulted in flattening, sparite filling, and enveloping. Remains of ostracods are aligned along bedding planes in certain thin sections. The macrofauna consists of crinoid debris, orthocones, rare trilobites (*Thysanopeltis*, phacopids), and gastropods. The abundant ammonoids include *Subanarcestes macrocephalus* (from the lower part), *Cabrieroceras*, and *Agoniatites*, dating the unit as upper Eifelian (MD I-E/F1 sensu Becker & House 1994). This age is well confirmed by the rich conodont association from the top of the unit (Beds 99-100) with *Belodella resima*, *Neopanderodus perlineatus*, *Linguipolygnathus linguiformis*, *Linguipolgnathus weddigei*, *Polygnathus praetrigonicus*, *Polygnathus angustipennatus* (Fig. 13.8), *Polygnathus.*

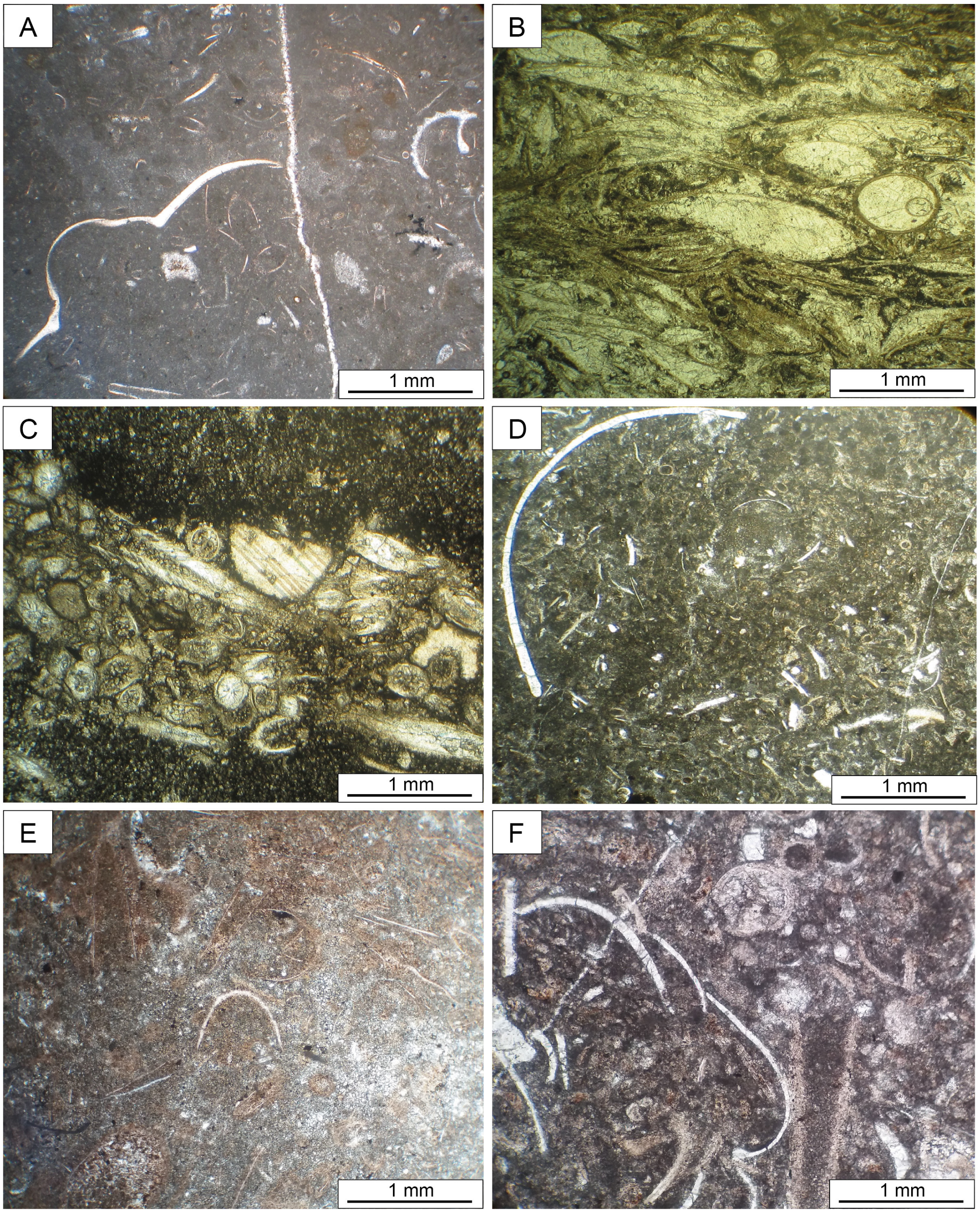


Figure 12. Microfacies of the Middle Devonian at Oued Ferkla. A. Bioturbated wackestone with fine mollusc debris, ostracods, recrystallized crinoid debris, cross-sections of a trilobite and spirally ribbed gastropod (upper right), and dense micrite matrix, Unit OF6, Eifelian. B. Strongly compacted, flattened and recrystallized shell packstone with sparite-filled bioclasts (mollusc shells), including an orthocone, Unit OF7, upper Eifelian. C. Alternation of dark-grey, microsparitic mudstone and a recrystallized, unsorted and non-graded dacryoconarid-crinoid grainstone, interpreted as deposited from a distal debris flow, Unit OF8, Kačák Event Interval. D. Peloidal and bioturbated bioclastic wackestone with unsorted mollusc debris, Unit OF9, lower Givetian. E. Bioturbated, bioclastic wackestone with poorly preserved, recrystallized dacryoconarids and mollusc debris; Unit OF10, middle Givetian. F. Recrystallized (micro- to pseudosparitic) bioclastic wackestone with poorly preserved dacryoconarids, bivalve and crinoid debris, Unit OF10, middle Givetian.

aff. *angustipennatus* (Fig. 13.4), *Polygnathus partitus* (Fig. 13.5), *Polygnathus robusticostatus* (Fig. 13.10), *Polygnathus angusticostatus*, *Polygnathus eiflius* (Fig. 13.9), *Polygnathus amphora*, *Polygnathus parawebbi*, *Tortodus kockelianus* (Figs. 13.6, 11), *Icriodus hollardi* (Fig. 13.7), and *Icriodus struvei*. The original upper Eifelian *Tortodus kockelianus* Zone (Weddige 1977) has been subdivided into three levels defined by the entries of *Tortodus australis*, *Tortodus kockelianus*, and *Plygnathus. eiflius* (Bultynck 1987, Belka *et al.* 1997). Recent data by Vodrážková & Suttner (2020) a subdivision of the *T. kockelianus* Zone based on either *Polygnathus eiflius* or *Polygnathus amphora*, is quetionable. In any case, the top of Unit OF7 represents well-oxygenated, pelagic limestones with a rich, diversified open shelf fauna of the immediate pre-Kačák Crisis interval (see Ward *et al.* 2013). In the northern Maïder, there are corresponding

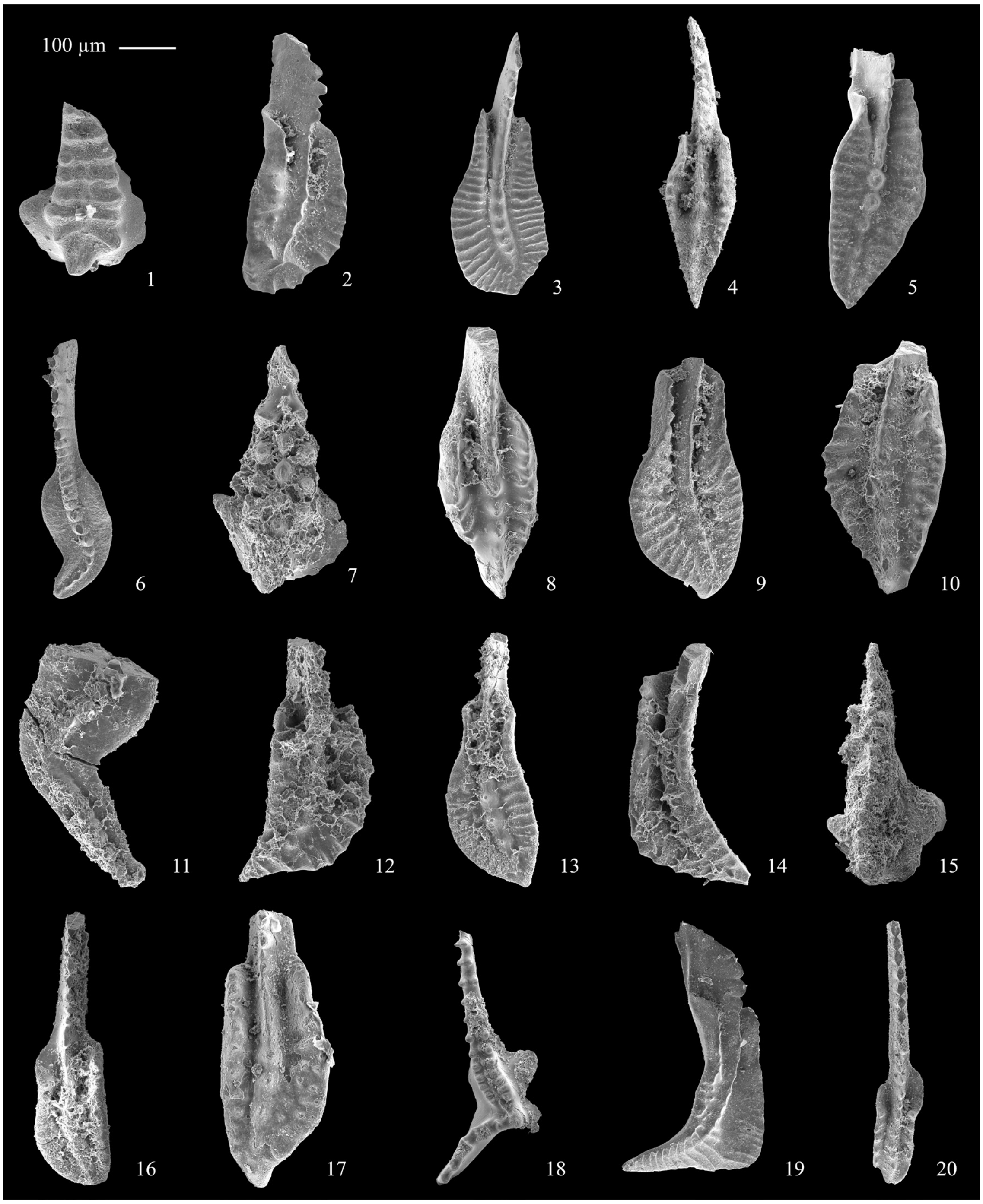


Figure 13. Representative Eifelian and Givetian conodonts from Oued Ferkla, 1-3 = Bed 30, *Polygnathus costatus* Zone (lower Eifelian); 4-6 = Bed 99, *Tortodus kockelianus* Zone (upper Eifelian); 8-11 = Bed 100, zone as Bed 99 (upper Eifelian); 12-14 = Bed 120, *Polygnathus ensensis* Zone (top-Eifelian); 15-17 = Bed 126, *Polygnathus hemiansatus* Zone (basal Givetian); 18-20 = Bed 152e, Upper *Pumilio* Bed, basal *Polygnathus ansatus* Zone (middle Givetian). 1. *Icriodus* aff. *orri*, 2. *Linguipolygnathus* aff. *alveolus*, 3. *Polygnathus costatus*, 4. *Polygnathus* aff. *angustipennatus*, 5. *Polygnathus partitus*, 6. *Tortodus kockelianus*, 7. *Icriodus hollardi*, 8. *Polygnathus. angustipennatus*, 9. *Polygnathus eiflius*, 10. *Polygnathus. robusticostatus*, 11. *Tortodus kockelianus* (broken specimen), 12. *Linguipolygnathus weddigei*, 13. *Polygnathus amphora*, 14. *Polygnathus parawebbi*, 15. *Icriodus obliquimarginatus*, 16-17. *Polygnathus hemiansatus*, 18. *Latericriodus latericrescens latericrescens*, 19. *Linguipolygnathus linguiformis*, 20. *Polygnathus timorensis*.

nodular limestones with *Agoniatites* in the Taboumakhloûf Formation (Bultynck 1987, and new data from the Bou Dib region). On the Tafilalt Platform, Unit OF7 correlates with the similar, flaser-bedded or nodular limestones of Unit N (upper Bou Tchrafine Formation; e.g. Walliser & Bultynck 2011, Hartenfels *et al.* 2018).

### *Unit OF8 (top-Eifelian Kačák Event Interval)*

The base of Unit OF8 (Unit VI of Ward *et al.* 2013) is marked by a rapid and drastic environmental change. The light-grey, micritic, oligotrophic, solid limestones of Unit OF7 are suddenly replaced by a ca. 6 m thick alternation of dark-grey, organic-rich, laminated limestone, fissile shale, and marl. The limestone beds consist of amalgamated alternations of fine-grained and detrital layers (Fig. 11D).

Wavy or normal lamination gives evidence for fluctuating current regimes. In thin sections, there are microsparitic mudstone layers and grainstones with abundant dacryoconarids, crinoid ossicles, ostracods, and fragments of orthocones (Fig. 12C). A diagenetic mineralization is observed by dark micritic-envelopes surrounding crinoids and dacryoconarids. Since thin dacryoconarid-crinoid coquinas lack sorting or current orientation, Ward *et al.* (2013, Fig. 7) suggested deposition by distal debris flows. The age of the unit is determined by the conodont association of Bed 120 (Fig. 8): *Belodella resima*, *Linguipolygnathus linguiformis*, *Linguipolygnathus weddigei* (Fig. 13.12), *Polygnathus pseudofoliatus*, *Polygnathus ensensis* Gp., *Polygnathus. eifelius*, *Polygnathus amphora* (Fig. 13.13), *Polygnathus robusticostatus*, *Polygnathus parawebbi* (Fig. 13.14), and *Polygnathus trigonicus*. It falls in the top-Eifelian *Polygnathus ensensis* Zone (see Walliser & Bultynck 2011), whose significance has been re-emphasized by Vodrážkova & Suttner (2020). The lower part of the unit still lacks *Polygnathus ensensis* in specimen-poor assemblages. Some black marl bedding surfaces are covered by masses of *Nowakia (Nowakia) otomari*, the index dacryoconarid of the *otomari* or Kačák Event. This two-phased global crisis has been described in the Tafilalt by Walliser *et al.* (1985), Ellwood *et al.* (1999, 2011), Crick *et al.* (1997, 2000), Walliser (2000), Klug (2002a, 2002b), and Hartenfels *et al.* (2018). At the top of Unit OF8, a black marl unit with concretions represents a final phase of the extended crisis interval (Fig. 11E) within the Upper Kačák Event sensu Walliser & Bultynck (2011). The peculiarity of Unit OF8 lies in the mixture of $C_{org}$-enrichment and extreme faunal blooms, reflecting strongly eutrophic conditions. The sudden onset of laminated shales and marls indicates anoxia and deepening/transgression, but with contrasting episodes of increased turbidity. The latter may reflect a steepening and adjacent unstable palaeoslope, with distal gravitational sediment and fossil transport. The Kačák Event Interval is developed in the northern Maïder at the top of the Taboumakhloûf Formation as a thick marl unit (e.g. Bultynck 1985, 1987), which has not yet been studied in detail. As noted above, it is not recognizable in the section Bou Tisdafine, which suggests that this allochthonous block does not have the same provenance as the Devonian of Oued Ferkla.

### *Unit OF9 (lower to middle Givetian)*

Unit OF9 (Unit VII of Ward *et al.* 2013) is 5 m thick and consists of massive limestone intercalated with nodular limestones without visible marl interbeds. At the base, Bed 126 shows convolute bedding with deformed laminae; the amplitude of the internal folds is centimetric, the hinges of folds giving the appearance of sediment balls. The convolute structures display a general NW-SE direction. In higher levels, they are replaced by horizontal laminations. Other limestones display bioturbation (Fig. 12D) and erosion surfaces, which mark short episodes of current-related non-deposition. The matrix of unsorted bioclastic wackestones is peloidal (Fig. 12D) or microsparitic. The fossil content consists of foraminifers, ostracods, dacryoconarids, crinoids, and mollusc debris. At the base (Bed 126), the association of conodonts corresponds to the basal Givetian *Polygnathus hemiansatus* Zone: *Belodella resima*, *Linguipolygnthus linguiformis* Morphotypes $\gamma$1-3, *Polygnathus. pseudofoliatus*, *Polygnathus ensensis*, *Polygnathus. hemiansatus* (Fig. 13.16-17), and *Icriodus obliquimarginatus* (Fig. 13.15). In complete agreement with the conodonts, Bed 126 shows cross-sections of an early maenioceratid, the lower/middle Givetian index goniatite group. The givetian marker species *Polygnathus hemiansatus* occurs also slightly higher, in Beds 128 and 130b. In the middle part of Unit OF9, the first *Polygnathus varcus* enter and indicate the base of the middle Givetian *Polygnathus rhenanus-varcus* Zone. At the top, an association from Bed 143 includes *Neopanderodus perlineatus*, *Icriodus arkonensis walliserianus*, *Icriodus regularicrescens, Icriodus difficilis, Linguipolygnathus linguiformis* (Fig. 13.19), *Linguipolygnathus mucronatus*, *Polygnathus timorensis* (Fig. 13.20), *Polygnathus pseudofoliatus*, and *Polygnathus varcus.* This fauna falls in the higher part of the *Polygnathus rhenanus-varcus* Zone (compare Bultynck 1987), still in the lower part of the middle Givetian.

### *Unit OF10 (higher middle Givetian)*

Unit OF10 (Unit VIII of Ward *et al.* 2013) is ca. 7.5 m thick and begins with a peculiar, lithologically distinctive marker level, the Lower *Pumilio* Bed (Bed 145, Fig. 11G). It is characterized by the mass occurrence of minute, often broken brachiopods (*Ense pumilio*) in black, organic-rich, crystalline limestone (brachiopod packstone). A similar second event marker, the Upper *Pumilio* Bed, lies in the middle of the unit (Bed 152e, brachiopod wackestone; compare Ward *et al.* 2013; Fig. 11j). Most of Unit OF10 is an alternation of nodular limestone and massive limestone, with a marly level (Bed 151) interspersed. Macrofauna is present in some samples, such as crinoid debris, dacryoconarids, and the goniatite *Sellagoniatites* sp. (Fig. 11F), which occurs commonly in the middle Givetian of the Tafilalt (e.g. Aboussalam & Becker 2011, Hartenfels *et al.* 2018). The microfacies texture is bioclastic and bioturbated wackestone with minor changes of the faunal association (Fig. 12E-F). The microfauna is represented by foraminifera, ostracods, dacryoconarids, fragments of molluscs, brachiopods, and crinoids. Conodont associations from the lower part (Beds 145, 147) fall in the upper parts of the *varcus-rhenanus* Zone: *Belodella resima, Icriodus difficilis, Linguipolygnathus linguiformis*, *Linguipolygnathus weddigei*, *Linguipolygnathus mucronatus, Polygnathus timorensis, Polygnathus varcus,* and *Tortodus bultyncki*. The Upper *Pumilio* Bed yielded *Belodella resima, Neopanderodus perlineatus, Latericriodus latericrescens latericresens* (Fig. 13.18), *Linguipolygnathus linguiformis* (Fig. 13.19), *Linguipolygnathus mucronatus, Linguipolygnathus weddigei*, *Polygnathus timorensis* (Fig. 13.20), *Polygnathus varcus,* and *Polygnathus ansatus.* The latter is the index species of the *Polygnathus ansatus* Zone (former Middle *varcus* Zone), comprising most of the upper part of the middle Givetian. The *Pumilio* Events are considered as benchmarks in the Middle Devonian time scale. They have been described in detail from the Tafilalt by Lottmann (1990) and occur also in the eastern Dra Valley (e.g. Aboussalam *et al.* 2004), Garcia-Alcalde & El Hassani (2020), but not in the shallower facies of the Maïder, nor, as noted above, in our section Bou Tisdafine (see above).

As in the top-Eifelian, there is a significant difference between the Tisdafine and Oued Ferkla Givetian.

At the top of Unit OF10, Bed 156 provided a restricted conodont fauna with *Linguipolygnathus linguiformis, Linguipolygnathus weddigei*, and *Polygnathus. timorensis*. An even higher, dark-grey, almost 40 cm thick crinoidal limestone (Bed 165), separated from Bed 156 by a ca. 3.5 m thick alternation of nodular marls, nodular limestone, and flaser-limestone (above the succession shown in Fig. 8), yielded *Belodella resima, Neopanderodus perlineatus, Icriodus difficilis, Linguipolygnathus linguiformis, Linguipolygnathus mucronatus*, *Polygnathus xylus, Polygnathus ansatus,* and *Polygnathus. timorensis*. Therefore, the *Polygnathus ansatus* Zone ranges until the top of Unit OF10.

### *Frasnian*

Above Bed 165, the outcrop of the roadcut becomes rather poor. However, ca. 50 m to the north, along the low western slope, separated by a stretch covered mostly by Quaternary deposits, there is an isolated, thin, dark-grey, detrital limestone with a Frasnian ancyrodellid fauna. We found *Ancyrodella. nodosa, Ancyrodella. gigas* (s.str. = M3) and *Ancyrodella. lobata* associated with various polygnathids and *Icriodus symmetricus*. This assemblage falls in the middle Frasnian (MN Zones 6-8, see Klapper 1997). Therefore, the marly mostly covered interval represents the upper Givetian to lower middle Frasnian. This basinal succession differs strictly from the Bou Tisdafine section and also from the contemporaneous strata of the northern Maïder (Bou Dib succession, Hollard 1967) or northern Tafilalt (e.g., Aboussalam & Becker 2007, Hartenfels *et al.* 2018). The facies change occurred obviously around the level of the global Taghanic Crisis (see Aboussalam 2003, Aboussalam & Becker 2007).

### Section Koudiat Inegh

The Koudiat Inegh section (Fig. 14) is located 8.2 km north of Tinejdad (GPS: N31°34'12.52"; W5°00'05.25"). It can be easily accessed just west of the road N32 connecting

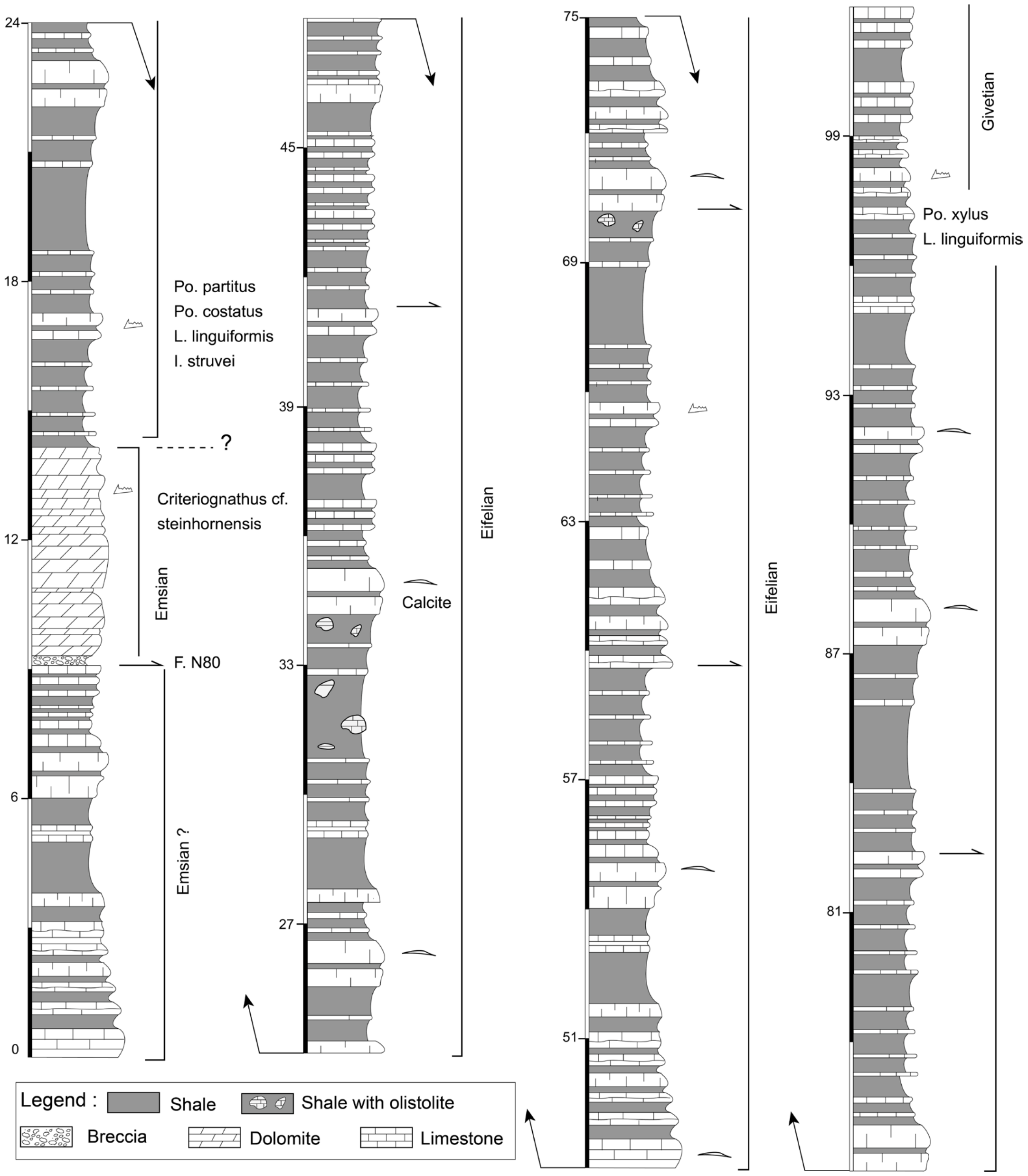


Figure 14. Schematic overview of Section Koudiat Inegh north of the Oued Ferkla, showing known conodont occurrences, the breccia bed at the base of the low dolomite cliff, and the level of small limestone olistolites.

Tinejdad and Goulmima (Figs. 3, 15A). The Devonian outcrop was noted in geological maps in Wendt & Belka (1991: thin dolomite of supposed upper Frasnian age) and Soualhine *et al.* (2003) and was briefly studied in the unpublished M.Sc. thesis of Hejja (2013: Section TIII). The landscape forms minor escarpments resulting from the dismantling of strata by minor alluvial fans guided by faults with NW-SE orientation. This is the imprint of Eovariscan structuring, even better known in the Tafilalt and Maïder (Baidder *et al.* 2008).

From south to north, a first Devonian succession consists of limestone and shale with a thickness reaching 9 meters. The dark limestone beds are highly recrystallized and thin sections show wackestone to packstone texture and some grainstone. The fossil content is moderately varied, with a microfauna represented by foraminifera and ostracods, and by bioclasts in fragmentary orthocones, dacryoconarids, and crinoids. As for brachiopods, Hadri (1997) recognized the genera *Uncinulus pila* and *Glossinulus mimicus*, which relate to the lower Emsian.

The second level is made up of yellow, cavernous dolomites, organized in sequences of thick- and thin beds (Fig. 15B). A polygenic breccia, consisting mainly of reworked limestone pebbles, forms the unit's base (Figs. 14, 15C). The same type of contact is observed 150 m further north, in the continuity of the outcrop. Clasts are subangular, of varying sizes, linked by carbonate cement with common encrustation

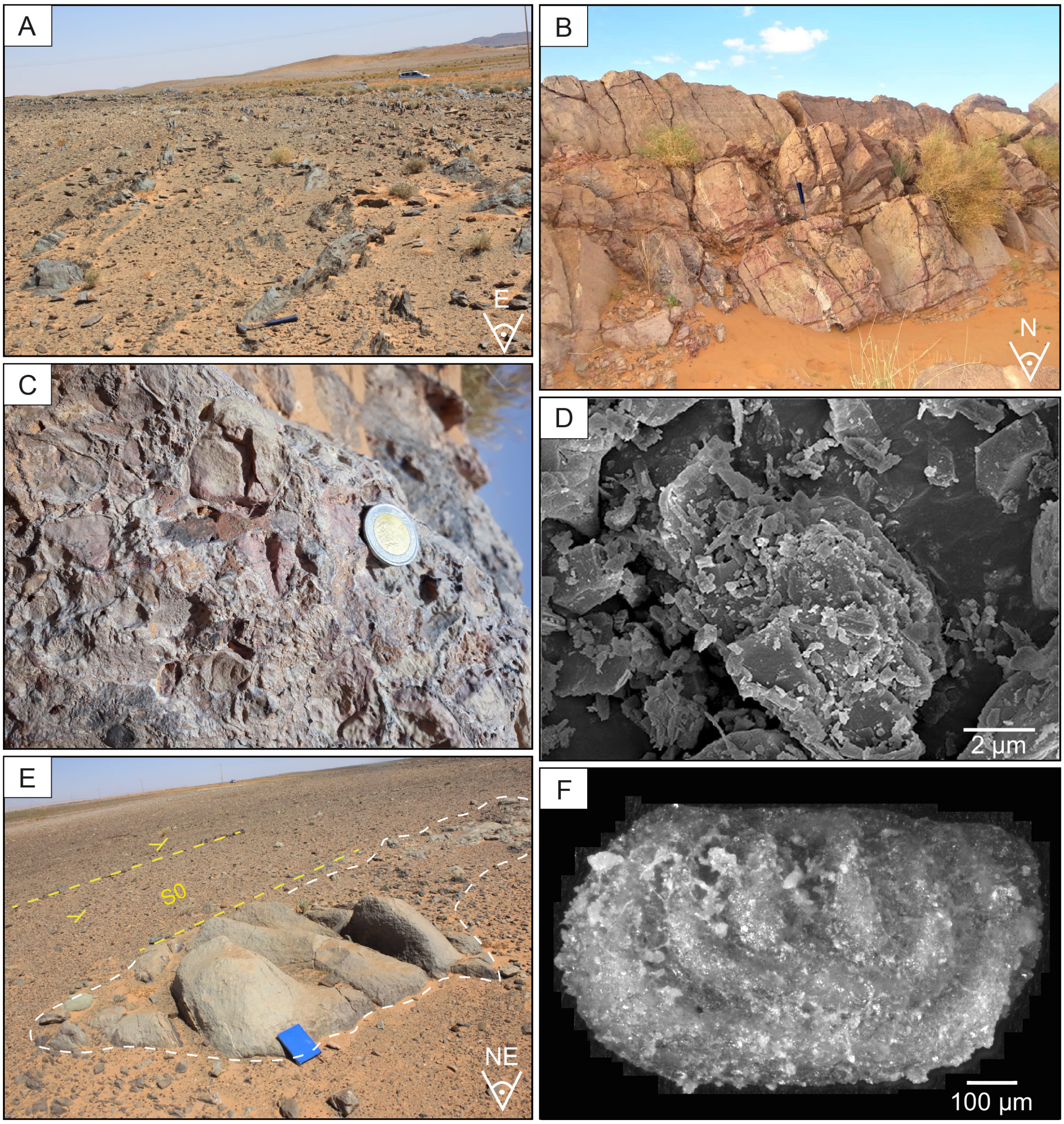


Figure 15. Field photos of Section Koudiat Inegh. A. Alternation of densely fractured/cleaved and dolomitized limestones and deeply weathered shales north of the marker dolomite. B. Marker dolomite member, organized in thick- to thin-bedded sequences. C. Polygenic breccia, which consists mainly of reworked limestone pebbles. D. SEM photo of automorphic dolomite rhombohedrals. E. Meter-sized, isolated limestone olistolite sitting within shale, succession north of marker dolomite. F. recrystallized benthonic ostracod *Polyzygia neodevonica neodevonica.*

features. Some levels show brecciation with ferromanganese oxide coatings. The matrix is micritic, recrystallized, peloidal, with quartz grains and shell fragments of turriform gastropods, ostracod valves, and crinoid stem pieces. Quartz grains have a diameter of less than 200 µm (fine sand fraction). They are poorly sorted, not graded, and angular in shape. This microfacies indicates a fairly advanced diagenetic state with recrystallization of the micrite to sparite, cementation, and dissolution.

A decimetric incision bed with bird-eyes carbonates, surmounted by a hard-ground surface, lies at the transition conglomerate – yellow dolomites. Bird-eyes dolomites are made of recrystallized carbonate mud with floating larger dolomite crystals. The yellow dolomites are formed by automorphic rhombohedral dolomite (Fig. 15D) with pseudomorphs of fossil remains. In the absence of any biostratigraphic data, the age of the dolomites is unclear. Hejja (2013) found an Emsian conodont (*Criteriognathus* cf. *steinhornensis*) at the top of these dolomites. The top of the dolomite unit on the northern slope is thin-bedded to lamellar; our conodont sample was barren.

After an outcrop gap of ca. 150 m distance, we found a ca. 90 m thick alternation of strongly cleaved and fractured, dark, dolomitic, sometimes laminated (turbiditic), unfossiliferous limestone and black schists, which dip uniformly (first with ca. 45°, then with ca. 70°) to the north. The lower part is shown in Fig. 14 and comprises levels with metric-sized carbonate olistolites (Fig. 15E). Close to the base, a conodont sample from a completely unfossiliferous mudstone yielded *Polygnathus partitus, Polygnathus costatus, Linguipolygnathus linguiformis*, and *Icriodus struvei*. This is clearly a lower Eifelian conodont fauna (*Polygnathus costatus* Zone/Subzone) from the upper water column of a hostile pelagic environment. The top bed of the upper succession contains a small conodont fauna with *Polygnathus xylus* and *Linguipolygnathus linguiformis.* This gives a lower/middle Givetian age but associated taxa are planktonic styliolinids and silicified valves (single and double) of a biostratigraphically meaningful, benthonic ostracod species. As the name implies, *Polyzygia neodevonica neodevonica* (Fig. 15F) is characteristic of the Frasnian, but the species enters already in the middle Givetian (e.g. Lethiers & Rachebeuf 1993). Therefore, the cleaved limestone-shale succession north of the dolomite represents most of the Eifelian to middle Givetian interval. The Kačák Event interval is not recognizable within its rather uniform basinal facies. The complete succession was hypoxic and strongly restricted environment for macro- and microfauna. The Middle Devonian of Koudiat Inegh contrasts very strongly with the closely adjacent, contemporaneous beds of Oued Ferkla (Units OF6-10), without evidence for a transitional zone (platform edge or slope). The distal turbidites show no evidence for bio- or lithoclats shed from the Oued Ferkla platform.

There is some support for Frasnian strata from a spot sample from 2010 with *Ancyrognathus tsiensi*, *Ancyrodella curvata* late form, *Palmatolepis hassi*, *Avignathus decorosus* s.l. (including the characteristic Pb element), and rare representatives of the *Icriodus alternatus* Group (close to *Icriodus kielcensis*), which suggest the MN 11 Zone (Klapper 1997) at the base of the Upper Frasnian. Unfortunately, the 2010 sample cannot be placed in the new section. Wendt & Belka (1991) showed a thin (only 0.2 m) dolomitized Kellwasser facies belonging to the Upper Frasnian in the north of the Oued Ferkla.

## CLAY MINERALOGY

The clay mineralogy in the Devonian of the Tinejdad region provides details concerning the clastic source as well as information on conditions of their deposition. It allows the characterization of environments and provides information in relation to the recognized sea-level changes (Ben Bouziane 1995, Han *et al.* 2000, Günal-Türkmenoğlu *et al.* 2015). The paragenesis of clay minerals is also able to indicate palaeoclimatic conditions during sedimentation (Millot 1964, Slansky 1980, Singer 1984, Curtis 1990, Thiry 2000).

X-ray diffraction was carried out on 145 samples that cover a time interval of 48.7 million years (Emsian to Famennian). The main clay minerals found are illite, kaolinite, and chlorite. The Emsian interval is marked by the dominance of illite with proportions varying between 71 and 82%. Kaolinite is present in low contents, between 10 and 16%; chlorite is estimated at between 8 and 13%. These general proportions are maintained with slight variations upsection. In the Eifelian, the clay fractions contain illite (67 to 80%), kaolinite (13 to 18%) and chlorite (7 to 15%). In the Givetian, the trend remains the same, still with a dominance of illite (69 to 77%) accompanied by low contents of kaolinite (13 to 16%) and chlorite (10 to 15%). The values for the Frasnian are 71 to 82% illite, 10 to 13% kaolinite, and 8 and 16% chlorite; and those for the Famennian are 68 to 79% illite, 11 to 13% kaolinite, and 10 to 19% chlorite.

The abundance of illite and the low proportions of kaolinite and chlorite suggest very arid greenhouse climate conditions (Slansky 1980, Singer 1984, Curtis 1990, Thiry 2000, Riquier 2005). The Tisdafine Basin received mostly a low detrital supply in a dry environment. Supports comes from the palynological content, which characterizes the Devonian basins of the Anti-Atlas and the Western Meseta (Snape 1993, Rahmani & Lachkar 2001). Unit OF2, the main siliciclastic interval of Oued Ferkla, appears to have been an exception.

## SYNSEDIMENTARY TECTONIC MOVEMENTS

The Devonian and Tournaisian of the Western Meseta and Anti-Atlas were characterized by intense Eovariscan tilted block tectonics, which preceded the main Hercynian compression phase. The Eovariscan phase, attributed to the Famennian-Tournaisian, is polyphase and relatively long (10 Ma) (Pique & Michard 1989). In the Western Meseta, it corresponds to extensional tilt blocks, uplift, and reworking sedimentation (Pique 1975, El Hassani 1990, Baidder *et al.* 2008). The Eastern Meseta corresponds to folding, thrusting, and metamorphism (Pique & Michard 1989, Hoepffner 1987, Michard *et al.* 2010). However, the reality of Famennian syn- metamorphic folding has been questioned recently by Ouanaimi *et al.* (2019). In our case, in the Tisdafine Basin, all syn-sedimentary crustal movements are consistent with an extensional regime of the Devonian age. At Bou Tisdafine, there are the three Givetian breccia units, which represent seismically triggered depositional events. In-situ brecciation, the formation of flat micrite slabs and clasts was followed by gravitational transport down the slope of an active fault scarp. Rapid lateral thickness changes are typical for debris flow talus. The strong iron impregnation of the third breccia (Fig. 7C) is not related to tectonism. However, it reflects long times of sea-level and climate-driven extreme condensation before and after the redeposition event. The sedimentary gaps at the base and top of the series are evidence of this tectonic instability. In addition, an iron solution of terrestrial origin accumulated in the marine environment during extreme sedimentary starvation. Less extensive iron incrustations are found in other disconformity horizons (Fig. 7B).

The Eovariscan breccias/conglomerates indicate the crustal fragmentation into tilted blocks as widely reported in the Western Meseta (e.g. Piqué 1979, El Hassani 1990, Fadli 1990, Tahiri 1991, Zahraoui 1991). More specifically, they represent the major synsedimentary block faulting episode that characterizes the Givetian to basal Frasnian of the Meseta (Becker *et al.* 2015, Becker & El Hassani 2020) and the northern Maïder to the south. In the latter region, it partly triggered massive rock falls, mass/debris flows and slump folds, followed by uplift that ended the sedimentary record in the Ouihlane-Tarherat belt (e.g. Hollard 1974, Wendt *et al.* 1984, Schröder & Kazmierczak 1999, Fröhlich 2004, Stichling 2013), which is the closest Devonian SE of the Jebel Ougnat (west of Mecissi, Fig. 2).

At Oued Ferkla, seismic activity is indicated by slumping and convolute bedding around the Kačák Event Interval, which is thought to record also distal debris flows indicative of an unstable palaeoslope. Earlier synsedimentary tectonics is indicated by normal faults with variable releases (centimetric to decimetric), mainly identifiable by significant on strike variations in thickness and facies (Fig. 10F). Such structures with distensive character are very abundant in the lower part of the section (Units OF4-6); their frequency decreases markedly at the top. The recorded deformations suggest compression in NE-SW to ENE-WSW directions, with which a NNW-SSE extension would be associated. It is a transtensive episode responsible for the movement of normal faults in the NE-SW accidents. As outlined below, the various faults are believed to be related to the opening of the Tisdafine Basin. Tectonic movements were also initiated in the Middle Devonian in the Tafilalt and Maïder (Wendt 1985, Baider *et al.* 2008, Baider *et al.* 2016).

At Koudiat Inegh, there is a breccia of uncertain age at the base of the dolomite unit. Isolated, meter-sized olistolites prove Eifelian-slumping in the lower part of the succession north of the dolomite. In addition, synsedimentary tectonics is further marked, particularly in the turbiditic facies. Finally, there are decimeter-scale slumps observed in beds with convolute-bedding structures. The latter are often affected by stretching and dilaceration.

## GEODYNAMIC TRENDS

The nature and age of Devonian sediments vary strongly from SW to NE within the Tisdafine Basin, from isolated, allochthonous olistolites in the SW, to the large glide block of Oued Ferkla, and to the autochthonous succession of Koudiat Inegh in the NE. Their different tectono-sedimentary history has to be interpreted separately. Due to the undetermined amount of displacements and outcrop isolation, it is not possible to provide palaeogeographical maps with meaningful spatial relationships.

### Bou Tisdafine glide blocks

At section Bou Tisdafine-West, the upper Pragian and lower Emsian are represented by a large, isolated glide block (olistolite) embedded together with minor detached lateral clasts within exposed silt- and sandstones of the Lower Carboniferous Aït Yalla Formation (Dal Piaz *et al.* 2007). They originated from a shallow pelagic carbonate platform/ramp, situated on the northern slope of the Jebel Sarhro-Jebel Ougnat axis (Figs. 2, 3) that was identified by Wendt (2021) as the original westernmost end of the Tafilalt Platform. Only the Cambro-Ordovician cover is partly still in place (Fig. 3); the Siluro-Devonian has been completely reworked and removed by Eovariscan block uplift. Abundant pieces of the same Emsian platform were shed as small clasts in the massive breccia at Taourirt n´Khellil (Hindermeyer 1955, Rytina *et al.* 2013). Above the massive breccia, just north of the Tinerhir-Tenjdad road, very similar olistolites as at Bou Tisdafine-West are imbedded in the greenish, fine siltstones of the lower Aït Yalla Formation (Rytina 2013, Rytina *et al.* 2013). There is no Bou Tisdafine outcrop that shows a transgression of Carboniferous strata onto an eroded lower Emsian platform or laterally younger Devonian strata, as suggested by Wendt (2021). All Devonian olistolites are embedded within the thick Carboniferous belt (Fig. 3).

The Bou Tisdafine section is another isolated olistolite embedded in the Aït Yalla Formation. It lies between the upper Tournaisian-Visean clastics of the Jebel Asdaf and the middle/upper Visean of Jebel Tisdafine (Graham & Sevastopulo 2007, Talih *et al.* 2022). However, both do not form a simple succession and the olistolite matrix is covered. The Devonian strata record the Eifelian to middle Famennian continuation of the allochthonous carbonate platform coming from the south. The Lower Devonian had been cut off before the olistolite emplacement. The Eifelian trilobite beds resemble the facies of the northern Maïder. At the same time, the condensed Givetian, with a very poor representation of coral limestones, is less similar to that region, apart from the breccia levels. The condensed Upper Devonian of the Bou Tisdafine section does not resemble the incomplete succession of Jebel Gherghiz (= Rheris; Fröhlich 2004), which is the only preserved Upper Devonian succession of the northern Maïder. Therefore, it can be assumed that the middle Givetian to basal Frasnian block faulting resulted in different platform developments southeast and northwest of the emerged Ougnat High. The rise of an Upper Devonian palaeo-island is supported by the reconstruction of currents and neodymium isotope distributions (Dopieralska 2009). It was recognized by Baidder *et al.* (2008) as the emerged Sarhro-Ougnat axis.

The timing of gravitational dislocation of the Devonian olistolites into the western Tisdafine Basin can be constrained by the ages of the youngest reworked strata (upper Famennian at Taourirt n´Khellil, middle Famennian at the Bou Tisdafine section) and the oldest ages known from the host beds of the Aït Yalla Formation (upper Tournaisian, Hindermeyer 1954). In the Tafilalt, the most distinctive synsedimentary tectonism phase of this interval was the middle Tournaisian (e.g., Kaiser *et al.* 2011, Tahiri *et al.* 2013). As a hypothesis, we postulate a similar timing of Eovariscan block tilting, resulting in reworking and redeposition, for the Bou Tisdafine area. It is hoped that further biostratigraphic constraints will become available in the future.

In other regions with Hercynian events, ante-Visean (Eovariscan) movements occurring between the Devonian and the upper Visean allowed the construction of an ante-Visean chain (in a broad sense) in the Western Meseta (Hoepffner 1987, Lahfid *et al.* 2019). They were responsible for the tectonic block tilting in central Morocco (e.g., Bouabdelli 1989, Fadli 1990, Tahiri 1991). It is possible to draw a parallel between our study region and the Meseta because in the terminal Devonian-Tournaisian, N70 is the direction of compression inducing the dextral setback of regional faults from NS to NE-SW, on which the basins opened (Zahraoui 1994).

### Oued Ferkla block

The Eifelian to Frasnian facies and faunas, the diagenetic overprint and the deformation style of the contemporaneous beds of the adjacent Oued Ferkla and Koudiat Inegh sections are incompatible. There is no transition of facies or diagenetic/tectonic overprint. Devonian of the Koudiat Inegh corresponds largely with the poorly studied basinal facies of the Touroug region some 30-40 km to the east (Hejja 2013), which represents an eastern expansion of the Tisdafine Basin. The most likely explanation is that the Oued Ferkla Devonian also represents an allochthonous unit that glided into the basin. Due to its strong similarity with the Tafilalt succession and the strong difference to the Bou Tisdafine glide blocks, it had a different provenance, from a northwestern extension of the Tafilalt Platform that reached perhaps the northern slope of the Jebel Ougnat (Fig. 2; see conclusions of Ward *et al.* 2013, and palaeogeography of Wendt 2021). Since the Oued Ferkla Devonian was spared from the dominant cleavage of Koudiat Inegh, its emplacement may have occurred late during the Variscan orogeny. The transport distance may not have been long, but the upper Emsian to middle Givetian congruence of Oued Ferkla and the Tafilalt is so complete that one would not hesitate to trace the original position in a NW continuation of the Erfoud-Jorf-Tantana belt of Devonian outcrops.

At Oued Ferkla, as in the northern Tafilalt, the complete succession belonged to a mixed carbonatic-siliciclastic outer shelf with pelagic conodont and ammonoid faunas and typical deeper-water (subphotic) benthos. The facies changed with eustatic sea-level changes between condensed shallow pelagic platform/upper ramp influenced by bottom currents (Units OF4, 6, 7, and 9), nodular to marly pelagic lower ramp or slope (Units OF 3, 5, parts of Units 6, 10), and pelagic shelf basin with restricted oxygenation (Units OF2 and 8). The presence of nodules is linked to common diagenetic processes, which characterize many Devonian basins, such as the Anti-Atlas, Western Meseta, and French Massif Central (e.g., Wendt 1985, Tahiri 1991, Cattanéo *et al.* 1993). Following an episodic cessation of sedimentation, calcareous surfaces with iron coatings, perforations, and crusts of endobiotic benthic organisms were formed.

In the Lower Devonian, the platform was suddenly drowned by the transgressive global Daleje Event, associated to an increased delivery of fine terrigenous material. This may reflect an associated episode of climatic change (humid interval) in the southern cratonic source region. Higher during the late Emsian, the platform first progradated basinward (Unit OF4), then retreated (Unit OF5). The Eifelian is characterized by variable facies, faulting, and cyclic facies fluctuations (Unit OF6). Deposition of black shales, specific to the Choteč and Kačák bio-events, occurred near the beginning and end of the stage.

The Kačák interval is characterized by eutrophication episodes, leading to mass occurrences of nowakiids, and contrasting evidence for deepening (laminated, basinal black shale deposition) and current-induced deposition. This is explained by a tectonic steepening of the basin slope, becoming unstable and releasing debris flows. The overall Eifelian tectonic instability can be correlated with the increasing disintegration of the Anti-Atlas south of the Jebel Sarhro-Jebel Ougnat axis into adjacent platforms and basins (e.g., Wendt *et al.* 1984, Wendt 1989), accompanied by synsedimentary normal faulting (Baidder *et al.* 2008).

In the lower and middle Givetian of Oued Ferkla, the pelagic carbonate platform was re-established by progradation during sea-level fall. The two *Pumilio* Events represent short-term interruptions with eutrophication as the cause for faunal blooms and increased organic matter accumulation. Later authors did not accept the tsunami scenario advocated by Lottmann (1990) (e.g. Aboussalam & Becker 2011, Hartenfels *et al.* 2018). The gradual deepening towards the top of the middle Givetian, accelerating during the late Givetian to middle Frasnian (change to pelagic marl facies), testifies to the transition from a pelagic platform into an outer shelf basin setting, which differs from the northern Tafilalt Platform. Since the eustatic curve displays no corresponding general or long-term sea-level rise, we suspect a terminal Givetian expansion of the Tisdafine Basin by accelerating subsidence at its southern margin. The similar timing of the seismic events recorded at the Bou Tisdafine section West and in the northern Maïder is unlikely to be a coincidence.

**Autochthonous Tisdafine Basin**

The early history of the autochthonous Tisdafine Basin is recorded in the Koudiat Inegh section and its equivalent strata a few km to the west and east (Fig. 3), extending further eastwards to the Touroug region (Hejja 2013). The age of the dolomite unit, which represents a strongly altered carbonate platform, is poorly constrained, but Hejja (2013) listed an Emsian conodont from the top. Furthermore, he found additional Emsian conodonts in strongly recrystallized, corresponding platform limestones of the Touroug area. Therefore, it can be postulated that the initiation of the Tisdafine Basin occurred close the Emsian-Eifelian transition. This is supported by our lower Eifelian fauna from the base of the thick black slate-limestone succession with turbidites north of Koudiat Inegh. The Tisdafine Basin originated obviously in the context of the platform-basin disintegration of all of the eastern Anti-Atlas (e.g., Wendt *et al.* 1984, Baidder *et al.* 2008).

The Tisdafine basin is characterized by synsedimentary tilted-blocks tectonics, deposition of distal turbidites, and the episodic slope collapses. This environment is thus responsible for rock falls and the accumulation of olistolites of variable size. Finally, no carbonate material resembling the Oued Ferkla platform is recognizable at Koudiat Inegh. This proves that the latter's lithological nature is foreign compared to the autochthonous sediments.

The basinal setting continued at least until the Frasnian. From the Touroug area to the east, lower Famennian reddish limestones with deep-water conodont faunas are known (Hejja 2013). However, due to the Quaternary cover, the contact between the autochthonous Middle/Upper Devonian early phase of the Tisdafine Basin and its main upper Tournaisian to Visean phase (Aït Yalla = Isfoul and Tinerhir Formations, e.g. Soualhine *et al.* 2003, El Boukhari *et al.* 2007, Graham & Sevastopulo 2008) is obscured.

The Eovariscan extensional premonitory stages of the Hercynian orogeny, resulting in narrow, rhombic block pattern, occurred in the Tafilalt and Maïder, mostly during the Eifelian to Famennian (Raddi *et al.* 2007, Baidder *et al.* 2008). The continuing Upper Devonian strong subsidence of basins with a predominance of fine detrital facies in the southeast (Tafilalt Basin) and center (Maïder Basin), and the emersion of the platform in the north (Margat *et al.* 1962, Wendt 1989, Dopieralska 2009), further attest unstable geodynamic conditions. Around the Devonian-Carboniferous boundary, strong subsidence shifted suddenly to the Amessoui Syncline and Jdaid regions in the southern Tafilalt (Kaiser *et al.* 2011), extending there to the upper Tournaisian (Oued Znaiguï Formation). But in the northern Tafilalt, a transgression resumed pelagic siliciclastic facies, lying disconformably on uppermost Famennian nodular limestones, only right at the end of Tournaisian and beginning of the Visean (Jebel Erfoud, Delepine 1941). This provides a possible link with the main sequence of the Tisdafine Basin. Furthermore, an upper

Tournaisian transgression may explain why Devonian basinal facies seems to be lacking from the western Tisdafine Basin.

## CONCLUSION

Despite its high significance at the transition from the cratonic Eastern Anti-Atlas to the Variscan Moroccan Meseta, the Devonian of the eastern Tisdafine Basin (wider Tinejdad region) has been incompletely studied in the past. Our combination of field logging, record of bed architectures, detailed macro- and microfacies analysis of lateral and contemporaneous sections, high-resolution biostratigraphic dating by conodonts and ammonoids, clay mineralogy, and the analysis of syn- and postsedimentary structural geology leads to a new synthetic reconstruction of the Devonian sedimentary evolution within and at the former southern margins of the basin. Our tectonic-sedimentary approach focused on changes of the paleoenvironment in time and space, especially on progradation-retrogradation patterns of the mobile carbonate platforms/ramps, reflecting relative sea-level changes, their correlation with known eustatic changes and well-known global events, and on changes of subsidence, reworking and redeposition related to synsedimentary block faulting and slope instability. The Devonian evolution of the Tisdafine Basin and its southern margins can be correlated precisely with the Devonian of the Anti-Atlas to the south and east of the palaeogeographically significant Jebel Sarhro-Jebel Ougnat axis. The main results are:

1. In the western part of the study area, two large olistolites embedded in the Lower Carboniferous (Sections Bou Tisdafine-West and Bou Tisdafine) provide a combined reconstruction of the upper Pragian to Famennian evolution of a carbonate platform/ramp that represents the original basin margin north of the Ougnat High.
2. The upper Pragian and lower Emsian of the Bou Tisdafine-West section represents a shallow pelagic carbonate platform/ramp. The trilobite-rich Eifelian of the Bou Tisdafine section resembles the neritic facies of the northern Maïder in the southeast of the Proterozoic belt. In the Givetian, the similarity decreases since there is only a very poor representation of reefal limestone. The overall setting returns gradually to a condensed shallow pelagic platform. Nevertheless, the middle Givetian to basal Frasnian seismic events, known widely from the Maïder and many Meseta regions, are well expressed by Eovariscan breccia beds.
3. Due to the uplift and emergence of the Jebel Ougnat, the Upper Devonian of the Bou Tisdafine area was isolated. However, the widespread Upper Kellwasser Limestone at the Frasnian/Famennian boundary could be identified, based on blooms of goniatites and records of index conodonts.
4. The Devonian of Oued Ferkla, just north of Tinejdad, had a very different facies history and belonged clearly to an extension of the northern Tafilalt Platform. It also represents an isolated, allochthonous glide block recording a pro- and retrograding carbonate platform originally forming the southern/southeastern basin margin. The facies remained shallow pelagic throughout the lower Emsian to Frasnian.
5. As in the Tafilalt, but unlike as at Bou Tisdafine, the transgressive and hypoxic episodes of the global Daleje, Choteč, and Kačák events are well expressed, as well as the two middle Givetian *Pumilio* Events that reflect short-termed eutrophication pulses.
6. Basinward carbonate platform progradation occurred during sea-level falls in the main part of the upper Emsian (Unit OF4), Eifelian (Units OF6-7), and in the lower Givetian (Unit OF9). The basal upper Emsian Daleje Interval is characterized by increased delivery of fine siliciclastics from the craton, perhaps due to climatically controlled erosion changes. However, the clay mineralogy indicates long-lasting arid conditions throughout the sequence. Eifelian limestones are rich in pelagic macrofauna and cyclic.
7. Synsedimentary tectonic movements can be recognized through the Eifelian and correlate with the general disintegration of the Eastern Anti-Atlas into platforms and basins at the same time, which represents increasing block tectonics (Wendt *et al.* 1984, Wendt 1989, Raddi *et al.* 2007, Baidder *et al.* 2008).
8. The Transformation of the middle Givetian pelagic platform at Oued Ferkla into an upper Givetian to middle Frasnian pelagic marl basin does not follow the known eustatic trend and, therefore, suggests increased subsidence, reflecting a southwards extension of the basin, and coinciding with the regionally widespread Givetian tectonic phase.
9. The thick succession of Koudiat Inegh records the sedimentary history of the true eastern Tisdafine Basin. It displays a very different, intense diagenetic overprint and deformation, including strong cleavage, in strict contrast to the allochthonous units. Obviously, it can be correlated laterally for several tens of km eastwards with the Devonian near Touroug.
10. The change from a dolomitized Emsian platform to a lower Eifelian hostile, outer shelf black shale basin with laminated, distal turbidites and small olistostromes, places the origin of the extensional Tisdafine Basin in the context of the Eifelian crustal segmentation of the Eastern Anti-Atlas. The strictly basinal facies, without transition to the Oued Ferkla platform, continued through the Givetian to Upper Devonian.
11. The Devonian basinal facies north of Tinejdad represents the tectonically controlled early history of the eastern Tisdafine Basin. There is no outcrop continuity with the thick, main upper Tournaisian to upper Visean basin sequence of the Bou Tisdafine region. There, knowledge of the Devonian is restricted to allochthonous platform remnants of the basin margin. This suggests an important role for the upper Tournaisian transgression that is also known from the northern Tafilalt.
12. While the facies history of remnants of the southern basin margin strongly resembles to Tafilalt and Maider platform/ramp successions, the strongly different, deep-water type autochthonous Devonian deposits of the Tisdafine Basin indicate a transition towards the less calcareous, even deeper, siliciclastic basin facies of the Eastern Meseta in the north and northeast. The intermediate nature in terms of geography and facies supports the view that there was no major structural boundary between the eastern Anti-Atlas and the Meseta in Devonian times.

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