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INTERNATIONAL FIELD COURSE

The kess kess mounds and
the Upper Ordovician carbonate formations of Eastern Tafilalt

Guide book

Upper Ordovician carbonate formations of Eastern Tafilalt: facies, depositional models and control

Author and leader
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CONTENT

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INTRODUCTION

During Ordovician times, the north and western margins of Gondwana was occupied by a large shallow oceanic domain: the Gondwanan platform that developed during a global extensional period and under the influence of the upper Ordovician Saharan ice cap

The Ordovician series of Morocco that belong to this platform, are widely exposed in all structural domains of Morocco (Fig. 1), especially in the Anti-Atlas where, there are some remarkable successions in both extent and quality of outcrop and where, here is a continuous succession from the Tremadoc to the Ashgill. These successions recorded during the upper Ordovician a glaciation and a major extensional tectonics.

In the eastern Anti-atlas, the extensional tectonic event resulted in the individualization of two sub basins: the « Khabt Lahjar sub basin » and the « Western Tafilalt sub basin », where exceptional environments developed under the interplay between tectonics and glaciation. In the « Khabt Lahjar sub basin » took place an isolated carbonate platform where bryozoan mounds nucleated and a mixed siliciclastic carbonate high energy peritidal littoral.

The objective of this field workshop is to illustrate the impact of these global events on the sedimentation processes and to specify the depositional systems of the upper Ordovician of the « Khabt Lahjar sub basin » with help of the successions of Khabt Lahjar and Rosfa El Hamra. It will provide the opportunity to examine facies and structures of cold (frost dominated) peritidal mixed siliciclastic carbonate high energy littoral, tide dominated delta, oolitic ironstone and glaciogenic deposits

GEODYNAMIC AND GEOLOGICAL SETTING

The Ordovician North -Gondwanan platform

The sedimentation during the Ordovician occurred in the north-Gondwanan platform. The reconstruction of this province is based on the geophysical (paleomagnetic) and geological (palaeontologycal, lithostratigraphic sedimentological, tectonic, volcanic) data.

Paleomagnetic reconstructions in the peri-Atlantic Palaeozoic belts during the Ordovician times (Scotese et al., 1979 ; Bambach et al.,1980 ; Scotese and Mc Kerrow, 1990 ; Van Der Voo, 1988, 1993), show three major continental blocks: Laurentia, Baltica and Gondwana separated by the lapetus and the mid-European oceans and numerous microcontinents or terranes (Fig 2 et 3). They also indicate that the Gondwana continent including the crotons of: South America, Africa, Arabian Peninsula, Madagascar, lndia, Antarctica and Australia, extended between the South Pole (where was located the western north part of Africa) and the Equator. The north-western peri-Gondwana terranes were detached from Gondwana by post-Panafrican peripheral introcratonic rifting and drifted away during the Ordovician (Erdtman, 1996). Those of central and south-western Europe and Appalachian eastern coast of United States (situated at the north periphery of Gondwana) being part of Armorica microplate (Van der Voo, 1979). Armorica became separated from the Gondwana at the end of the Ordovician, underwent a rapid drift towards the north (Fig. 4) and collided Laurusssia by Middle or late Devonian times (Perroud, 1985 ; Perroud et al., 1983, 1984). This displacement would have resulted in the opening of the Proto Tethys ocean. However, sedimentary and faunal evidence does not support this hypothesis (Babin et al., 1980 ; Noblet and Le Fort, 1990 ; Robardet et al., 1990).

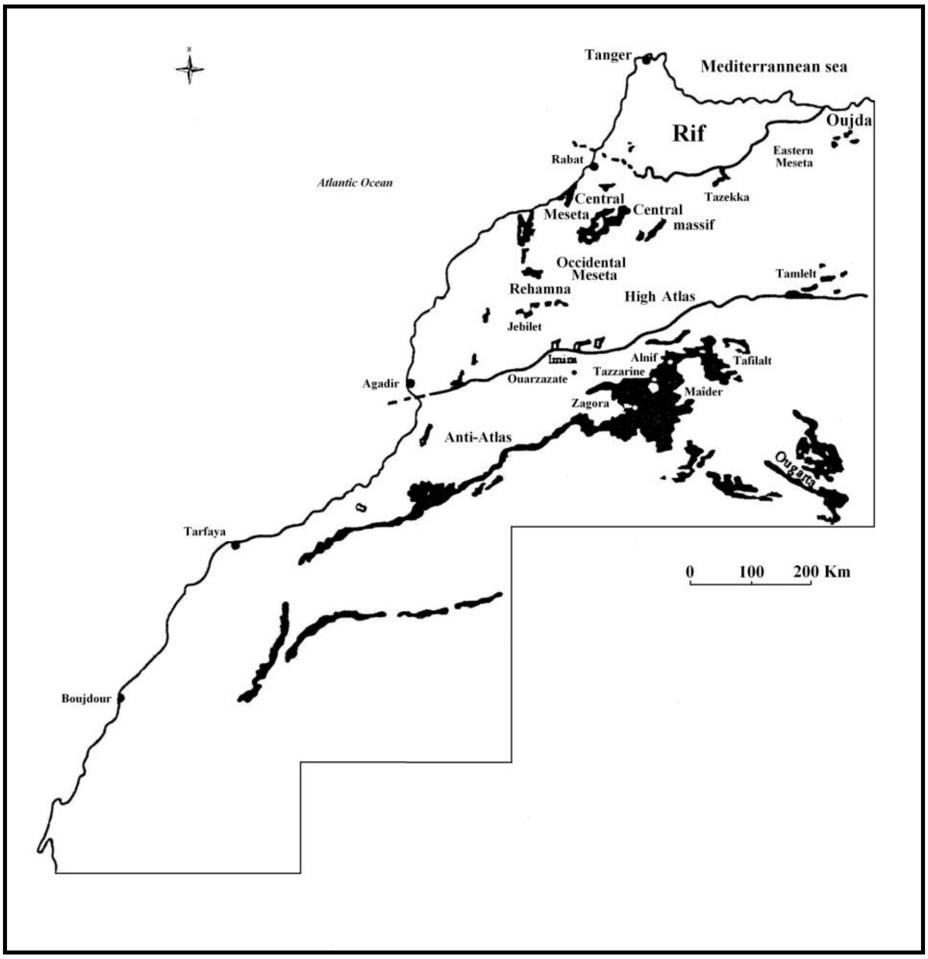


Fig. 1: Ordovician outcrops of Morocco

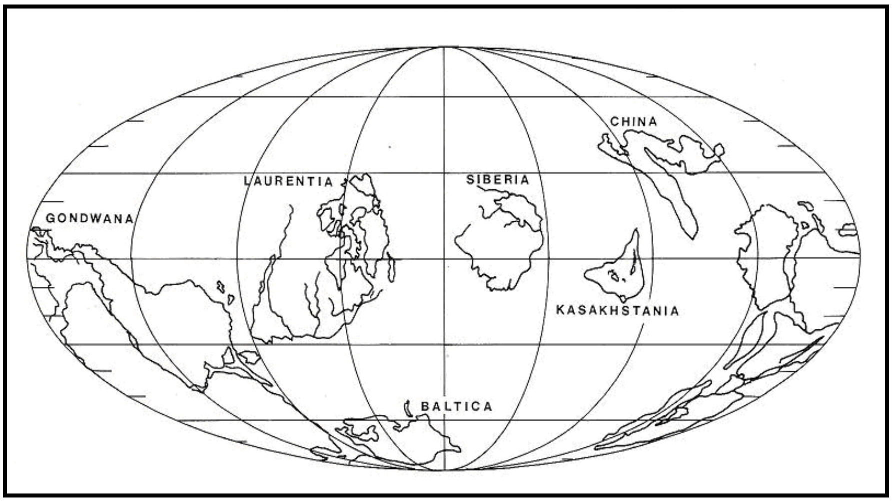


Fig. 2: Continents position during middle Ordovician (Bambach & al., 1980)

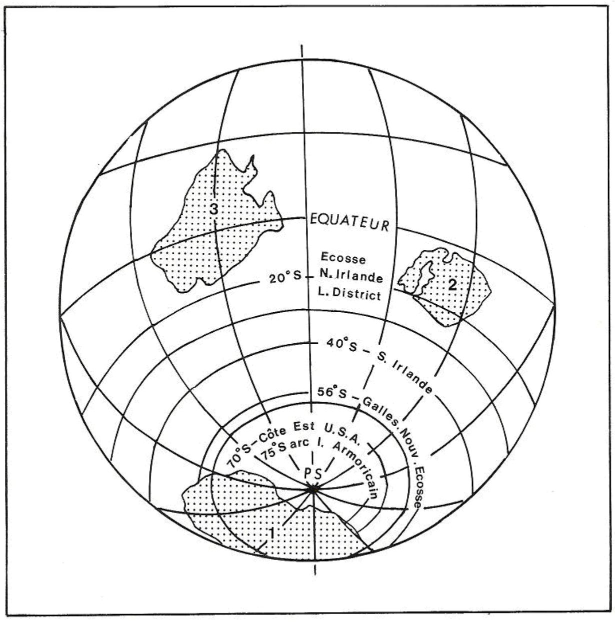


Fig. 3: Ordovician paleolatitude of Palaeozoic terrane s and cratonic zones (Perroud et Van Der Voo, 1984)
1 Gondwana
2 Baltica
3 Laurentia

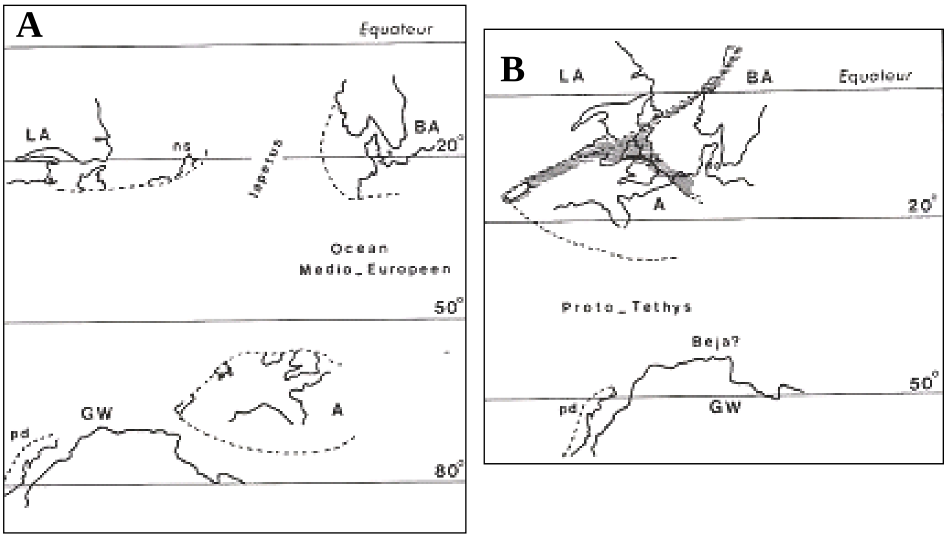


Fig. 4: **A**/ Ordovician paleogeographic reconstruction et **B/** Devonian paleogeographic reconstruction after the closing of Iapetus and midi-European oceans
A= plaque Armorica ; GW= Gondwana ; LA= Laurentia ; BA= Baltica (Perroud, 1985)

The pattern of climatic zones based on faunal provinces for Ordovician times indicate that North Africa and the Southern-Central Europe belonged to the same cold-water high latitude "Mediterranean province" (Spjeldnaes, 1961 and 1981) that extended from Florida up to the middle East (Whittington, 1953 ; Fortey and Morris, 1982). This outline was corroborated by numerous studies (Chauvel 1966 a ; Dean, 1967 ; Whittington and Hugues, 1972 ; Whittington,1973 ; Skevington 1973; Williams 1973 ; Havlicek, 1976 ; Babin et al, 1980 ; Cocks and Fortey, 1982, 1988, 1990 ; Paris, 1990 ; Paris and Robardet, 1990 ; Martin 1982 ; Mergl,1983 ; Mc Kerrow and Scotese, eds 1990 ; Linan et al., 1996; Le Menn and Spjeldnaes, 1996 ; Destombes and Babin, 1990 ; Villas, 1985, 1995 ; Villas et al., 1999). This "Mediterranean province was separated from Baltic province by the Mid European Rheic ocean (Paris and Robardet, 1990) or by two successive oceans: the Tornquist's sea, which closed during Upper Ordovician and the Rheic sea, which opened during the middle Silurian and persisted until Carboniferous (Cocks and Fortey, 1982, 1988, 1990 ; Fortey and Cock, 1992). The biofacies and the ichnofacies identified in the Gondwanan province are consistent with shallow platform environments (Crimes and Crosseley, 1968; Babin, 1966, 1993; Babin et al.1980; Destombes et al. 1985; Vidal, 1996).

Inside the "Mediterranean province, the Ordovician fauna of the Anti-Atlas (Morocco) that present some affinities with that one of Bohemia and "Montagne Noire", are characterize by numerous endemic species: Brachiopods (75%), Echinoderms (70%), Trilobites (70%), and Mollusks (10%) (Havlicek, 1970, 1971 ; Mergl, 1981 ; Chauve l, 1966 a, b and c, 1969 a and b,1971 a and b, 1977, 1978; Destombes 1963 c, 1966, 1967 a and b, 1972; Destombes et al. 1985 ; Horny, 1997 a and c).

The available lithostratigraphic and sedimentological data show strong similarities between the Ordovician series of the Gondwanan platform. The successions that are mainly siliciclastic, include intercalations of limestone or mixed siliciclastic/carbonate sediments, horizons of oolitic ironstones or glauconies and localized volcanic flows. During the upper Ordovician, some series can also include sediments of glacial origin and / or carbonate build-ups.

The sedimentary facies indicate storm and tide-dominated shelf environments in: Spain, France, Morocco, Portugal and Bohemia (Bradshaw 1963, 1966; Hamoumi, 1981, 1988; Guillocheau, 1983 ; Eshard, 1984 ; Noblet, 1984; Durand, 1985 ; Dafir, 1986; Hamoumi and Hoffert, 1986 ; Brencheley et al, 1986 ; Young, 1988 and 1990; Ouanaimi, 1989 ; Aramburu, 1989 ; Aramburu et al, 1992 ; Attou, 1992 ; Khoukhy, 1993, 2002 ; Chakrone, 1996, 2000; Kraft et al., eds., 1999).

The Ordovician period was characterized by a great production of oolitic ironstones (Fig; 5) especially in the margin of the Gondwana and it's different micro blocks (Petranek, 1964; Chauvel, 1971 ; Destombes ,1976; Chauvel and Massa, 1981 ; Joseph, 1982; Van Houten and Bhattachariyya, 1982; Guttierrez Marco et al, 1984; Van Houten, 1985 ; Guerrak, 1987 ; Van Houten and Hou, 1990; Hamoumi, 1988 ; Chauvel and Guerrak, 1989 ; Young 1989, 1992 ; Zahidi, 1994 ; Benbouida, 2002). These special deposits that occur often at the boundary of stratigraphic units and underlined major unconformities are excellent markers for paleogeographic reconstructions and basin analysis.

The record of the upper Ordovician glaciation is well documented in: Africa, Middle East, Europe, South America, Nova Scotia and Newfoundland (Fig.6), in numerous publications whose references can be find in some synthesis (Hambrey and Harland eds., 1981 ; Spjeldnaes,1981 ; Caputo and Crowell, 1985 ; Deynoux, 1985 ; Hambrey, 1985 ; Hamoumi, 1988 ; Robardet and Doré, 1988 ; Vaslet, 1990 ; Brencheley et al, 1991 ; Peralta and Baldis, 1992 ; Abed et al., 1993 ; Eyles, 1993 ; Ghienne, 1998). But there is still now two conflicting point of view concerning the duration of this glaciation. It has been considered to range from the Caradoc to Wenlock times (Frakes et al, 1992; Hamoumi, 1999 b), however, Brencheley et al, (1994), Paris et al, (1995), Marshall et al. (1997) and Sutcliffe et al, (2000), suggest a more short duration (1 to 2 My) according to the paleontological and sea level evidence and oxygen and carbon stable isotope data.

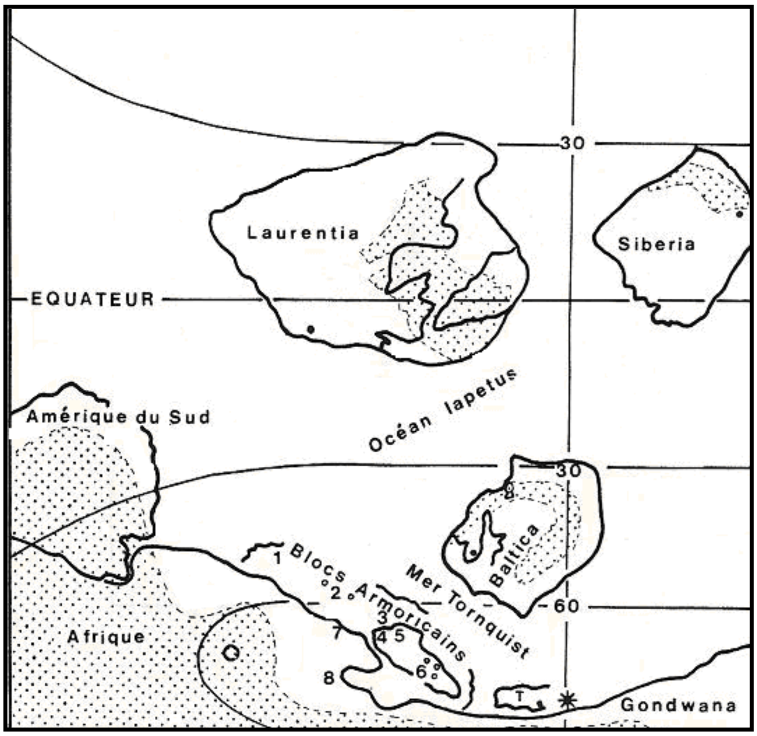


Fig. 5: Ordovician distribution of oolithic iron ore (Van Houten, 1985)

1. Upper Ordovician of Piedmont
2. base of Upper Ordovician in New Foundland
3. Lower Middle and Ordovician in Wales
4. Lower to Upper Ordovician in Spain and Portugal
5. Lower and Middle Ordovician in Britany and Normandy
6. Lower Ordovician of Cologna, Middle and lower Ordovician of Thuringia and lower and middle Ordovician in Bolivia
7. Lower and middle Ordovician in western Morocco
8. Lower and upper Ordovician in Libya

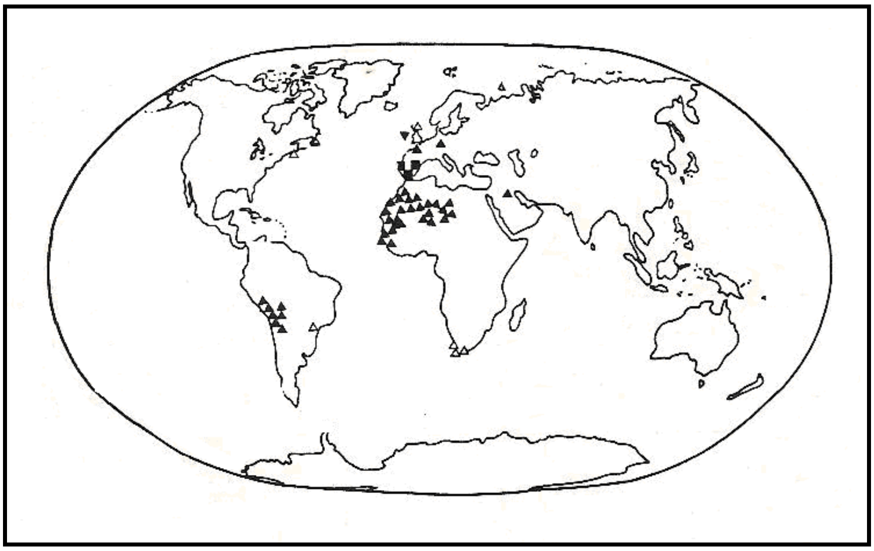


Fig. 6: Ordovician glacial deposits (Hambrey and Harland, 1981)

Upper Ordovician bryozoan carbonate formations arc known in some regions related to the north Gondwanan platform, in: Libya (Massa, 1988), Algeria (Le Grand, 1985), Morocco (Hamoumi ed., 1994 ; Hamoumi, 1996 a, 1997, 1998), Portugal (Young, 1988), Spain (Fuganti and Serpagali, 1968 ; Fortuin, 1984 ; Vennin et al, 1998), France (Lindstrom and Pelhate, 1971 , Weyant et al., 1977 , Hamoumi, 1981 ; Hamoumi et al., 1989 ; Hamoumi and Thonon, 1991 ; Havlicek, 1981 ; Paris et al., 1981), Thuringia (Knüpfer, 1967) and Carnic Alpes (Serpagli, 1967 ; Keppie ed., 1994). All these formations present strong similarities in facies and faunal associations (Bryozoans, Conodontes, Trilobites, Brachiopods) as well as in their relationships with glacial deposits and their stratigraphic position.

Volcanic rocks of upper Ordovician and Silurian age associated with distensiv conditions, are known, in: the Carnic Alps, Spain, France, Sardinia, Morocco and Portugal (Maillet, 1977 ; Boyer, 1979 ; Hamoumi, 1981 ; Cornée et al, 1985 ; Chalouan, 1986 ; Ribeiro et al, 1987 ; Young, 1988 ; Hamoumi et al, 1989 ; Hamoumi and Thonon, 1991 ; Leone et al, 1991 ; Guttierrez Marco, Saavadea and Rabano eds., 1992 ; Keppie ed., 1994 ; Guttierrez Marco et al, 1996; Truyols et a1., 1996). On the other hand, extensional movements of upper Ordovician age, has been recognized in: the Sahara (Fabre, 1976 ; Beuf et al, 1971), Mauritania, (Deynoux, 1980) and Morocco (Hamoumi 1988 ; Ouanaimi, 1989).

However, the Gondwanan platform was probably subdivided into several isolated basins as indicated by :

- the specificity of the Anti-Atlas fauna that includes several endemic species,
- the existence of two distinct paleogeographie eras in the Armorican Massif during the upper Ordovician (Hamoumi, 1981),
- the existence of two epeiric shelf basins in Morocco during the Ordovician: the "Atlasic basin" and the "Mesetian basin" (Hamoumi, 1995),
- the existence of basins separated by emerged land in Spain during the Ordovician (Aramburu, 1989).

The structural domains of Morocco

From south to north, Morocco is subdivided (on the base of outcrop materials and their tectonic history) into five structural domains (Fig. 7) corresponding to the: Saharan, Anti-Atlas, Atlas, Meseta and Rif domains (Piqué, 1983 ; Piqué and Michard, 1989).

The Saharan domain is a part of the West African shield. It corresponds to the Lower Proterozoic series that were highly deformed and metamorphosed during the Eburnean Orogeny (about 2000 Ma) and covered by non-deformed series varying from Paleozoic age (Tindouf Basin) to Mesozoic-Cenozoic (coastal plains).

The Anti-Atlas domain is constituted of: 1) the old basement and its Late Proterozoic cover that were overprinted by the Panatrican orogeny, 2) the Paleozoic cover which has been gently folded and uplifted during Hercynian Orogeny, 3) the thin and undeformed Mesozoic-Cenozoic cover.

The Atlas domain is composed of: the Middle-Atlas that runs across the Meseta and the High-Atlas that is situated between the Meseta and the Anti-Atlas. Its basement that outcrops as large Hercynian inliers are overlained by a relatively thick Permo-Mesozoic and Cenozoic cover that was folded during the Alpine Orogeny.

The Meseta domain corresponds to the Hercynian chain. uncomformably and partially covered by undeformed Mesozoic-Cenozoic deposits. It’s separated by the Middle-Atlas Tertiary foldbelt into two parts: the Moroccan Meseta and the Oran Meseta.

The Rif domain at the northern part of Morocco, belongs to the west alpine Mediterranean chain. It corresponds to an allochtonous alpine belt trusted southward and displaying Paleozoic terranes, which recorded Hercynian Orogeny.

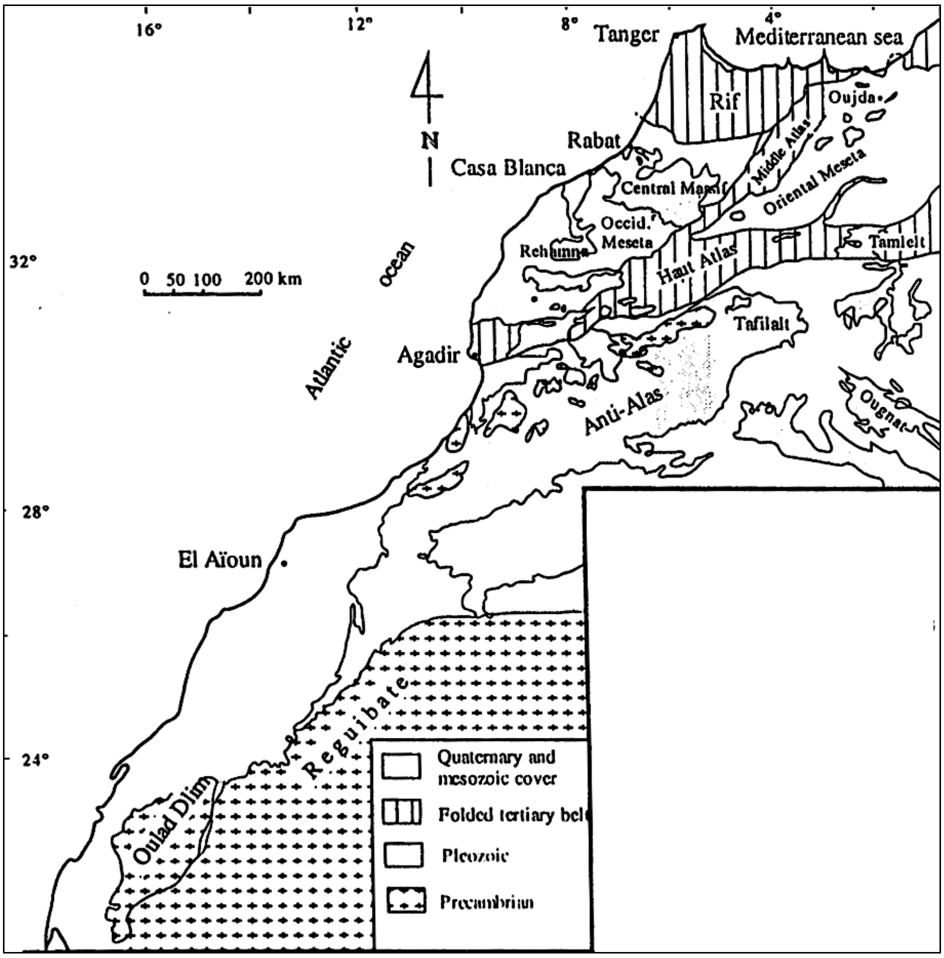


Fig. 7: The structural domains of Morocco (Piqué, 1982)

The Anti Atlas

The structural domains

The Anti-Atlas extent from the Atlantic ocean to the west until the “Hamadas du Guir” to the East (Figs. 8). It is separated from the High Atlas by the south atlasic fault and it is bounded to the south by “Hamada du Kem Kem”, “Hamada du Drâa”and Tindouf Basin. This domain is an anticlinorium formed by a Precambrian basement deformed during the Pan-African orogeny between -680 and -570 My (Hassenforder, 1987) and a cover of late Proterozoic to Palaeozoic age, weakly affected by Hercynian Orogeny (see references in Michard, 1976 and Piqué, 1994).

The Pan-African cycle comprises a major deformation episode that deformed the Quartzite series and reactivated the Eburnean basement. It induced the formation to depth along ductile zone of shearing and overlapping structures. It continued with a later orogenic phase with reverse faulting, tangential deformation and gravity y slides.

The Variscan episode has few effects in the Anti-Atlas. Distortions and orientations of folds have been guided by vertical movements of the basement's components (Le Blanc, 1975, 1981 ; Le Blanc and Lancelot, 1980). The cover was affected differentially by the basement structure and the movements produced remarkable disharmonic structures to the base of this cover. Some brittle or plastic distortions occurred in the Precambrian units. The effects of the Hercynian Orogeny were concentrated in narrow zones with the schistosity and epizonal metamorphism (Michard, 1976 ; Michard and Piqué, 1980 ; Jeannette, 1981 ; Jeannette and Piqué, 1981 ; Piqué and Michard, 1981 ; Hassenforder, 1987). Mega-folds of kilometric scale, are organized according to three structural directions (Choubert, 1963): 1) an Ougartian direction (NW-SE), 2) an atlasic direction (NE-SW to E- W) and 3) a submeridian mesetian direction.

The age of folds and the brittle tectonics is difficult to determine, Hollard (1978) signaled one phase in the Devonian and geochronology revealed an event of 400 My (Esquevin and Mendez, 1974). The last shortening is considered to have an age of 240 ± 10 My (Le Blanc, 1981). The Stephanian emersion corresponds to the Asturian phase and the later movements on the south Atlasic fault belonging to the Autunian -Saalian phase.

The Anti-Atlas is subdivided into three structural domains (Choubert, 1963): the western Anti-Atlas, the central Anti-Atlas and the eastern Anti-Atlas. The western Anti-Atlas spreads from the Atlantic Ocean to a transverse zone going from the Souss plain to Kheneg Tirerfa in the Bani. It is characterized by regular folds and disharmonic folds of submeridian mesetian directions, associated with schistosity and weak metamorphism.

The central Anti-Atlas extents from the eastern boundary of western Anti-Atlas to the Draa cross-valley and includes the massif of Siroua and the Bou Azzer-El Graara belt. It is characterized by box folds of atlasic direction (NE-SW to E- W) and a complex tectonic zone: the major Anti-atlasic fault of EW to ESE- WNW direction that induced thrusting and a graben structure toward Zagora. It is related to the basement structure that was removed during the Hercynian Orogeny (Choubert, 1947 ; Choubert and Marçais, 1952 ; Choubert et al.,1971 a and b ; Choubert et al, 1980 a and b ; Destombes and Hollard, 1989).

The eastern Anti-Atlas extends to the Guir Hammada it includes the Precambrian massifs of Jbel Sargho, Ougnat and Garat El Anes, the Paleozoic hills and belts of the Todhra-Maider and the Tafilalt- Taouz, the Hammada and the quaternary depressions and plains (Tafilalt, Tizini, Maider). The Jbel Sargho and the Ougnat are formed mainly of Proterozoic series that were subdivided into three units: the PC III cover (post- Pan-African Orogeny), the Aq n'Tanzit serie comparable to Anzi and Tidiline formations in terms of lithology and deformation and the PII unit (Clariond and Gubler, 1937 ; Le Collé et al., 1989 ; Rgimati et al. 1992). ln the Jbel Sargho, the Pan-African tectonics is complex with two folding events: a first phase with flexible fold axis trending N 30 to N80, warped toward the south with a regional epizonal metamorphism and a second late orogenic phase that gives a transverse structure to the chain trending N 130 with intermediate to acid magmatism and a post orogenic event with brittle tectonics (reworking of previous direction N70, N110-120). The Paleozoic hills and belts of Todhra-Maider and Tafilalt- Taouz consists of Cambrian, Ordovician Silurian, Devonian and carboniferous series that forms large anticlines and synclines of NE-SW, N-S, W-E or NW-SE trending, affected by NE-SW faulting (Clariond, 1935, 1944 ; Destombes et al, 1988, Destombes et al, 1986). The Hammada domains (Meski , Aoufous, Kem Kem, Bou Laouaiche..) consists of Cretaceous or Oligocene deposits.

Lithostratigraphy of Upper Ordovician

The upper Ordovician of the Anti Atlas is subdivided into two groups and six formations (Fig. 9), (Choubert et al, 1955 ; Choubert and Faure-Muret, 1956.; Destombes, 1960, 1962, 1971 ; Destombes et al, 1985).

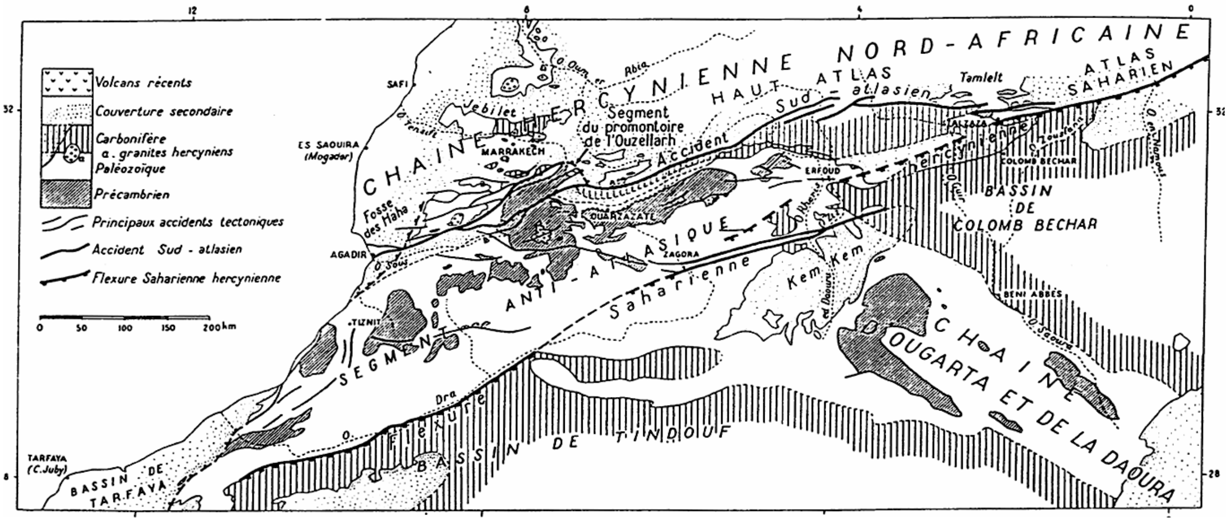


Fig.8: The structural domains of the High-Atlas and the Anti-Atlas
(Choubert and Faure-Muret, 1962)

The Ktaoua Group

The Ktaoua Group is entirely clayey at Ktaoua where its thickness reaches 300 to 400 m. Elsewhere, it can include 2 or 3 sandy bars. This group is divided into 3 formations: the lower Ktaoua Formation, the Tiouririne Formation and its lateral equivalent the Rouid Aissa Formation and the upper Ktaoua Formation.

<u>The Lower Ktaoua Formation (lower middle Caradoc):</u> It consists locally of grey- white argillite with thick carbonated spheroids rich in fossils. In certain regions it can includes two sandstone members. Its lower limit is marked, in places, by an oolitic ironstone level or ferruginous microconglomeratic sandstone.

<u>The Upper Tiouririne Formation and it's western equivalent: the Rouid Aissa formation (upper Caradoc):</u> These two formations are well developed in all the Anti-Atlas except in the plain of the Ktaoua and to the south of Jbel Tafenna. The Tiouririne formation is recognized in Maider and Tafilalt where it shows lateral variations of facies and thickness.

- At Jebels Amougueur, Assarnar and Isk-n'brahim, it is formed of dark micaceous sandstone with some intercalations of lumachellic limestones
- In the western Tafilalt, it rests on a conglomerate and it consists of dark green micaceous sandstone with pelitic-carbonate nodules
- In the region of Taguerroumt, the dark sandstone become thicker and well stratified and they terminate either in more argillaceous levels with pelitic-carbonate nodules or thin lumachellic and conglomeratic limestones with bryozoans
- At Jbel Imzezoui and to the north of Timjarfouine, it is entirely conglomeratic
- ln the Tinjdat-Ouzina axis, it is formed by thick flysch type arenaceous sediments
- At Khabt El Hajar, it includes the biostrome of bryozoan limestones localy at its base.

The Rouid Aissa Formation is restricted to the west of Tagounite where it is developed considerably. At Rouid Aissa it is constituted of pink quartzite (40 m) with some conglomeratic intercalations. Towards the west near Zaouia Sidi Abd en Nebbi, the Quartzites shows some Tigillites, inclined bedding and coarse horizons. Towards Tata and Akka, the coarse horizons develops, become highly ferruginous and include lumachellic levels. Farther to the west, towards Foum El Hassan they contain bryozoans and at their top several horizons of oolitic ironstones.

The upper Ktaoua Formation (Ashgill p.p): It is formed mainly of green grey arenaceous argillites with fossiliferous carbonate nodules and few rare levels of oolitic ironstone or phosphatic nodules. These facies appear in two regions of NNW- SSE orientation: The basin inbordered by the Grara and the Ougnat-Ougarta rise and the basin situated between Jbel Tschila and Adrar Zougar-Akka. Its thickness in these two domains doesn't exceed 90 to 100 m. Between Agadir-Tissint and the lriqui, it thins and includes coarse sandstone and in the Tafilalt-Maider, it is either greatly reduced or it disappears locally. It may overlain conformably the lower sandstone of the 2nd Bani or uncomformably by the upper sandstone of the 2nd Bani.

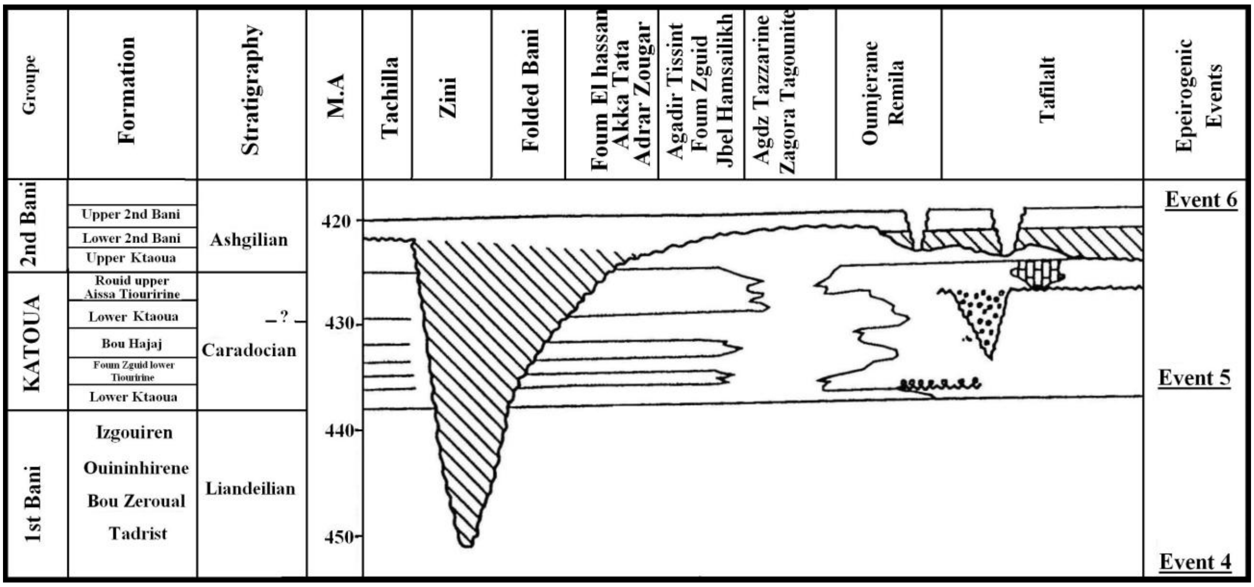


Fig. 9: Simplified lithostratigraphic shem of Anti-Atlas (after Destombes et al., 1985)

The 2nd Bani Group

The 2nd Bani Group is divided into 2 formations: the lower 2nd Bani Formation and the upper 2nd Bani Formation. This last formation lies on the first one through an unconformity in the whole central And-Atlas and may rest on any of the previous Ordovician formation depending on the region.

The lower 2nd Bani Formation (upper Ashgill): This formation is clearly individualized only in the central Anti-Atlas along a north south to north-west/south-east through. Hs maximum thickness may reach 160 m to south-east of Tagounite and at the Ait Isioul. It consists of bioturbated argillaceous sandstones with brown pelitic carbonate lenses, quartzoze sandstones and bedded quartzites with Tigillites, arenaceous argillites, coarse sandstones and coarse ferruginous sandstones with Tigillites. Its lower boundary is commonly marked by an interval of bryozoan limestone and its upper boundary is commonly eroded by the upper 2nd Bani Formation.

The upper 2nd Bani Formation (lipper Ashgill): The upper 2nd Bani Formation may forms a :

- thin coarse arenaceous cover (15 to 100 m), resting on unconformity over the old formations and composed of conglomerate (with granite and rhyolite pebbles), microconglomeratic arenaceous clay with exotic pebbles, coarse and fine sandstones and massive quartzite with soft pebbles,
- thick infillings of glacial channels (80 to 100 m) as coarse sandstones, reddish argillaceous sandstone, conglomerates with exotic pebbles and the ferruginous quartzites. The lower boundary of this formation, where sediments reflect the trace of glaciation, shows a glacial striation of north-south orientation at Jebel Hamsailikh. Its upper boundary is marked by the first facies with Silurian graptolites: the lower Llnndovery green argillites or more commonly, the middle Llandovery- upper Llandovery flaggy graptolitic sandstones

ORDOVICIAN DEPOSITIONAL SYSTEMS IN MOROCCO

Introduction

In north Gondwanan Moroccan margin, the Ordovician sedimentation occurred in two sedimentary basins the "Mesetian basin" and the "Atlasic basin"(Hamoumi, 1995). The "Mesetian basin" of north-east/south west direction, includes the Ordovician successions of: the Moroccan Meseta, the western High-Atlas and the northwestern Anti-Atlas domains. The "atlasic basin" of east-north-east/west-south-west direction, includes the Ordovician succession of the southwestern Anti-Atlas, the central and eastern Anti-Atlas and the central and eastern High Atlas domains. These two basins were initiated by peripheral intracratonic rifting of the north Gondwana margin.

The "atlasic basin" was initiated during early Infra-Cambrian (possibly late Precambrian) (Bensaou, 1990 ; Bensaou and Hamoumi, 1999 ; Bensaou and Hamoumi, 2003 ; Hamoumi ed., 1994) and the "Mesetian basin" was initiated only during the Lower Cambrian (Hamoumi, 1995). The rifting processes continue diachronously in these two basins according to three directions succeeding one to another in time and space: SW-NE, WSW-ENE and NW-SE. During the middle Cambrian, the "Atlasic basin" evolved as an epeiric shelf where the sediments (mainly siliciclastic) accumulated in storm and / or tide-influenced shelf (offshore to upper shoreface), under the control of sea level fluctuations, subsidence and episodic tectonics. In the south-east and east of the Ougnate massif, extensional tectonics was accompanied by intensive volcanic activity.

The mesetian basin was a highly subsiding graben limited by major normal faults (Cornée et al., 1987 ; Bernardin, 1988 ; Bernardin et al, 1988 ; Mayol, 1987 ; Corsini, 1988), where, the depositional systems (storm-dominated-offshore, transition zone and shoreface, and tide-dominated and storm-influenced littoral) occurred on a complex topography in horst blocks and graben limited by faults, which have influenced the sedimentation (Habibi, 1992 ; Habibi et al, 1992 ; Hamoumi and Habibi, 1992). Tectonic movements that were accompanied by volcanic flows produced: soft sediment deformation, variations in the sediment thickness, eustatic fluctuations and paleogeographic inversions.

The Ordovician evolution could be divided into two major stages: the lower and middle Ordovician period and the upper Ordovician period. During the lower and middle Ordovician period, the "Mesetian basin" and the "Atlasic basin" (Fig. 10 et 11) acted both of them as epeiric shelf that were dominated by shifting of the subsidence centres (reactivation stage), block faulting associated some times with basic volcanic flows and eustacy. The deposits, which accumulated in these two basins are mostly silicilastic with minor intercalations of mixed carbonate/ siliciclastic sediments, glauconitic sand, oolitic ironstones and localized volcanic and pyrocIastic rocks.

The Upper Ordovician period

The upper Ordovician was initiated by a transgressiv phase following a major sea level falling and emersion associated with an uplift and the ice cap growth. This period was marked in the two epeiric shelves by a glacial climate, subsidence and major extensional events. The glacial climate induced eustatic sea level fluctuations and new transport mechanisms (cold wind and glaciers) and controlled the rate and the nature of the sediment supply. The tectonic created a NE-SW through in the "Mesetian basin" (eastern Jebilet) and two NW-SE through and a mosaic of carbonate platform in the "Atlasic basin" (eastern Anti-Atlas). The sediments are mostly siliciclastic deposits (mainly coarse sand derived from the Saharan ice sheet) and carbonate from bryozoan mounds and banks (in the atlasic basin) associated with minor intercalations of oolitic ironstones and glauconitites. The paleogeography was widely modified by these two events and various environments occurred in the two basins.

Caradoc

During the Caradoc, in the "Mesetian basin" the sedimentation took place in glacial fjord, marine outwash fjord, fluvial system and mesotidal to macrotidal estuary, tidal flat, high energy wave and/or storm-dominated beach, wave and storm-dominated delta, tide-dominated littoral, storm-dominated shelf (offshore, transition zone and shoreface), fan delta and deep sea fan. In the "Atlasic basin", the ice cap growth leads to the formation of glacial surfaces: a glacial pavement situated at the Llandeilo-Caradoc boundary and a surface remnants at the corrie heads of lower Caradoc age (central Anti-Atlas), a frost dominated cold tidal flat at the Middle Caradoc-Ashgill boundary, in Eastern Anti Atlas and a shore ice dominated cold tidal flat at the Upper Caradoc - Ashgill boundary, in the Eastern Anti-Atlas. The sedimentation occurred in cold (frost dominated) peritidal mixed siliciclastic/carbonate high energy littoral, carbonate platform, storm-dominated continental shelf ice distal (muddy shelf, offshore transition zone, shoreface), fluvial and tide-influenced delta, fan delta and deep sea fan.

Ashgill

During the Ashgill in the "Mesetian basin", the important sea level falling associated with the ice cap growth at the Caradoc-upper Ashgill boundary produced an important emersion (coastal Meseta). Deposition realized in wave-dominated littoral, wave and tide-dominated littoral, intertidal flat, deep sea fan, storm and tide-influenced upper shoreface, wave and storm-dominated delta and macrotidal estuary. ln the "Atlasic basin", the ice cap growth leads to the formation of shore ice-dominated cold tidal fiat during the Hirnantian and at the Ashgill- Silurian boundary in the central Anti Atlas. Deposition occurred in: cold (frost dominated) peritidal mixed siliciclastic/carbonate high energy beach, carbonate platform, tide and storm- influenced temperate shelf (prograding sand ridge, tidal banks), mixed wave and tide-influenced temperate nearshore (ebb tidal delta), storm-dominated temperate shoreface, macrotidal estuary, tide dominated shelf (tidal banks), fan delta and deep sea fan.

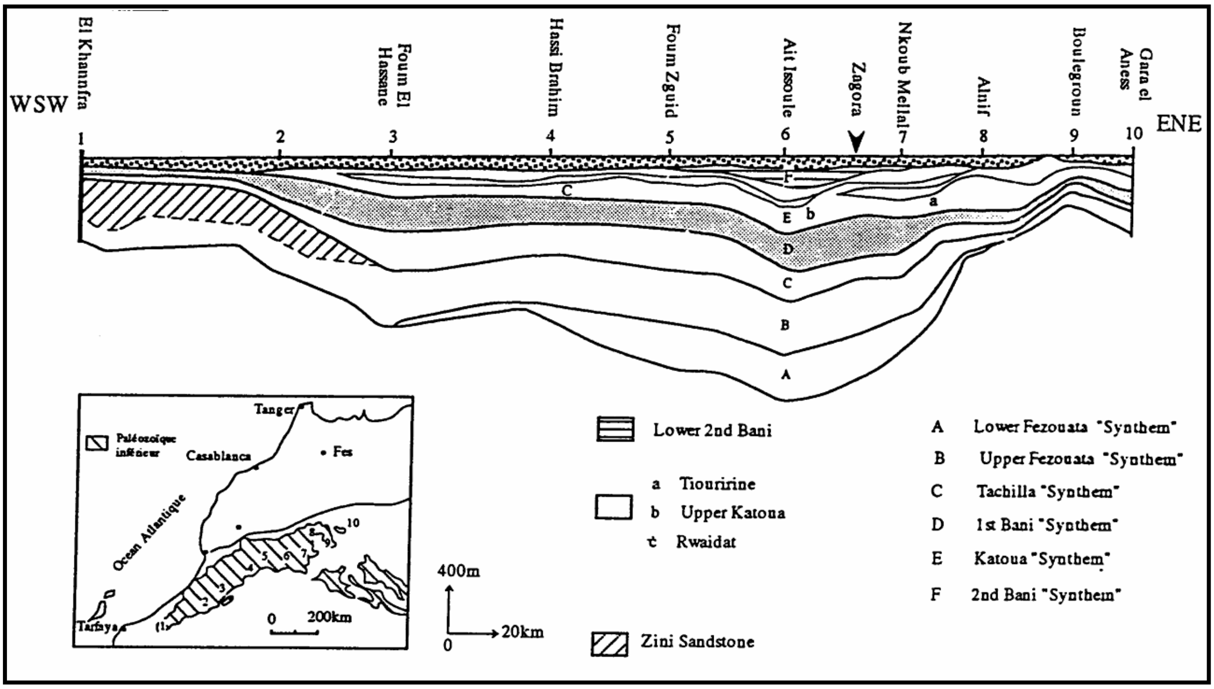


Fig. 10 : The Ordovician basin in the Anti-Atlas (after Destombes et al, 1985)

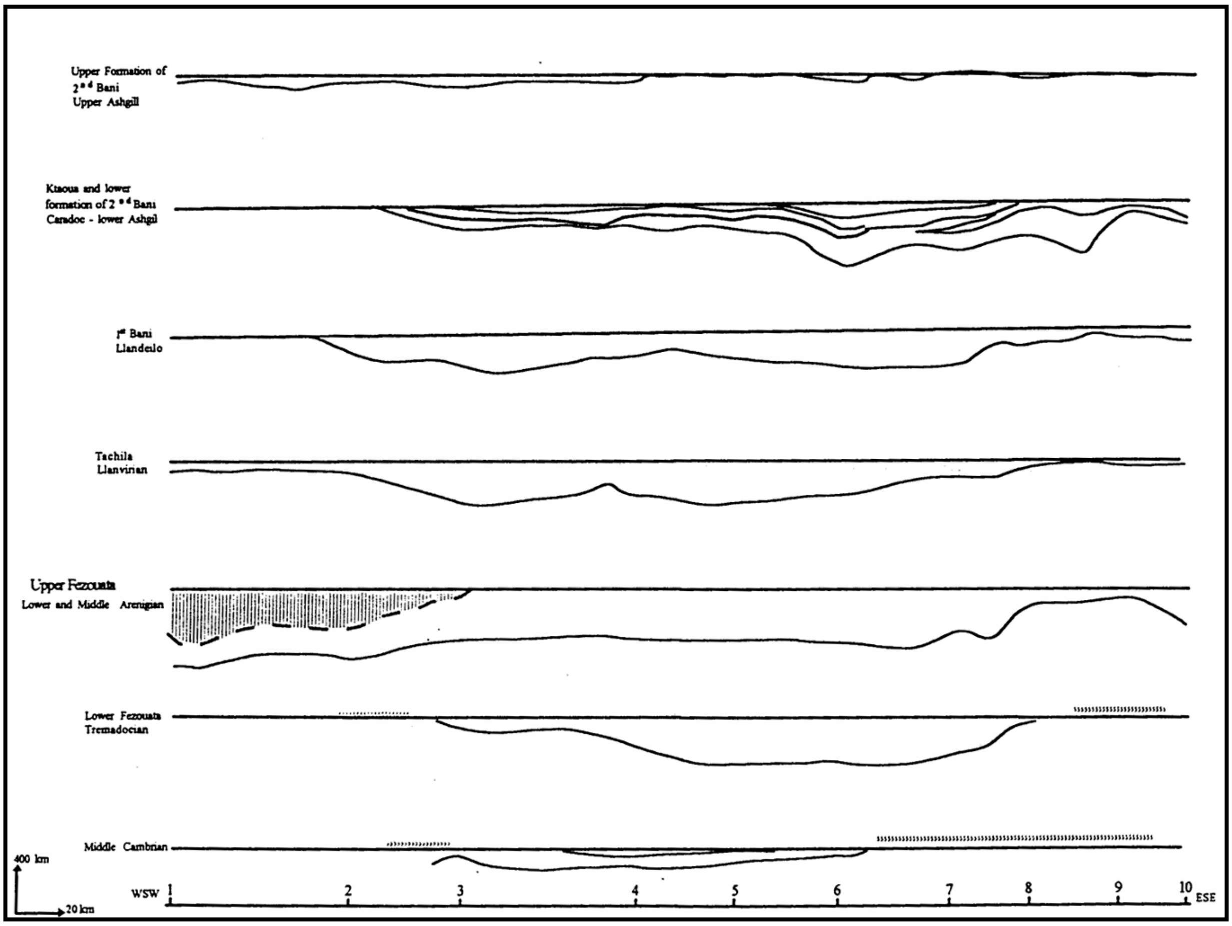


Fig. 11: Unfolded systems of the "Atlasic basin" (Hamoumi, 1988)

THE UPPER ORDOVICIAN OF EASTERN TAFILALT (EASTERN ANTI ATLAS)

Introduction

The eastern Tafilalt display exceptional and unique upper Ordovician series that are really different from the other Moroccan upper Ordovician successions and from the most upper Ordovician successions of the north Gondwanan platform. Indeed this domain where the oldest outcrops are of Lower Caradoc age, the upper Ordovician successions are characterized by a low thickness and extent ant a high contents of carbonate and bryozoans. The carbonate formations ("Biostrome à bryzoaires") of this upper Ordovician (Fig. 12) were discovered first by Clariond (1935, 1944), and their lithostratigraphy and biostratigraphy were studied by H. and G. Termier (1947, 1950) and Destombes (1987) Sedimentological studies (Hamoumi, 1996a, 1997, 1998, 2001 and El Mazzouz, PhD in progress) allow to recognize sedimentary facies and environments, mineralogic composition and sources and to understand the sedimentation control of this part of North Gondwanan platform.

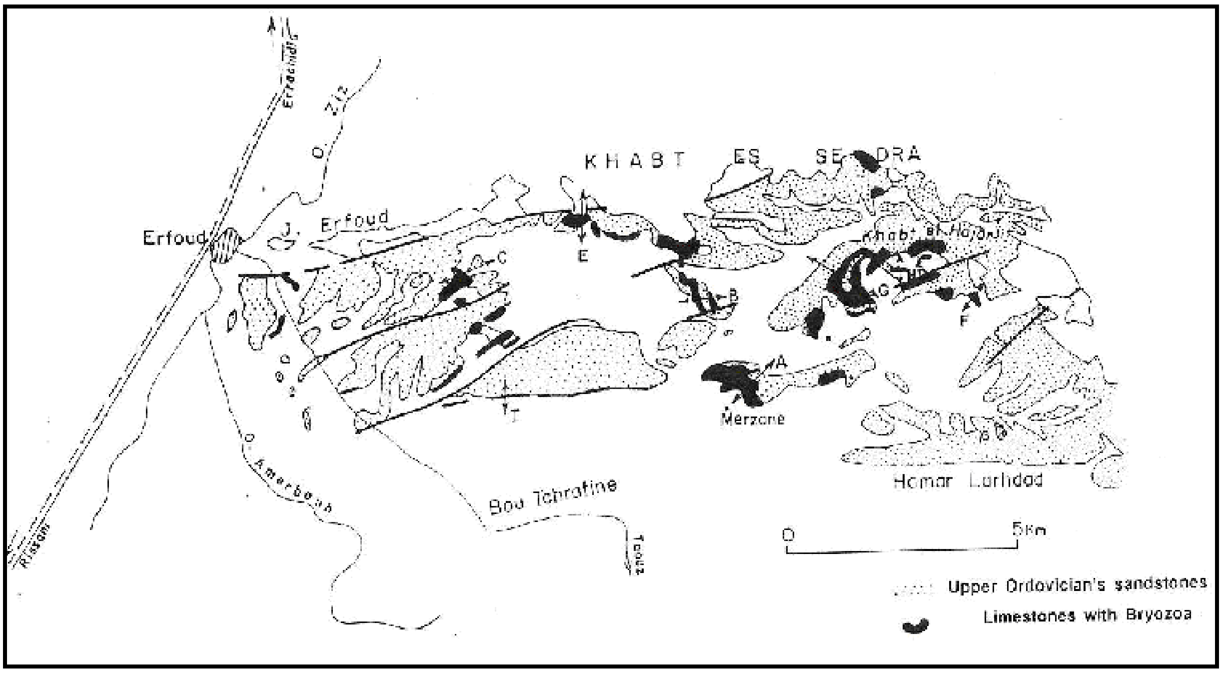


Fig. 12: Geological schematic map of Khat Lahjar area (eastern Tafilalt) showing the geographic distribution of Upper Ordovician Bryozoan limestones (Destombes, 1987)

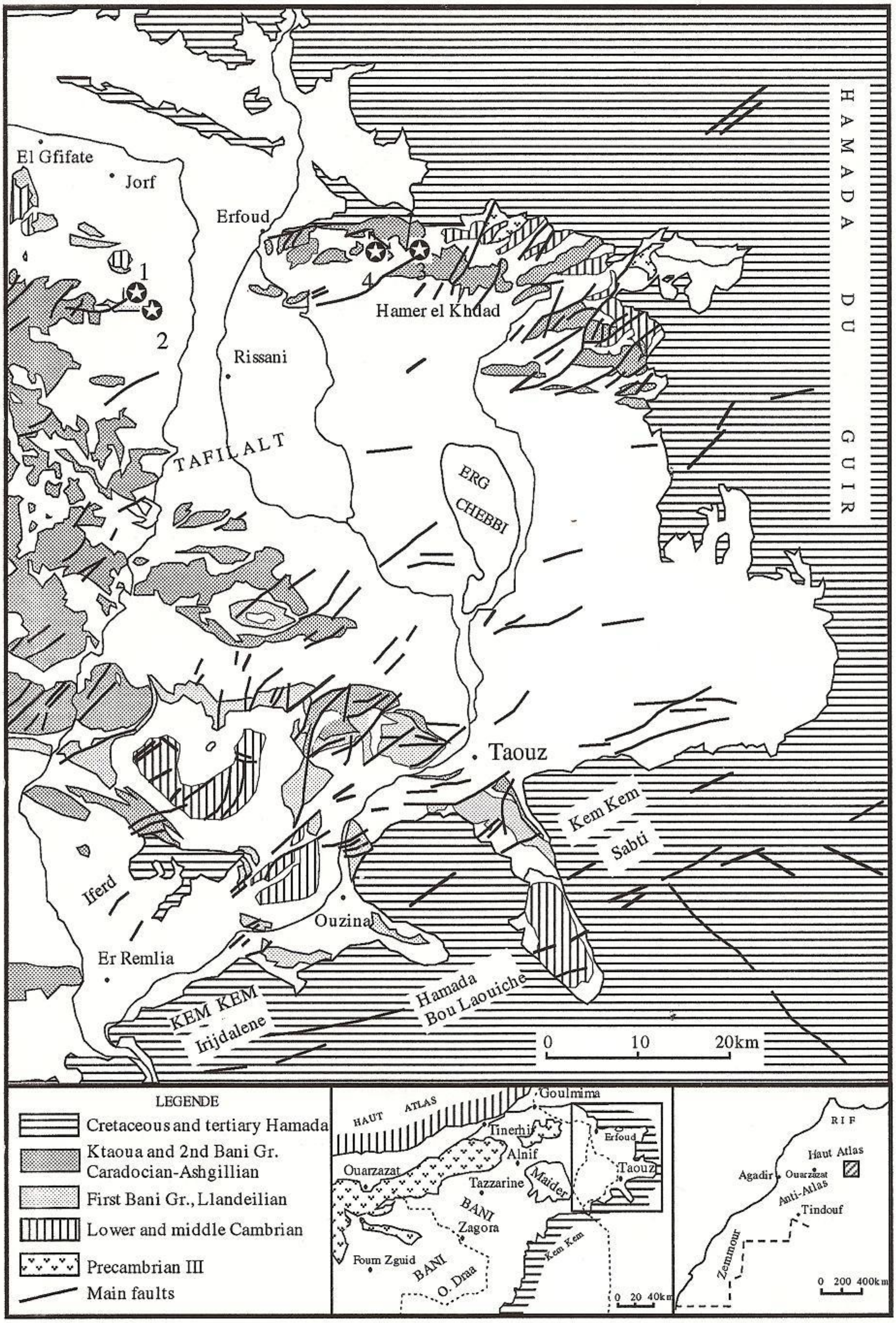


Fig. 13: Geological map of Tafilalt Taouz (Destombes, 1987)

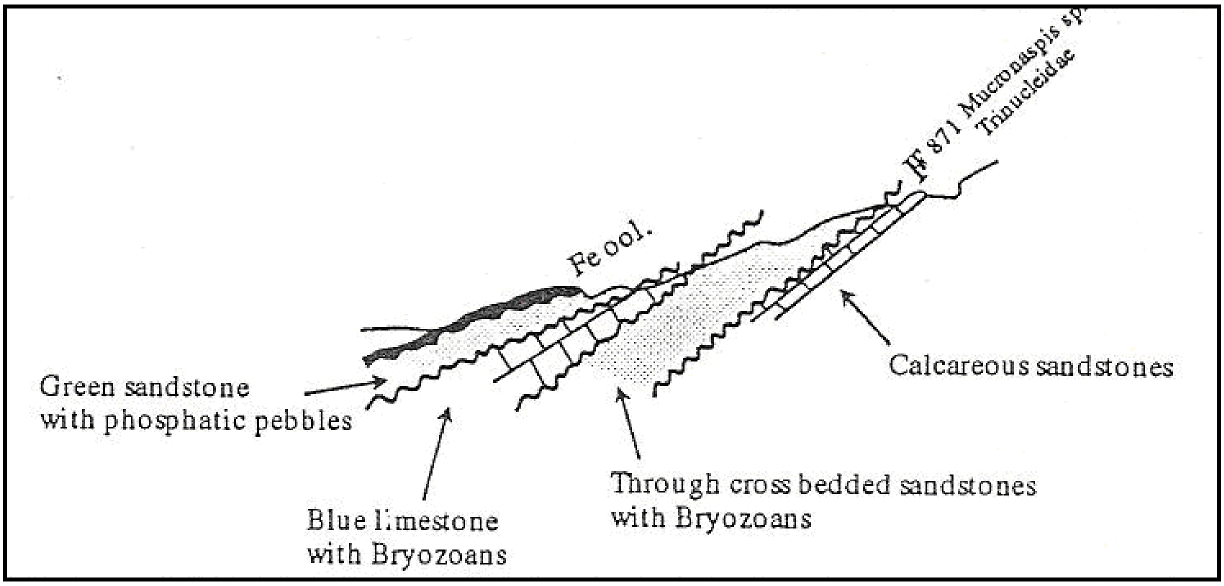


Fig. 14: Upper Ordovician of Khabt Lahjar succession (Destombes, 1987)

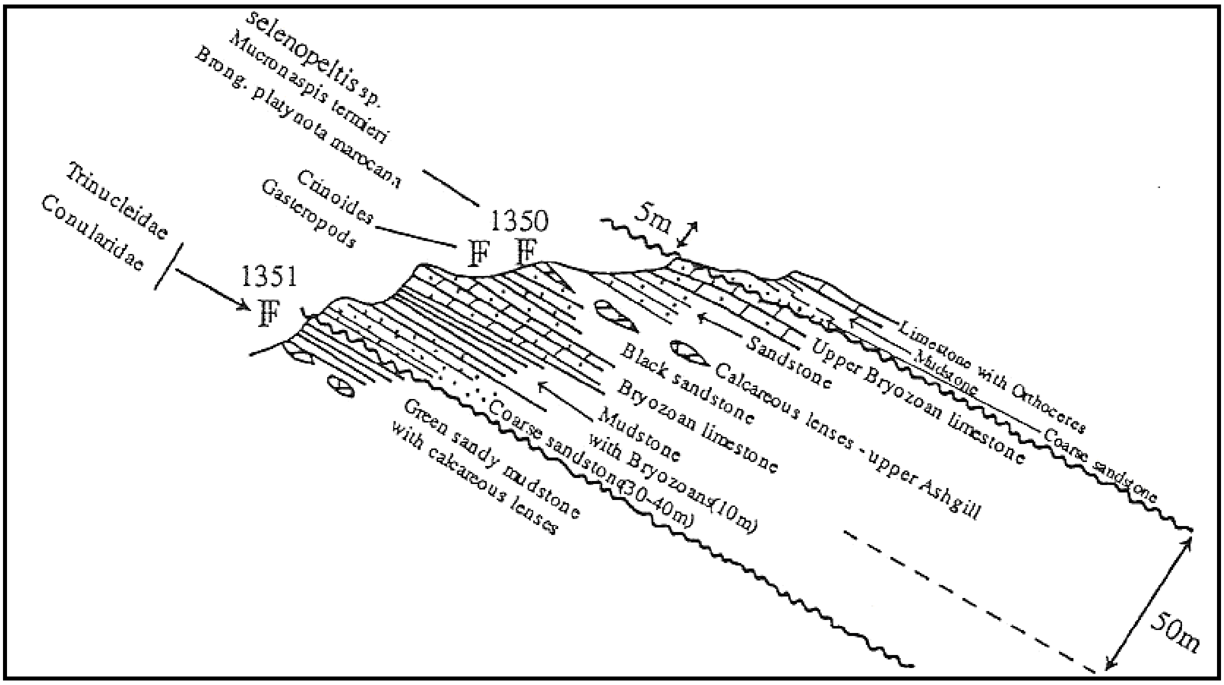


Fig. 15: Upper Ordovician in the East of Oued Er-Rosfa-Al Hamra, (Destombes, 1987)

The khabt Lahjar succession

Situation and lithostratigraphy

The Upper Ordovician Khabt Lahjar succession crops out in Khabt Lahjar areas 21 Km Eastern of Erfoud city (Fig. 12 and 13). It's subdivided into three lithostratigraphic units separated by angular unconformities (Destombes et al. 1985 ; Destombes 1987): calcareous sandstone of Lower Caradoc age (corresponding to the oldest deposits in this areas), Bryozoan sandstones of Middle Caradoc age and Blues Bryozoan limestones of Upper Ashgill age. The uppermost unit of Ashgill age is capped by Silurian oolitic ironstones (Fig. 14)

Facies and sequences

Facies association of Khabt Lahjar succession (Fig. 16) reflect deposition in cold (frost dominated), peritidal mixed siliciclastic/carbonate high energy littoral, adjacent to a carbonate platform. Coastal biogenic sediments are largely derived from this carbonate platform and limited terrigenous influx are realized from the Saharan ice sheet

The six facies (KA to KE) recognize in this succession are grouped into two high energy peritidal shallowing upward successions (Pratt et al. 1992) where evidence of tidal action is strongly suggested by facies association and typical sedimentary structures as : tidal bundle, mud couplets, neap/spring cycles, reactivation surfaces, sigmoidal and through cross bedding, channels, flaser beddings and wavy beddings (see references in Terwind, 1981 and 1988 ; Nio and Yang 1991 , Dalrymple 1992). The basal sequence have a subtidal lower member (facies KA) deposited under action of tidal currents and episodic storm events. This is followed by lower intertidal succession consisting of mixed flat dissected by tidal gullies and influenced by frost action (facies KB). This unit is in turn overlain by a thin Bryozoan biostrome (facies KD) and intertidal sandwaves capped by thin muddy supratidal deposits where frost action produce field of dessication polygones during period of emergence (facies KC). The second sequence begin with Bryozoan biostrome (facies KD) associated with glacial terrigenous sediments. The bryozoan biostrome is overlain by a tidal estuarine point bar deposits (facies KE) which in turn is overlain by intertidal sandwaves (facies KC).

The Rosf Al Hamra succession

Situation and lithostratigraphy

The Rosfa El Hamra succcession crops out in Khabt Lahjar areas 19 Km Eastern of Erfoud city (Fig. 12 and 13). It’s lithostratigraphy is established by Destombes (1987) who recognized two lithostratigraphic units separated by angular unconformities (Fig. 15): 1) Green sandy mudstones with calcareous lenses of lower Caradoc age, 2) coarse sandstones, mudstones with Bryozoans and Bryzoan limestones of middle Caradoc and Upper Ashgill age capped by Silurian deposits through an unconformity

Facies and sequences

The facies association of the Rosfa El Hamra succession (Fig. 17) reflect sedimentation in tide dominated delta under control of ice sheet dynamic

The Lower Caradocian deposits are grouped into two facies sequences. The basal sequence is composed of argillites, siltites, calcareous sandstones and lumachelic lenses exhibiting typical facies and structures of waves and tides dominated subtidal environment. The facies associations of the second sequence indicate tides dominated subtidal environment

The Middle and Upper Caradocian sediments consist of Bryozoan rich siltites and argillites, bryozoan limestones and coarse sandstones sandwaves, exhibiting tidal facies of estuarine subtidal sandwaves, soft sediments deformations, shattering and topography in ruins, they are overlain by sandy to muddy intertidal flat deposits. Soft sediments deformation is typical of melting deformation and shattering is induced by thermic fluctuations. Melting related soft sediments deformation and frost shattering suggest that after transport and deposition by glacier and reworking by tidal currents, these sediments were subjected to a multiphase periglacial evolution

The Upper Ashgill deposits consist of sandflat and coarse sandstones exhibiting facies of intertidal sandwaves

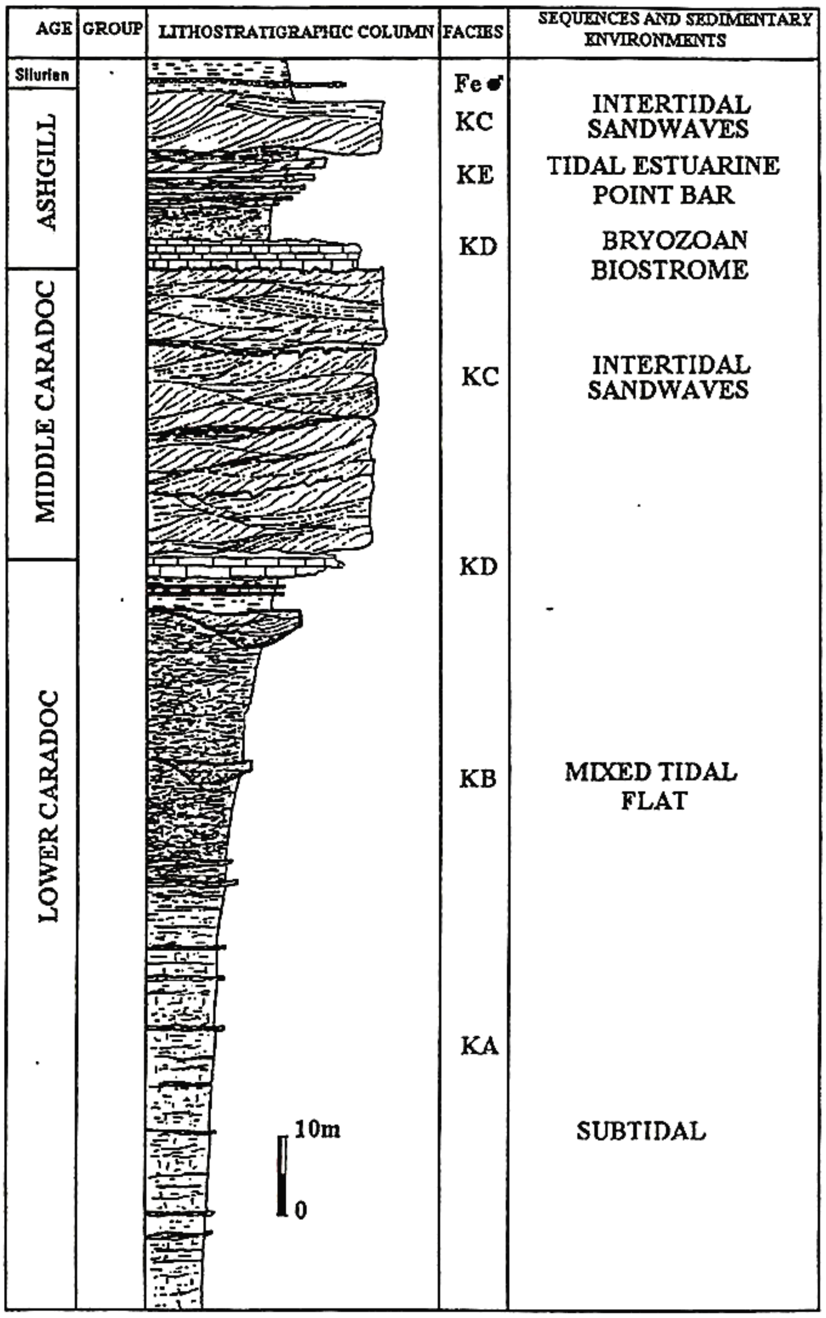


Fig.16 : Upper Ordovician succession of Khabt Lahjar with stratigraphic framework (after Destombes 1987), location of facies described in the text and environmental interpretations

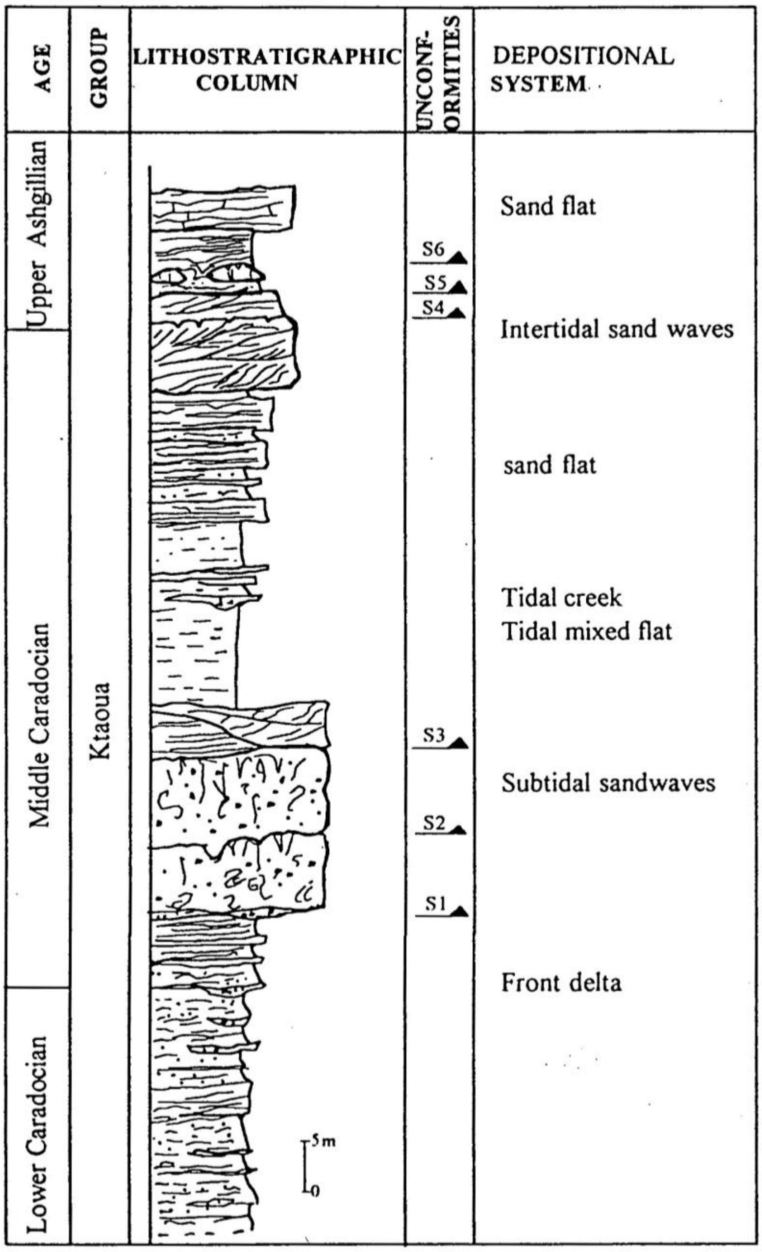


Fig. 17: Depositionnal Systems of Rosfa El Hamra succession

Nature and provenance of the sediments

The sediments of upper Ordovician successions are bafflestones, rudstones, limestones, carbonate cemented coarse arenites, sandy limestones, calcareous marls siltstones and mudstones. Petrographic analysis of these rocks indicates the existence of a terrigenous clastic phase and a carbonate phase in different proportions. In the silisiclastic phase, quartz grains are the dominant component and display textural shapes and polymodal grain size distribution of glacial evolution. This phase correspond to the limited terrigenous influx realised from the Saharian ice sheet and is associated with glacio fluvial discharges during melt phase. The carbonate phase is dominated by Bryozoan fragments ranging in sizes from fine sand to boulders, often well preserved, with accessory Brachiopode, Lamellibranch and Echinoderme intraclasts and in some case zoned dolomite rhombs and sparitic or microspartic cement . This phase resulted from the abrasion of Bryozoan mounds.

Depositionnal models and sedimentation control

The facies and the trend of Khabt Lahjar and Rosfa El Hamra successions and the other upper ordovician successions, of eastern Anti Atlas (Hamoumi ed., 1974 ; Hamoumi, 1996, 1997, 2001 and El Mazzouz, PhD in progress) reflect a complex history of this part of the Moroccan North Gondwanan platform. The eastern Anti Atlas that corresponded to a storm and / or tides dominated epeiric silisiclastic sea with E-W to ENE-WSW trending isopachs, during the lower and middle Ordovician such as the whole « anti -atlasic basin» ” (Hamoumi 1995), recorded major paleogeographical changes during the Upper Ordovician (Hamoumi, 1996, 1998, 2001 and El Mazzouz, PhD in progress). An extensional tectonic event resulted in the individualization of two sub basins: the « Khabt Lahjar sub basin » and the « Western Tafilalt sub basin », where the sedimentation developed under the interplay between autogenic (intrabasinal processes) and allogenic (glaciation and tectonic) controls. Intrabasinal processes includes reworking by tidal currents and storms. Changes in climate and glacial mass influence rate and nature of sediment supply, relative sea level fluctuation and isostacy.

During the Lower Caradoc, extensional tectonics related to the reactivation of the pan African faults movements according to NE-SW direction leads to the creation of a ramp type basin at the North Eastern edge of the “Anti Atlasic basin” (Eastern Anti Atlas), a break up and structuration in half graben of the epeiric siliciclastic shelf and a change in the isopach direction that became NW-SE. Rising of sea level due to glacier retreat allows the water depth over the platform to be right for the development of carbonate platform in the ramp type basin, where bryozoan mounds nucleate. Carbonate formed in subtidal areas and terrestrial glacial clastics sediments were reworked and redeposit as tidal flat either by accretion and progradation sedimentation in the shoreline.

Evidence of the development of a carbonate platform is attested by the existence of individual Bryozoan mud mounds and mud mounds in many successions of eastern Anti Atlas (Eastern Tafilalt and Thodra Maider). These carbonate mounds are often of small size, from decimetric to metric scale. They are mostly of flat lense shape but conical or ovoid form my also exist. They have generally flat base and their top is often affected by a strong karstification and in some case may correspond to a condensed surface with high concentration of phosphatised shells. Skeletal mounds correspond to Bryozoan mud mounds with Bryozoan in living position and mud mounds are composed of mud with variable amount of unsorted and weakly eroded bryozoan fragments together with trilobite, echinoderm and brachiopod debris. The common preservation of these bryozoans indicate mound initiation and growth bellow fair-weather wave-base in the subtidal environment of an isolated platform where the currents are able to keep water moving, oxygenation high, turbidity low and sufficient input of nutrients. Bryozoan mound are skeletal builder that act as mud baffler, trappers, binders and stabilizers. Skeletal mound are known in deep water around seeps of petroleum, methane and vent of hydrothermal water (Callender et al., 1990). While mud mounds are the result of the accumulation of lime mound produced in situ from degradation of delicate skeletons and microbial production.

Modern Skeletal rich banks with coarse shell debris (sand and gravels) derived from Bryozoans occur in cool water in zone of 200-300 m depth, low terrigenous clastic influx and active nutrient uppwelling swepts by storm waves and tidal currents (Jones and Desrocher, 1992). Shallow marine Bryozoan carbonates occur in many modern cold sea at high latitude: the Threee Kings Plateau, New Zeland (Nelson and al., 1982), Southern Australia (Boreen and al., 1993) and Tasmania (Marshall and Davis, 1978). They are known also in ancient cold water carbonates: in Oligo- Miocene Abrakurrie limestones, southern Australia (James and Bone, 1991) and Permian shelf of Tasmania (Rao, 1981), where they are associated with glacial terrigenous sediments. Mud -mound complexes were also identified in the Cystoid Formation of Ashgill age (Vennin et al., 1998)

During Middle and Upper Caradoc, the « Khabt Lahjar sub basin » was controlled by subsidence and glaciation. Rising of sea level due to glacier retreat allows the water depth over the ramp to be right for the development of bryozoan mounds and sedimentation in peritidal zones that was alimented by biogenic gravels and sand derived from subtidal areas and glacial siliciclastic sediments derived from pan African shield and it's old cover. Falling of sea level induced the stop of bryozoan mounds build up and their degradation by mechanical erosion. In the « Western Tafilalt sub basin » that corresponded to a longitudinal trough of NW-SE direction sediments from Saharan ice sheet and carbonate platform accumulated in fan deltas systems ("upper conglomerate" defined by Destombes et al., 1985)

During the Upper Ashgill the two sub basins were controlled by glaciation and subsidence. In the « Khabt-El-Hejar sub basin » the previous paleogeography in the most part, except in the central part (NW de Merzane), where subsidence was more important than sea level rise and induced the development of bryozoan mounds. In the « Western Tafilalt sub basin ».sea level rise induced deposition of terrestrial glacial clastics sediments in littoral environments. The lost regressiv / transgressiv stage which occur at the upper Ordovician/ Silurian boundary, leads to the formation of Early Silurian oolitic ironstones and deposition of graptolitic shales during the sea level rise associated with the melting of the glacier .

ANNEXE 1

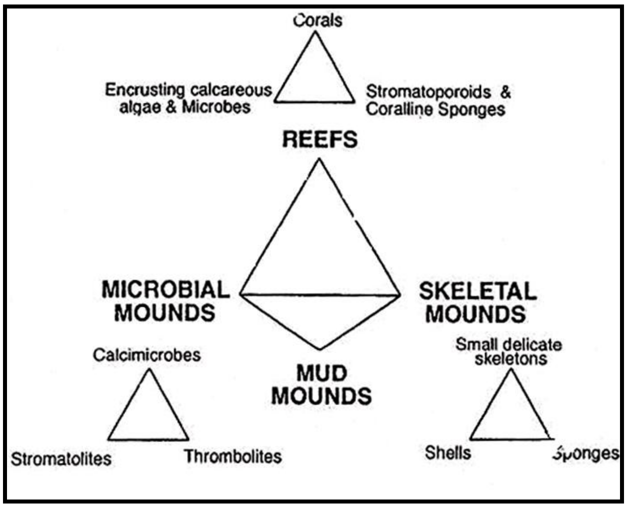


Fig. 18: A conceptual classification of reefs and mounds (James, 1991)

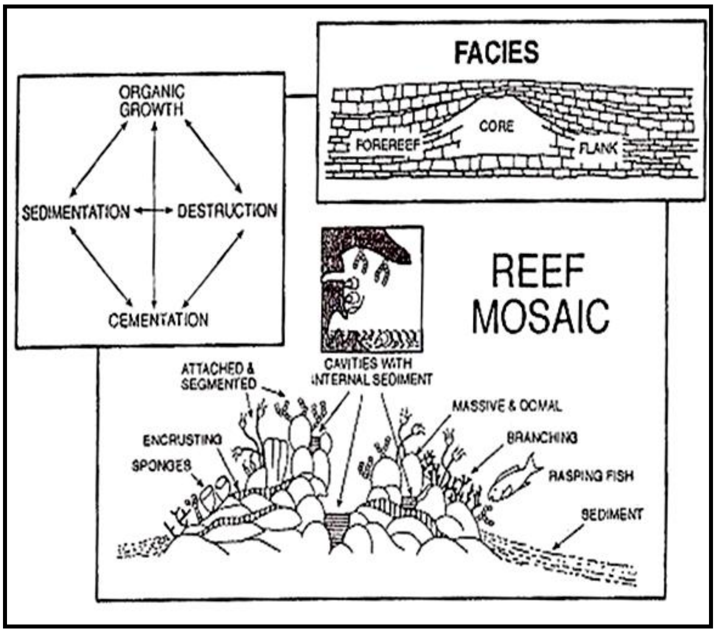


Fig. 19: A diagrammatic cross section of the different facies which comprise a reef or mound (upper right) and the organism-sediment mosaic which typifies a growing reef, whose composition is the sum of the four different processes illustrated in the inset. (James, 1991)

ANNEXE 1

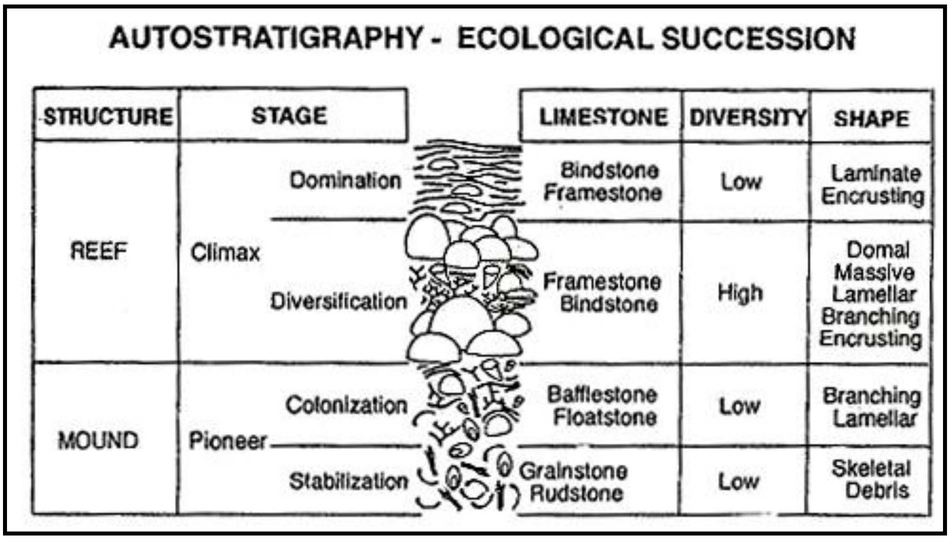


Fig.20: A sketch of the four divisions of the core facies that can be generated by ecological succession of the reef builders. (Right) the most common types of limestone, relative species diversity and shape of the reef builders found in each part. (Left) the stages of growth. Mounds typically exhibit the first two stages while reefs can exhibit all four stages (James, 1991).

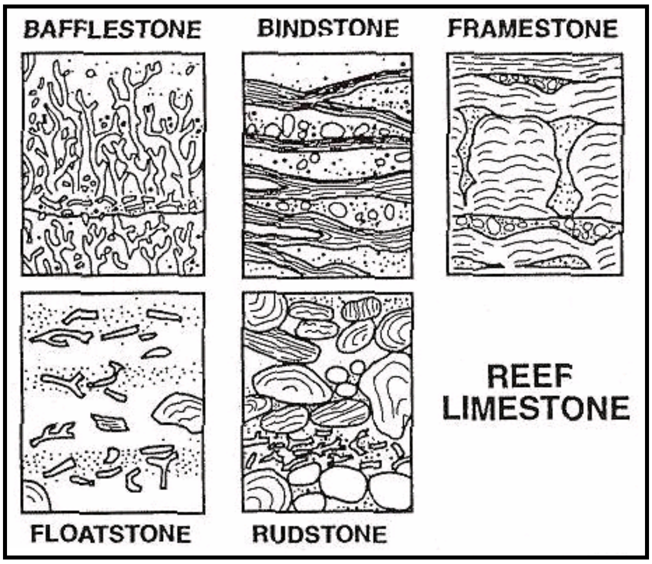


Fig. 21: An interpretative sketch of the different types of reef and mound limestone recognized by Embry and Klovan (1971). Autochthonous limestones (top) are found in the core facies while allochthonous limestones (bottom) typify the flank or fore reef facies (James, 1991)

ANNEXE 1

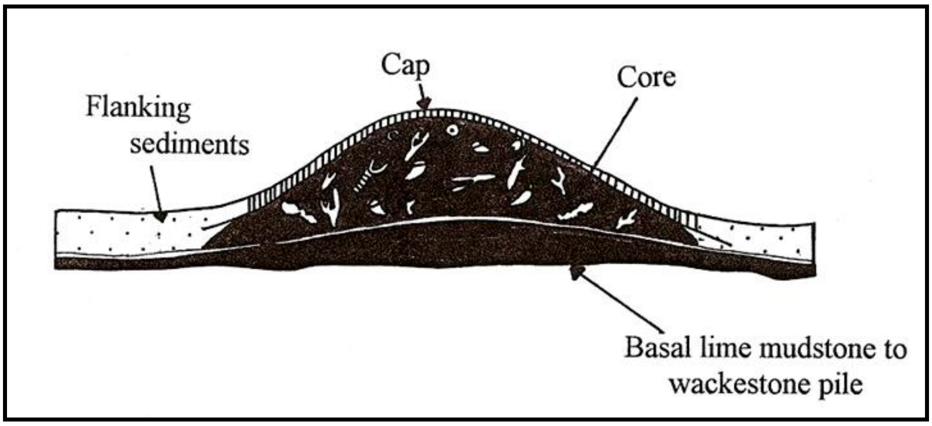


Fig. 22: Mud mound structure

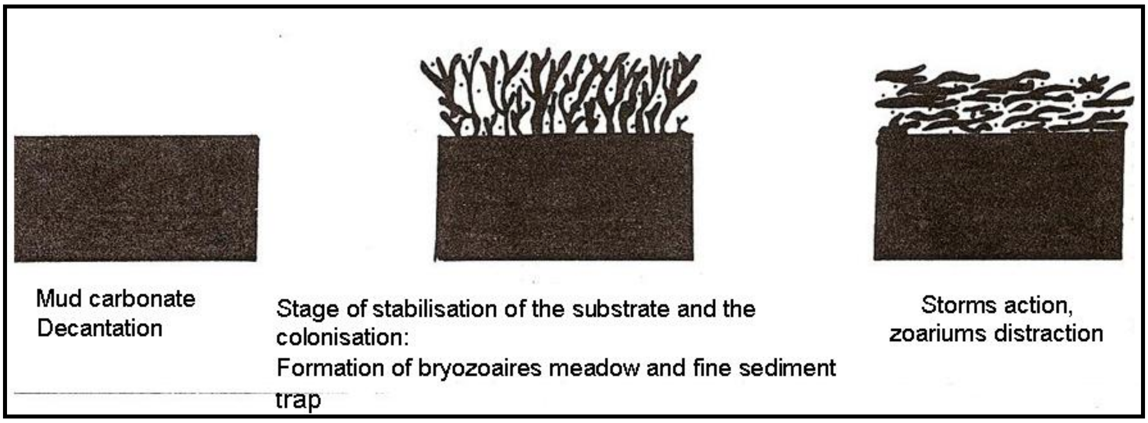


Fig. 23: Formation of mud mound

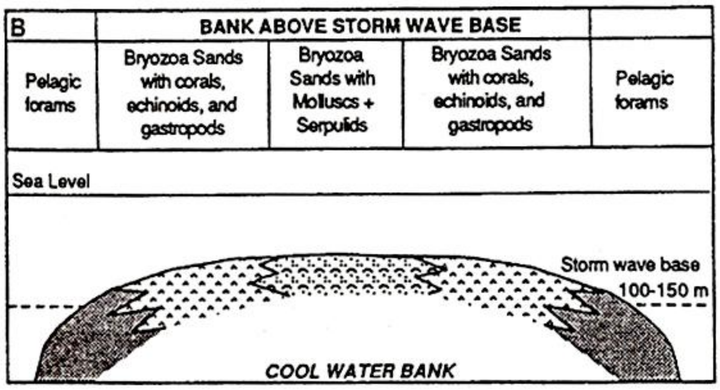


Fig.24: facies models for carbonates on cool water banks (Desrocher, 1991)

ANNEXE 1

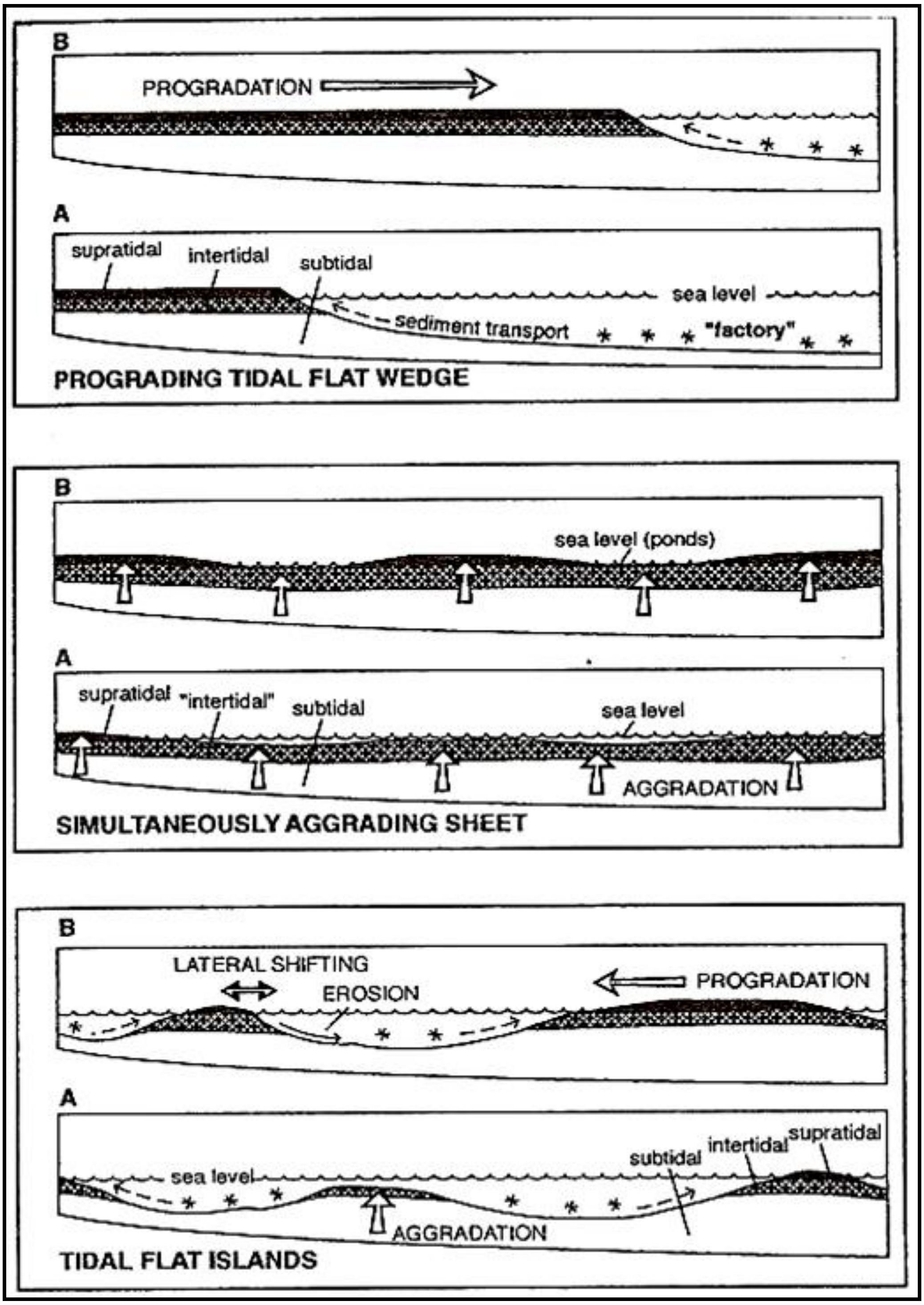
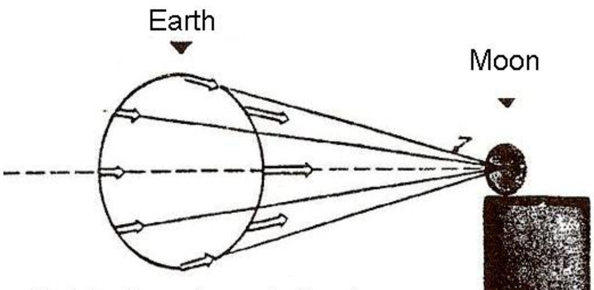
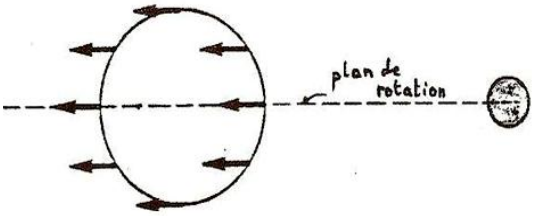


Fig. 25: Diagrams illustrating various ways in which a metre-scale, peritidal, shallowing-upward succession can form. A prograding wedge is generated by sediment transported onto the tidal flat from the offshore carbonate factory. A simultaneously aggrading sheet accretes vertically to sea level and the whole platform becomes sequentially intertidal and then supratidal. Tidal flat islands nucleate and accrete by aggradation and progradation and shift in response to hydrographic force, (Pratt et al., 1991)

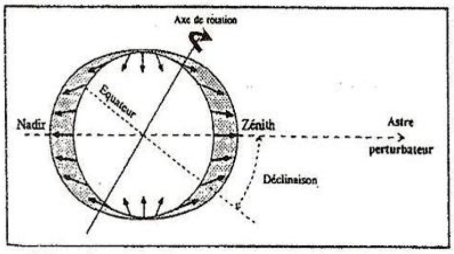
ANNEXE 2



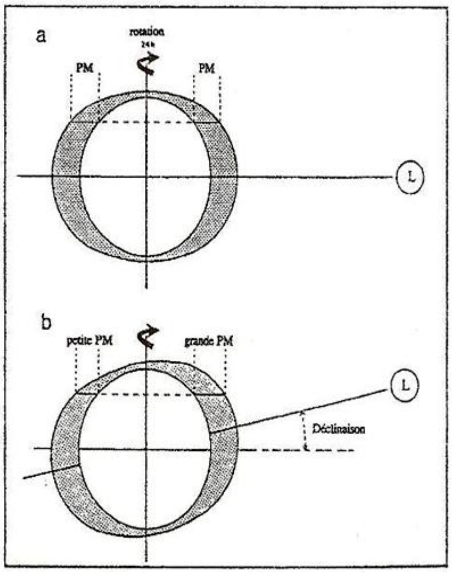
Distribution of gravitation forces



Distribution of centrifuge forces



Generative forces of tide



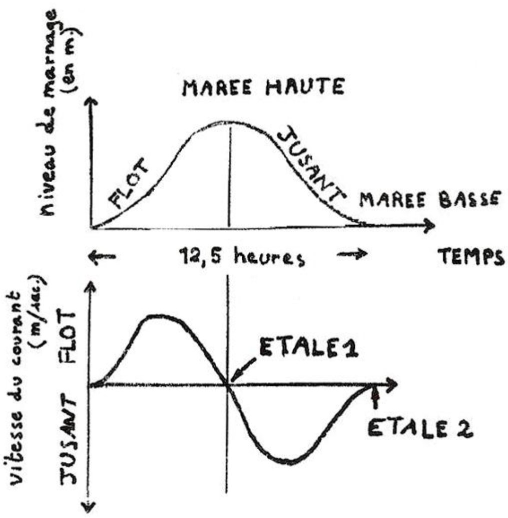
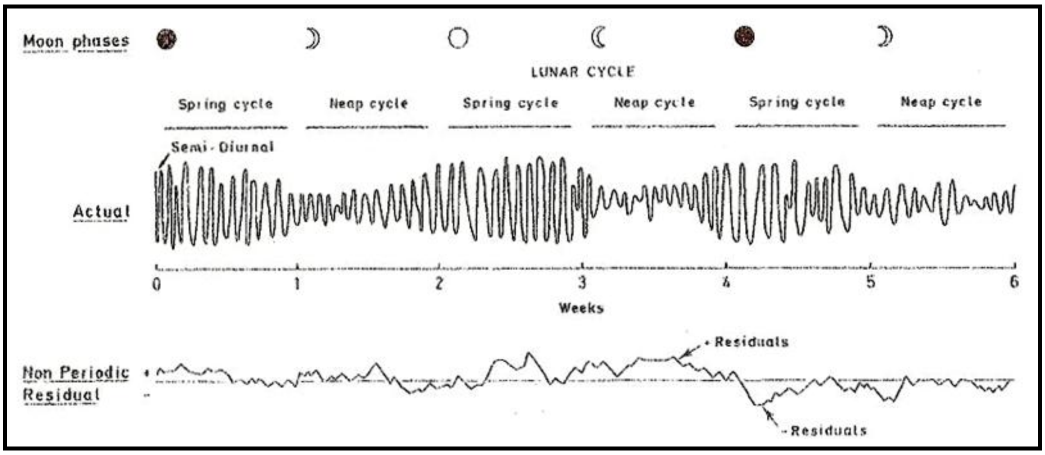
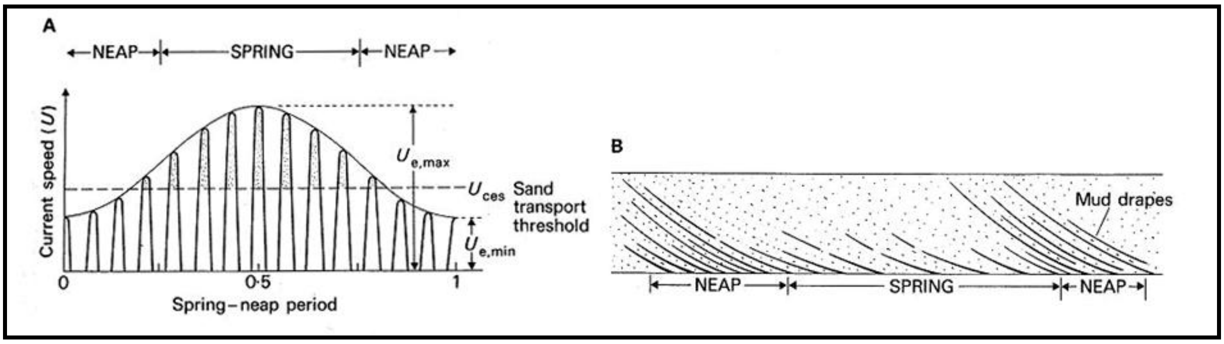


Fig.26 : Tidal cycle

ANNEXE 2





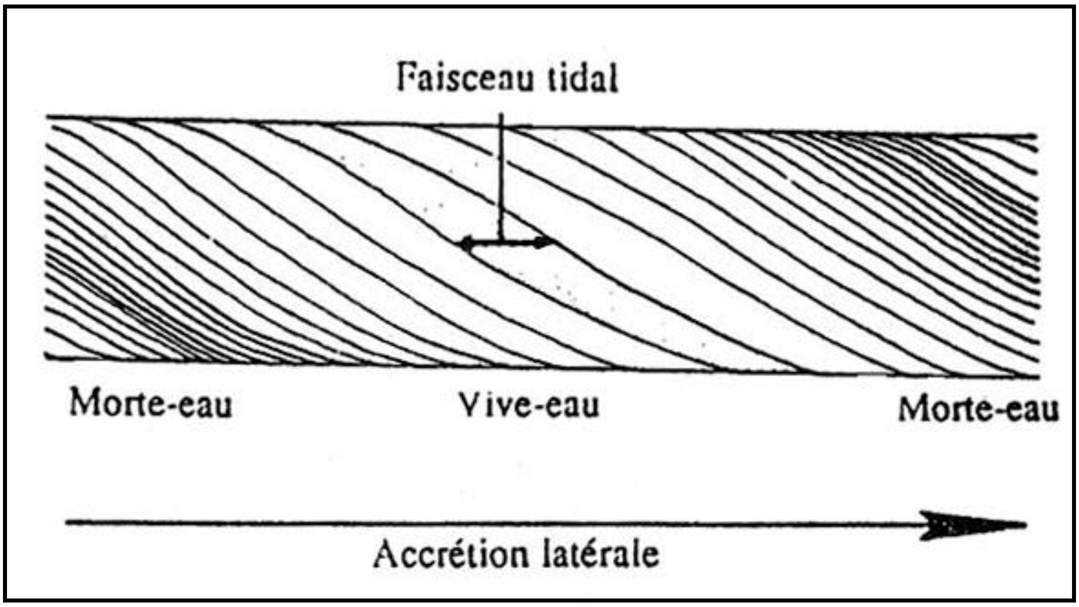
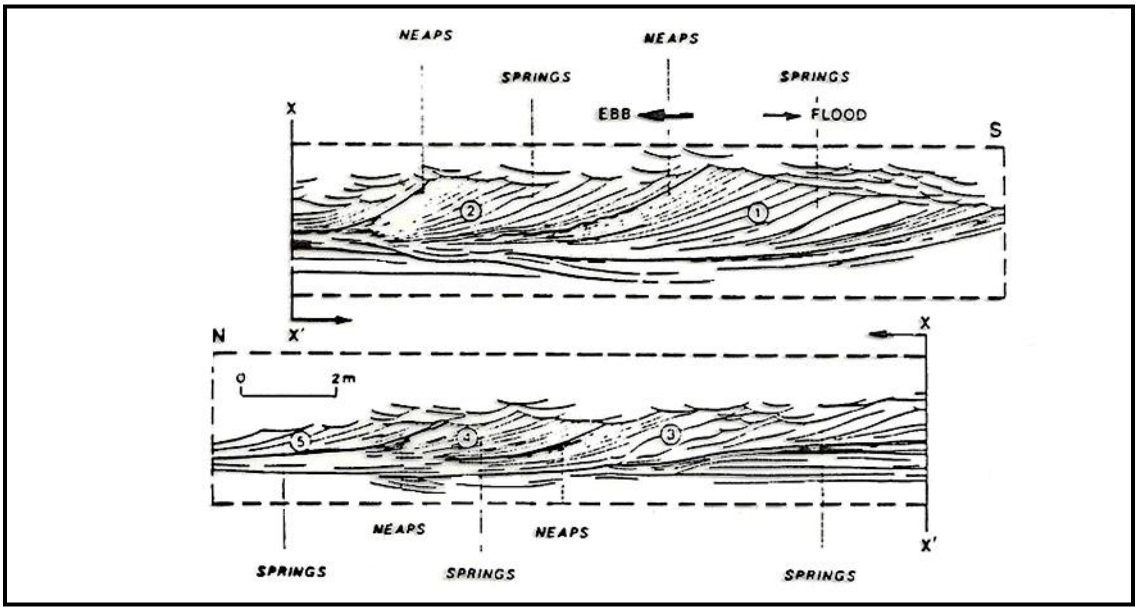
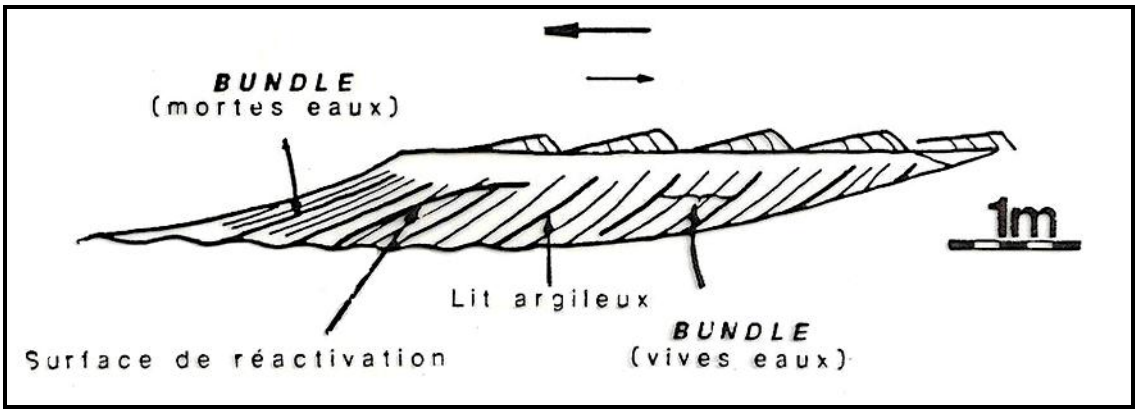


Fig. 27: Neap-spring cycle (Yang et Nio, 1985)





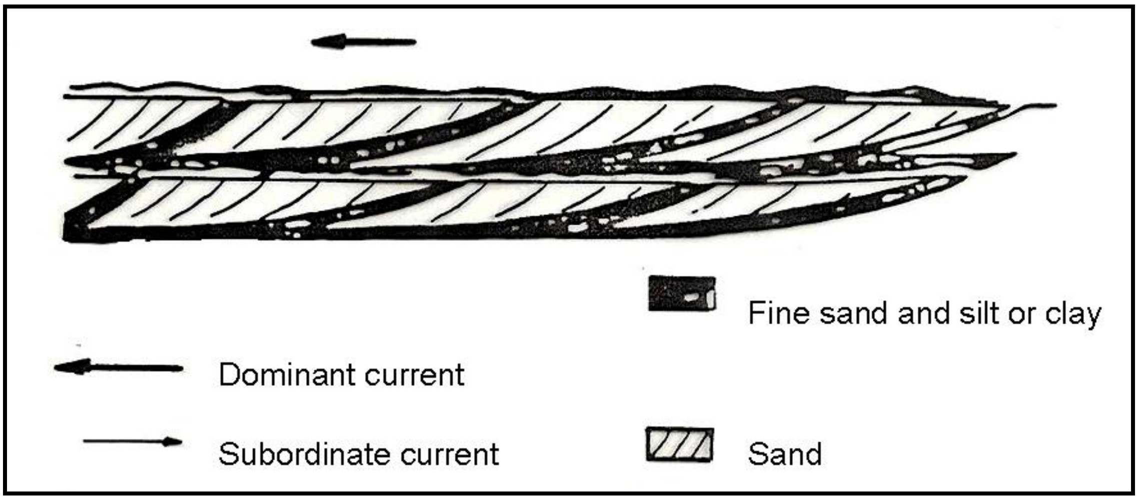


Fig. 28: Tidal bundle

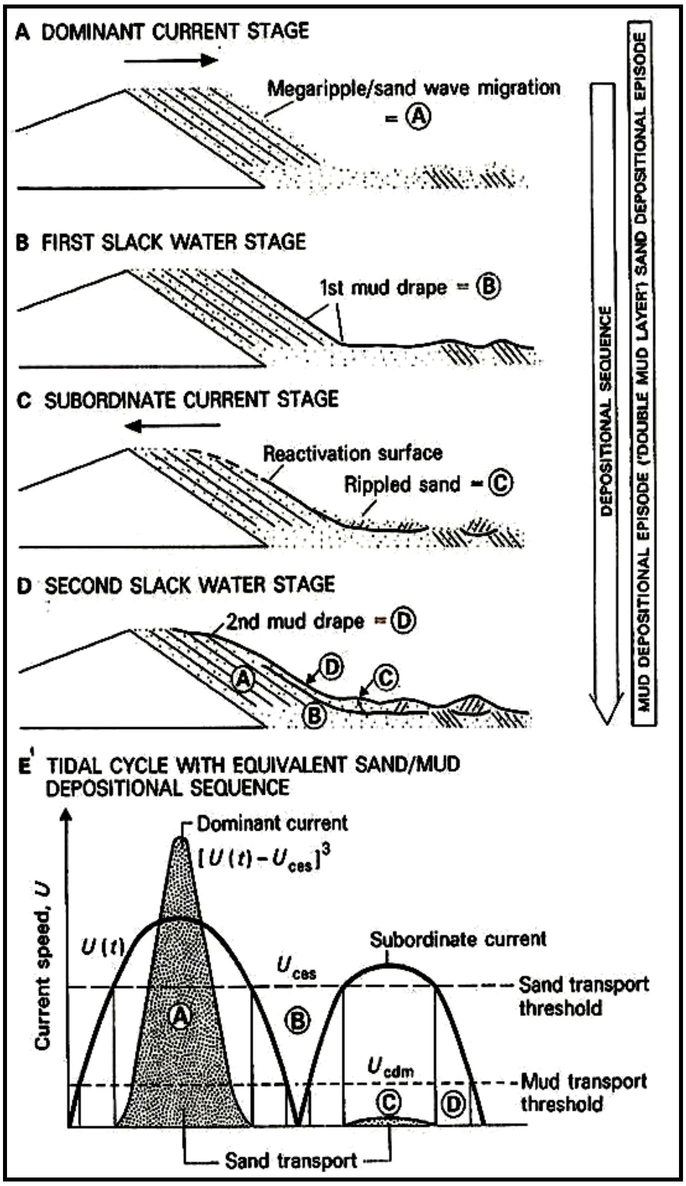


Fig. 29 : Interpretation of a bundle of cross-bed forests in terms of an ebb-flood tidal cycle with pronounced velocity asymmetry: A-D represent the stage of deposition associated with an ebb- flood cycle (from Visser, 1980); E represents a tidal velocity curve which would produce the structures depicted in A-D(from Allen, 1982b)

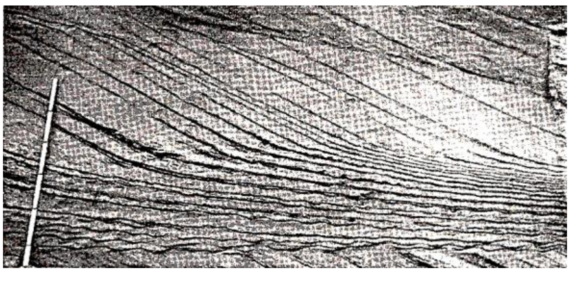


Fig. 30: Part of an exposure. Groups of ebb-oriented foreset laminae are separated by mud drapes. At top of preserved lee side only one mud drape occurs, whereas in bottomset there are two more or less parallel mud drapes. Scale divided into 20cm lengths.

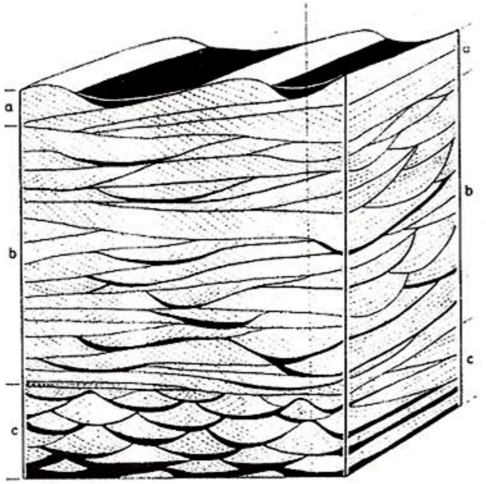


Fig. 1 Block diagram showing various types of flaser bedding. a flaser bedding associated with straight-crested small-current ripples; b flaser bedding formed from small ripples with curved crests; c flaser bedding in association with wave ripples. (After Reineck and Wunderlich 1968b)

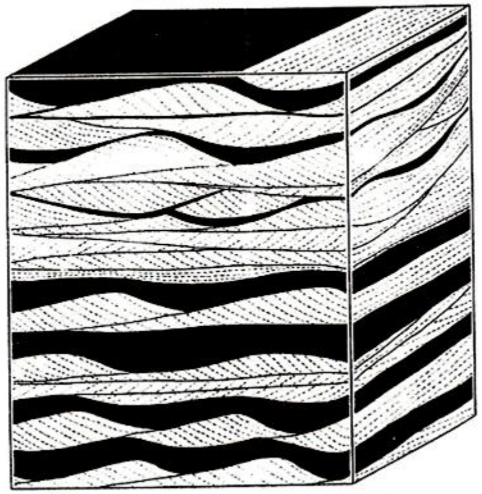


Fig. 2. Wavy bedding. Various types of wavy bedding are produced because of different thicknesses of mud layers. (After Reineck and Wunderlich 1968b)

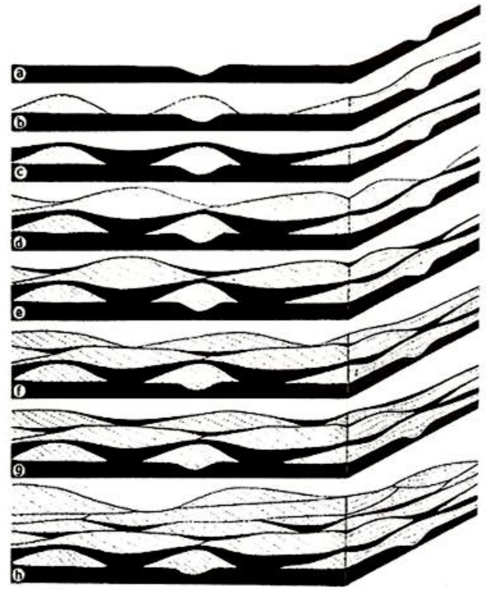


Fig. 5. Scheme showing the genesis of lenticular bedding and flaser bedding (in lower part). a–d shows the genesis of lenticular bedding as a result of formation of incomplete ripples on a muddy substrate, later covered again by a mud layer, followed by deposition of sand in the form of ripples. Periods of current activity alternate with stillstand periods, resulting into deposition of sand and mud, respectively. Only unidirectional current has been considered, from left to right. In tidal environments ripple-bedded units usually show two current directions. (After Reineck 1960a)

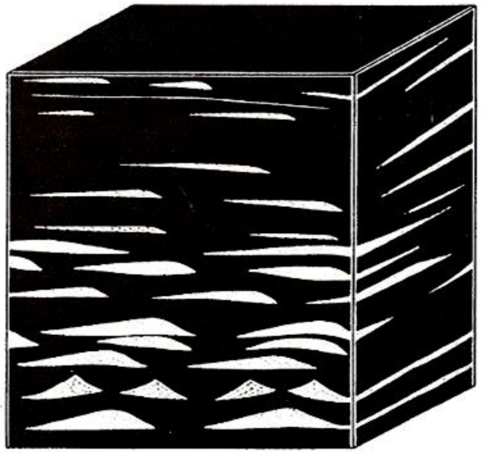


Fig. 3. Lenticular bedding with isolated lenses, the upper part with flat isolated lenses. (After Reineck and Wunderlich 1968b)

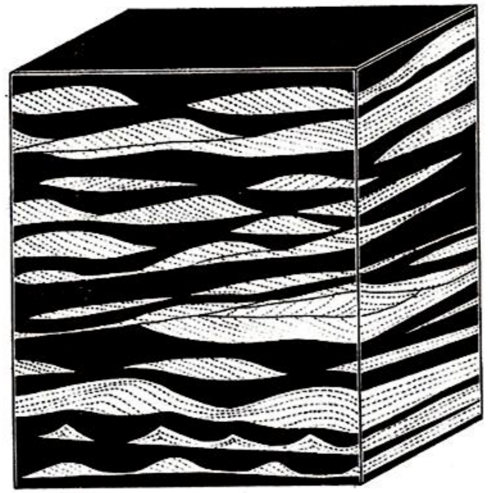


Fig. 4. Lenticular bedding with thick connected lenses. Ripple-bedded lenses of the upper part are of current origin. Sand lenses of the lower part are deposited as wave ripples, both symmetrical and asymmetrical. (After Reineck and Wunderlich 1968b)

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