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DEPOSITIONAL HISTORY AND DIAGENESIS OF QUARTZ-SAND BARS AND LIME-MUD ENVIRONMENTS IN THE DEVONIAN BASIBÉ FORMATION (CENTRAL PYRENEES, SPAIN)

BY

M. A. HABERMEHL

ABSTRACT

The Basibé Formation, of Lower Devonian age (Emsian) according to conodonts, consists, in the area between the Esera River and Mañanet River, of nodular weathering limestones, dolomites, silty to sandy argillaceous dolomites, quartzites and limestones. Thickness variations of the lower member (nodular limestones) and the upper member (limestones) are of minor importance, whereas the middle member (quartzite-dolomite alternation) is wedge-shaped: its thickness decreases over a distance of about 35 km from about 100 m (50 m of which are quartzites) in the west to 0 m in the east. In the surrounding areas the Basibé Formation consists solely of limestones.

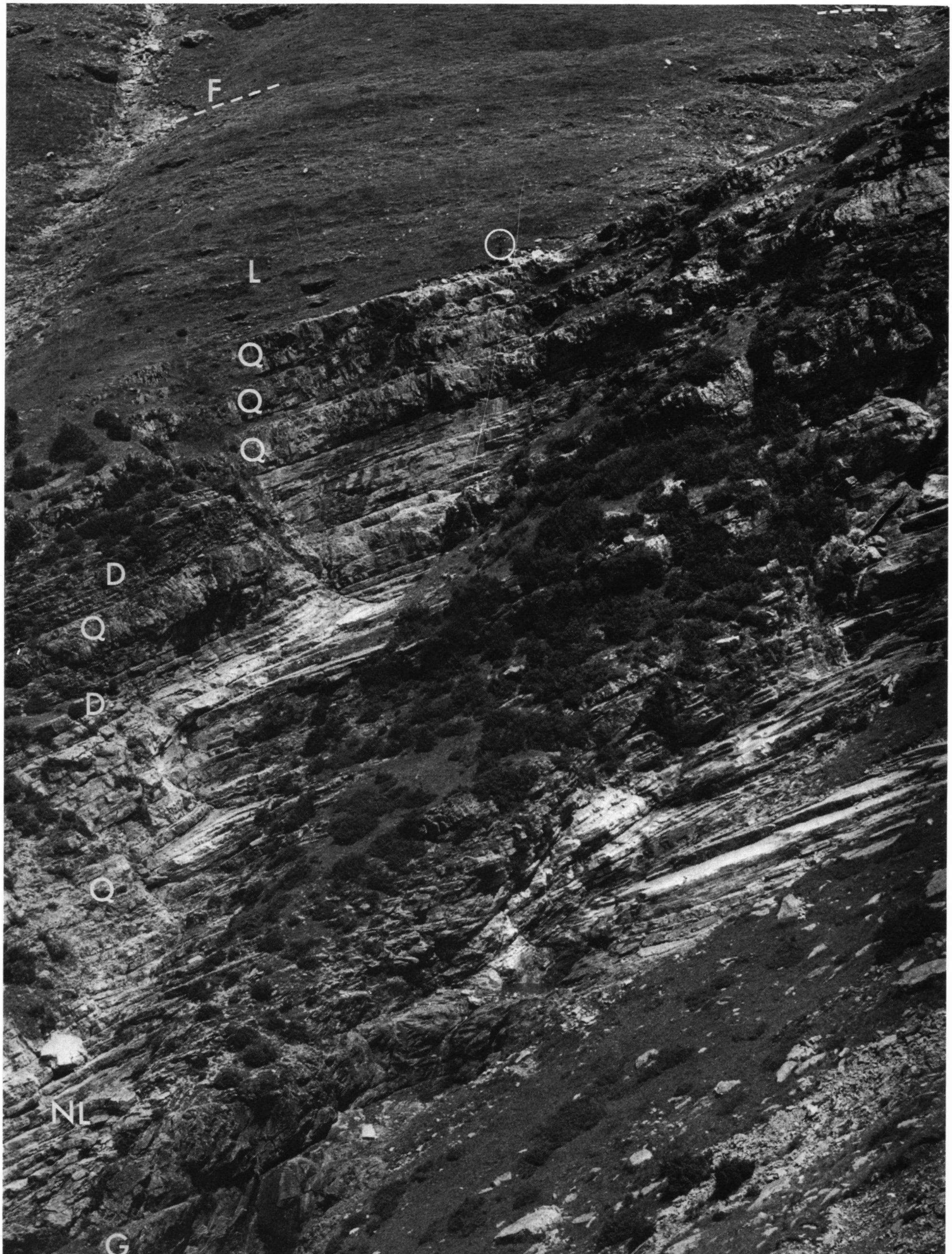
Carbonates were deposited as lime muds in shallow, open marine environments. Mature quartz sands, probably brought into the area by longshore currents, were accumulated by wave and current action as rather stationary bars or barrier islands, with a NW-SE direction. Leeward of bar complexes, highly bioturbated silty to sandy argillaceous lime muds were deposited. These back-bar deposits are, however, open marine sediments. Bars were buried by shallow, open marine lime mud sediments, due to subsidence and/or lack of clastic supply.

The carbonates of the lower and middle member were originally fossiliferous micrites or biomicrites and were changed by diagenetic processes into nodular weathering limestones. This nodular appearance is caused by numerous stylolites and solution stringers which originated after lithification. Mg-enriched interstitial water, driven out of the quartz sandstones, during cementation, caused the limestones interfingering with these quartz sandstones to be replaced by dolomites.

The pure quartz sandstones underwent quartz cementation, while pressure solution led to various types of contacts and to high pressolution of the quartz grains. Burial was not the only cause, but also tectonic pressure and a rise in temperatures due to the intrusion of the Maladeta batholith. Replacement of quartz grains and secondary quartz by dolomite is a rather important late diagenetic process.

The limestones of the upper member show recrystallization and stylolitization, while some original pelmicrite beds were preferentially dolomitized.

CONTENTS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CHAPTER I

INTRODUCTION

SCOPE OF THE STUDY

The Basibé Formation of Devonian age in the southern Central Pyrenees consists mainly of limestones, but in part of its distribution area it contains a peculiar quartzite-dolomite association. This is found in the area between the Esera and Mañanet Rivers (provinces of Huesca and Lérida, Spain), the area with which our study is concerned. The surrounding areas, where the Basibé Formation consists solely of limestones, are only briefly mentioned.

The formation is a persistent and readily recognizable, often cliff forming unit in the area studied (Figs. 1 and 5). It has a striking appearance amidst the Devonian rocks of the southern Central Pyrenees, which are generally composed of limestones and slates (Fig. 3).

The quartzites interfinger with dolomites, and together form the middle member of the Basibé Formation. The lower and upper members consist of limestones, those of the lower member having a nodular weathering appearance. In the lower and middle members silty to sandy argillaceous carbonate beds occur between the limestones, dolomites and quartzites.

The geometry of the Quartzite-Dolomite Member is wedge-shaped with a thickness of about 100 metres, in the west, near the Esera River, which diminishes to 0 metres in the east, near the Mañanet River. The quartzites of this member have an aggregate thickness of about 50 metres in the west, which also decreases towards the east.

The purpose of the present study is to establish the origin and to reconstruct the depositional environment of the sediments of the Basibé Formation. The necessary data are provided by an analysis of the sediments with respect to sedimentary structures, mineralogical and textural aspects, primary lithology, diagenetic changes, probable energetic conditions in the depositional environment and the distribution pattern of the various lithologies.

The limestones, the lower ones of which greatly resemble the "Flaserkalke" of other workers, are interpreted as deposits of a shallow marine environment. The quartz sandstones will be shown to represent bar, barrier-island, or beach sediments. The silty to sandy argillaceous carbonates are in our opinion sediments of a shallow, low energy environment, between bars or in sheltered areas behind bars or barrier-islands.

Dolomitization of the limestones, which are intercalated with the quartzites and genetically correspond to the lower limestones, occurred in our opinion in a later diagenetic stage by Mg-enriched interstitial water which was driven out of quartz sandstones during silicification.

Fig. 1. Type section of the Basibé Formation. Section 25 in the Barranco de San Silvestre. Stratigraphic top is at top of photograph. G: Gelada Formation, NL: nodular limestones, Q: quartzites, D: dolomites, L: limestones, F. Fonchanina Formation. Looking north. Figure in circle for scale.

The earliest stratigraphic studies of this particular area are those by Dalloni (1910, 1930) and Schmidt (1931).

The lithostratigraphy of this area was established by Mey (1967b, 1968), and published as part of a mapping programme of the Paleozoic of the Central Pyrenees initiated by Professor L. U. de Sitter. From these mappings arose the need for a sedimentological investigation of the Basibé Formation in this area.

The structural geology of the area, described in detail by Mey (1967b, 1968), Wennekers (1968) and Roberti (1970, 1971), shows that the area was strongly affected by the various phases of the Hercynian orogeny and was affected a second time by Alpine movements.

METHODS

The location of 39 measured outcrop sections within the area studied are shown in Fig. 2 and on the index map of Plate I.

Use was made of the geological maps prepared by Mey (1967b, 1968), Wennekers (1968) and Roberti (1970).

Due to the rugged relief and scarce vegetation, and also to the resistant character of the Basibé Formation in regard to other rocks, the formation is fairly well exposed. All sections given in Fig. 2, were measured on mountain slopes or in rivers. Plate I shows 7 measured outcrop sections, representative of, and evenly distributed over, the area. A total of 38 sections were measured, while 26 additional, more detailed examinations, accompanied by drawings and photographs, were carried out in some of these sections. Data and samples of Section 65 were provided by Mr. K. J. Roberti.

About 1000 samples were taken. Colours were compared by means of the Rock Colour Chart distributed by The Geological Society of America (2nd. ed., 1951). From about 500 samples thin sections were prepared and examined. All thin sections were partly stained with alizarin red-S and with a mixture of alizarin red-S and potassium ferricyanide, according to Dickson (1966), for identifying the carbonate minerals.

Petrographic microscopical investigation of the thin sections provided data on primary lithology and diagenetic changes.

The volumetric calcite-dolomite proportions were also obtained from analyses of the stained thin sections. X-ray analyses of 25 selected samples provided some additional mineralogical data.

Insoluble residues of 24 limestone samples were examined for conodonts by Mr. K. T. Boersma. Some of these residues were mounted as polished sections and were examined in reflected light in order to identify the iron minerals.

Sizes of quartz grains are given according to Wentworth's scale (in Pettijohn, 1957), crystal sizes of (chiefly strongly recrystallized) carbonates according to Folk (1959, 1962, 1965). The composition of limestones and dolomites as given on Plate I is classified according to Cayeux (1935) and Carozzi (1960).

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CHAPTER II

GEOLOGICAL SETTING

The Basibé Formation forms part of the Paleozoic rock sequence of the Axial Zone of the Central Pyrenees. The Paleozoic rocks of the Axial Zone are strongly affected by the Hercynian and to a lesser degree by the Alpine orogeny. Instrusion of granodiorite batholiths and accompanying dykes in a late-Hercynian phase caused aureoles of contact metamorphism in some of the Paleozoic rocks.

OUTLINE OF STRATIGRAPHY

The following lithostratigraphic and structural data have mainly been taken from the works of Dalloni (1910, 1930), Schmidt (1931), Misch (1934), Zwart (1960, 1964), Boschma (1963), de Sitter (1964, 1965), Nagtegaal (1966, 1969), Souquet (1967), Mey (1967a, b, 1968), Mey et al. (1968), Wennekers (1968), van Hoorn (1970), Hartevelt (1970) and Roberti (1970).

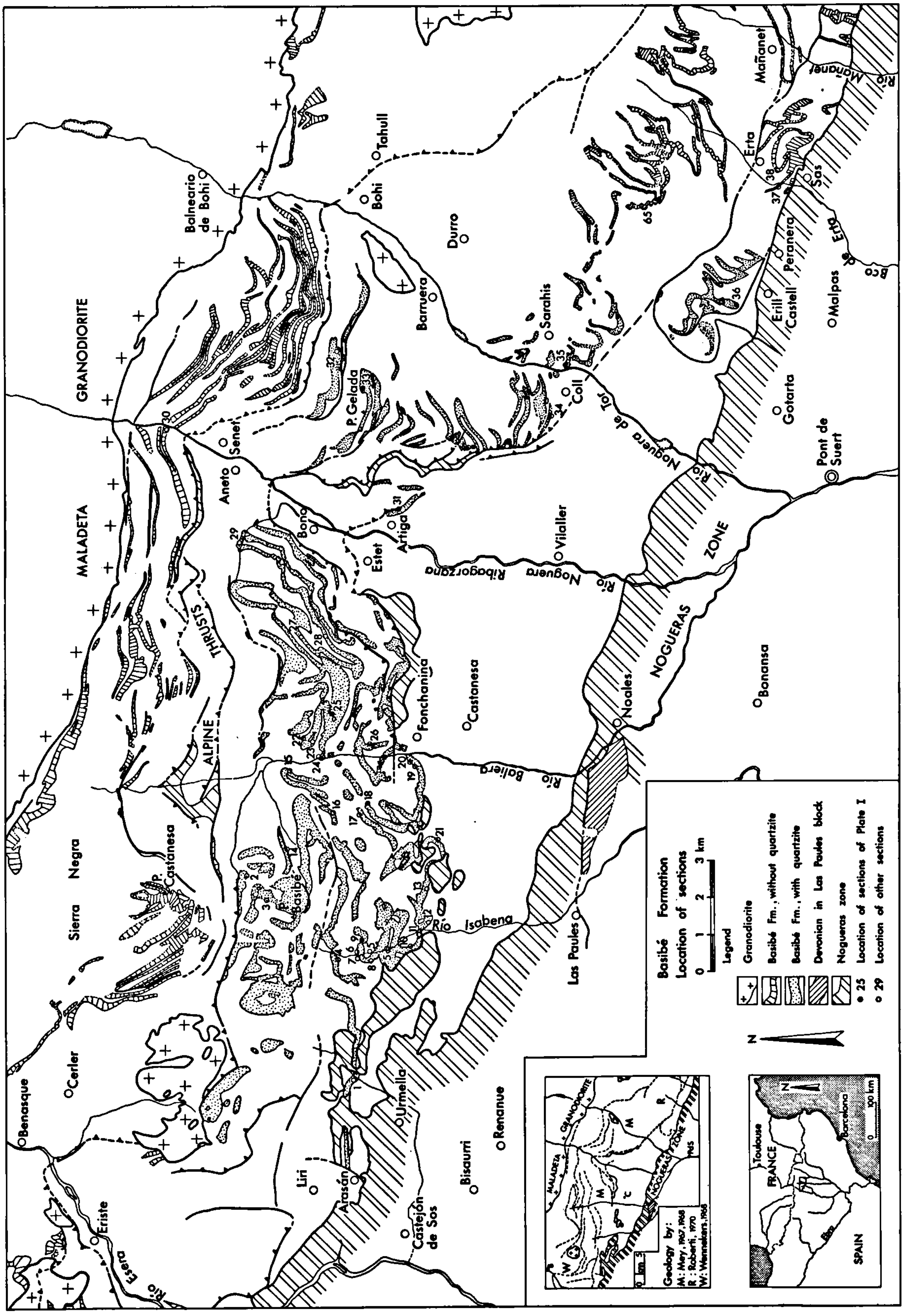
In the area studied, as in the whole of the Central Pyrenees, the pre-Hercynian Paleozoic rock sequence consists almost entirely of shallow marine sediments, which can be divided into a number of formations of only moderate thickness. Slates predominate, with carbonates in the second place, whereas only a few coarser clastic intercalations occur, mainly quartzites.

In the area studied the oldest outcropping rocks are an alternation of slates and thin quartzite bands, with one marly limestone intercalation in the upper part. The upper part of this sequence was lithologically correlated by Mey (1967b, 1968) with a sequence in the Segre Valley, which was dated by earlier workers as Upper Ordovician age. In the Segre Valley and many other localities in the Central Pyrenees a coarse conglomerate of Ordovician age is found, though not in the area studied. Below this conglomerate an alternation of slates with quartzites exists, while the conglomerate is overlain also by slates and quartzites, representing fluvial deposits according to Hartevelt (1970 and pers. comm.). The entire succession of Cambro-Ordovician rocks is several hundred metres thick in the Baliera area. These sediments are overlain by slates and an alternation of thin-bedded quartzites and impure limestones. These are followed by graptolite-bearing carbonaceous slates, which are very uniformly developed in the Pyrenees, the Cantabrian Mountains, Catalan Mountains and the Montagne Noire, and have been dated by several workers as Silurian. The upper part of this approx. 100 metres thick formation contains an Orthoceras limestone.

The Devonian rocks in the southern part of the Central Pyrenees consist mainly of an alternation of limestones and slates (see Chapter III and Fig. 3). The Devonian sequence has been subdivided into four (in the area studied five) formations (Mey, 1967b, 1968). In the area studied a conspicuous wedge-shaped member occurs, consisting of quartzite and dolomite, which is interbedded in limestones and which is the main subject of this paper.

The Lower Carboniferous deposits consist of micaceous slates and calcareous sandstones, together several hundred metres thick. Typical marker beds occurring elsewhere in the Pyrenees at the base of the Carboniferous, such as bedded chert, calcium phosphate nodules and ash layers, are not present in the area studied. The original thickness of the Lower Carbon-

Fig. 2.


Benasque
Cerler
Sierra Negra
P. Castanesa
MALADETA
GRANODIORITE
ALPINE THRUSTS
Balneario de Bohi
Tahull
Bohi
Durro
Barruera
Sarahis
Coll
Senet
Aneto
P. Gelada
Bono
Estet
Artiga
Vilaller
Río Noguera Ribagorzana
Río Noguera de Tor
Pont de Suert
Gotarta
Malpas
Erill Castell
Peranera
Sas
Erta
Mañanet
Río Mañanet
Bco Erta
NOGUERAS ZONE
Castanesa
Fonchanina
Bonansa
Noales
Río Baliera
Las Paules
Río Isabena
Basibé
Urmella
Renanue
Bisaurri
Liri
Arasán
Castejón de Sos
Eriste
Río Esera
Basibé Formation
Location of sections
0 1 2 3 km
Legend
Granodiorite
Basibé Fm., without quartzite
Basibé Fm., with quartzite
Devonian in Las Paules block
Nogueras zone
25 Location of sections of Plate I
29 Location of other sections
N
Geology by:
M: Mey, 1967, 1968
R: Roberti, 1970
W: Wennekers, 1968
Toulouse
FRANCE
SPAIN
Barcelona
0 100 km

iferous deposits is not known, as they are unconformably overlain by post-Hercynian rocks. The Lower Carboniferous deposits are exposed south of the Devonian outcrops and also north of the Alpine thrust zone in the area studied.

The post-Hercynian deposits need only concern us in so far as they are found, or once possibly extended, into the area studied.

The first post-Hercynian formations are continental deposits (Nagtegaal, 1969), the clastic material of which is derived from the degradation of the folded Paleozoic rocks of the Axial Zone. These Upper Carboniferous, Permian and Lower Triassic deposits outcrop in the border of the Nogueras Zone, south of the area studied. The Upper Carboniferous deposits consist of coarse conglomerates, which are the fillings of fossil valleys in a steep paleo-relief, and coal-bearing mudstones and other rock types of Stephanian age, with a maximum cumulative thickness of about 780 metres.

These rocks are overlain by red mudstones and sandstones of Permian age, with a thickness of 0 to 700 metres. The Upper Carboniferous rocks are of very local origin, and it is doubtful whether they covered the area under consideration entirely. They were deposited within steep valleys and probably did not extend far upstream from the localities where they are now found.

The Permian of the Nogueras Zone represents, according to Nagtegaal (1969, p. 163), the downstream end of alluvial fans, which were situated more to the north and therefore may have extended into the area studied.

The Triassic rock sequence has an angular unconformity as its lower boundary. The probably Lower Triassic sediments are also continental deposits, generally red sandstones, siltstones, and mudstones, with a conglomerate at the base of these deposits. These Lower Triassic rocks, generally of fluvial origin, must have covered at least large parts of the Axial Zone of the Pyrenees, also in the area studied, where slabs have been preserved, unconformably overlying Devonian and Carboniferous rocks. Thicknesses range from 180 to 220 metres. The Middle and Upper Triassic deposits, consisting of lagoonal sediments (mainly dolomitic limestones and dolomite, with thicknesses of 200 to 270 metres) must also have covered our area since they, too, are found north of the Alpine thrust zone, in the upper valley of the Baliera River. In the Nogueras Zone they contain gypsum, which strongly affected the structure of the Nogueras Zone, by acting as a lubricant for the sliding structure and diapyrical intrusions (see also Chapter IV and Mey, 1968).

The remaining Mesozoic sediments are only known from the area south of the Nogueras Zone. These Jurassic and Cretaceous deposits consist mainly of carbonates, limestones and marls, which locally attain great thicknesses in the southern marginal basins, which were located parallel to the present strike of the Hercynian axial zone of the Pyrenees (de Sitter, 1964, 1965, Souquet, 1967, Mey, 1968, Mey et al. 1968, van Hoorn, 1970). It is doubtful whether the axial zone was ever covered by the Jurassic and Cretaceous sediments. Information concerning the Jurassic and Cretaceous can be found in Souquet (1967), while van Hoorn (1970) gives information relating to the Upper Cretaceous. Souquet (1967) stated that the north-Pyrenean and south-Pyrenean Jurassic and Lower Cretaceous are very well comparable and that locally connections must have taken place across the slightly submerged Axial Zone. During the main part of the Upper Cretaceous, sedimentation was fully marine as far north as the Nogueras Zone, although the deposits gradually grow thinner in a northerly direction (van Hoorn, 1970). It is therefore possible that they at one time at least partly covered the area studied, which is only some 0–15 km north of the Nogueras Zone. The total thickness of these Jurassic and Cretaceous sequences can, in local basins, amount to several thousand metres (Mey, 1968; van Hoorn, 1970).

After the deposition of these marine shelf-sediments of Jurassic and Cretaceous age and the deeper marine turbidites of Upper Cretaceous age (van Hoorn, 1970) farther south a regression took place in Maastrichtian time and sedimentation in a continental environment occurred. In Paleocene and Eocene times, marine, littoral and continental sediments were deposited on the border of a marginal basin (van Eden, 1970), south of the Nogueras Zone of the area studied. The detrital material had been supplied by the Axial Zone, that already began to rise during the Cretaceous folding phase, while marginal basins developed outward of the Lower Cretaceous basins (de Sitter, 1965). The Late Eocene Pyrenean folding phase mainly affected the marginal basins of the Pyrenees, whereas it had a limited influence on the Axial Zone. The Axial Zone continued to rise after this orogenic phase, and fluvial piedmont deposits were spread out, consisting of conglomerates and sandstones, which derived mainly from Jurassic and Cretaceous rocks (Nagtegaal, 1966). These deposits are of an Upper Eocene to Oligocene age and are exposed south of the Nogueras Zone of the area studied, overlying folded Mesozoic rocks, Paleocene and Lower Eocene deposits, but more to the east they locally extend over the Axial Zone. It is not known whether they also covered the area studied. Thicknesses of these fluvial fans range from 300 to 700 metres.

During Miocene times a planation surface developed in the Central Pyrenees. The post-Miocene uplift and erosion led to the present Pyrenean mountain chain. During the Pleistocene, glaciers formed which left glacial erosion marks and morainal deposits in some of the valleys of the area studied (Mey, 1967b, 1968, Höllerman, 1968, Wennekers, 1968).

STRUCTURE

The Hercynian diastrophic movements which probably already began in early Carboniferous times, increased in strength towards the Westpalian B (Mey, 1967 b, 1986). The following Hercynian deformation phases can be distinguished, according to Mey's words (1967b, p. 153; see also Boschma, 1963 and Zwart, 1964):

... "The first deformation produced concentric, open to tight asymmetric folds without cleavage development. Their axial planes have a general E-W to ESE-WNW trend in the north (Sierra Negra Unit)[1], a constant NE trend in the centre (Baliera Unit)[1], and an E-W and NW-SE trend in the south (Ribagorzana Unit)[1]"...

The main outcrop pattern of the competent Basibé Formation in the area studied is largely dominated by the large W-E and SW-NE striking asymmetric concentric folds of this first deformation phase. These folds are sometimes even overturned (Mey, 1967b, 1968). The shapes of these folds are only slightly disturbed by the following Hercynian deformation phases, which compressed the existing structures. North of the Alpine thrust zone, where the quartzite of the Basibé Formation is lacking, isoclinal folds were formed during the first deformation phase, with W-E to WNW-ESE striking axial planes.

Mey (1967b, 1968) describes the second deformation as the main folding phase, which, to quote Mey's text (1967b):

... "was caused by a N-S compression and is characterized by tight to isoclinal folds with a steep northward-dipping axial plane cleavage in the north, the dip becoming more moderate in the centre and south".

This was followed by a third deformational phase, that:

..."bent the entire structure around a NNE-trending axis coinciding with the bed of the Ribagorzana River. The Maladeta granodiorite and accompanying dykes intruded parallel to the general cleavage trend and caused a metamorphic aureole of moderate width. Near its southern border, gravity folds were formed locally"...

The Maladeta granodiorite cut off the rocks of the Basibé Formation in the area north of the Alpine thrust zone.

Again quoting Mey's description (1967b):

... "A fourth deformational phase was produced by a renewed N-S compression, causing local folding of the first or main-phase cleavage, also showing a weak secondary axial plane cleavage (fracture or crenulation cleavage), and local thrust movements along the earlier cleavage plane. This deformation might be a late Hercynian or an Alpine phase"...

Generally it has been accepted (Boschma, 1963; Zwart, 1964) that this granodiorite intrusion predated the actual deformation of the W-E refolding, which resembles the fourth deformation phase of Mey (1967b). However, conclusions regarding the relative age of this intrusion by Mey (1967b, 1968) are rather contradicting.

An uplift of the original Hercynian core occurred along important fault zones, i.e. the North-Pyrenean and South-Pyrenean faults, probably starting immediately after the Hercynian folding. Marginal basins in the two external zones on the borders of the Axial Zone were formed during the Jurassic and Lower Cretaceous (de Sitter, 1964, 1965), while the marginal basins of Upper Cretaceous to Eocene age developed outward from these Lower Cretaceous basins. Strong folding of a concentric type occurred in the marginal basins during the Late Eocene Pyrenean folding phase. The Axial Zone was shortened during the major Alpine deformation by means of north to south upthrusts and overthrusts along the Hercynian cleavage or fault planes, which at the same time caused asymmetric folding of the post-Hercynian strata above the unconformity (Mey, 1967b, 1968). Hence the competent Basibé Formation was also affected by the Alpine deformation, mainly by shear and fault movements. The amount of throw of the different thrusts was evaluated by Mey (1967b, 1968) to be several hundred metres.

The Nogueras Zone, which is a narrow zone between the Axial Zone in the north and the Jurassic and Cretaceous folded basins in the south, consists of large slabs of normal and overturned strata of Upper Carboniferous to Upper Triassic age, in which pre-Hercynian rocks also occur (Mey, 1968, see also Chapter IV). This Nogueras Zone, which extends from the Esera Valley up to the Segre Valley, is interpreted by Mey (1968) as a steep flexure zone that collapsed due to the diapirism of the incompetent Upper Triassic series, causing the irregular structure of this fault zone. The overlying Mesozoic calcareous rocks are thought to have moved to the south by sliding over the marls and gypsum of Upper Triassic age.

RELIEF

The area of the Basibé Formation outcrops has a steep relief, ranging in altitude from about 1000 to 2700 metres. The Basibé Formation appears as a very competent rock unit composed of limestones, dolomites and quartzites, which together form one single conspicuous ridge in the area (Fig. 5). The underlying and overlying slates of respectively the Gelada Formation and Fonchanina Formation form smooth denudation forms and depressions, generally covered with grass vegetation. The Lower Carboniferous shale in the triangle between the Basibé Formation outcrops and the Nogueras Zone also forms smooth slopes.

The courses of the main rivers are almost perpendicular to the most prominent structural features. The Upper Isabena and Baliera Rivers cross the resistant Paleozoic rocks, especially the competent Basibé Formation, in narrow V-shaped valleys. The course of the Upper Noguera Ribagorzana River has been influenced by a brecciated zone in the Maladeta Massif, which also separates two petrographic units (Charlet, 1968).

[1] The Sierra Negra Unit, as given by Mey corresponds to the area north of the Alpine thrust and fault zone, the Baliera Unit with the main outcrop area of the Basibé Formation containing quartzites (Fig. 2) and the Ribagorzana Unit with the area south of the foregoing area and north of the Nogueras Zone.

Mey (1968) pointed out that the Noguera Ribagorzana River and the Noguera de Tor River in the area studied are situated where the cleavage shows sharp bends within these valleys, and therefore supposed fundamental fractures at greater depths.

The direction of the present rivers may have been dictated by the shape of the Miocene relief, which was the final product of the steep denudation forms of Late Eocene-Oligocene times, when a southward directed stream pattern existed (Nagtegaal, 1966). A doming-up in Upper Miocene times of the Miocene planation surface, reactivated the run-off and the erosion of the slopes of the Pyrenees. The planation level is a well-known feature in the Central Pyrenees (Nussbaum, 1935, 1938; García Sainz, 1940a, b; Kleinsmiede, 1960; Zandvliet, 1960). Mey (1967b, 1968) mentioned some flat surfaces at about 2000–2400 metres altitude in the area studied, that may belong to this denudation level. As Rijckborst (1967) pointed out, the Miocene water divide coincides very well with that of the present main rivers of the Central Pyrenees.

CHAPTER III

STRATIGRAPHY OF THE DEVONIAN

NOMENCLATURE

The Devonian of the Central Pyrenees consists mainly of limestone and slate alternations. The sediments and the fossil content indicate deposition in a mainly shallow marine environment. Intense deformation during the Hercynian and Alpine orogenies and paucity of fossils make it difficult, however, to interpret the detailed stratigraphy of the Devonian of the Central Pyrenees. The basic (bio)stratigraphic work in the area studied was carried out by Dalloni (1910, 1930) and Schmidt (1931). For the southern Pyrenees, between the Esera River and the Noguera de Tor River, Mey (1967b) performed detailed lithostratigraphic work as a result of which he proposed a subdivision into a Sierra Negra and a Baliera facies area. The main argument for this distinction between two different lithofacies areas was the existence of the Basibé quartzite-dolomite member and the thick amount of slates in the lower part of the sequence in the Baliera facies area.

In this so-called Baliera facies area five formations were distinguished by Mey (1967b, 1968; see also Fig. 3), from top to bottom:

Mañanet Griotte	- nodular limestone, limestone and calc-schists
Fonchanina Formation	- slate with rare limestone intercalations
Basibé Formation	- limestone, quartzite, dolomite and nodular limestone
Gelada Formation	- sandy slates, quartzwackes and few impure limestones
Aneto Formation	- mainly shale and slate with few marly limestone intercalations

In the so-called Sierra Negra facies area, north of the Alpine thrust zone, the Basibé quartzite and dolomite is missing, and the Gelada and Aneto Formation can no longer be separately mapped. For the inseparable lower unit, Mey (1967b) introduced the name Rueda Formation. The earlier introduced name Castanesa Formation (Mey, 1967b), for the Basibé Formation in the so-called Sierra Negra facies area, was dropped (Mey, 1968).

Mey (1967b) named the Basibé Formation after the Basibé massif (2725 metres), a mountain ridge which limits the Upper Isabena Valley in the north, and which contains a vast outcrop of sediments of this formation. However, he did not give a type section from this area, probably because the Basibé Formation is rather incomplete and tectonically disturbed at that place.

We propose our Section 25 in the Barranco de San Silvestre (see Fig. 1 and plate I) as a type section for the Basibé Formation, but retain the name Basibé Formation, this mountain being a well-known feature in the type area. As the lowermost part of the type section is tectonically somewhat disturbed, we propose to supplement the type section with a reference section. Section 26 in the Barranco de Ponferrat was chosen as reference section for the Nodular Limestone Member (International Subcomm. Stratigr. Termin., 1961; American Comm. Stratigr. Nomencl., 1961).

The characteristics of the Basibé Formation and a distinction into members is given in Chapter IV.

A better name for this Basibé Formation, which is mainly composed of calcareous rocks and can be traced over a large part of the southern Central Pyrenees, would have been a formation name derived from its calcareous development between the Esera River and the Segre River. The name Basibé could then have been reserved for the Quartzite-Dolomite Member in the area between the Esera River and the Noguera de Tor River. However, in view of the well-established use of the name Basibé Formation, priority is given to this term.

The Baliera and Sierra Negra (litho-) facies areas that were distinguished by Mey (1967b, 1968) do not give any clarification of the differences in lithology, which are better expressed by the use of the term Quartzite-Dolomite Member of the Basibé Formation in the Esera-Tor area, south of the Alpine thrust zone (see also Krumbein and Sloss, 1963, p. 321).

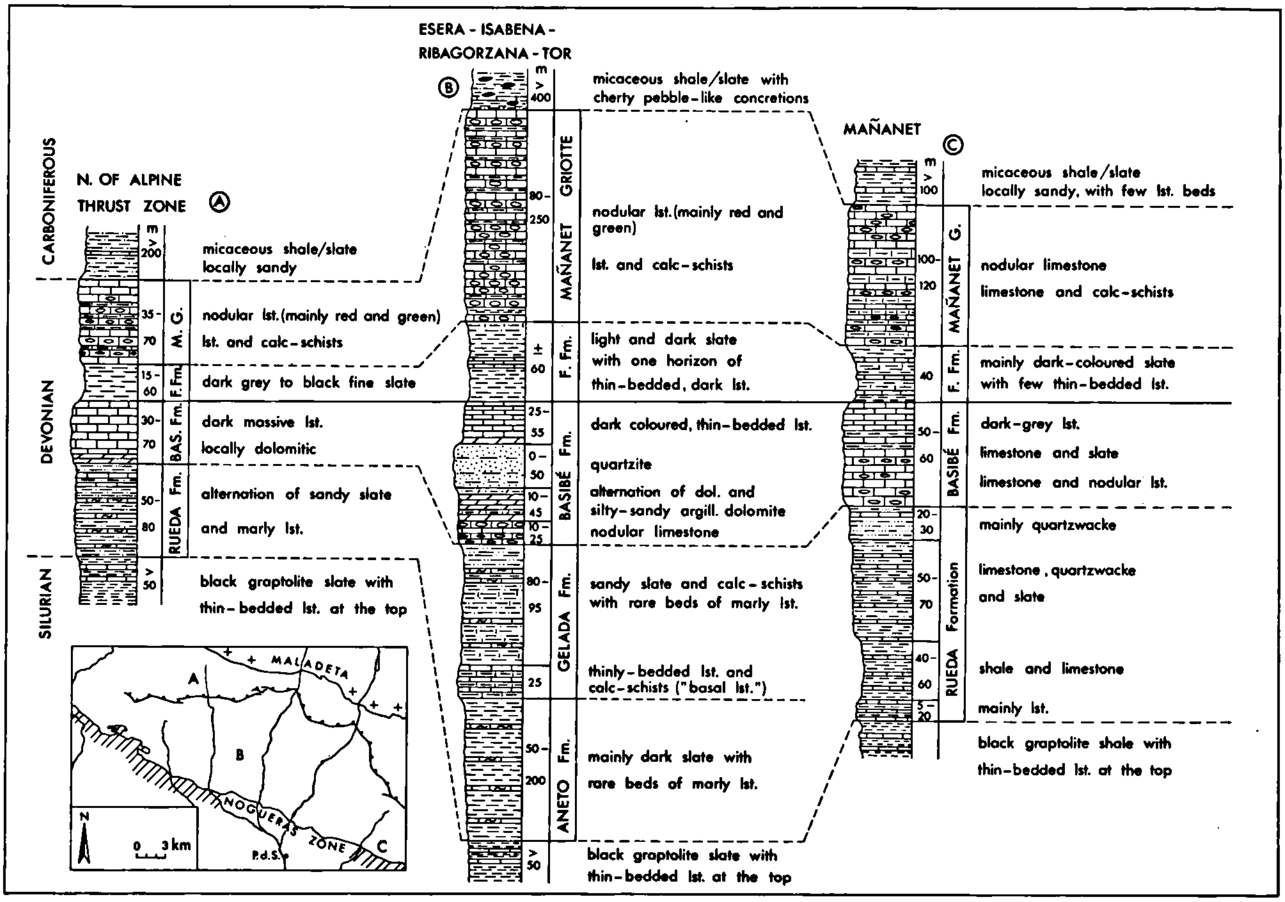


Fig. 3. Schematic lithostratigraphic sections of the Devonian in the area. (Modified after Mey, 1968).

AGE

The age of the Basibé Formation in the area studied is difficult to establish, since none of the observed fossils (Brachiopoda, Crinoidea, Orthocerida, Trilobita, Tentaculitida and isolated Tabulata) are sufficiently well preserved to permit of accurate identification. Mey (1967b) gives a detailed fossil list, derived from Dalloni (1910) and Schmidt (1931), mainly consisting of Brachiopoda, Rugosa, Tabulata, Trilobita and Orthocerida, which according to Dalloni and Schmidt, indicate a Middle Devonian age, confirmed by Dalloni in his 1930 paper. Unfortunately, it is not clear from Schmidt's description whether the assemblages of fossils collected in the western part of the area studied were taken from rocks of the Basibé Formation or from the Gelada Formation. Both Dalloni (1910) and Schmidt (1931) mention the quartzite on the Pico Basibé, and call it a Carboniferous quartzite.

More exact locations were given by Schmidt (1931) for fossil collections from rocks of the Gelada Formation (see also list in Mey, 1967b). Schmidt assigned a "Lower Coblentzian" age to the fossils of this formation, which agrees with the age given by Dalloni.

The Fonchanina Formation has only a very few identifiable fossils. Earlier workers mentioned no fossils from this formation; the same applies to the Mañanet Griotte. Correlation with the well-dated nodular limestone (griotte) formations in the adjoining areas makes it probable that the age of the Mañanet Griotte is Frasnian to Famennian (and locally Lower Tournaisian) (Dalloni, 1910, 1930; Schmidt, 1931; Destombes, 1953; Keizer, 1953; Ziegler, 1959; Mirouse, 1962, 1967; Mey, 1967a, b, 1968).

A conodont investigation of the Basibé Formation in the area between the Pallaresa and Segre Rivers (eastern Central Pyrenees), and of the outcrops in the Devonian Freixa and Castells blocks (Roberti, 1971) by K. T. Boersma (pers. comm.) indicates an Emsian age for the upper limestones of this formation. Samples from the upper limestones of the Basibé Formation from the Flamisell area (east of the area studied, see Roberti, 1970) also give an Emsian age.

A search for conodonts by K. T. Boersma in our samples taken from the Nodular Limestone Member and the Limestone Member of the Basibé Formation in the area studied, resulted in some specimens from Sections 7 and 8 (both situated in the Upper Isabena

Valley), whereas other samples did not contain any conodonts. The conodonts from Sections 7 and 8:

Icriodus latericrescens bilatericrescens
Spathognathodus cf. steinhornensis

(K. T. Boersma, pers. comm.) indicate an Emsian age. They were taken from limestone samples at a level of the Limestone Member which can be correlated with the 90–95 metre level of Section 9, Plate I.

According to these conodont data, the upper part of the Basibé Formation was deposited during Emsian (upper Lower Devonian) time. This age contradicts the Middle Devonian age given to these deposits of our area by Dalloni (1910, 1930), Schmidt (1931) and Mey (1967a, b, 1968).

CORRELATION WITH SURROUNDING AREAS

A paleogeographic interpretation of the Devonian of the Pyrenees can only be given in a very schematic manner since detailed lithostratigraphic data only exist from a few areas and these rocks have not furnished satisfactory biostratigraphic information. Most lithological data are derived from regional tectonic studies, whereas stratigraphic or sedimentological studies are scarce or not available at all. Paleontological data come mainly from the Basque Pyrenees, western Pyrenees and Montagne Noire, while the western part of the Axial Zone, the northern and eastern Pyrenees and also the southern Pyrenees supplied some isolated data; their chronostratigraphic usefulness however is reduced by the fact that macrofauna and microfauna occur in different rock types. The central part of the Pyrenees (Valle de Arán and the areas described by Zandvliet (1960) and de Sitter and Zwart (1962)) lack fossil data, so that Conodontophorida investigations in the limestones of these areas would probably be useful.

Mirouse made an attempt, in 1962 and also in his review on the Devonian of the western and central Pyrenees (1967), to collect the lithostratigraphic and biostratigraphic data of these areas, as did Ovtracht (1967) for the eastern Pyrenees, Mouthoumet massif and the Montagne Noire, whereas the latter area is described in more detail by Avias et al. (1967). These workers all give schematic paleogeographic interpretations of these areas. Mirouse (1962, 1967), Mey (1967a, b) and Zandvliet (1960) remark on the variations in lithology and thickness of the Devonian rocks, perpendicular to the Hercynian axis, whereas these features are constant in a W-E or WNW-ESE direction, parallel to the Hercynian orogene. These variations in thickness, mainly of Middle Devonian and Upper Devonian rocks in the western and central Pyrenees are similar to the variations existing in our area (see Fig. 3), and both Mirouse (1962, 1967) and Mey (1967a, b) explain these by assuming different facies areas, caused by more or less subsiding basins, which are directed W-E and separated by more stable or less subsiding ridges. The subsidence of these units was a prelude to the approaching Hercynian orogeny. The maximum thickness of the Devonian rocks of the Hercynian core is nearly constant from the Basque Pyrenees and western Pyrenees up to the Upper Salat area (1000–1400 metres), whereas the thicknesses in the northern Pyrenees (about 600 metres) and in the southern Pyrenees (about 300 and 600 metres) are much less, but also nearly constant in a W-E direction.

Almost all the Devonian sediments and faunas of the Pyrenees indicate a shallow marine deposition. The deposits in the Lower Devonian consist mainly of detrital sediments, while the younger deposits are mainly calcareous, with only some detrital sequences in the Middle Devonian or Upper Devonian. In the central area, however, the Devonian rocks are mainly detrital. The overall calcareous sedimentation which began in the eastern Pyrenees in upper Lower Devonian times and in the western Pyrenees on the Lower Devonian – Middle Devonian boundary is in sharp contrast with the underlying detrital sediments.

As shown by Boersma's datings (pers. comm.), the carbonate sedimentation in the southern Pyrenees already began in Lower Devonian times, so that quite a different picture arises as compared with that given by Mey (1967a, Figure 3c and 3d) for the distribution of the lithofacies areas, in which Mey places the carbonate sediments of the Basibé Formation in the Middle Devonian.

The nodular weathering limestones and the gray limestones of the Basibé Formation, can lithologically, be correlated in the southern Central Pyrenees with the nodular weathering limestones and the gray limestones occurring up to the Flamisell valley (about 10 km east of our map-Fig. 2), where the nodular limestones disappear, although gray limestones continue up to the Segre valley (about 40 km SE of our map: Roberti, 1970, 1971; Hartevelt, 1970).

Biostratigraphically the upper deposits of the Basibé Formation of these areas can be correlated with the aid of Conodontophorida (K. T. Boersma, pers. comm.), so that we may assume all these upper limestones of the Basibé Formation to have been deposited during Emsian time.

The underlying Rueda Formation is also present up to the Segre Valley, whereas the overlying Fonchanina Formation and Mañanet Griotte exist in these areas and the Llavorsi area, but disappear east of the Valira River (about 48 km E of our area) and are replaced by the Vilech Formation, consisting of multicoloured calc-schists, and the Compte Formation, consisting from bottom to top of limestone, nodular limestone and limestone. In the Devonian outcrops that exist up to 80 km east of the Valira River the Basibé Formation is scarcely present or absent, while the Rueda, Vilech and Compte Formations show the same appearance as in the Valira area, except that the Rueda Formation contains a remarkable limestone unit, which, however, cannot be correlated with the limestone of the Basibé Formation (Hartevelt, pers. comm.).

In the Paleozoic area north of the Maladeta granodiorite batholith, the Valle de Arán, situated 12 km north of our area, Kleinsmiede (1960) studied the Devonian rocks. The north-western part of that area was also described by Destombes (1953), the south-western part by Waterlot (1965, 1969). Kleinsmiede (1960) gives a detailed description of the Devonian deposits (with a total thickness of 530–1060 metres), which, however, do not contain any fauna. From bottom to top he distinguished: (a) limestones with intercalations of chert, sandstone and slate, overlain by (b) slates with occasional limestones and thin sandstones, (c) limestones, (d) slates with detrital limestones, (e) slates and sandstones, followed by (f) sand/shale alternations (sediments of a "littoral facies"), (g) orthoquartzites (h) non-graded sandstones and shales, (i) graded sandstones and slates, above which occur (j) very fine-grained slates and sandstones. The sequence above the basal limestone with intercalations of chert, sandstone and slate, represented by limestones and up to and including the orthoquartzites, (b up to including g), with a total sequence thickness of 110–285 metres, was described by Kleinsmiede as an orthoquartzite-limestone association (Pettijohn, 1957), which he compared with the Quartzite-Dolomite Member of the Basibé Formation of our area, the orthoquartzite possessing the same characteristics of thinning in an easterly direction. The non-graded and graded sandstones with slates (h and i) of the upper part of the deposits are interpreted by Kleinsmiede as a turbidite association with an eastward directed sediment transport.

East of this area, about 40 km NE of our area, Zandvliet (1960) studied the Devonian rocks in the Upper Salat and Pallaresa Valleys, from which he mentions only lithologic data. He compared the Devonian rocks of the areas studied by him, which consist from bottom to top of an alternation of slates and limestones, and griotte limestones, with the stratigraphy given by Dalloni (1930), Schmidt (1931), Destombes (1953) and Keizer (1953). He also studied the Devonian rocks in the above mentioned Llavorsi area (about 15–40 km NE-E of our area), from which he did not give any paleontological data either. Zandvliet called attention to the marked differences in total thickness between the deposits in the Salat area (about 1400 m) and those of the Llavorsi area (about 330 m), which latter compare very well with the total thickness of the Devonian rocks in our area (see also for the Devonian of the Llavorsi area Hartevelt, 1970 and Roberti, 1971).

North and north-east of the Valle de Arán, de Sitter and Zwart (1962, 1958) describe the lithology of the Devonian deposits in the areas of the Garonne, Salat and Ariège Rivers, where they began their Central Pyrenean mapping project. The southern part of these Devonian outcrops (at about 30 km N and NE of our area) belongs entirely to the Axial Zone, which is limited in the north by the North-Pyrenean fault zone (de Sitter, 1965; de Sitter and Zwart, 1962). North of this fault, Devonian outcrops exist in the so-called satellite massifs: the Arize and St. Barthélemy massif. Detailed descriptions of these massifs (which are about 60 km NNE and 80 km NE of our Fig. 2) are given by Keizer (1953) and Zwart (1953). Keizer (1953) supplies a detailed lithology and paleontological data of the Devonian rocks, and mentions no paleontological data of earlier workers. Perseil and Tourenq (1963) dated the lower deposits of the Devonian of the Arize massif, which consist of Mg-limestones (not dolomites according to these authors) and calcschists, as Lower Devonian, on account of a trilobite fauna. These deposits are directly overlain by calcschists, limestones and nodular limestones (griotte) probably of Upper Devonian age, with conglomerates on the top. The authors supposed the Middle Devonian to be absent. In the description by Zwart (1965) of the Devonian outcrops south of the North-Pyrenean fault zone (about 80 km NE of our area) and outcrops in Andorra (about 50 km ENE of our area) no fauna is mentioned.

Further east, Devonian outcrops occur at about 120 km and 150 km east of our area, in the valley of the Têt River and SW of Perpignan, which outcrops were studied by Cavet (1957). Without mentioning fossils, he described calcschists and limestones which are overlain by limestones dated as Eifelian on account of their trilobite fauna; nodular limestone (griotte) of Upper Devonian age is also found.

Ovtracht (1967) mainly reviews lihological data of the Devonian rocks of the eastern Pyrenees (area described by Cavet (1957)), the Massif de Mouthoumet (about 140 km ENE of our area) and the Montagne Noire (about 220 km NE of our area). The schists, limestones and dolomites which are considered to represent Lower Devonian deposits lack characteristic fossils, whereas the calcareous Middle and Upper Devonian deposits can be dated by their macro- and microfauna. Avias et al. (1967) describes the lithostratigraphic composition of the Devonian of the Montagne Noire, with Conodontophorida data for the Middle and Upper Devonian. From that area, Maurel (1965) also mentioned limestones and dolomites of Lower Devonian age and limestones of Middle Devonian age, with some fossils.

West of our area, the Basibé Formation can be correlated lithologically, according to Wennekers (1968), with the gray sandy limestones and limestone-dolomite alternations occurring up to 15 km W and NW of the Esera River, which are locally high-grade metamorphic. For the north slope of the Axial Zone, 15–30 km N to NNW of our area, in the surroundings of Luchon, Destombes (1953) provides lithological data concerning the Devonian rocks, with datings derived from biostratigraphic work of earlier workers (Bresson, 1903, Dalloni, 1910). His later work (Destombes, 1959) indicates that schists, calcschists and some limestones represent the Lower Devonian and that the overlying schists according to their trilobite fauna, belong to the lower part of the Middle Devonian.

Pelissonnier (1958, 1959) presented lithological data on the Devonian of areas west and east of the Garonne River (about 30 km NNW of our area, also described by Destombes, 1953; de Sitter and Zwart, 1958, 1962), which consist of an alternation, from bottom to top of schists, calcschists, limestones, dolomites, schists and nodular limestones (griotte), without, however, any paleontological data.

West of the areas described by Destombes and Pelissonnier, Clin (1959/1964) studied the Devonian deposits of the Axial Zone in an area which is about 35 km NW to 20 km N of our area, but only gives lithological data. In the southern part of the area he mentions limestones and dolomites at the base of the Devonian deposits, whereas in the northern part mainly schists occur, with thin quartzites and some limestone.

From the upper Cinca area, on the south slope of the Axial Zone, about 35 km WNW of our area, van Lith (1965) reports shales and greywackes with some limestones as Coblentzian deposits. He collected faunas but also mentions the fauna found by Bresson (1903) and made use, too, of correlations of bio- and litho-stratigraphic data of Dalloni (1910, 1930), Laverdière (1930), Clin (1959), Mirouse (1962) and Wensink (1962). The transition from the Lower Devonian to Middle Devonian is represented there by a thin (2–4 m) calcareous zone with Brachiopoda, considered by Schmidt (1931) and Wensink (1962) to be Lower Devonian and by Dalloni (1910) and Laverdière (1930) to be Middle Devonian.

Wensink (1962) provided a detailed stratigraphy for the Devonian rocks of the Upper Gallego and Ara Valleys (75–55 km WNW of our area), which mainly consist of greywackes, shales (with in the upper part marls with intercalations of some limestones) and sandy shales, overlain by the above-mentioned calcareous zone. The sediments below this calcareous zone contain an abundant fauna, consisting mainly of Brachiopoda, Trilobita, Anthozoa and Bryozoa, which indicates a "Coblentzian" age. The Middle Devonian deposits are, for the greater part, composed of well bedded limestones, which are sometimes dolomitic or marly and often alternate with reef limestones, which both supplied many fossils.

West of this area, in the area north of Canfranc (about 85 km WNW of our area) and in the Upper Aragón Subordan Valley (about 105 km WNW of our area), van der Lingen (1960) and Schwarz (1962) mention greywackes and shales of "Coblentzian" age, with a thin sandy limestone in the top. These Lower Devonian deposits were dated by means of Brachiopoda and other fossils mentioned by Schmidt (1931), and show a great resemblance to those of the area studied by Wensink (1962), which is also true of the Middle Devonian deposits.

The Devonian outcrops of these three areas and of the area west of the Gave d'Aspe and east of this river up to the Gave de Pau, on the north slope of the western Central Pyrenees, have been thoroughly studied by Mirouse (1962/1966). The latter area lies north of the areas described by Wensink (1962), van der Lingen (1960) and Schwarz (1962) and about 45–110 km WNW and 70 km NW of our area. Mirouse (1962/1966) gives detailed lithological sections, the rocks of which are well dated by an abundant macro- and microfauna. Sandy shales, greywackes and shales with in the lower part some quartzitic layers and in the upper part sometimes sandy argillaceous limestones have been dated as Siegenian and Emsian (together Coblentzian), whereas the argillaceous limestones in the top of this sequence generally represent the transition from the Lower Devonian to the Middle Devonian (base of the Eifelian, Mirouse, 1962, 1964, 1967). The Middle Devonian deposits are entirely calcareous, with reef limestones in the north and south. The different lithofacies areas which Mirouse (1962, 1967) distinguished perpendicular to the WNW-ESE directed hercynian structure, are confirmed by Nicolaî (1963), who made a detailed stratigraphical analysis of the Devonian rocks (which can be dated by fossils) in the valley of the Gave de Pau (about 65 km NW of our area).

The Lower Devonian of the western part of the Central Pyrenees is thus characterized by shallow marine detrital sediments. The Emsian sequence consists mainly of sandy shales and greywackes (with some few limestone). Mirouse (1962) presumed an emergent area to be situated south of the area he studied, during the Emsian and the transition to the Middle Devonian.

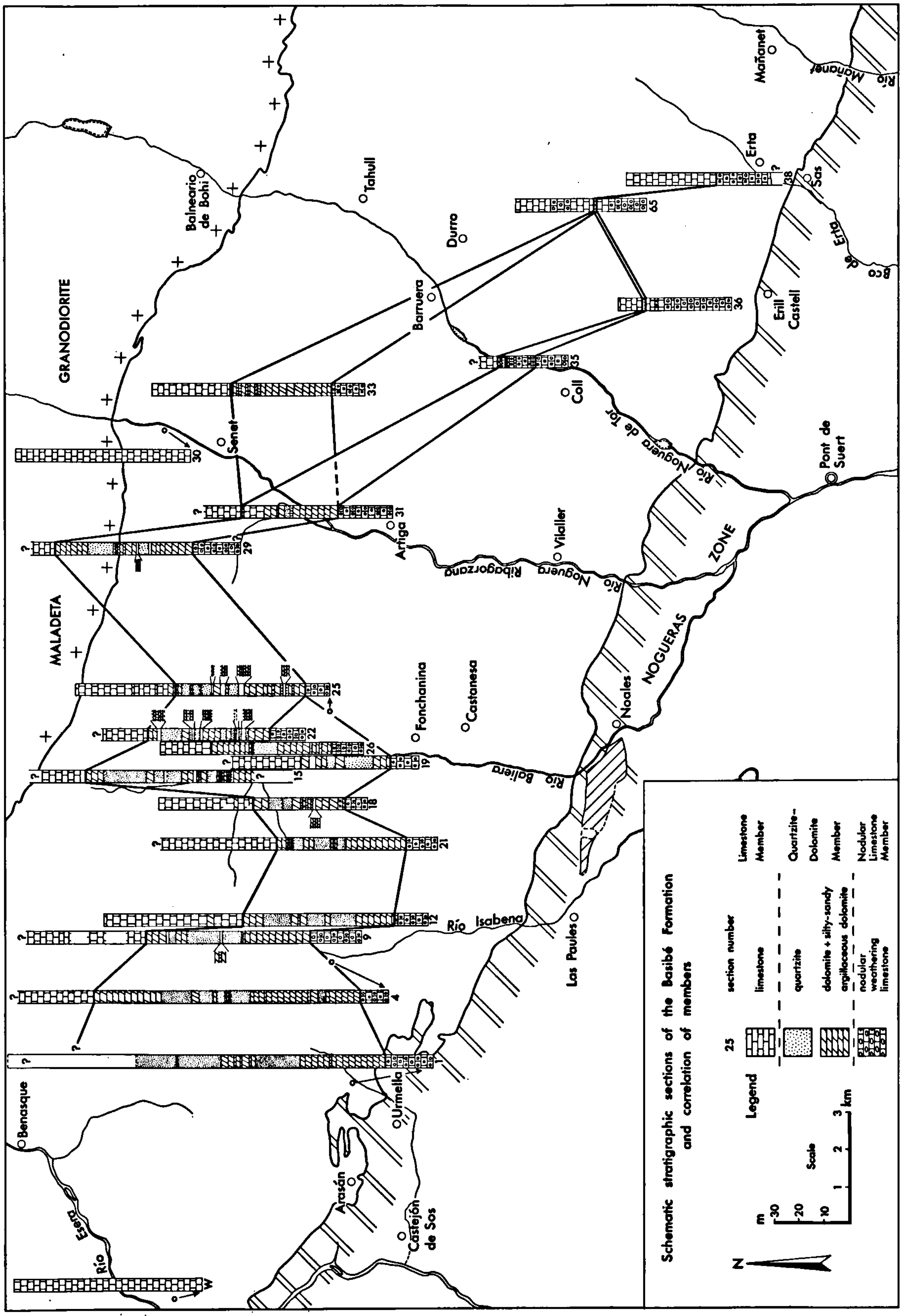
From the Basque Pyrenees, Heddebaut (1965, 1966) describes Devonian outcrops in the Aldudes massif, about 160 km WNW of our area, where a fauna of Emsian age is found in dolomites overlain by reef limestones of Eifelian age. Below the Emsian dolomites occur, from top to bottom: limestones, some quartzite layers, dolomite and slate, overlying quartzites, which are dolomitic or alternate with dolomitic layers, and probably have a Siegenian age (the quartzites had already been mentioned by Laverdière, 1930). Similar lithological descriptions are given by Damestoy (1961) of the Lower Devonian of the Aldudes massif and by Lagny (1964) of an area NW of this massif, where the deposits can also be dated as Coblentzian (Siegenian and Emsian).

Since the upper limestones of the Basibé Formation in the southern Central Pyrenees have been dated as deposits of Emsian age, and since data from the eastern and northern Pyrenees, the Mouthoumet massif and the Montagne Noire also indicate limestone or dolomite deposition during Emsian times, a continuous area of calcareous sedimentation must have occupied the entire southern, eastern and northern part of the Pyrenees at that time.

At that time the western Pyrenees still showed a sedimentation of detrital terrigenous deposits, which must have ceased in Middle Devonian times, since reef limestones developed in those areas. In the Basque Pyrenees some calcareous sedimentation occurred locally during Emsian time.

The increase or decrease in the supply of terrigenous siliciclastic detrital material would appear to have

Fig. 4.



Schematic stratigraphic sections of the Basibé Formation and correlation of members been the main reason for the differentiation of lithofacies of these areas, since the fauna does not suggest a considerable change in bathymetric conditions.

However, as was already stated above, a large part of the available stratigraphic information has only been collected for mapping purposes and cannot be used for sedimentological interpretations. For an interpretation of the paleogeography and the sedimentology of the Devonian deposits a different approach will be necessary, and our study intends to be a contribution to such an approach.

CHAPTER IV

FIELD CHARACTERISTICS OF THE LITHOLOGIC UNITS

The Basibé Formation in the area studied may be subdivided into the following members [1] (from bottom to top):

Nodular Limestone Member
- nodular weathering limestone
- silty-sandy argillaceous limestone
- dolomitic nodular weathering limestone

Quartzite-Dolomite Member
- quartzite
- dolomite
- silty-sandy argillaceous dolomite

Limestone Member
- limestone

The lateral relationship of these members within the area studied is shown in Fig. 4.

The Basibé Formation in its most characteristic development, as it occurs in the area of the Rio Isabena, Baliera and Ribagorzana, is very evenly bedded, with layers of 10–30 cm thick. Detailed sections of the formation, taken from the entire area, are given in Plate I, Sections 9, 18, 25, 31, 33, 35 and 36.

The Basibé Formation is underlain by the Gelada Formation, which consists of sandy shales to slates (according to Mey, 1967b, 1968), dark yellowish brown (10 YR 4/2) when weathered, and medium gray (N 5) when fresh. The Gelada Formation locally contains thin calcareous beds near the top and carbonaceous black streaks on the bedding and cleavage planes (see also Mey, 1967b, 1968). The boundary with the overlying nodular weathered limestone of the Basibé Formation is always sharp and conformable.

NODULAR LIMESTONE MEMBER

The "nodular weathering" limestones are very fine crystalline limestones the colour of which, when fresh, varies from very light gray (N 8) and light gray (N 7) to light bluish gray (5 B 7/1) and weathering to a grayish orange (10 YR 7/4) to yellowish gray (5 Y 7/2) colour. The uneven weathering pattern that brings about the nodular appearance, is the result of wavy argillaceous laminae, which are yellowish brown and sometimes greenish coloured (Fig. 6).

Locally these limestones alternate with high argillaceous limestones, with the same weathering pattern.

Beds are sometimes difficult to distinguish; the alternation of the light gray limestones with the brown argillaceous limestones is the only indication of bedding.

The thicknesses of the light gray and brown argillaceous limestones beds range from 10–20 cm, and are constant over great distances.

The limestones contain fragmentary fossil material, including crinoid columnals and brachiopods. The crinoid fragments are sometimes dark-coloured. Pyrite crystals occur frequently in the light gray limestones.

Sedimentary structures have not been recognized.

Upwards, these nodular weathering limestones are interbedded with medium dark gray (N 4) argillaceous limestones (to sometimes calcareous shale). The argillaceous limestones beds, with a light brown (5 YR 6/4) weathering colour, are about 5–15 cm thick, and contain quartz grains from silt-size to very fine sand-size. Brownish, wavy shale laminae generally surround the abundant crinoid fragments.

The silty-sandy argillaceous limestone beds are less resistant than all other lithologic units in the formation.

Cleavage is usually strongly developed in these beds, but its direction with regard to the bedding depends on the orientation of the section within the tectonic structures. In many places, the difference in angle of cleavage in this silty-sandy argillaceous limestone and in the more competent limestone is obvious: the cleavage is more perpendicular to the bedding in competent beds.

[1] We propose the following names for these members:
Ponferrat Nodular Limestone Member,
San Silvestre Quartzite-Dolomite Member and
Llaviero Limestone Member.

These geographic names have been derived from the brooks in the Baliera Valley, where typical sections of these members are developed, respectively Section 26,25 and 18. Since the latter section is not completely exposed, reference sections are proposed for the Limestone Member, viz. Sections 8, 12 and 25.

However, farther in this study only use will be made of the descriptive lithologic term of these formal names.

Fig. 5. Section 11, west slope of the upper Rio Isabena Valley. Stratigraphic top at the left. G: Gelada Formation, NL: nodular limestones, Q: quartzites, D: dolomites, B: Bunter Formation (Triassic). The boundary of the Basibé Formation and the Bunter Formation is an angular unconformity (see also Mey, 1968, p. 244). Looking northwest.

Bedding planes are generally sharp and conformable (Fig. 7). Beds, like those of all other lithological units in this formation, are very constant in thickness over long distances.

Sedimentary structures have not been recognized.

The upper layers of the Nodular Limestone Member are partly to completely dolomitized (Fig. 7). In an upward direction the nodular weathering character is lost, although the wavy argillaceous laminae or shale laminae with a brown-greenish colour still exist. This member is excellently exposed in Sections 4, 6, 9, 23, 26, 29, 31, 34 and 38, whereas the contacts with the underlying Gelada Formation and the overlying Quartzite-Dolomite Member are shown very nicely in Sections 26 (Barranco de Ponferrat) and 31 (Barranco de Artiga).

QUARTZITE-DOLOMITE MEMBER

The Quartzite-Dolomite Member is characterized by an alternation of evenly bedded quartzite, dolomite and silty-sandy argillaceous dolomite layers (Figs. 5 and 8).

The boundary of this member is given by the first appearance of dolomite containing quartz veins.

This dolomite directly above the Nodular Limestone Member is medium gray (N 5), finely, medium or coarsely crystalline with a grayish orange (10 YR 7/4) weathering colour.

In the Rio Isabena area and west of the Rio Baliera, dolomite and also other lithologic units can locally be reddish.

All dolomite beds contain numerous white quartz veins, which are perpendicular to the bedding. As a result of selective erosion, the quartz veins project above the surface of the weathered dolomite. The wavy argillaceous or shale laminae that occur in the dolomite have the same pattern as in the underlying nodular weathering limestone. In a few cases the rare wavy argillaceous laminae grade into heavily argillaceous, laminated dolomite.

Most contacts, however, are sharp and conformable. The thickness of the dolomite beds is 5–30 cm, between the quartzites even up to 1 m. Crinoid fragments occur abundant. Upwards, the first quartz sandstone or quartzite beds occur, 20–50 cm thick, with sharp, normal contacts with mainly silty-sandy argillaceous dolomite. In most sections these quartzite beds alternate with dolomite and silty-sandy argillaceous dolomite. Higher up in the formation, the number and thickness of the quartzites increases. Thick-bedded (up to 2.5 m) massive quartzite beds together form 7–10 m thick complexes in the Rio Isabena and Baliera area (Sections 9, 18 and 25; Plate I and Figs. 4 and 31).

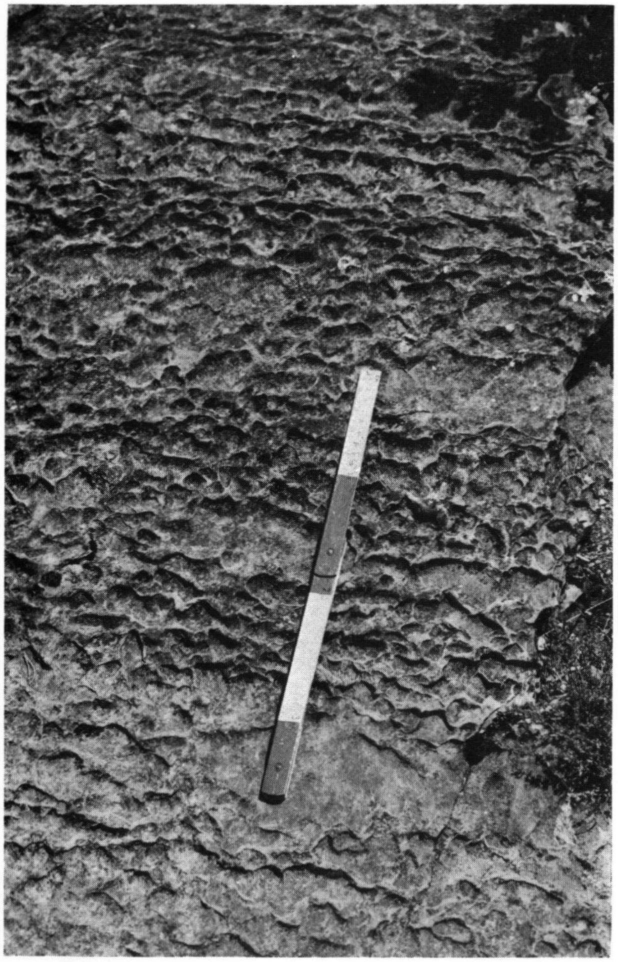


Fig. 6. Weathering pattern causing a nodular appearance in the Nodular Limestone Member. Section 35; ruler perpendicular to the bedding. Coloured parts are 10 cm long.

The quartzites are light gray (N 7) to medium light gray (N 6), with a very pale orange (10 YR 8/2) to grayish orange (10 YR 7/4) weathering colour. Clear white quartz veins occur abundantly (Fig. 8). The quartzite is a very resistant and hard rock. Most quartzite beds are very homogeneous, devoid of any structure. They are composed of well-rounded and well-sorted quartz grains, chiefly of fine and medium sand size, cemented by quartz. As the quartzite thickness increases, the grain size seems to increase as well.

Some isolated or basal quartz sandstones are very soft and brownish when weathered; they have a carbonate cement. Between some quartzite beds, laminated to very thinly bedded black shale is present (Fig. 8). The contact with upper and underlying beds is nearly always sharp and abrupt. Grading from shale to quartzite or the reverse is exceptional.

Sedimentary structures are scarce, and visible only on a smooth weathered or fresh surface. Small-scale cross-bedding and parallel lamination are the most common structures and are only found in some outcrops, mainly in the Urmella, Basibé and Baliera area (Fig. 29). Streaks of heavy minerals, such as occur in an outcrop in the Basibé massif, accentuate small-scale cross lamination very well. Burrows are visible on some weathered planes, perpendicular to the beds, or as holes in the bedding planes, as a result of selective weathering and erosion (Fig. 28).

Within one single very thick quartzite bed, horizons of homogeneous and burrowed rock may alternate. Fossils have never been found within the quartzites.

Other sedimentary structures such as load casts, or erosion features are absent. The quartzite layers are very well bedded, without any change in thickness within an outcrop.

Correlation with outcrops at a distance of some hundreds of metres is almost impossible. Within such a distance separate quartzite beds meet or beds are lost.

Nevertheless, the sections as a whole can generally be compared fairly well.

Joints, mainly quartz-filled, are abundant in the quartzite beds, but cleavage never occurs in the quartzites. In an outcrop near Arasan (Esera Valley), coarse fracture cleavage was observed, not, however, in a quartzite but in a carbonate-cemented quartz sandstone.

In the western and central part of the area, the quartzite beds of the Quartzite-Dolomite Member are overlain by an alternation of argillaceous dolomite with silt to very fine sand-sized quartz grains, and dolomite, below the medium-gray limestones of the Limestone Member.

In the eastern part, east of the Rio Tor, the dolomite and the silty-sandy argillaceous dolomite are nearly absent, and the quartzite beds are directly overlain by the upper Limestone Member. In some outcrops the quartzite beds are inter-bedded with the medium gray limestone of the upper member, e.g. in Sections 31 and 36.

The silty-sandy argillaceous dolomite is the equivalent of the silty-sandy argillaceous limestone of the Nodular Limestone Member, containing the same characteristics and fossils. Feeding trails occur in a thin layer of silty-sandy argillaceous dolomite on the top of a dolomite bed (Section 18).

In the entire area the Quartzite-Dolomite Member is almost completely exposed, excellent outcrops are formed by Sections 4, 9, 11, 12, 18, 23, 24 (quarry above the weir), 25, 26, 31.

Fig. 7. Transition from the Nodular Limestone Member to the Quartzite-Dolomite Member in Section 31 (approx. 20—25 m); north slope of the Bco. de Artiga. Top photo is stratigraphic bottom. Hammer and 1 metre ruler for scale. NL: nodular limestone, D: dolomite, S: silty to sandy argillaceous limestone/dolomite. Note the quartz veins in the dolomites, perpendicular to the bedding.

Fig. 8. Part of the Quartzite-Dolomite Member of Section 24. Stratigraphic bottom is top of photo. Q: quartzite, B: black shale, other abbreviations as in Figure 7. Length of ruler 1 metre.

Fig. 8

Fig. 7

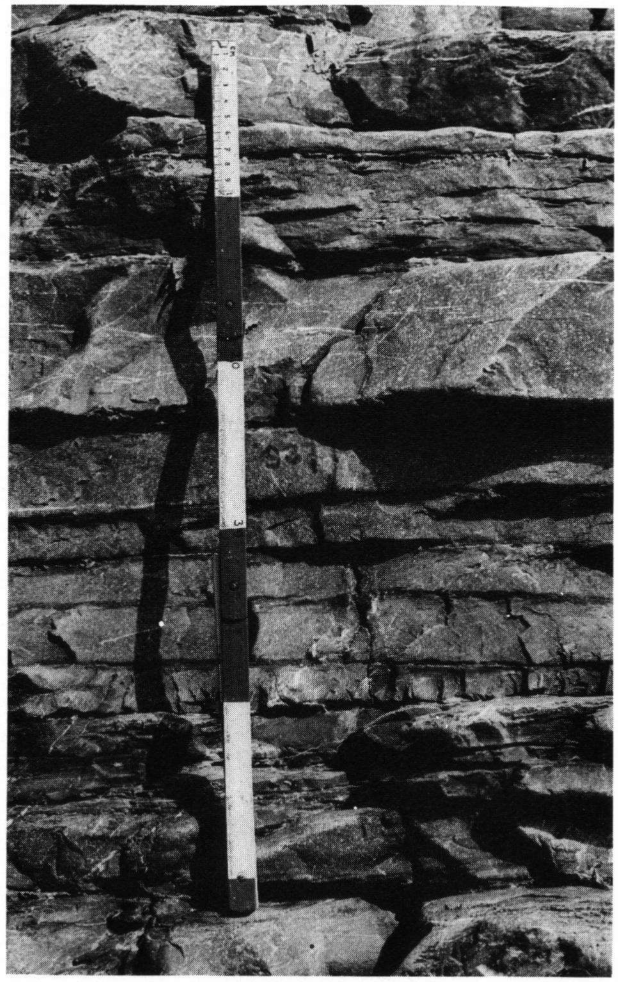
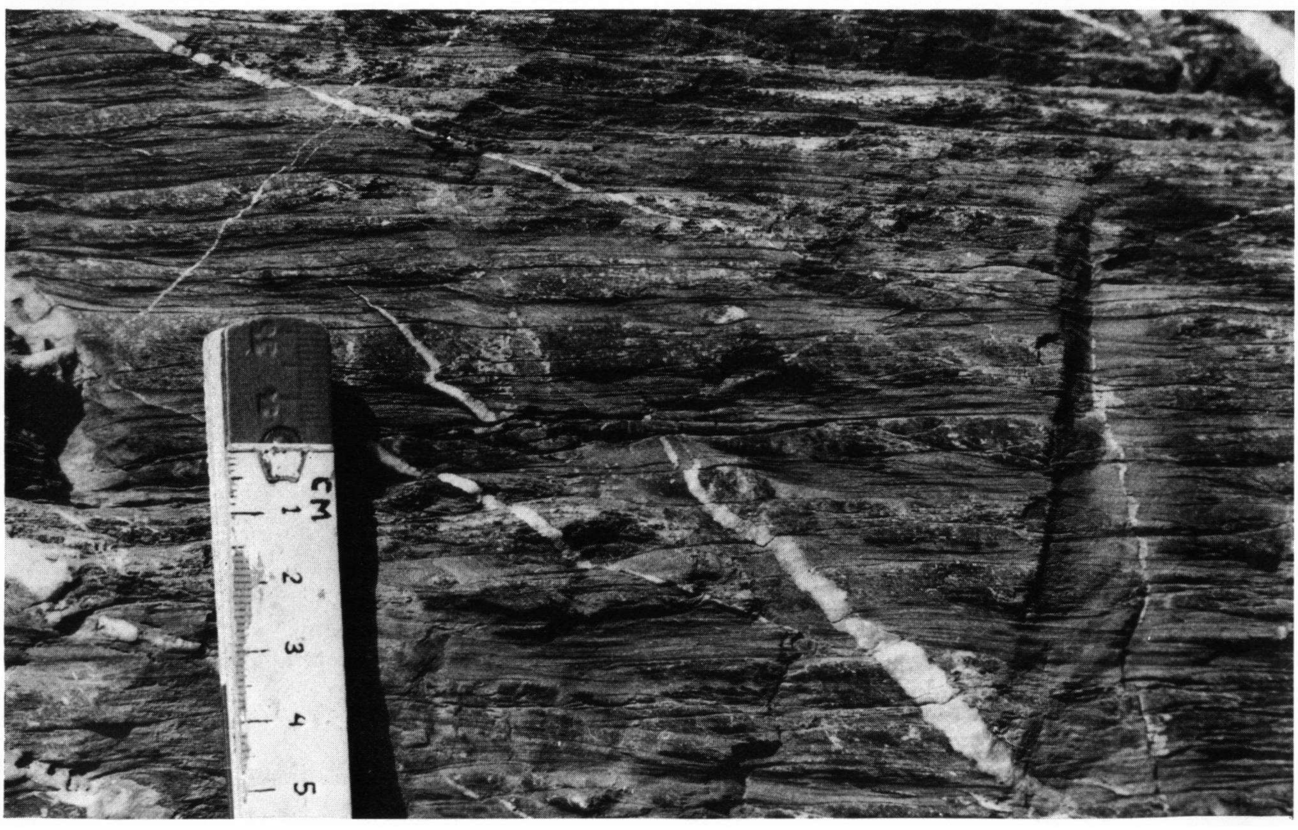


Fig. 9. Limestone Member of Section 24. Stratigraphic top is bottom of photo. Lower part of this member consists of thinly bedded, medium dark gray limestones.

Fig. 10. Alternation of laminated to very thinly bedded limestones and argillaceous limestones in the upper part of the Limestone Member of Section 26. Note displacement of calcite veins.



LIMESTONE MEMBER

The upper member of the formation consists of medium gray (N 5) to medium dark gray (N 4), thinly bedded limestones. Thicknesses of these very finely to finely crystalline limestone beds decrease near the top, where laminae occur. The limestone beds are 5–10 cm thick, and are frequently interbedded by limestone laminae (Fig. 9). In the upper part, thin beds sometimes consist of piled up laminae (Fig. 10). Well-developed cleavage often causes a lustrous appearance on the surfaces of the limestone beds.

The thicknesses of the layers are very constant, and the contacts are flat.

The limestones weather smoothly or in laminae. Weathering patterns indicate an alternation of medium gray pure limestone and brownish argillaceous limestone (Fig. 10).

Sedimentary structures occur in some outcrops. Load casts of silt within a limestone layer, a small gully, cross-bedding, and probably small ripples were noted (Sections 9, 12, 25 and 31).

In the top of the formation, where there is a transition from the calcareous laminae to the slates of the Fonchanina Formation, flaser-linsen structures were recognized. It is, however, sometimes difficult to decide whether features of this kind are of sedimentary or tectonic origin.

In thick, massive limestone layers (macro-) stylolites occur, filled with blackish shale residue and some pyrite crystals.

Macrofossils are found in some outcrops, e.g. isolated Tabulata, Brachiopoda, Orthocerida, and Crinoidea fragments.

The limestone contains many calcite-filled joints, which frequently show displacement along the intersections with the bedding planes, which often are the cleavage planes (Fig. 10).

Dolomitized layers within this limestone unit are very scarce, and usually isolated. Recognition is easy, since they have a yellow brownish or grayish orange weathering colour, in contrast to the medium gray of the limestones. These dolomite layers also contain quartz veins perpendicular to the bedding.

The upper boundary of the Basibé Formation has either a sharp contact with the overlying Fonchanina Formation, or shows a gradual transition. In the latter case, the boundary is laid, where the laminae are no longer calcareous. The Limestone Member, together with the contact with the Fonchanina Formation, occurs completely in Sections 7, 8, 12, 16, 23, 24, 25, 26, (31), 36 and 38.

The Fonchanina Formation is composed of fine, fissile, medium gray (N 5) to medium dark gray (N 4) slates, with lustrous cleavage planes. Higher up in this formation, a few thin-bedded gray limestones may occur. The Fonchanina Formation is a very incompetent and badly exposed unit in the area.

BASIBÉ FORMATION NORTH OF THE ALPINE THRUST ZONE

The Basibé Formation, north of the Alpine thrusts and faults, and south of the intrusive Maladeta granodiorite batholith, is solely composed of medium gray (N 5) to medium light gray (N 6) limestones and dolomites. More argillaceous limestone beds are scarce. Dolomitic parts are irregularly distributed within the limestones.

This fine to medium-grained carbonate rock generally has sharp contacts with the underlying and overlying slates of the Rueda and Fonchanina Formation respectively.

The rocks are thick-bedded to massive; the cleavage is well-pronounced. Crinoid fragments are often found.

Contact metamorphic influence is found in the area along the Maladeta granodiorite between the Rio Baliera and Rio Tor. This has caused a marked recrystallization and a strong accentuation of the alternation of medium gray limestones and darker gray argillaceous limestones. Coarse crystals of quartz, pyrite and micas can be seen on the surfaces or, if they are eroded, numerous small holes occur.

Mey (1968) mentioned, in the isoclinal synclines south of the Maladeta and between Senet and Bohi, a few sandy dolomites as well as nodular weathered limestones, in the lower part of the formation. This shows that there is probably a (more) gradual transition into the southern area of the formation.

THICKNESS AND DISTRIBUTION

The Basibé Formation shows a gradual change from west to east in the area studied, as is shown by Figure 4 and by Sections 9, 18, 25, 31, 33, 35 and 36, Plate I.

Broadly speaking, the thickness decreases from west to east, i.e. from the area between the Esera and Isabena Rivers to the Tor Valley.

The greatest thickness, probably amounting to about 170 m, is found in Section 1, situated near Urmella, west of the divide between the Esera River and the headwaters of the Isabena River. (In this Section 1, the two lower members measure up to 120 m; the upper limestone member is badly exposed but is evaluated at about 50 m).

The area of greatest thickness of the whole formation coincides with the area of greatest thickness of the middle member, i.e. the Quartzite-Dolomite Member. The decrease in thickness of the whole formation toward the east, the region of the Tor Valley, runs parallel with a similar decrease in thickness of this middle member, Figs. 4 and 32. Moreover, Section 1, where the whole formation is thickest, also contains the greatest accumulated thickness of quartzite layers : 52 m. This value also tends to decrease towards the east, although more irregularly. This general trend can be demonstrated by the following values, arranged from west to east, see also Figs. 4, 30, 31 and 32. Ten sections, made in the Rio Isabena Valley, give a mean thickness value of about 140 m for the Basibé Formation, in which is included about 20–30 m of quartzite layers. This mean thickness value for the Rio Baliera Valley, according to 13 sections, is about 80–90 m, with about 10–20 m of quartzite; for the Rio Noguera Ribagorzana Valley (west slope, 3 sections) the values are about 80 m and 12–18 m; for the east slope (1 section) about 75 m and about 3 m; for the Rio Noguera de Tor Valley (4 sections) about 45 m and about 2–4 m. A section east of the Tor River (Section 36) is about 45 m thick and contains 0.85 m of quartzite. Sections made still further east (in the Barranco de Erta Valley) show only 0.35 m of quartzite on the west slope, whereas outcrops on the east slope do not contain any quartzite at all.

In the east of the area studied the Basibé Formation consists only of the basal member (the nodular weathered limestone) and upper member (the well-bedded gray limestone), together about 40 to 50 m. Still more to the east, this lithology remains the same, with nearly uniform thickness, up to the Rio Flamisell, and is only composed of gray limestone up to the Rio Segre. Distinguishing in members is no longer possible (Roberti, 1970, 1971; Hartevelt, 1970).

West of the area studied, i.e. in the Esera Valley, the Basibé Formation has completely lost its Quartzite-Dolomite Member. West of the Rio Esera the Basibé Formation is composed of thick gray sandy limestones, overlain by gray limestones, sometimes with an alternation of yellowish brown weathered sandy dolomite and light gray argillaceous limestone, with a thickness of up to 75 m, according to Wennekers (1968).

From these data it follows that in a section from west to east, the Quartzite-Dolomite Member is wedge-shaped, which means that all the lithological units of which it consists, especially the quartzite, pinch out towards the east, Figs. 4, 30 and 32.

In a north-south direction between the Alpine thrust and fault zone in the north and the Nogueras zone in the south, differences in formation thickness and thicknesses of the lithologic components are not conspicuous.

In the north, between the Alpine thrust and fault zone and the intrusive Maladeta granodiorite batholith, the Basibé Formation is composed solely of gray limestone and some dolomite, with a thickness of 30–60 m (Mey, 1967b, 1968), up to 70 m in the area of the Rio Noguera Ribagorzana (Section 30).

The Alpine thrust and fault zone is the northern limit of the area where the formation consists of members containing nodular weathered limestone, silty-sandy argillaceous limestone and dolomite, quartzite, dolomite and limestone. The southern boundary is not known, since the formation is covered there by Carboniferous and Mesozoic deposits.

More to the south, however, in the Nogueras Zone, which mainly consists of Stephanian, Permian and Triassic rocks, some Paleozoic blocks occur which mainly consist of Devonian (Dalloni, 1910, 1913, Mey, 1968). Quartzite is completely absent in these blocks, which in the lower parts contain gray limestones, lithologically comparable with the Basibé Formation of the area studied.

There has been a discussion as to whether these blocks are autochthonous, i.e. have risen up from below by faulting and thrusting within the Nogueras zone, or derived from the Axial Zone in the north by sliding. Mey (1968) interpreted the Nogueras zone as a collapsed steep flexure zone in which at least the Paleozoic Las Paules block and the Gotarta-Malpas block are of sub-autochthonous origin. His main argument is that there are a number of differences between the Paleozoic rock characteristics of the Axial Zone and those of the Nogueras Zone, mainly differences in lithology of the Devonian, presence or absence of a slaty cleavage in the Paleozoic blocks, differences in the nature of igneous bodies intrusive in Paleozoic rocks and differences in the Stephanian to Lower Triassic rocks. The explanation that the majority of the blocks is derived from within the Nogueras Zone itself, is in accordance with the opinion of de Sitter (1965), Zwart and Mey (1965), and Roger (1965), who all advanced an autochthonous origin.

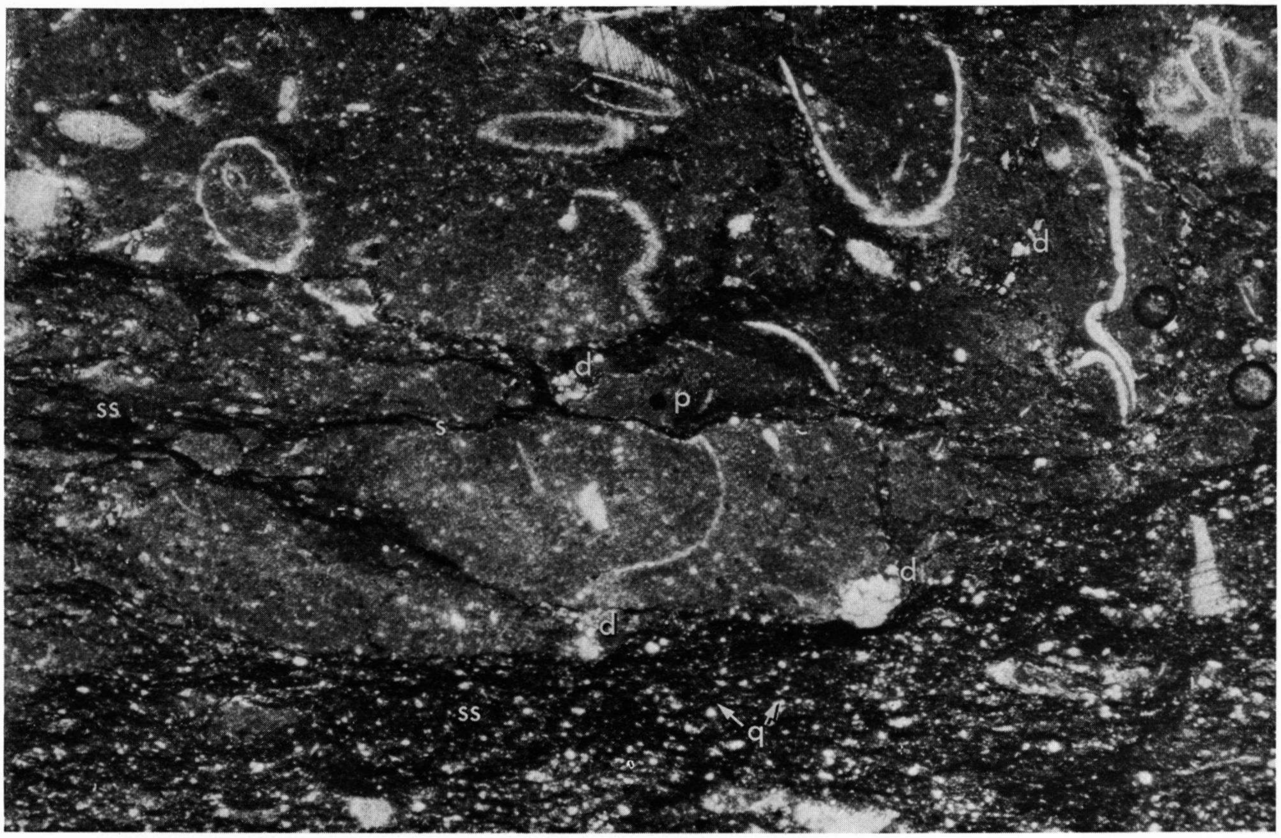
In Mey's (1968) review the interpretation of workers who advocated an autochthonous origin, including also Jacob et al. (1927), Misch (1934), Jacob (1935), Almela and Rios (1947) is placed against the opinion of those, Dalloni (1910, 1913, 1930), Nagtegaal (1962), Seguret (1964, 1966), Wennekers (1968), who explained the Paleozoic blocks as having come from a northern area by sliding.

If we accept a sub-autochthonous origin of the Las Paules block, south of the area under study, an approximative southern boundary for the occurrence of quartzite, dolomite and silty-sandy argillaceous dolomite of the Basibé Formation can be given. The clastic wedge represented by the quartzite and accompanied by dolomites must in that case occupy a very limited area, not extending beyond the present Rio Esera Valley, the northern Alpine thrust and fault zone, the Barranco de Erta, and the Nogueras Zone.

The W-E extention is then about 35 km, the N-S extention about 6 km; flattening out of the existing folds, in order to reconstruct the original lateral extent, will only slightly increase this value.

Fig. 11. Photomicrograph of thin section of nodular limestone consisting of very finely crystalline calcite, with a low dolomite (d), quartz silt (q) and pyrite (p) content and fossil fragments. Stylolites (s) and solution stringer (ss) are shown, in which concentrations of insoluble residues such as clay and quartz occur. These cause the nodular weathering appearance of the rock. Sample 398 (Section 31), nicols crossed, 20 x.

Fig. 12. Very finely crystalline nodular weathering limestone. Displacement of calcite veins (c), partly dolomitic (d), by stylolites (s). Crinoidea (cr), Trilobita (t), Echinodermata (e) and Bryozoa (b) fragments. Sample 51 (Section 12), nicols crossed, 50 x.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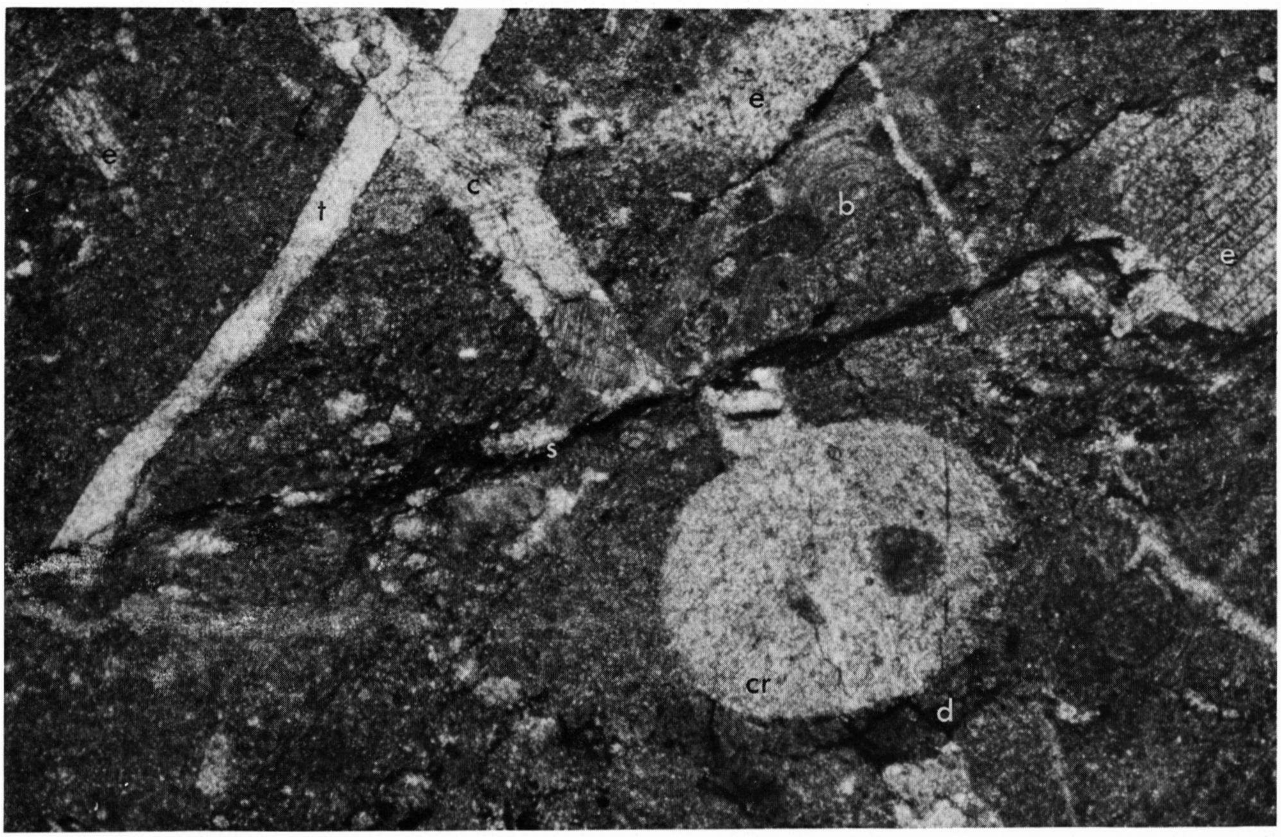
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CHAPTER V

PETROGRAPHY AND DIAGENESIS

The petrographical composition of a sedimentary rock is the result of:
a. the original composition of the sediment, and
b. the changes it has undergone during diagenesis.
One of the aims of the petrographical study of sedimentary rocks is to distinguish between the effects of these two features. According to Nagtegaal (1969) diagenesis can further be subdivided into:
a. changes directly related to the depositional environment,
b. changes after deposition leading to lithification, and
c. changes in the lithified sediment.
Study of more than 500 thin sections from the sections sampled (Fig. 2) showed that the distinction in lithologic units, as given in Chapter IV, is confirmed by microscopical analysis. The microscopical characteristics of these units are very uniform vertically and laterally as are the field characteristics, so that we may confine ourselves describing three main sediment types which can be distinguished:

1. an originally fossiliferous micrite or biomicrite, with a low silt and clay content, changed by diagenetic processes into:
   nodular limestone ("griotte"),
   dolomite and
   medium gray limestone.

2. fossiliferous micrite or biomicrite, with a high quartz-silt and quartz-sand, muscovite, and clay content, changed by diagenetic processes into:
   silty-sandy argillaceous limestone or
   silty-sandy argillaceous dolomite.

3. quartz-sandstone,
   diagenetically changed into:
   orthoquartzite or
   dolomitic orthoquartzite.

NODULAR LIMESTONE

The light gray, very finely crystalline limestones acquire, exposed to weathering a nodular appearance due to uneven, wavy yellowish-brown or greenish shaly laminae. Microscopical analysis of thin sections of these limestones, stained according to Dickson (1966) with alizarin red-S and alizarin red-S with potassium ferricyanide, demonstrated that they are calcitic limestones in different stages of recrystallization, while the upper parts of the member are partly dolomitized. This very finely crystalline calcite, microspar according to Folk (1965), may have derived from fossiliferous micrite or biomicrite, according to the terminology of Folk (1959, 1962), or from carbonate mudstone or wackestone according to Dunham (1962), and contains a varying amount of fossils, mainly remnants of Crinoidea, Trilobita, Ostracoda, Tentaculata, Bryozoa, Anthozoa, and Gastropoda. Detrital quartz grains of silt size and some of very fine sand size, and muscovite flakes are scarce (less than 5%).

The most striking feature, however, are the numerous stylolites, which occur isolated, at least 1 micron thick, and 0.1 mm long, or pile up into bands some millimetres thick and some centimetres long, which can split up and come together again, thus forming an interconnecting network (Figs. 11, 13 and 14). These irregularly undulating solution stringers (Schmidt, 1965), are mainly orientated parallel or subparallel to the bedding. In these solution stringers the insoluble residue consisting of clay, detrital quartz silt, muscovite flakes and iron minerals is concentrated. The clay and iron minerals are illite (illite 2M > 1M) and pyrite respectively, according to X-ray analyses by R. O. Felius (pers. comm.). Fossil remnants in the microspar are cut off very sharply by the solution stringers, an indication that these stringers are mainly true solution features and not sedimentary shale laminae. Some solution stringers may have originated from sedimentary shale laminae or detrital quartz grains, since inhomogeneities in the sediment favour pressure solution (Park and Schot, 1968; Füchtbauer et al., 1970).

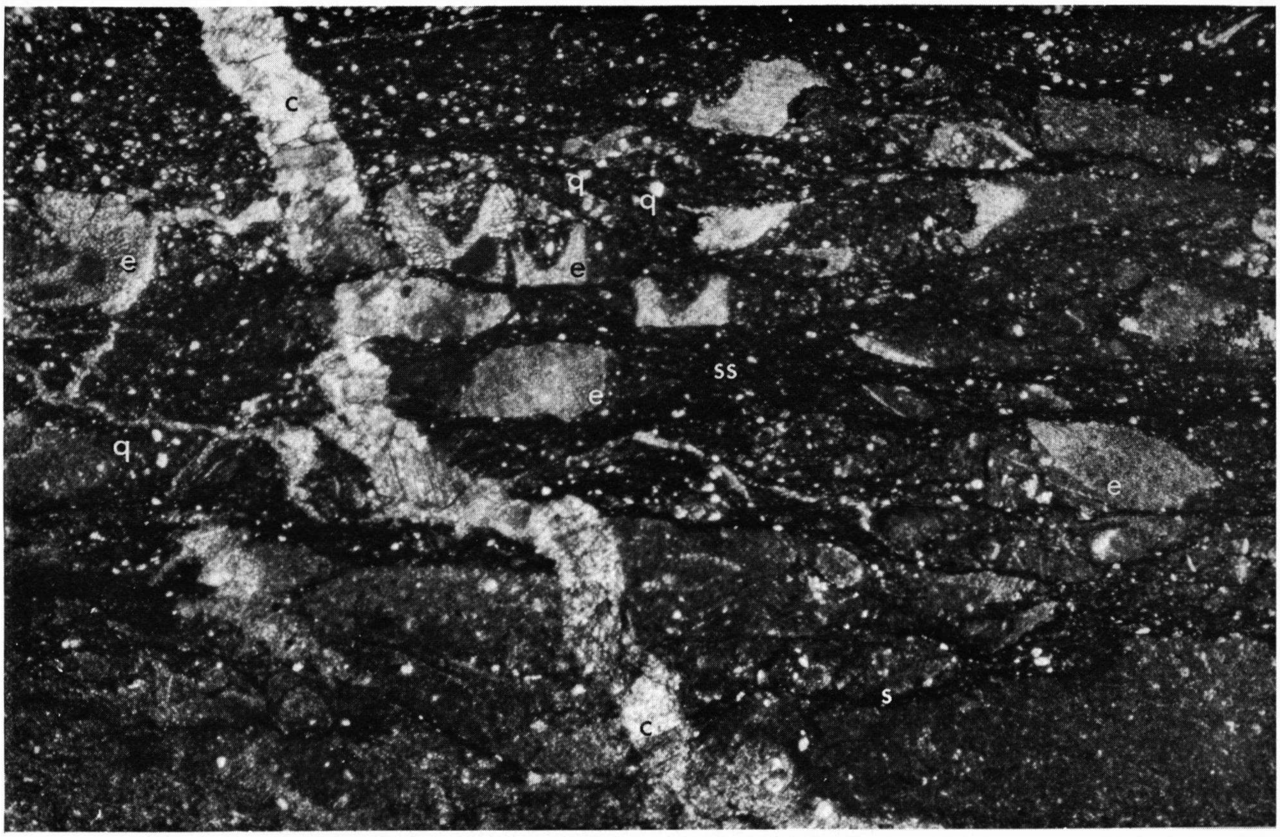
Within these solution stringers fossil remnants and lenses of undissolved calcite have been preserved (Fig. 13). They often have a high dolomite and microcrystalline quartz content and also have a rim of finely crystalline dolomite, which is bordered by neomorphic finely crystalline fibrous calcite, the latter forming the transition zone to the very finely crystalline (0.004–0.016 mm) calcite sediment. Since the stylolites and the solution stringers interrupt and displace calcite filled fractures and also some fossil fragments, it is assumed that these stylolites and solution stringers have mainly been formed by pressure solution during a later diagenetic stage, after lithification of the sediment (Figs. 11, 12 and 13).

Deformation of fossils has not been observed; the fossils consist of neomorphic finely crystalline equant calcite (Folk, 1959, 1965).

The original lime mud, which may have originated from several processes (Cloud, 1962; Chilingar et al.,

Fig. 13. Nodular weathering limestone composed of neomorphic very finely crystalline calcite (microspar). Stylolites (s) combine to form solution stringers (ss), which dislocate calcite vein (c) and cut off fossil remnants (mainly Echinodermata (e) fragments). Quartz silt (q) concentrated in solution stringers. Sample 48 (Section 12), nicols crossed, 20 x.

Fig. 14. Stylolites (s) and solution stringers (ss) in a very finely crystalline limestone. Solution stringers oblique to the bedding have rims of neomorphic bladed and fibrous calcite (f). Some Trilobita (t) and Brachiopoda (br) fragments occur as well as quartz silt (q). Sample 786 (Section 18), nicols crossed, 20 x.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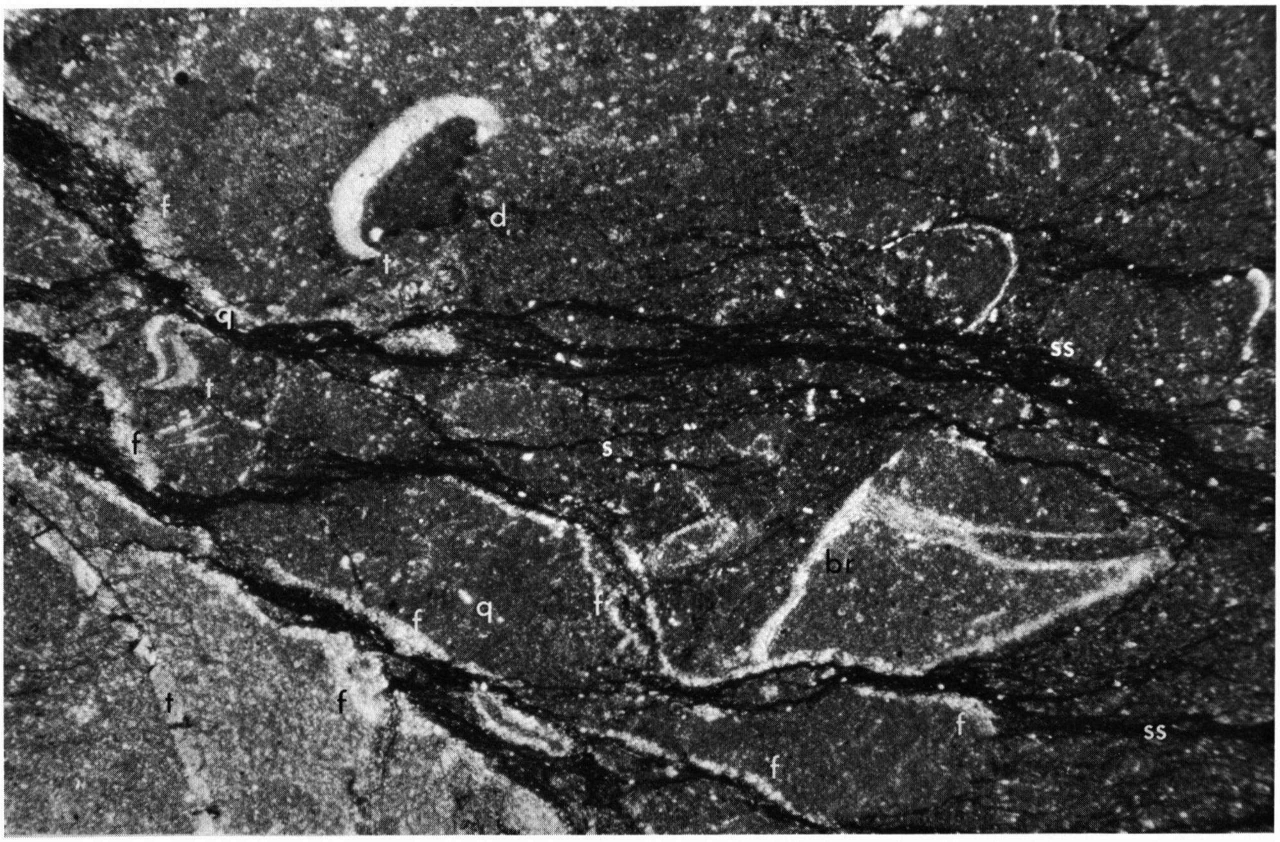
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1967; Bissell and Chilingar, 1967; Sanders and Friedman, 1967) viz. organical, chemical and detrital, accumulated in a shallow marine environment, where some detritic components were introduced, as indicated by the quartz-silt, the muscovite flakes and fragments of shallow marine fossils. After deposition, in the early diagenetic stage, the sediment did not undergo much distortion by organic activity, hardly any burrowing being observed. One spiral-shaped burrow was recognized in a thin section (655).

Later diagenetic processes, such as compaction and cementation, leading to lithification, also included the change of calcareous mud into micrite (Folk, 1965, Chilingar et al., 1967), and authigenesis of pyrite. The pyrite is evenly distributed in the limestone, mainly as idiomorphic crystals, but the pyrites sometimes surround fossil fragments. In the solution stringers concentrations of pyrite occur, indicating accumulation by the late diagenetic pressure solution process of these earlier diagenetic or even synsedimentary pyrite (Krumbein and Garrels, 1952, Chilingar et al., 1967). Recrystallization of the microcrystalline calcite to microspar calcite, called neomorphism by Folk (1965), also occurred, followed by some further local coarsening of the crystal size and the development of neomorphic fibrous sparry calcite which sometimes borders the solution stringers (Fig. 14).

Replacement of calcite by dolomite started from isolated spots. Dolomitization is a later diagenetic process, as is shown by the occurrence of isolated euhedral crystals of very fine crystalline clusters of dolomite in the microspar, and of dolomite in fossil fragments. When the amount of subhedral dolomite crystals increases, an interlocking dolomite mosaic is formed. Euhedral dolomite crystals are mainly brown in colour and sometimes show a brown outer rim. In the upper part of the member the dolomitization is often more intense, and has mainly affected the crystalline ground mass, whereas the fossils, especially those composed of larger calcite crystals, are more resistant to this replacement. Detrital quartz grains often show replacement by calcite; the microcrystalline quartz in the solution stringers could be the result of deposition of this replaced quartz (Park and Schot, 1968, Chilingar et al., 1967). Pyrite crystals show replacement by iron oxides (mainly hematite) according to reflected light analyses of some polished samples of insoluble residues.

The nodular weathering limestones strongly resemble the "Flaserkalke" described by Hollmann (1962), Weber (1965) and Jurgan (1969), which show the same characteristics and must have the same late-diagenetic origin. The nodular weathering limestones of the Basibé Formation were called "griotte" by Mey (1967 a, b, 1968), a term originally used for cherry-red nodular limestones of Upper Devonian age in the French Pyrenees. The term "griotte" is often used in literature for the red-green coloured nodular limestones with varying shale content, or any vividly coloured nodular limestones, occurring for instance in the Pyrenees (Upper Devonian) and Cantabrian Mountains (Cambrian). According to Oele (1964), van der Meer Mohr (1969) and Nagtegaal (1971), the nodular appearance of these limestones mainly has an early-diagenetic origin.

The origin of the nodular texture of nodular limestones has been studied by many workers, and several modes of origin have been suggested. Sedimentary, diagenetic and tectonic causes have been brought forward to explain the nodular character. Reviews of possible modes of origin of nodular limestones are given by Oele (1964), Dooge (1966), van der Meer Mohr (1969), Jurgan (1969), Garrison and Fischer (1969), and van Hoorn (1970). Jurgan (1969) made a genetic differentiation in what is generally termed nodular limestones, which comprises a great many of different types.

DOLOMITE

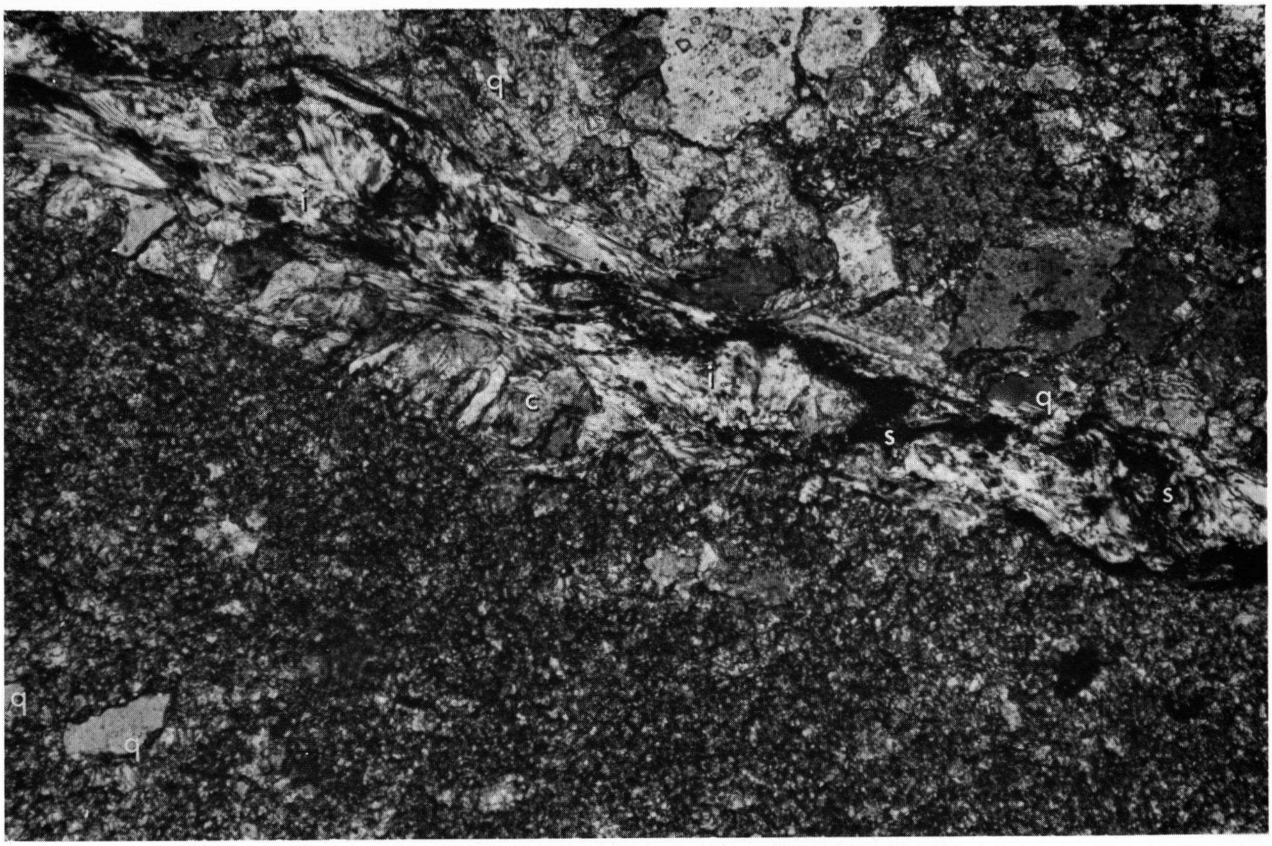
The dolomite occurring above the nodular weathering limestones in the stratigraphic column are coloured medium light gray and have a finely, medium or coarsely crystalline texture. They are pure dolomites, according to staining of thin sections with the methods proposed by Dickson (1966), and generally consist of a mosaic of anhedral and subhedral dolomite crystals. The dolomites are characterized in the field by abundant white quartz veins, mainly perpendicular to the bedding.

In the field the dolomites sometimes show a pattern similar to that of the nodular weathering limestones. In thin sections these structures are very clear: the very finely or finely crystalline dolomites often contain stylolites and solution stringers (Figs. 15 and 16). Fossil remnants, converted into medium crystalline dolomite, are also recognizable in the finely crystalline dolomite.

In some samples parts of the dolomite are recrystallized into a medium crystalline dolomite mosaic, while the remaining parts consist of very fine or fine-grained dolomite. Other samples have completely recrystallized into a mosaic of medium or coarse dolomite crystals. These partly and completely recrystallized dolomites are, in most sections, irregularly distributed over the column.

Fig. 15. Dolomite, very finely crystalline in the lower part, subhedral, finely to medium crystalline in the upper part. These areas of different crystal sizes are separated by a pre-dolomitization stylolite (s), near which authigenic illite (i) occurs. The stylolite and illite are bordered in the very finely crystalline dolomite by a rim of coarser bladed dolomite crystals (c). Quartz grains (q) occur, showing replacement by dolomite. Sample 291 (Section 25), nicols crossed, 200 x.

Fig. 16. Medium crystalline (m) and finely crystalline (f) dolomite, with a gradual transition. In the finely crystalline dolomite stylolites (s) displacing a dolomite vein. In this vein medium to coarsely crystalline dolomite (d) and quartz (q) occur, showing replacement (see also inset). Zoning (z) in dolomite rhomb. Sample 302 (Section 25), nicols crossed, 50 x. Inset 300 x.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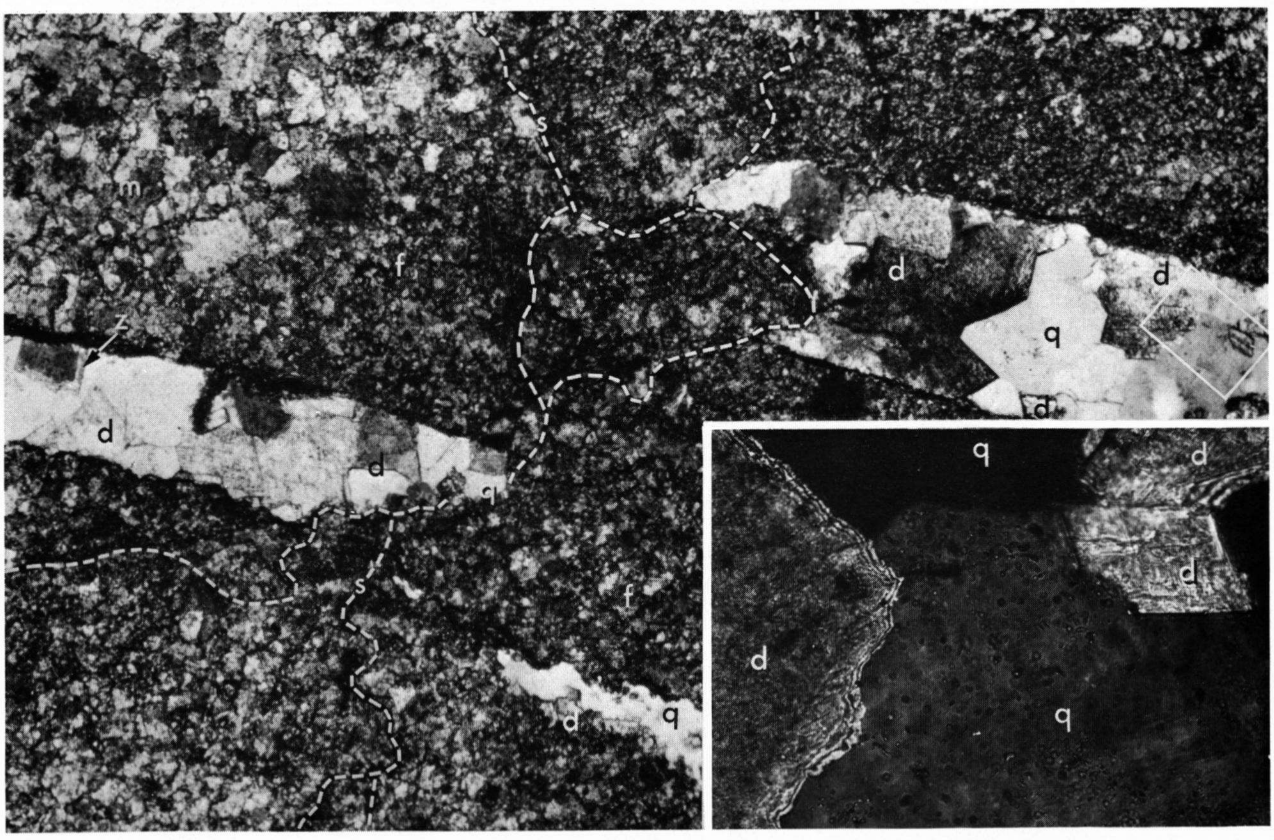
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In the medium crystalline dolomite mosaics, fossil remnants are generally obliterated, although stylolites and solution stringers are still visible. The stylolites and stringers, which mainly contain illite and some ferric oxide, according to X-ray analysis, and authigenic illite, sometimes form the boundaries between the fine-grained and the medium crystalline dolomite mosaics in the partly recrystallized dolomites (Fig. 15). A gradual transition can, however, often be observed (coalescive neomorphism according to Folk, 1965). We assume the recrystallization from fine to medium crystalline dolomite to be a later phase in the dolomitization process, since dolomitization of very finely crystalline calcite is common. Fractures filled with dolomite are dislocated, (Fig. 16), and fossil fragments are cut off, by the stylolites in the finely crystalline dolomite. In the medium crystalline mosaics the smaller dolomite fractures or veins are generally no longer perceptible, in contrast to the veins filled with coarser crystals.

Detrital quartz of silt size, heavy minerals, and muscovite flakes scarcely occur in the finely crystalline dolomite. In the medium crystalline dolomite mosaics silt-sized quartz grains are often wholly or partially replaced by dolomite.

The abundant fractures or veins are filled with lamellar dolomite, medium to coarse subhedral dolomite crystals, brown-coated dolomite crystals (these euhedral or subhedral crystals often show well-developed zoning by hematite, according to X-ray analyses), lamellar, very undulatory, interlocking anhedral quartz, (Fig. 17), and coarse subhedral quartz crystals. Some veins contain both subhedral dolomite and subhedral quartz crystals (Fig. 16).

The lamellar anhedral quartz often shows shearing effects on both sides of the fracture, while the coarse subhedral quartz is generally bordered by bladed, medium to coarse dolomite crystals, thus forming a transition to the finely or medium crystalline dolomite (Fig. 17). Replacement of quartz by dolomite, and/or dolomite by quartz have been observed.

From the intersecting patterns of different fractures or veins filled with dolomite and quartz, and of the stylolites or solution stringers, a distinction into several fracture generations, can be made, using dislocations, differences in mineralogy, texture and replacement features. The dolomite-filled veins are always dislocated by, and thus must have preceded, the stylolites and the quartz-filled fractures.

In some isolated spots, microcrystalline anhedral quartz (chert) occurs, which usually seems to be related to stylolite remnants. Subhedral (ferroan) calcite sometimes surrounds these chert patches. The main part of the quartz, however, postdates the stylolitization and also the dolomitization. The latter is supposed to be a rather late diagenetic process. The date of quartz precipitation is indicated by the fact that quartz-filled veins are crossed by stylolites and dolomite veins without dislocations. That the quartz precipitation occurred after dolomitization is also shown by small euhedral dolomite crystals enclosed within these quartz crystals (Carozzi, 1960; Walker, 1962; Dapples, 1967). Solution of detrital quartz grains in the overlying and interfingering quartz sandstones is the most probable source of this quartz precipitation in the joints and faults of the dolomites, and/or in the pore spaces, created by the dolomitization. A similar secondary precipitation of silica has been described by, for instance, Krauskopf (1959), Siever (1959, 1962), Dapples (1959, 1967) and Chilingar et al. (1967).

In the lowermost part of the dolomite member, calcite veins occur locally within the dolomite. The (ferroan) calcite (Dickson, 1966; Evamy, 1969) contains anhedral and euhedral quartz and euhedral dolomite crystals, the latter often showing an iron oxide zoning in the core of the dolomite rhombs. The following replacement sequence can be inferred: (ferroan) calcite is replaced by dolomite, while quartz replaces both calcite and dolomite.

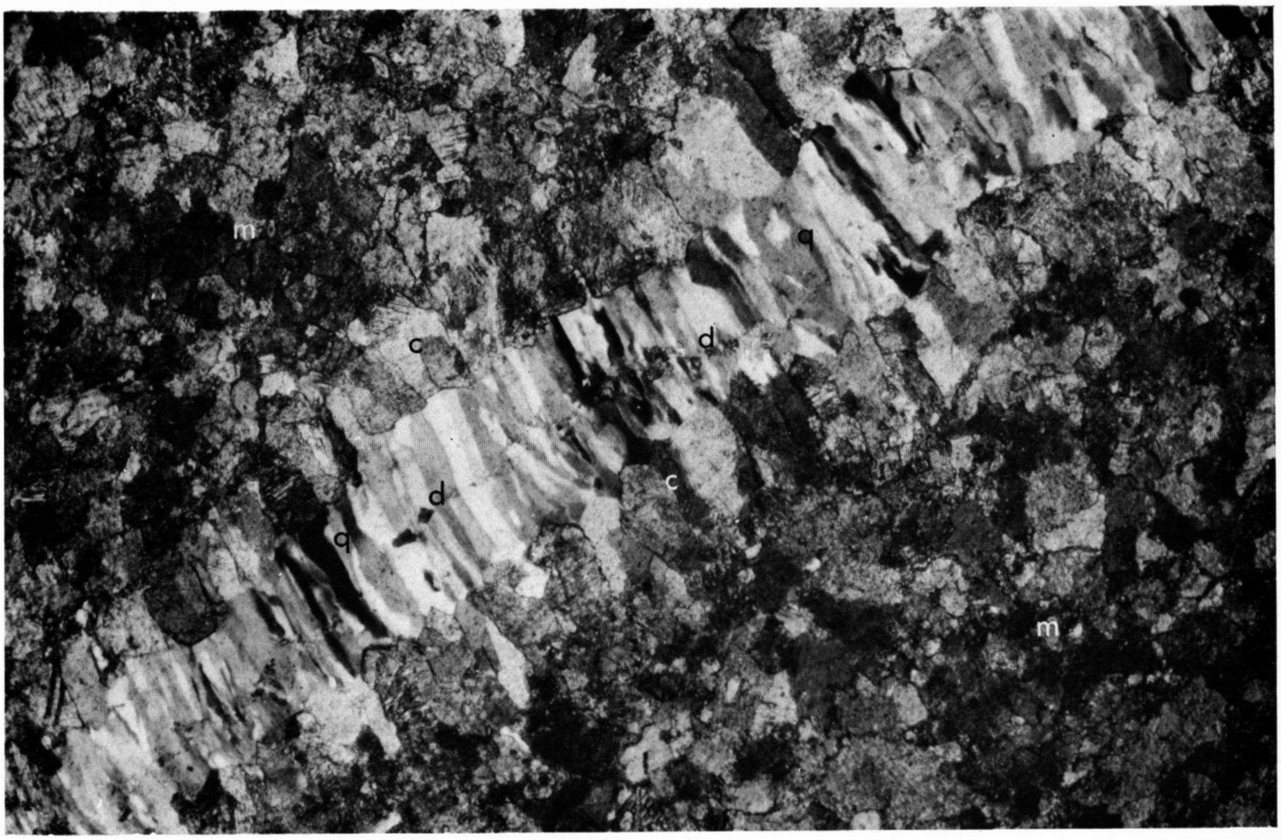
The dolomites of the Basibé Formation apparently came about by the replacement of calcium carbonate during a rather late diagenetic stage. Outlines of original calcitic fossil fragments, now composed of medium to coarse crystalline dolomite, and the preserved textures of finely crystalline dolomite with fossil relics and solution stringers clearly indicate the diagenetic origin of the dolomite, which must have replaced the nodular weathering limestone. The occurrence of insoluble residual stylolite material within single dolomite crystals also indicates that the dolomitization postdated the formation of most stylolites and solution stringers, hence a rather late diagenetic stage (postlithification) is apparent. This is also shown by the patchy distribution of the finely and medium crystalline dolomite in some beds. The dolomite can be classified as diagenetic dolomite according to Friedman and Sanders (1967) or as anadiagenetic dolomite according to Fairbridge (1967).

Dolomitization only occurs where the dolomitized limestones are associated with quartz sandstone layers, so that a genetic relationship is assumed.

The movement in a lateral and/or vertical direction through the limestone of water with a high concentration of Mg^{2+} is a process by which the replacement of the original calcium carbonate by dolomite may have been achieved. A downward migration of saline water from a shallow marine environment or a lagoon, causing dolomitization of limestones, was called seepage reflux by Adams and Rhodes (1960). Similar explanations, based on this theory were given by Deff-

Fig. 17. Lamellar quartz filling (q) vein in medium crystalline dolomite (m). Vein is bordered by coarser bladed dolomite crystals (c). In the centre of the quartz vein, little dolomite (d) is present. Sample 283 (Section 25), nicols crossed, 50 x.

Fig. 18. Pellet relics (p) in a dolomitized limestone of the upper member, in which also some pre-dolomitization stylolites are preserved (dashed lines). Sample 760 (Section 18), nicols crossed, 80 x.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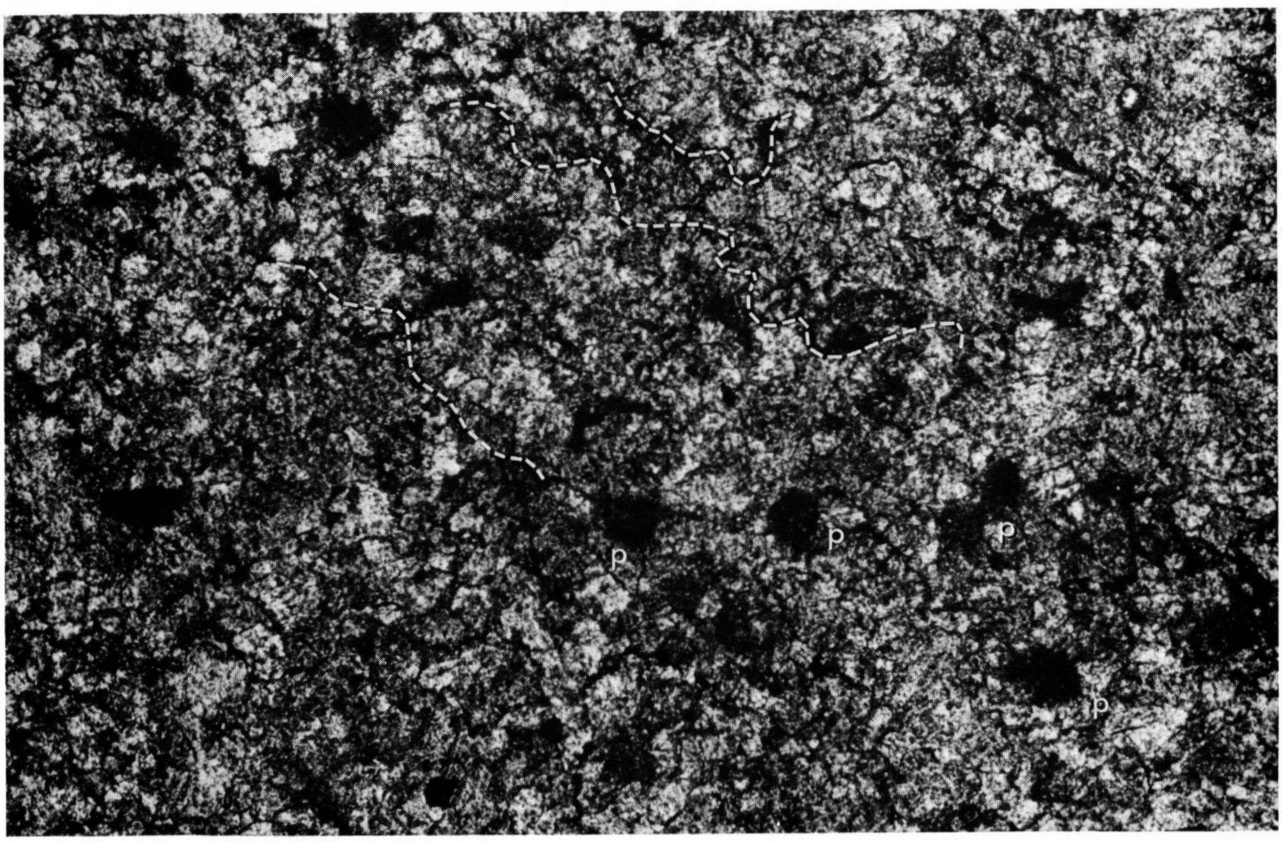
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eyes et al. (1965) and Murray (1969) for the dolomites occurring on the island of Bonaire. Hsü and Siegenthaler (1969) concluded after experiments that only a hydrodynamic movement opposite to the seepage reflux and induced by loss of interstitial water due to evaporation could account for this dolomitization process.

These mechanisms could have been active in limestones, overlain by, and/or interfingering with, highly permeable quartz sandstones, as was the case in the Basibé Formation. Concentration of Mg^{2+} by evaporation could have occurred in very shallow marine environments, "lagoons", limited by the quartz sandstone beach barriers or bars.

The present author was initially inclined, on the basis of his field work, to accept this explanation for the dolomitization in the Basibé Formation until microscopic examination of the samples provided evidence, set forth above, in favour of a rather late dolomitization, which must have taken place after lithification and stylolitization of the limestones.

Two possibilities can be considered:

a. Sea water may have ascended in the quartz sand barriers (the present quartzites) due to surface evaporation (reversed reflux mechanism or evaporative pumping or capillary concentration), resulting in supersaline connate water in the sand barriers. This connate water, with high concentrations of Mg^{2+}, was driven out of the quartz sandstone bodies by cementation during a deep burial stage. Expulsion of this connate water was mainly initiated by silicification of the quartz sandstone, which also created the driving forces for this movement.

Some incipient dolomitization may have already taken place in the limestones at the bottom of the lagoon, but any traces of it will have been obliterated by the secondary dolomitization.

b. Another possibility, though less probable, is that saline water collected in the sandstones bodies by compaction of underlying sediments.

Upward mixing of pressed-out connate waters through alternations of carbonate, argillaceous and quartz sediments, with different ion-exchange properties and permeabilities, may lead to curious anomalies in the pressure and (increasing) salinity (Bredehoeft et al., 1963; Siever et al., 1965; Fairbridge, 1967; Usdowski, 1967, 1968). Trapped connate water can also be liberated during fracturing. Fracturing is favoured by dolomitization, and these fractures will be added to the joints brought about during compaction. During or after this anadiagenetic stage, silica solutions must have descended from the quartz sandstones and precipitated quartz into the dolomites.

LIMESTONE

The medium gray, thinly bedded limestones which form the upper member of the Basibé Formation are composed of very finely or sometimes finely crystalline calcite. Staining of the thin sections by the methods proposed by Dickson (1966) showed the material to consist mainly of calcite, with only very few fine to medium-sized anhedral dolomite grains. More dolomitic limestones and dolomites occur locally and will be dealt with below. According to Folk's classification (1959, 1962), the limestones are micrites or biomicrites, or, following Dunham's nomenclature (1962), mud – or wackestones (Fig. 19). Microscopical analyses show that they consist of micrite and microspar (Folk, 1965).

Some fossil fragments are enclosed in the very fine crystalline calcite, although beds with abundant fossil debris also occur (Fig. 20). The fragments are parts of Crinoidea, Trilobita, Echinodermata, Cricoconarida, Ostracoda, Foraminifera, Bryozoa and Tabulata. These fragments are all replaced by neomorphic finely to medium crystalline equant calcite. Spiral-shaped burrows, (consisting of brownish coloured dolomite and 0.05 mm in thickness), were recognized in thin sections (265, 261, and 253), while in sample 260 a fossil remnant was found at the end of a burrow (Fig. 19).

Detrital quartz, of silt size, is very scarce in these limestones, and always shows replacement by calcite. Muscovite flakes sometimes occur. In the sections east of the Ribagorzana River, quartz grains are sometimes more abundant in the Limestone Member, especially near intercalated quartzites, as in Sections 31 and 36.

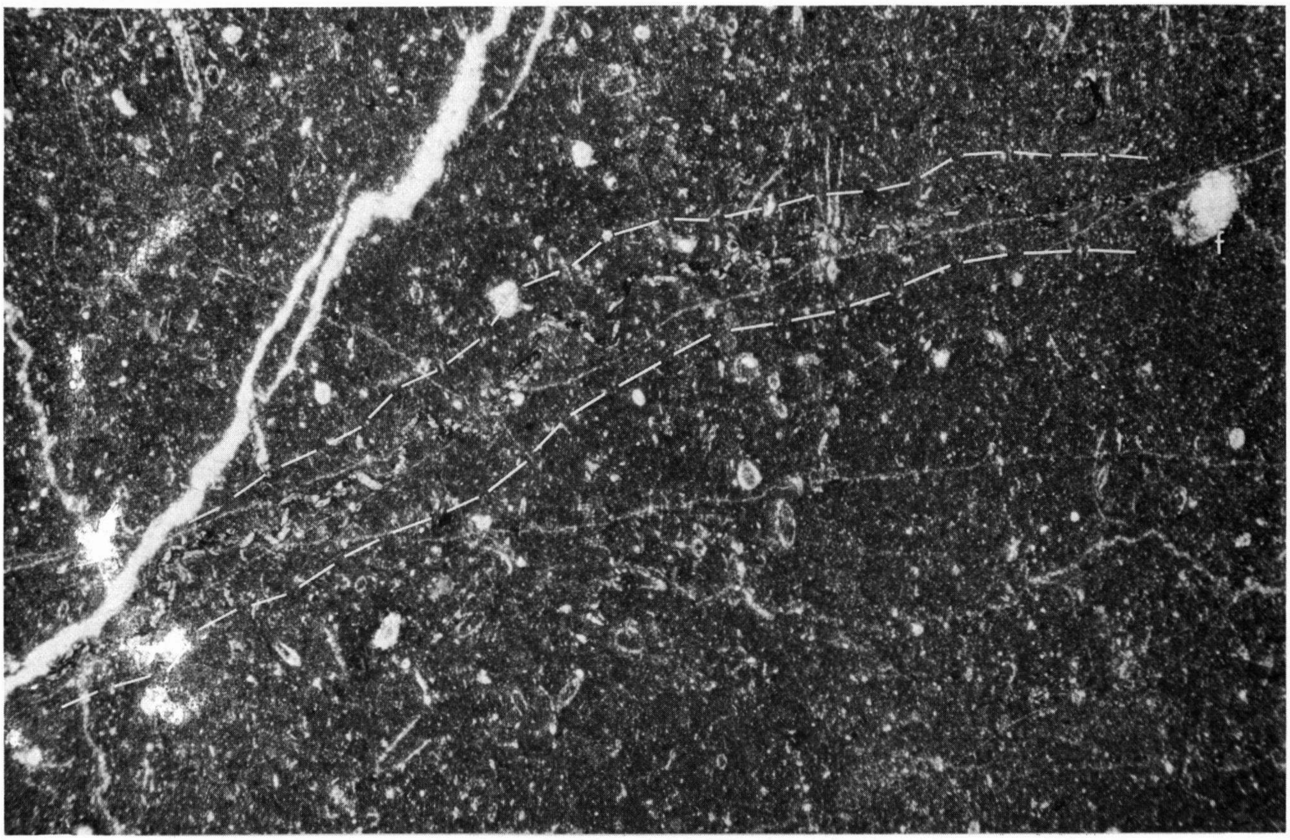
Small euhedral pyrite crystals are frequent, usually associated with fossil remnants.

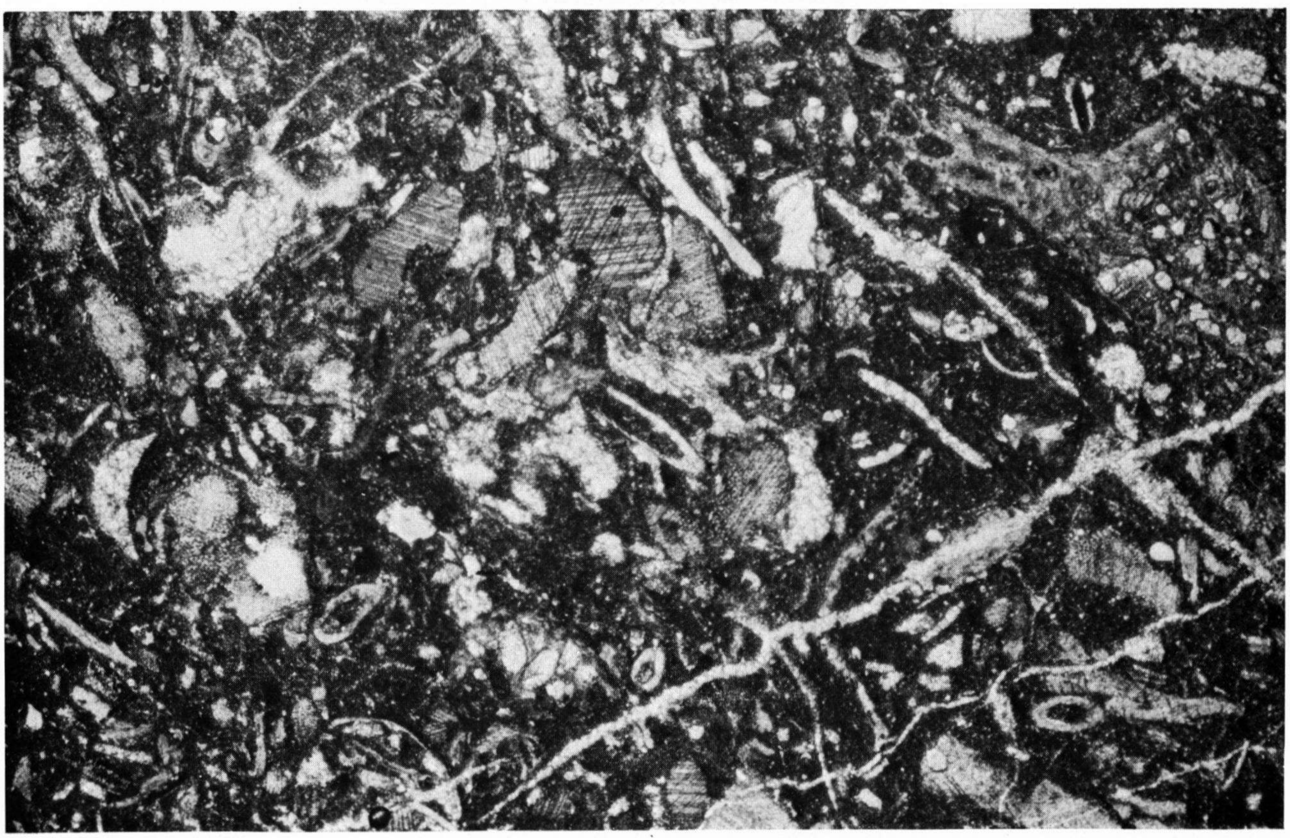
In more dolomitic limestones, the crystal size of the calcite tends to increase. Crystal mosaics of sparry calcite are found, while with the dolomite content the microcrystalline silica content increases as well. Subhedral dolomite crystals sometimes occur as fossil fills. The lowermost part of the Limestone Member is generally dolomitized (see Sections of Plate I). Thin sections show a mosaic of anhedral and subhedral, finely and medium crystalline dolomite. Near the transition into the Fonchanina Formation, the limestones are also often more dolomitic.

Dolomite beds, intercalated in the limestones (e.g. Plate I, Section 25, approx. 68 m and 78 m, Section 18, approx. 56 m, and Fig. 4, Section 12), could already be recognized in the field by the different colour and the quartz veins. These quartz veins are restricted to

Fig. 19. Very finely crystalline limestone of the upper member, with some fossil fragments, mainly Cricoconarida. The lime mud with this fossil content is characteristic for an open marine low energy environment of deposition. Between dashed lines a spiral-shaped burrow, ending at the resting place of the fossil (f). Sample 260 (Section 25), nicols crossed, 20 x.

Fig. 20. Abundant fossil fragments (mainly Echinodermata, Bryozoa, Cricoconarida, Trilobita and Crinoidea), deposited in a lime mud environment, indicate higher energy conditions. Sample 255 (Section 25), nicols crossed, 20 x.





the anhedral and subhedral, medium to coarse crystalline dolomites, and do not occur in the limestones. Most dolomitic limestones show isolated dolomitic spots, whereas the fossil fragments must generally have been dolomitized at a final stage, since some still consist entirely or partially of calcite.

Stylolites are common in the limestones, although their amount and sizes vary. Wave-like seams, up to about 0.2 mm thick, of aggregated stylolites are most frequent while sutured micro-stylolites (Park and Schot, 1968), about 0.005 mm thick, with relatively thick insoluble seams in the crests and valleys also occur in these limestones. In thick-bedded limestones, larger, sutured type stylolites appear with thicknesses of some millimetres and amplitudes up to 2 cm. Most stylolites are parallel to the bedding and sometimes dislocate vertical or subvertical calcite veins and cut off fossils. The stylolites must have originated at a later diagenetic stage as a result of pressure solution; they contain the insoluble residue of the limestones, mainly clay (illite according to X-ray analyses). In the upper part of the member the amount and sizes of the stylolites increase. With increasing amounts of clay, pyrite and micro-crystalline quartz also appear in the stylolites, which locally show some fibrous calcite rims. Subvertical stylolites, due to tectonism according to Park and Schot (1968), also occur.

Calcite veins mainly consist of medium crystalline calcite, sometimes of ferroan calcite.

In general, original carbonate muds may have derived from various processes, whether biological, biochemical, physicochemical or detrital, according to Folk (1965), Chilingar et al. (1967), and Sanders and Friedman (1967). During diagenesis it was converted by neomorphism into very finely to finely crystalline calcite (Folk, 1965). This process can also be assumed for the Limestone Member. Early diagenetic fractures were filled with calcite, and at a later diagenetic stage were dislocated by stylolitization due to pressure solution. At a later diagenetic stage, ferroan calcite precipitated in some fractures, while dolomitization occurred in some isolated levels.

The preferential dolomitization of the isolated beds in Sections 12, 18 and 25 may be due to primary textures and structures of the sediment. Thin sections of samples from these layers contain relics of grains, 0.10–0.15 mm large, composed of small crystals (0.002–0.004 mm), within the anhedral and subhedral, medium to coarse crystalline dolomite (Fig. 18). This indicates that the original sediment was a carbonate grainstone or calcarenite, interbedded with lime mudstone, and that preferential dolomitization of single beds was favoured by their greater permeability.

According to the seepage-reflux model, the magnesium-enriched, refluxing brines migrated through the initial permeable carbonate sands, causing exclusive dolomitization of these layers, while the surrounding lime mud was not affected. A similar feature was described by Fisher and Rodda (1969). However, these findings more or less contradict those reported by Murray (1960) and Murray and Lucia (1967), who stated that lime muds are preferentially dolomitized, rather than the more permeable carbonate sands. They believe that the permeability of the sediments was reduced by predolomite cementation and compaction, and that cementation has been more rapid in the carbonate sands than in the lime muds. Greater reactivity to dolomitizing water of fine grained sediment may also be a reason.

Taylor and Illing (1969) and Shinn (1969) described alternations of cemented and uncemented carbonate sediments in the Persian Gulf. Cementation is mainly controlled by the rate of sedimentation and the grain size (Shinn, 1969). Subtidal cementation in sediments only takes place where the rate of sedimentation is sufficiently low to allow the cement to form.

A very low sedimentation rate is necessary for fine sediments to become cemented. Taylor and Illing (1969), Shinn (1969) and also De Groot (1969), point to the high-magnesium calcite in subtidal cemented limestones and the high Mg/Ca ratio of the interstitial water. The presence of high-magnesium calcite may also be a condition favouring later dolomitization.

The limestones of the Basibé Formation north of the Alpine thrust and fault zone are metamorphosed near the contact with the intrusive Maladeta granodiorite.

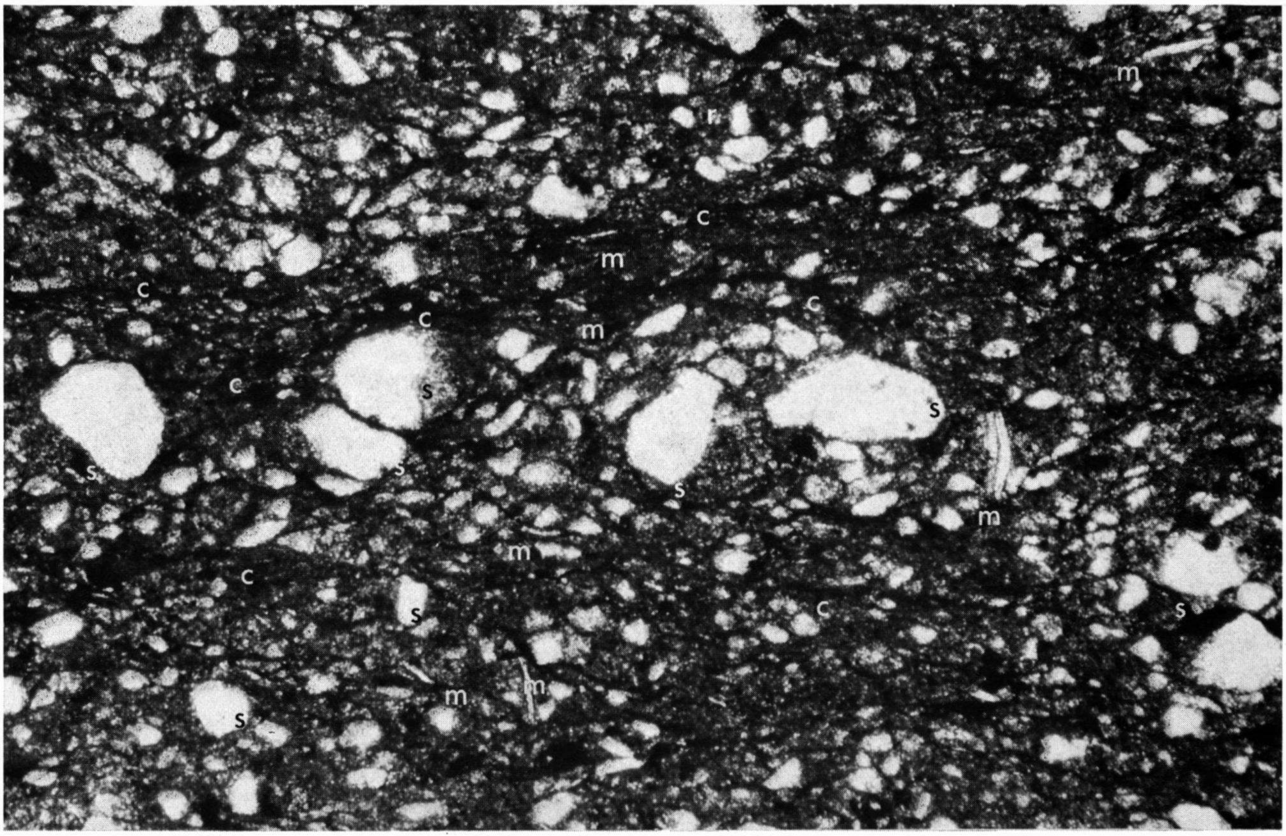
The limestones often possess a banded appearance, caused by a fine alternation of limestones and limestones with some pelitic material. Detrital quartz of silt size and some very fine sand occur in minor quantities.

Calcite is the main component, but dolomite also occurs, mainly as porphyroblasts. Relics of fossils, mainly crinoids, can be recognized, although generally heavily deformed. Stylolitization is common in these rocks, but stylolite seams are nearly straight. Residues in these seams are clay and pyrite.

Most samples of Section 30 show metamorphic characteristics. Calcite is recrystallized into a crystalloblastic mosaic, while idioblastic scapolite (according to X-analyses by R. O. Felius, pers. comm.) and mica have been formed. The idioblastic scapolite crystals show poikiloblastic textures, cleavage and often rotation. Lens-shaped microcrystalline quartz fillings occur in the more impure limestones or slates, which are metamorphosed into phyllites.

Fig. 21. Finely crystalline dolomite with quartz silt, quartz sand (s) and muscovite flakes (m). Clay (c) and iron oxides mainly are concentrated in stylolites. Rutile grain (r). Sample 608 (Section 9), plain light, 80 x.

Fig. 22. Silty to sandy argillaceous dolomite, with hematite (h), partly replaced by quartz (q). Hematite pseudomorph after pyrite (?). Most quartz grains show replacement by dolomite. Muscovite flake (m). Sample 305 (Section 25), nicols crossed, 200 x.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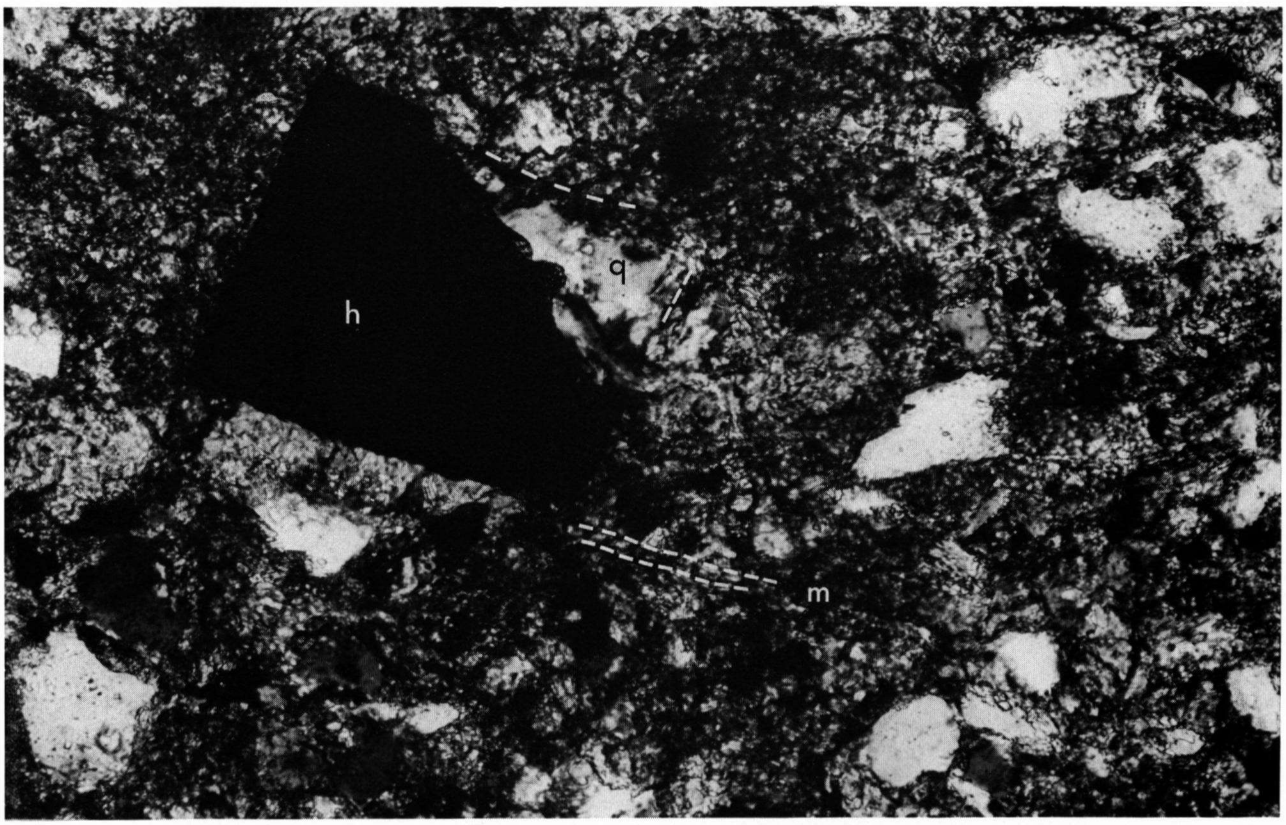
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SILTY TO SANDY ARGILLACEOUS LIMESTONE AND DOLOMITE

The silty to sandy argillaceous limestones are intercalated between the nodular weathering limestones of the Nodular Limestone Member, and have been transformed into silty to sandy argillaceous dolomites in the Quartzite-Dolomite Member. In spite of the dolomitization, the structure and texture of the original limestone have been preserved. Locally the silty to sandy argillaceous limestone within the Nodular Limestone Member is also dolomitized, usually in preference to the nodular limestone.

The dolomitic type is the most common and consists of very finely crystalline, or sometimes medium crystalline dolomite, often of a brownish colour in the thin sections. Some fossil relics can be recognized and generally appear as clusters of medium crystalline dolomite. Fossil remnants in the silty to sandy argillaceous dolomite within the Nodular Limestone Member are often still calcitic, according to staining with alizarin red-S.

More striking than the fossil content, however, is the amount of siliciclastic detrital material. Quartz grains of fine silt size up to fine sand size occur in varying quantities of up to about 40%, while locally very few quartz grains of medium sand size are present (Fig. 21). The quartz grains of silt size are mainly angular to subangular, whereas the quartz sand grains are better rounded. The sorting of the quartz grains is fairly poor.

Muscovite flakes are present, with a maximum lenght of 0.2 mm. The orientation of the muscovite flakes tends to be subparallel to the bedding. Heavy minerals, usually tourmaline (greenish, bluish and brownish varieties), rutile (yellowish and reddish brown varieties) and zircon (colourless and purple varieties), of silt to fine sand size and rather rounded, are fairly common.

Dispersed clay (illite according to X-ray analyses, 2 M > 1 M, R. O. Felius, pers. comm.) is always present in the sediment and is sometimes concentrated in stylolitic seams. The clay-filled solution seams, from 0.001–0.01 mm thick and some tens of millimetres long, often enclose medium crystalline dolomite clusters, representing dolomitized fossil relics and quartz grains. Within semiparallel aggregates of these seams, the longer-shaped quartz grains and the muscovite flakes usually show an orientation parallel to these stylolite sets. Very finely crystalline quartz sometimes occurs near the seams, and illite-muscovite within the seams. Idiomorphic hematite (identified by X-ray analyses) is also often present in and near the seams.

Small, very finely crystalline and finely crystalline subhedral hematite is evenly distributed in the sediment, while some large euhedral crystals have also been met with. In the slides the euhedral, up to 0.3 mm large, hematite crystals mainly appear as rhombs, originating from authigenic growth as is shown by zoning, (some from six-sided crystals). These hematite rhombs often show replacement by quartz (Fig. 22). Outlines of partly replaced hematite rhombs can be recognized by dust rings. The hematite rhombs perhaps are pseudomorph after original pyrite crystals.

Replacement of quartz by carbonate is common, while tourmaline is also affected. No veins, either of dolomite or quartz, are present in these sediments. In parts not influenced by stylolitization, the distribution of the quartz grains, especially the larger ones, indicates bioturbation.

The original sediment would appear to have been a lime mud or fossiliferous lime mud with a high quartz and a low clay content. Bioturbation has greatly influenced the texture of the sediment after deposition. Recrystallization in the sense of neomorphism transformed the mud into micrite and micro-spar, and some stylolitization occurred. Dolomitization appears to have taken place simultaneously in the lower two members. The silty to sandy argillaceous limestones in the lowermost member are dolomitized preferentially to the limestones, as mentioned above.

Samples from the uppermost part of the Gelada Formation, near the transition into the Basibé Formation, often show the same composition as the silty to sandy argillaceous limestones. The samples, however, are mainly calcitic, and often show a nodular structure in thin sections, due to stylolitization. Samples from lower stratigraphic levels contain more quartz.

QUARTZITE

The quartzites of the middle member of the Basibé Formation are light gray to medium light gray and are pure quartz sandstones according to their microscopical aspect. Since these sandstones are silica-cemented, usually by secondary overgrowth of the quartz grains, and the rock breaks across the grains, the term quartzites may be applied in accordance with the descriptions of Pettijohn (1957), Carozzi (1960) and Füchtbauer et al. (1970). Pettijohn (1957) preferentially uses the term orthoquartzite.

The diameter of the quartz grains varies from 0.03 to 1.0 mm, i.e. from medium silt size up to coarse sand size, according to Wentworth's scale (in Pettijohn, 1957). The medium grain diameters of particular samples range from medium to fine sand size. The maximum quartz grain diameters of some stratigraphic sections, measured directly in thin sections, are given in Fig. 23, while values of the quartz grain sizes also appear in Plate I.

Roundness of the quartz grains has been estimated in thin sections from comparison charts (Müller, 1964). Silt and fine sands are mainly subangular, medium and coarse sands subrounded to wellrounded. Most grains are equidimensional, a minor proportion are elongated in thin sections. Elongated quartz grains are orientated parallel or subparallel to the bedding. Sorting within single samples is fairly good.

Quartz grains are the predominant detrital component in these quartz sandstones. Their quantity varies from about 70% to more than 95% of the present total rock volume. The only other detrital grains are some heavy minerals, while the sand is

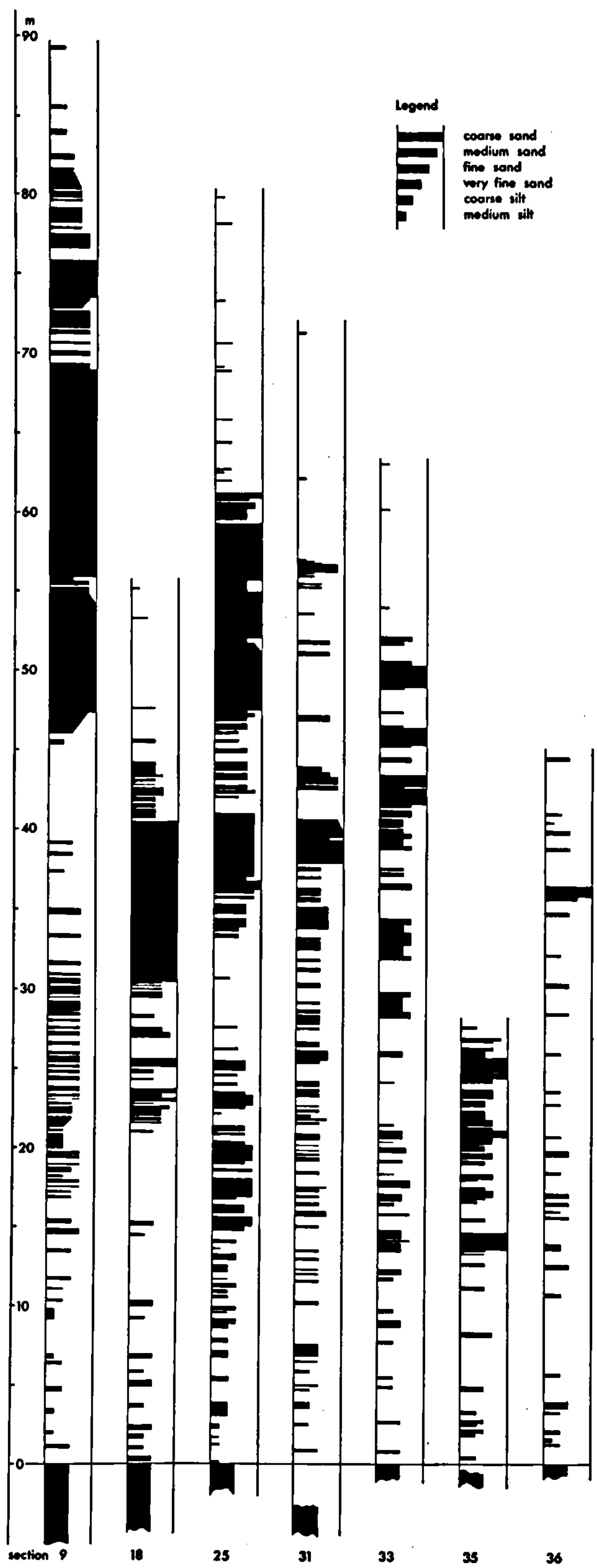

Fig. 23. Maximum sizes of quartz grains in the sediments of the sections given on Plate I. Grain sizes are given in accordance to Wentworth's scale (see also legend of Plate I).

nearly free of clay and of carbonate grains. Feldspar and rock fragments are totally absent.

The quartz grains are almost completely monocrystalline; only very few polycrystalline grains are present. Nearly all quartz grains are undulatory and show secondary quartz overgrowth. Pressure solution and replacement by carbonate have affected the quartz grains in varying degrees.

Secondary quartz and dolomite are present as cement between the quartz grains. Dolomite mainly occurs as replacement mineral of quartz. In some sections east of the Rio Noguera de Tor, however, the carbonate cement of some quartz sandstones is calcite.

Heavy minerals are common in the quartzites and have very fine sand to medium sand sizes, while the main grain size is fine sand size. Most frequent are tourmaline, rutile and zircon. The tourmaline is generally confined to green, yellow and blue varieties, the rutile to reddish-brown varieties and the zircon to colourless variety. Tourmaline predominates and is often present as fine sand to medium sand-sized grains, well rounded and nearly always showing cracks. Many quartz grains contain abundant inclusions composed of tourmaline, rutile needles and extremely fine dustlike material. Clay is only found in thin films on some detrital quartz grains and also on sutured contacts of quartz grains in highly pressolved zones of the orthoquartzites. Only very little authigenic, idiomorphic (six-sided) hematite is present in some quartzites. Pyrite concretions, met with in some quartzite samples of Section 31, may have a relation with the dyke swarm of the metamorphic area around Bono and Artiga.

Some recrystallized dolomite represents relics of fossils, which are, however, very scarce.

In general, two types of quartzites were distinguished:

a. quartzites which consist solely of quartz grains, tightly packed or interlocked with varying degrees of pressure solution. Cement is nearly absent or scarce and consists of secondary quartz, while minor quantities of dolomite are present (Fig. 24).
b. quartzites in which the main part of the cement consists of dolomite, with some secondary quartz (Fig. 25).

Transitions between these two types occur widely. The term orthoquartzites is used in Plate I for the quartz sandstones of type a, while type b is mentioned as dolomitic orthoquartzites, in accordance with Pettijohn (1957).

As will be shown below, the present great amount of dolomite in type b is mainly due to replacement by this mineral of original detrital quartz grains and their secondary quartz cement. Some sedimentary carbonate may have been present in the original pore spaces, although no proof of this is available.

Replacement always started from the original pore centre and advanced preferentially along the grain contacts (Fig. 25). Small euhedral dolomite rhombs are often observed on these contacts, in a later stage followed by larger subhedral or anhedral dolomite patches. Larger dolomite rhombs and patches often contain brown-coloured iron mineral matter. This iron mineral matter sometimes is the cause of excellent zoning in the dolomite crystals. Replacement of the quartz by dolomite can become so important that only small remnants of quartz are left. The original grain is sometimes recognizable by a dust ring or ghost structure. Type a orthoquartzites have been transformed by this replacement process into type b dolomitic orthoquartzites.

The type of contact between the quartz grains varies greatly, as mentioned above. Grains with only some point contacts occur as well as all other types of interpenetration features: tangential, concavo-convex and sutured contacts (Fig. 27). The latter three contact features are most frequent in all samples. In the remaining pore spaces, secondary quartz is nearly always precipitated. It also occurs as an overgrowth in optical continuity with most detrital quartz grains. Dust rings, indicating the original grain outlines, can often be recognized (Figs. 24 and 25).

In the very scarce dolomite veins, brown-coloured euhedral or subhedral dolomite rhombs occur, which often show striking zoning. Euhedral quartz crystals are often present within these dolomite crystals. This authigenic quartz replaces the dolomite as is shown by inclusions of dark iron oxide matter derived from the dolomite and the interruption of the zoning of dolomite rhombs, while this zoning is visible within the quartz in the form of inclusions.

Quartz veins consist mainly of anhedral to subhedral quartz crystals, often with a lamellar orientated interlocking texture.

The original sediment, consisting of fairly well rounded detrital, mainly medium and fine sand sized quartz grains were deposited in an environment, in which the conditions for rounding and sorting were excellent. Only in the high-energetic littoral environment do such conditions prevail (see Chapter VI). Good sorting is accomplished by the waves breaking on the beach, which winnow out all fine-sized grains; repeated abrasion results in some rounding of the sand grains. A more complete rounding may, however, be the result of recycling of the sands (Kuenen, 1959; Pettijohn et al., 1965). In the quartzites only the most resistant detrital minerals are present, viz.: quartz among the light minerals, and tourmaline, rutile and zircon among the heavy minerals. Clay, muscovite flakes and ferric oxides, frequently present in the other lithological units of the Basibé Formation, are practically absent in these quartz sandstones.

A specific source of the predominantly monocrystalline undulatory quartz grains cannot be given, and the few polycrystalline quartz grains seem to be even less indicative of a specific source rock (Blatt and Christie, 1963; Blatt, 1967; Sippel, 1968). The amount of polycrystalline and undulatory quartz present in a quartz sandstone is a function of the grain size according to Conolly (1965); this is also true in our case. The high amount of specific mineral inclusions in many quartz grains is possibly a significant tool for the interpretation of the provenance (see Pettijohn, 1957). Samples or data from older rocks are, however, not available for establishing such a relationship.

Although the quartz grains or the heavy minerals do not give any information on a particular source bed, the mineral association and the grain shapes do in a general way give evidence on the former history of these sediments. The fact that feldspar, rock fragments, and even mica are absent, and that only quartz has been left, indicates that the source material must have undergone prolonged weathering during one, or rather more than one, cycle. The same appears from the heavy minerals, of which only the three least weatherable, and therefore ubiquitous, species have been preserved. This, too, indicates repeated cycles of weathering and redeposition. The same conclusion was drawn above from the high roundness values of the mature quartz sand and the well-rounded heavy minerals of the sediment (Krynine, 1946; Pettijohn, 1957; Hubert, 1962).

The source rock must therefore have once, or more than once, formed an extensive subaerial lowland, where this weathering could proceed. It must have been a lowland without vertical erosion of the rivers, because such erosion would have been supplied new unweathered minerals and mixed these with the weathered material. Once the material was in the marine environment, no mechanism was available for removing feldspar or other light minerals from the quartz sand, or for separating the three heavy mineral species from the others.

In the stratigraphic column, possible source rocks are the Cambro-Ordovician quartzites and the quartzites at the Ordovician to Silurian transition (see Chapter II and Mey, 1967b, 1968; Mirouse, 1962; Hartevelt, 1970). The latter quartzite shows a considerable thickness in the western Pyrenees, up to 100 metres according to Mirouse (1962), while the thickness in the eastern Pyrenees is up to 20 metres (Mey, 1967b, 1968; Hartevelt, 1970). Outcrops of these rocks in the present Axial Zone during the Devonian are less probable. It is unknown whether land masses existed, during Emsian times, north and south of the present Axial Zone, where these rocks could have outcropped. Mirouse (1962) stated that an emergent

Fig. 24. Quartzite; moderately pressolved quartz grains with secondary quartz overgrowth (c). Dust rings (r) show the outlines of the original grains. Some small dolomite rhombs (d) are present. Sample 705 (Section 1), nicols crossed, 200 x.

Fig. 25. Dolomitic quartzite; dolomite (d) has replaced much of the authigenic quartz overgrowth and cement and also parts of the detrital quartz grains. The dustring (r) of a quartz grain is preserved in the replacement dolomite. Sample 307 (Section 25), nicols crossed, 200 x.


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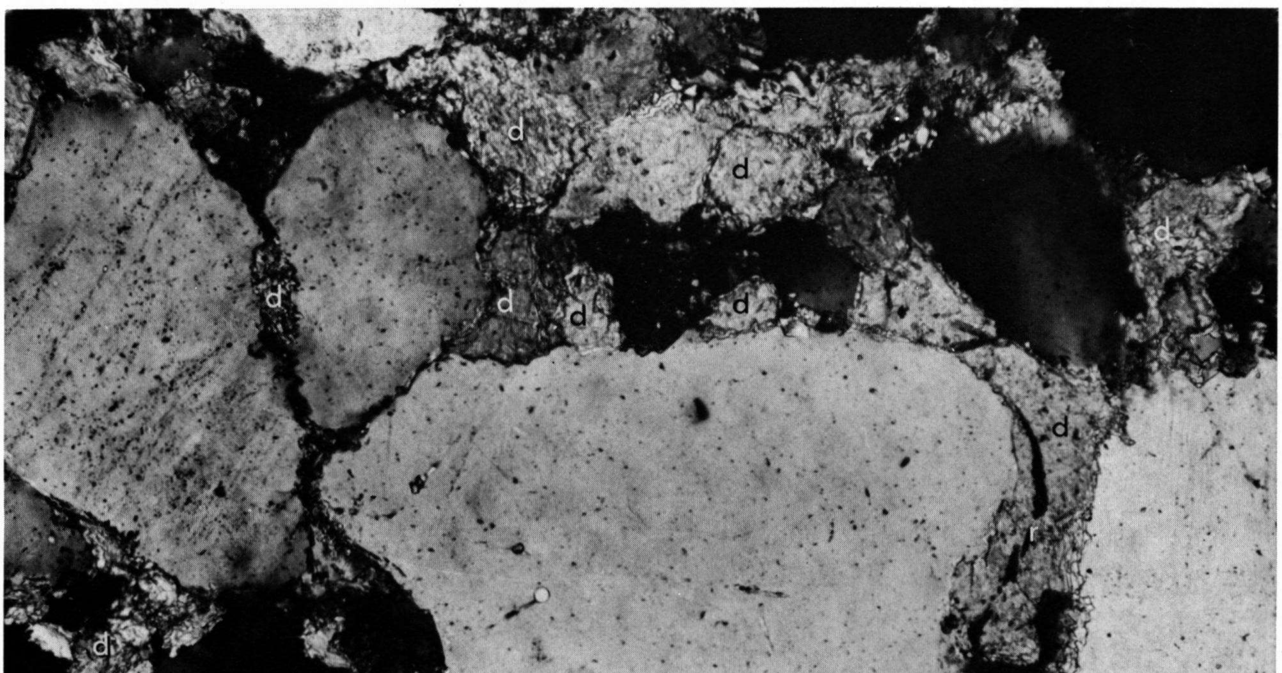
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area south of the western Pyrenees was present during the Emsian and beginning of the Middle Devonian.

In contrast to the association of the light and heavy minerals, which was due to conditions, affecting the source material, the sorting of the quartz sands is the direct result of the littoral environment in which the sediment was transported before it definitively came to rest.

After deposition these quartz sands have locally undergone bioturbation, as is shown by field observations (Fig. 28 and Plate I) and in thin sections. Moreover, in these sandstones of high porosity and permeability, an intensive interstitial water circulation must have occurred. Evaporative loss of interstitial waters could in many parts have raised the salinity, the loss being replaced by seepage flow of sea water, which could already be concentrated if it had descended from shallow lagoons (see under Dolomite).

Increasing depth of burial and compaction led to an increase in the number of grain contacts and a decrease in the porosity. For the explanation of the intense interpenetration of the quartz grains, burial alone was of minor importance, because the amount of overlying sediments not exceeded approx. 1000–1500 metres (see Chapter II and Mey, 1968), and hence the temperature rose only slightly. Present knowledge of the stratigraphy and paleogeography of the Mesozoic and Cenozoic rocks which can possibly have covered the Axial Zone, does not permit to admit greater values for the overburden. In the unlikely case that this overburden amounted to much more, it could give an alternative explanation for the quartz grain interpenetration. We will use, however, the presently most probable data concerning burial.

According to many authors there exists a general relationship between the degree of pressure solution and depth of burial. Taylor (1950) noted that sutured contacts begin to appear with a depth of burial of about 2000 m in Wyoming sands. Therefore the strong interpenetration of the quartz grains of the Basibé Formation must be attributed to additional factors. Burial together with the presence of intergranular solutions, tectonic pressure and heating may be suggested. Solution of quartz at the grain contacts appears to have been accompanied by simultaneous precipitation of silica as quartz overgrowth on quartz grains and/or pore fillings (Waldschmidt, 1941; Taylor, 1950; Heald, 1955, 1956; Pettijohn, 1957; Dapples, 1959, 1962, 1967; Siever, 1959, 1962; Thomson, 1959; Maxwell, 1960; Scherp, 1963; Sharma, 1965; Renton et al., 1969; Füchtbauer et al., 1970). Initial porosity must have almost completely disappeared in most places due to the quartz precipitation (Waldschmidt, 1941; Taylor, 1950; Siever, 1959, Thomson, 1959; Adams, 1964). During this stage the saline interstitial water must have been driven out, which in our opinion dolomitized the intercalated limestone beds. It is known (Maxwell, 1960; Von Engelhardt, 1967; Renton et al., 1969) that moving fluids greatly affect the solution, transport and redeposition of mineral matter, and also favour the above-mentioned processes. Solutions, having become supersatured in silica by pressure solution of quartz grains, can even have precipitated their silica in the adjoining dolomitized limestones.

At a later stage, a renewed solution of the quartz will have taken place, caused by tectonically induced pressure and elevated temperature, the latter being due to the intrusion of the Maladeta granodiorite and its accompanying dykes (see Chapter II). Interpenetration of the quartz by pressure solution gave rise to various types of contact (as described by Taylor, 1950; Heald, 1955, 1956; Siever, 1959; Thomson, 1959; Carozzi, 1960; Trurnit, 1968), see Fig. 27. Rise of temperature and tectonic pressure must also have caused deformation of grains and, at least partially, the undulosity and some fracturing of grains (Fig. 27), (Borg et al., 1960; Conolly, 1965). Since changes in temperature greatly control the solution of quartz (together with pH values), according to Krauskopf (1959); Siever (1959, 1962) and Dapples (1967), this will have been the main reason for the considerable pressure solution of the quartz.

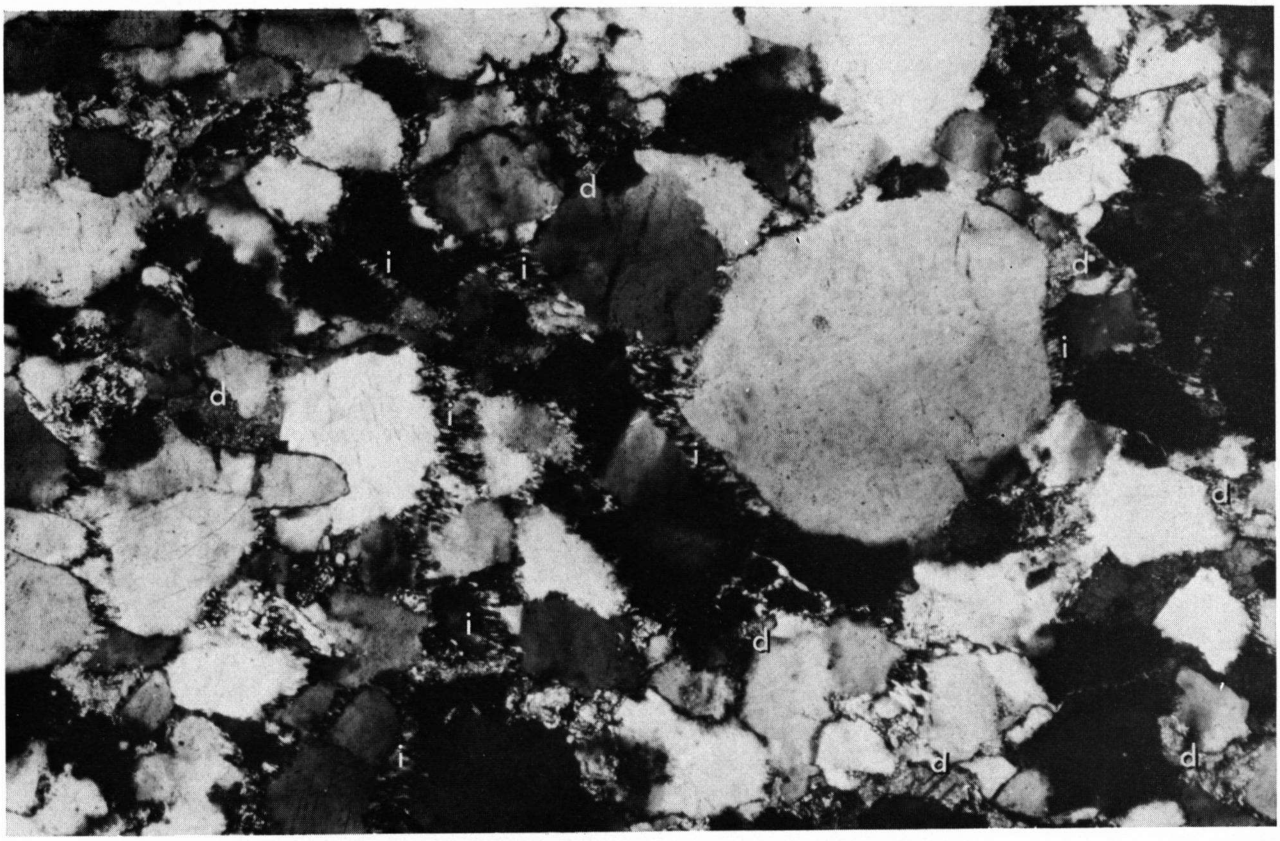
Heald (1956) and Thomson (1959) state that the presence of clay between quartz grains seems to generate considerable pressure solution resulting in sutured, microstylolitic contacts. Renton et al., (1969), on the other hand, reported no observable increase in the rate of pressure solution when mica or clay minerals were present. This may, however, be due to the duration of their experiments, although Thompson (1959) also observed in fossil sediments, quartz grains with clay coatings where no pressure solution developed.

Sutured, microstylolitic kinds of contacts are found between the quartz grains of the Basibé Formation, although very little clay is present. More often, highly stylolitic up to fibrous contacts are observed between the quartz grains (Figs. 26 and 27). Rather little clay occurs within this fibrous microcrystalline quartz. The clay fibres are generally directed perpendicular to the contacts and mainly exist in the secondary quartz. On concavo-convex contacts, however, thicker clay coatings occur. According to Dapples (1962, 1967), (such) border zones of authigenic micaceous minerals are common in strongly folded quartz sandstones.

The occurrence of various stages of pressure solution within one sample, or within a small distance in one quartzite bed, could be due to the orginal presence of some interstitial clay in the case of the more highly sutured and elongated quartz grains, in accordance

Fig. 26. Highly sutured, stylolitic contacts between quartz grains. Within these sutured zones, illite fibres (i) often are present. Dolomite (d), replacing quartz, also occurs. Sample 345 (Section 15), nicols crossed, 80 x.

Fig. 27. Tangential (t), concavo-convex (v) and sutured (s) quartz grain contacts in quartzite. Secondary quartz (q), clay coatings (c) and illite fibres (f) are present. Replacement by dolomite (d), sometimes occurring as rhombs. Rutile needles (r) in quartz grain. Sample 303 (Section 25), nicols crossed, 200 x.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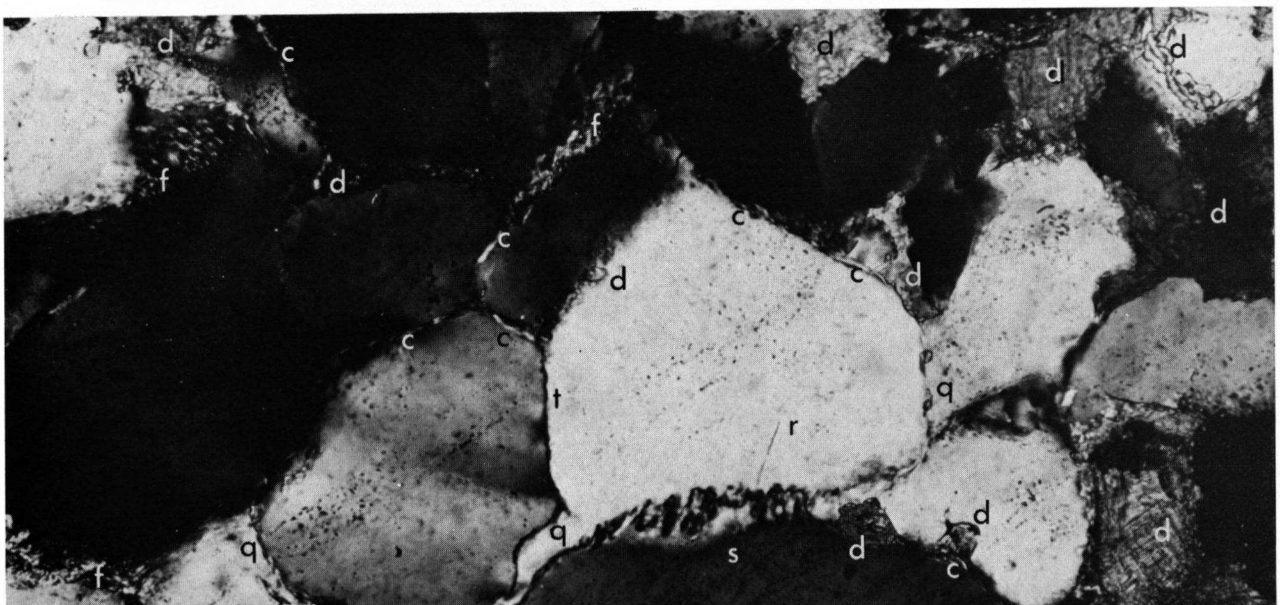
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with similar features described by Gilbert (1954); Heald, (1955, 1956, 1959); Siever (1959); Thomson (1959) and Dapples (1967).

A rather important diagenetic process in many quartzites is the replacement of quartz by carbonate, mainly dolomite (Figs. 25 and 27). This replacement initially started from the original pore center, as mentioned above, perhaps due to residual porosity, as far as residual pores were still present. Replacement advanced preferentially along the grain contacts, affecting, at a later stage, both the secondary quartz rims and the detrital quartz grains. In quartzites with highly sutured, stylolitic and deformed quartz grains, dolomite replacement is rare or absent. A possible cause may be the lack of original pore space from which the replacement by carbonate could begin.

According to current opinion (Walker, 1960, 1962; Siever, 1962; Dapples, 1962, 1967; Sharma, 1965), replacement of quartz by carbonate is mainly the result of a change in the pH of interstitial water (rising above pH 9) but also of a change in temperature. An increase in temperature can cause an increase in pH.

Sandstones which laterally grade into carbonate rocks, as occur in the eastern part of the area studied, tend to have a higher original carbonate content than the uniform (pure) quartz sandstone beds. However, this is often difficult to prove, since the quartz grains in samples from that area are often largely replaced by the carbonate. Original outlines are often recognizable.

Brownish iron oxide patches within the carbonate are abundant in some of these sandstones. Sutured pressure solution contacts and contacts with clay coating between the quartz grains are also met with. These features, together with dust rings of partially replaced quartz grains indicate that abundant replacement of quartz by carbonate has taken place. Some quartz grains floating in carbonate cement are present in some sandstones. These quartz grains seem to be less or not corroded, which is in accordance with Dapples' (1959) statement, that this carbonate addition took place during an early burial stage.

As mentioned above, replacement of quartz by dolomite occurs in some samples (the detrital quartz grains being affected), while replacement of dolomite by quartz also takes place (in the dolomite veins). It seems reasonable to assume that the replacement dolomite is of a later generation than the vein dolomite. The replacement carbonate cannot have come from the pressure solution processes in the carbonate rocks, since most of the solution features in the nodular limestones occurred before the dolomitization and the introduction of silica (see under Dolomite).

The uppermost limestones also show abundant pressure solution phenomena, and may also have contributed to the carbonate cement, although generally speaking the amount of carbonate in the top of the quartzite units is not higher than in the lowermost or middle part of these units.

Introduction of silica mainly took place during and after the dolomitization of the nodular limestones (see under Dolomite).

SUMMARY OF PETROGRAPHY AND DIAGENESIS

Summarizing, the three different lithologies, as given on page 22, are composed of carbonate and quartz although the amounts of these components are quite different. These differences are mainly due to the energy conditions in the depositional environments of each of these lithologies (see Chapter VI). The physical and chemical changes which took place in these sediments subsequent to deposition, i.e. the diagenesis, have many features in common.

Lime mud recrystallized and underwent compaction, cementation and replacement processes, which led to lithification. Pressure solution gave rise to stylolites, while replacement by dolomite occurred locally due to addition of Mg-enriched interstitial water. Another type of replacement was the corrosion of quartz by carbonate. Authigenic pyrite crystals originated near former fossil remnants.

The silty to sandy argillaceous limestones or dolomites show, in general, a similar diagenetic history. Prior to lithification these deposits were affected by strong organic disturbance. The only authigenic iron mineral is hematite.

The quartzites are also heavily affected by diagenetic processes which changed the original detritic quartz sandstones into the present quartzites. The main phenomenon was pressure solution, by which silica became available for grain overgrowth and cementation. Some of the silica migrated to the dolomitized limestones. The latter rocks were mainly dolomitized by saline interstitial water which was driven out of the quartz sandstones during the silicification process. Authigenic hematite is scarce.

The detrital components of all lithologies of the Basibé Formation are the most resistant ones, viz. quartz, three heavy minerals, and muscovite. This mature assemblage must have been the result of weathering of the source rock and/or the composition of this source rock. Transport and energetic conditions in the depositional environment led to sorting of the material. The water depth during sedimentation was perhaps somewhat different for the three lithologies, but they all belong to the shallow marine environment (see Chapter VI). The differences are mainly due to the different energy conditions.

The climate must have been warm (Shirley, 1964), as indicated by the fauna. Seasonal evaporation may have been a factor in the diagenetic changes (dolomitization), although the area was situated outside the Devonian evaporite zone, as indicated by Lotze (1964).

Other factors which influenced the diagenetic processes in the sediments are burial, tectonic activity and rise in temperature, due to intrusion of the Maladeta granodiorite and its accompanying dykes.

CHAPTER VI

CONDITIONS AND ENVIRONMENT OF DEPOSITION

Sedimentary structures characteristic of particular environments are scarce in the sediments of the Basibé Formation, as are faunal indicators of water depth or environments. Therefore the environmental interpretation of the deposits had to be based mainly on other data. These data, assembled in the preceding chapters, concern the vertical and lateral distribution patterns of the deposits and the petrography, which gives information on the mineralogy and texture, the primary lithology and the diagenetic changes. Comparison with analogous fossil and recent sediments and depositional environments facilitated a better understanding of the dynamic complex of conditions that prevailed at the depositional site of the sediments.

GELADA FORMATION

The uppermost sediments of the Gelada Formation which were deposited before the sediments of the Basibé Formation contain a relatively high amount of detritic components. Addition of quartz, muscovite, heavy minerals and clay occurred in an environment of predominant carbonate deposition. No indications of wave and current activity were found. Fossils are almost completely absent, but burrowing by organisms was common, as shown by the orientation and concentration of the quartz grains. Influxes of siliciclastic material led to the mixed sediments.

Black streaks, described by Mey (1967b, 1968) and also found by the present author in the uppermost part of the Gelada Formation, may represent plant remains, but also feeding trails (Chondrites). These streaks correspond to the description given of Chondrites by Simpson (1957) and Osgood (1970), who stated, however, that Chondrites are present in a wide range of depositional environments. Apart from these burrows, some oscillating tracks were also observed.

A shallow sea with a moderately subsiding to stable bottom is assumed to be the depositional environment of these sediments.

BASIBÉ FORMATION

Nodular weathering limestone

The nodular weathering limestones of the Basibé Formation originated in lime muds which had only a low clay and quartz silt content. Fossils are not frequent, and are generally only present as fragmental remains, scattered in the sediment. No indications were found of strong organical or wave and current disturbance of these deposits. Features demonstrating solution in the depositional environment were not found either. All present detectable structures are of later diagenetic origin, as mentioned in Chapter IV.

Low energy conditions in the environment of deposition of these limestones must be assumed, according to the depositional texture classification by Folk (1959, 1962), Dunham (1962) and Plumley et al. (1962). In this quiet water area, where lime mud was deposited, some very slight agitation or currents may occasionally have occurred, strong enough to transport the quartz silt and muscovite flakes, and some of the fossil fragments. Such quiet to slightly agitated water conditions may be found below wave base in an offshore shallow marine, as well as in a deep marine environment and in a lagoon, where wave and current energy are reduced by a remote barrier or hampered by a wide expanse of very shallow water. From these three possible environments a choice has to be made.

The fossil content indicates a shallow marine environment, while the quantity of pelagic forms is also in favour of an open shallow sea. The warm water fauna is in accordance with the warm climate (see Chapter V and Shirley, 1964) assumed for this area during the Devonian.

The predominantly light colours of the sediments indicate deposition in a well-aerated environment and a scarcity of organic material (Ginsburg, 1957). The presence of pyrite is not an indication that the water above the sediments was poorly oxygenated or that benthonic life was impeded according to Bathurst (1964). Not much can be said regarding the pH and the Eh; there are no indications of their being abnormal or different from the values in normal sea water.

Dolomite

The fossil remnants, stylolitic structures and various crystal sizes in the dolomites of the upper part of the Nodular Limestone Member and in the Quartzite-Dolomite Member indicate that the dolomites are dolomitized nodular weathering limestones (Chapter V).

In Chapter V we pointed out that there are no reasons to admit a penecontemporary or early diagenetic dolomitization by processes such as described by Skinner (1963), Illing et al. (1965), Shinn et al. (1965) and Atwood (1970). Other arguments are that the lime muds were, in our opinion, not deposited in a restricted lagoon but in an open shallow marine environment, since intertidal and supratidal structures and textures, such as algal mats, stromatolites, solution features, breccias, desiccation cracks, birdseye structures and vertical burrows as described by Roehl (1967), Laporte (1967), Germann (1968), Shinn (1968), Kendall and Skipwith (1968) and Shinn et al. (1969) are also completely lacking. An epigenetic origin as defined by Friedman and Sanders (1967) for the replacement of limestone by dolomite, being localized by post-depositional structural elements as faults and fractures, is also excluded. Such dolomites related to faults, described by Pélissonnier (1959) in the Devonian rocks of the northern Central Pyrenees, are not found in the Basibé Formation of our area. The arguments given in Chapter V for the later diagenetic stage of dolomitization of the limestones provide, in our opinion, a conclusive proof. Dolomitization by the the reflux mechanism as described by Adams and Rhodes (1960) and Deffeyes et al. (1965) is less probable than the explanation given.

Deposition of the original lime muds within this shallow sea must have been fairly uniform over large areas, as is shown by the evenly bedded layers which can be traced over considerable distances. Differences in total thickness of these carbonates (Figs. 4 and 32; Plate I) may be due to irregularities in submarine topography, but may also have been caused by differential subsidence or erosion.

The absence of similar carbonates in the area north of the Alpine thrust zone was probably caused by non-deposition.

Silty to sandy argillaceous limestone

A merely slight increase in the supply of terrigenous material, composed of quartz silt and sand, muscovite flakes, heavy minerals and clay, to the carbonate depositional environment could give rise to sediments described as silty to sandy argillaceous limestones. Changing current or wave conditions or differences in the submarine topography may be responsible for the influx of these clastics into the carbonate environment. Currents may have transported the terrigenous material and most of the fossil fragments into the marine environment. The clastics may have been supplied by rivers from a relatively stable lowland (see Chapter V) or from distant littoral or submarine structures.

It is also shown by crinoid stem and shell fragments that most fossil remnants were transported. Bioturbation indicates that the rate of deposition was not extremely fast.

Wave or current energy in the depositional environment was certainly weak since no sorting and winnowing of these sediments was found.

As the silty to sandy argillaceous limestones which are replaced by dolomite occur mainly directly below the quartzites, which were interpreted as offshore bars or barrier island complexes, these argillaceous carbonates were considered as lee-side deposits of these bars. The lee-side or back-bar areas are similar to lagoonal environments or small depression zones between ridges on the shallow sea floor, areas where quiet water conditions permitted the sedimentation of lime muds while intermittent water agitation may have brought in the clastic material. Variations in the amount and/or direction of the supply of clastic material may also be the cause (Sanders and Friedman, 1967). Grading from pure carbonate to silty and sandy argillaceous carbonate, observed in some sections, demonstrate such increases in the supply of detrital sediments.

Quartzite

The quartzites were deposited as clean, well-sorted and rounded quartz sands. According to their structures, texture, lithology, grain size distribution and their geometry these sediments are interpreted as complexes of offshore bars and barrier islands. The few fossils and the interbedding with marine and probably back-bar sediments also strenghten this interpretation.

As stated in Chapter V, the pure quartz sands will have been derived from older quartz sandstones or quartzites, probably outcropping on a strongly weathered lowland. However, erosion of submarine or island outcrops of pure quartz sands or sandstones should also be considered.

The main transport will probably have taken place by longshore currents and wave action. Longshore transport of sediments is at present a common feature in many shallow marine seas and along coasts all over the world. By this agent clastic material can be transported over considerable distances and easily brought to the site of deposition of other sediments. Eolian transport of some sands could also be expected in this environment, but such deposits were not found.

As the transporting power decreased, quartz sands came to rest in the shallow marine carbonate environment. This may have taken place due to a change in hydrodynamic conditions, e.g. by entering shallower water and/or by trapping near low ridges or elevations on the sea floor (Ball, 1967). The rate of sand supply may also have been a factor (McKee and Sterrett, 1961).

The lowermost, mainly thin, quartzite layers in some sections often consist of a poorly sorted mixture of finer sand sizes (Fig. 23), which may indicate lower energy conditions of deposition, e.g. deeper offshore deposition, according to Moore and Scruton (1957), Bernard and LeBlanc (1965), van Straaten (1959, 1965) and Allen (1967).

Stratigraphically higher quartzites often tend to have fine grain sizes at the base and to show an increase in grain sizes towards the top. These sands which coarsen upwards are generally better sorted and winnowed, which may be due to conditions of greater energy caused by waves and currents higher in the depositional environment (Logvinenko and Remizov, 1964; van Straaten, 1965; Hoyt, 1969). The usual grain size in these beds is medium sand size.

After building up shallow sand ridges, breaking waves will have begun to spill over these ridges, causing strong turbulence (McKee and Sterrett, 1961). By this mechanism, different grain sizes can be deposited together. In this way offshore bars may be transformed into barrier islands. Lateral accretion of the offshore bars may also have occurred, thus forming blanket or sheet sands. Whereas the offshore bars are subaqueous, the barrier islands even rise slightly above sea level (Shepard, 1963). Breaking waves on these

Fig. 28. Burrows in a quartzite bed. Section 5, east slope of the upper Rio Isabena Valley.

Fig. 29. High-angle cross-bedding in a quartzite bed of Section 3. Basibé massif. Photograph perpendicular to the bedding. Head of hammer measures 13 cm.

bars and barrier islands or beaches keep the depositional interface agitated. Sorting and winnowing of the sands according to their grain size are mainly the result of waves. Part of the material was taken up and retransported after initial deposition, so that the sand bars could migrate.

The occurrence of coarser grains sizes in an upward direction in these sediments is in accordance with the characteristics given by Pettijohn et al. (1965), Oomkens (1967), Potter (1967), Shelton (1967) and Visher (1965, 1969a, b) for bar and barrier island complex deposits. The high energy conditions in this particular environment also caused low angle and occasionnally high angle cross-bedding. The structures normally found in such deposits have been described by Thompson (1937), McKee (1957, 1964), van Straaten (1959, 1965), Werner (1963), Soliman (1964), Logvinenko and Remizov (1964), Bernard and LeBlanc (1965), Klein (1967), Masters (1967) and Andrews and van der Lingen (1969). These workers indicate that the gently inclined cross-bedding dips seawards and is generated by the swash and backwash, while the high angle cross-strata dip shorewards and are due to deposition of sand on the shoreward side of a bar by waves spilling over the sand ridge. The parallel laminae often present are also characteristic of the high energy beach environment and breaker zone and are formed when sand settles from suspension clouds of sand in advancing waves (Reineck, 1967), and was afterwards transported and sorted by swash and backwash (van Straaten, 1959; Imbrie and Buchanan, 1965).

In the Basibé Formation cross-stratification (Fig. 29) is so scarcely present that it could not serve to determine which side of a certain quartzite layer is the seaward and which side is the shoreward direction.

However, in the deposits of the Basibé Formation homogeneous sandstones, without cross-bedding, predominate (Fig. 8). Such a homogeneous appearance could be due to the original depositional conditions, e.g. rapid and sheet sedimentation, but by such a process poorly sorted deposits will be formed if not derived from uniform source material. More often, however, homogeneity will result from intensive disturbance by organisms.

Burrowing is common in offshore parts of sand deposits seawards of the breaker zone, and much less in the subaerial parts. This bioturbation can already occur during deposition near the depositional surface, but also at a later stage as long as the deposit is exposed at the interface. Sand laminae may have been disturbed by micro-organisms, while complete homogenization will be due to larger burrowing fauna (Werner, 1963). Burrowing may also be favoured by a low sedimentation rate or an interruption in sedimentation (Moore and Scruton, 1957; van Straaten, 1959 and Curray, 1960), but these are not necessary conditions.

Burrowing animals will have occurred in these sediments and some layers show burrows (Fig. 28), but few fossil fragments are preserved. The high energy conditions were not favourable for preservation, while possible percolation by meteoric water will have led to complete leaching of the predominantly carbonate fossil content. Organic matter, if originally present, will have been removed by oxidation.

Eolian deposits, on the top of a barrier island or intercalated within bar deposits, are to be expected but were not found. They were either not formed or not preserved in these rocks or could not be recognized. They may easily have been eroded or have migrated in the depositional environment under consideration.

Whether some parts of the bars were also eroded after deposition could not be ascertained either; if it occurred it left no traces.

The attempt at an interpretation of the quartzites made above applies to all quartzites taken together. The separate quartzite layers are so similar in texture and in the scarcely present sedimentary structures that they cannot be distinguished. Moreover, as we previously pointed out, separate layers can hardly be correlated from one section to another, even if they are only some hundreds of metres apart.

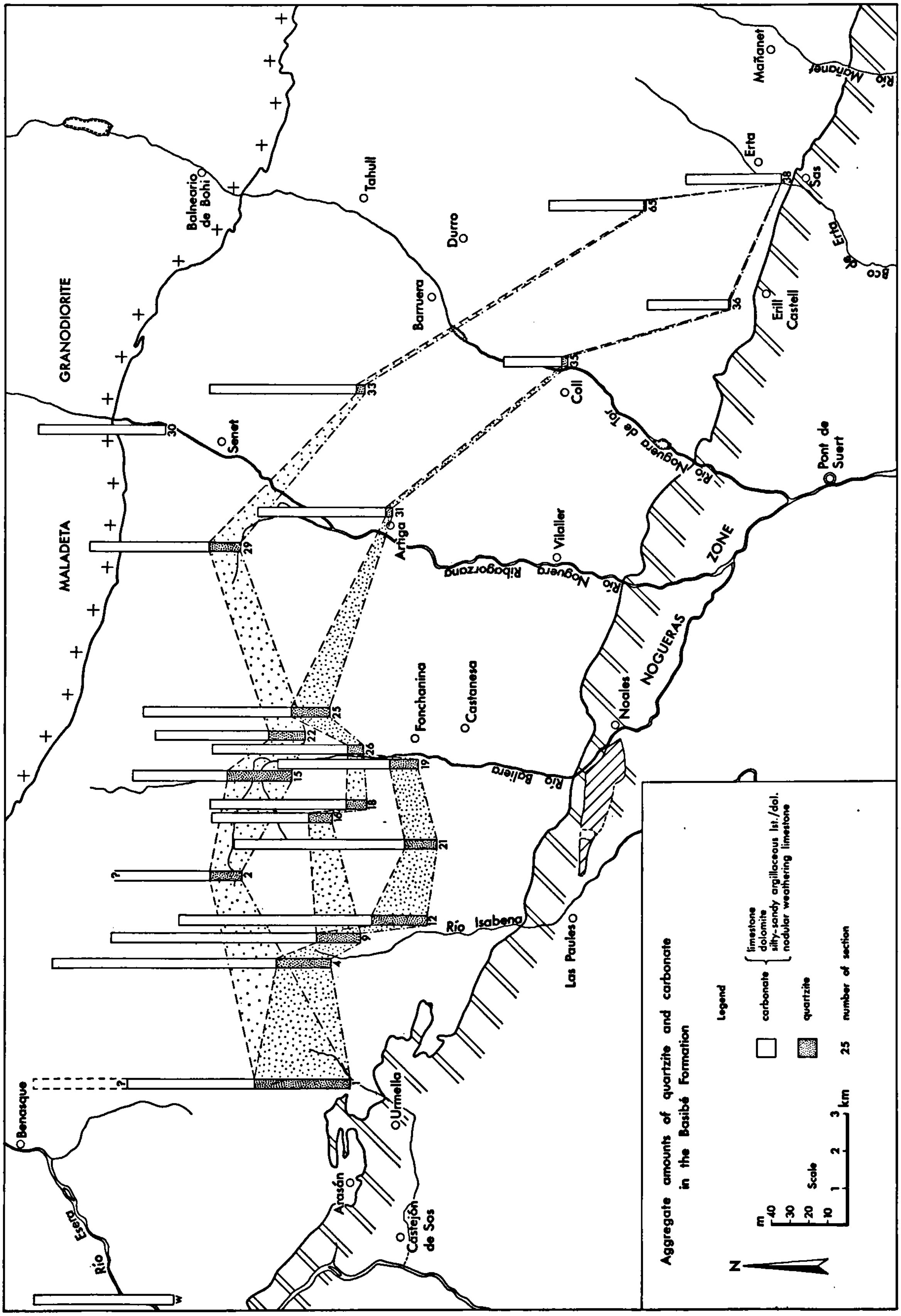
In general, the characteristics mentioned above agree with those given of bars and barrier islands by Thompson (1937, 1949); Shepard and Moore (1955); McKee (1957); Moore and Scruton (1957); King (1959); van Straaten (1959, 1965); Shepard (1960); Bernard and LeBlanc (1965); Pettijohn et al. (1965); Potter (1967); Shelton (1967); Zenkovich (1967); Johnson and Friedman (1969) and Visher (1969).

Relation Quartzite-Silty to sandy argillaceous dolomite

The silty to sandy argillaceous dolomites, which generally underlie the quartzites, are considered to be low energy sediments. The contacts with the quartzites are abrupt and straight without any erosion features (Fig. 8). Load cast structures were not found either, which indicates an early diagenetic cementation of the carbonate sediments. Thus a certain lapse of time must have passed between the end of the carbonate sedimentation and the influx of quartz sands. As stated above, grading from pure carbonate to silty and sandy argillaceous dolomite, due to influxing of fine clastic material can sometimes be recognized in the deposits. These argillaceous dolomites (back-bar deposits) often became overrun by quartz sands, migrating as sheets, and probably derived from the bars or barrier islands. Such a spreading or washover of bar or barrier island sands over the back-bar environment may have been caused by increased wave energy, storm activity, increased sand supply or local relative subsidence (Fisk, 1959, Rusnak, 1960). However, also the lack of sand supply can have induced migration of the bar sands (Dillon, 1970).

The argillaceous carbonates show many characteristics of back-bar environment deposits, but do not have characteristics indicative of a restricted back-bar or lagoonal environment (Sabins, 1963). According to

Fig. 30.


Aggregate amounts of quartzite and carbonate in the Basibé Formation
Legend
carbonate
limestone
dolomite
silty-sandy argillaceous lst./dol.
nodular weathering limestone
quartzite
25 number of section
Scale
m
40
30
20
10
1 2 3 km
N
GRANODIORITE
MALADETA
NOGUERAS ZONE
Benasque
Río Esera
Arasán
Castejón de Sos
Urmella
Las Paules
Río Isabena
Río Baliera
Noales
Fonchanina
Castanesa
Río Noguera Ribagorzana
Vilaller
Artiga
Senet
Coll
Río Noguera de Tor
Pont de Suert
Barruera
Durro
Tahull
Balneario de Bohí
Erill Castell
Bco de Erta
Sas
Erta
Mañanet
Río Mañanet

their structures, texture, lithologies and fossil content they are shallow open-marine sediments, deposited in a sheltered, slightly agitated, environment (Davies, 1969).

A dark, silty to sandy argillaceous carbonate layer (about 5 cm thick) in Section 24 contains shale flasers and small sand lenses. This may, according to Reineck (1967), indicate an alternation of current activity and quiet water conditions. This could be expected to occur in shallow environments, such as a back-bar, or in a deeper offshore environment.

A few black shale layers (Fig. 8) may represent deposits in a more restricted lagoon. Thinner shale laminae found in some quartzites may have been deposited as mud layers after periods of storm, as large amounts of sediment are deposited, or at the turn of tides (van Straaten, 1965).

Paleogeography of the Quartzite-Dolomite Member

Unfortunately, the data obtained from a careful analysis of the sediments, which have undergone many diagenetic changes, provide no information on some important questions concerning the paleogeography, such as the position of a possible shoreline, the source area and the transport direction of the quartz sand. The transport direction cannot be concluded from the scarce structures. Migration of the bars themselves was rather restricted, in accordance with the accumulation of quartzite units in the western and central parts of the area studied (see Figs. 4, 30 and 31).

From the horizontal and vertical distribution of the quartzites it may be concluded that they were deposited in a moderately stable environment, in which the bars or barrier islands once formed, remain rather stationary, in view of the vertical and lateral appearance in the area studied. As mentioned above, lateral accretion and migration certainly can have occurred within this area, forming blanket or tabular shaped sand bodies.

The geometry of the present quartzite units, considered separately as well as together (Figs. 4, 30, 31 and 32), is also well in accordance with recent examples of bars and barrier islands. All these recent accumulation forms are more or less wedge-shaped and parallel to the coast line along gently inclined shelves. They are generally bordered on their landward side by lagoons or marshes with fresh, brackish, normal or hypersaline marine water. Seawards, in the direction of greater depth, these sandy deposits usually pinch out, giving way to finer grained sediments.

As we only found marine sediments surrounding the bars, these will have occurred rather isolated in a shallow marine area, forming offshore bars or a barrier island complex, relatively far from the shore on a slightly inclined sea bottom. The back-bar environment may have been present leeward of these isolated bars or barrier islands, but may also have existed as shallow quiet water environments between some bars or barrier islands.

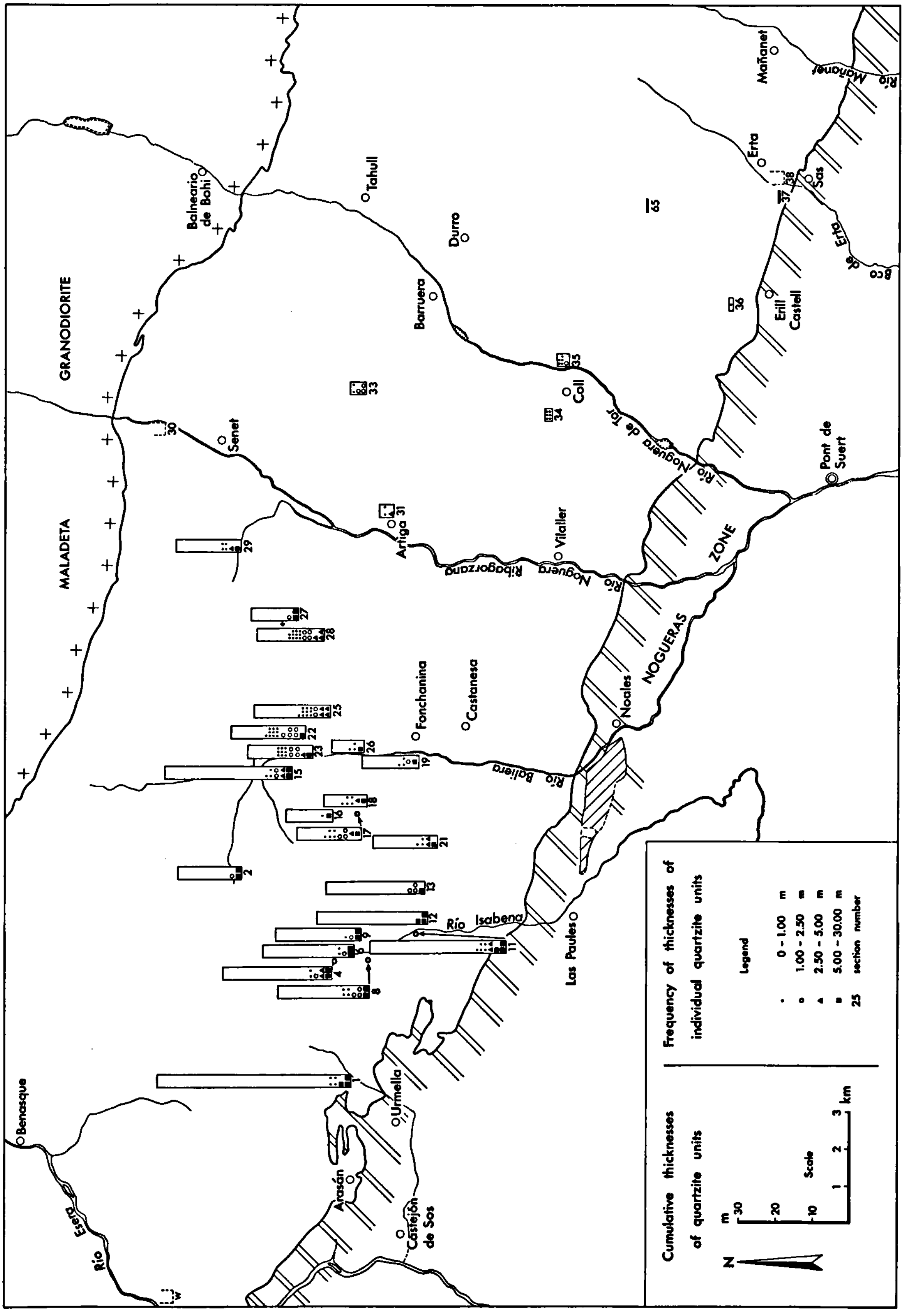
In general these bar or barrier island deposits resemble recent examples described by Fisk (1959); LeBlanc and Hodgson (1959); Shepard (1960); Rusnak (1960); Phleger and Ewing (1962); Allen (1964, 1965 a, b); Curray and Moore (1964); Bernard and LeBlanc (1965); van Straaten (1965); Bird (1967) and Kwon (1969).

Fossil examples such as those described by Sears et al., (1941); Hollenshead and Pritchard (1961); Weimer (1961); Sabins (1963); Lane (1963); Oele (1964); Boyd and Dyer (1966); Thomas and Mann (1966); Masters (1967); Shelton (1967); Davies (1969) and Johnson and Friedman (1969) also show many similar characteristics. Various conclusions drawn by many of these authors, such as a differentiation between the seaward and landward or back-bar environment on account of the sediments and their fauna content, the source of the sediments, their transport direction and the transporting medium, and also the differentiation between regressive or transgressive sequences, cannot be given for the Quartzite-Dolomite Member of the Basibé Formation with the present data.

From a comparison of the descriptions of fossil and recent bars and barrier islands and the data on the quartzites of the Basibé Formation in the previous chapter, with special regard to their geometry, some statements can, however, be made. Figures 4, 30, 31 and 32 (and Plate I) show that the quartzite units occur mainly in blanket or tabular shape, that they pinch out in an easterly direction, that their boundaries with overlying and underlying rock types are generally abrupt and that they hardly contain any sedimentary structures. Their general direction is W-E to NW-SE, which may represent the pattern of the original bar and barrier island deposits. The Hercynian folding moderately influenced these directions. The greatest increase in the thicknesses of the separate quartzite units (Figs. 4 and 31) and in the aggregate thickness (Fig. 30) is from ESE to WNW. Normally to these approximately W-E directions the number of quartzite layers gradually diminishes, both to the north and the south (Fig. 31), and a nearly symmetrical pattern exists. In the northern and southern parts of the area (see figures) no quartzite is present (see also Chapter IV). A possible transition between these parts and the central part may be expected, but was not found. Mey (1968), however, mentioned such a probable transition (see Chapter IV).

The transporting longshore current was probably directed parallel to the bars and barrier islands, according to most of the recent examples (Klein, 1967). It is expected that the current also ran parallel or with a small angle to the possible coast line. Such a longshore current could have been modified considerably by submarine topography and tidal action. Whereas a longshore current is assumed to have been the large-scale transporting agent, wave action in the shallow water environment was the predominant factor in the local modification of the bars and barrier islands. Whether this current, which transported the clastic material, came from a NW or from a SE direction, cannot be

Fig. 31.


GRANODIORITE
MALADETA
Balneario de Bohi
Tahull
Durro
Barruera
Mañanet
Río Mañanet
Erta
Sas
Bco de Erta
Erill Castell
Coll
Río Noguera de Tor
Senet
Artiga
Vilaller
Río Noguera Ribagorzana
Pont de Suert
NOGUERAS ZONE
Fonchanina
Castanesa
Noales
Río Baliera
Río Isabena
Las Paules
Urmella
Benasque
Río Esera
Arasán
Castejón de Sos
Cumulative thicknesses of quartzite units
m 30 20 10
Scale 1 2 3 km
N
Frequency of thicknesses of individual quartzite units
Legend
0 - 1.00 m
1.00 - 2.50 m
2.50 - 5.00 m
5.00 - 30.00 m
25 section number

determined. The dip direction of the few cross-laminations does not give any information, while the back-bar side cannot be indicated either, since the bar or barrier island deposits are all uniformly underlain by these back-bar sediments and no difference in the deposition on either side of the quartzites can be made.

The thickness of the underlying sequence of nodular weathering limestones and their dolomitized counterpart (Figs. 4 and 32) gradually increases towards the west. Only in the Baliera area, where these rocks show a decrease in thickness, do some quartzite layers occur in these lower parts of the sections.

The bars and barrier islands, once developed, must have remained rather stationary in accordance with their vertical and lateral occurrence in the area studied. Some migration has, however, taken place, as is shown by the intercalated carbonate sediments. From what we know of present-day sand bars, the rate of sedimentation of the sands can have been relatively high. A considerable lapse of time may have passed between the deposition of the various quartzite layers. A slow and nearly uniform relative subsidence took place in the area.

In the south-eastern part of the area, lime mud sedimentation of the upper member began locally before the quartz sand sedimentation came to an end. These lime muds are again deposited in a low energy environment, probably a shallow open-marine environment similar to that of the lower member limestones. However, according to their present colour and bedding characteristics, some sedimentation conditions and (early) diagenetic conditions were different. Locally, higher energy conditions occurred in this depositional environment, as is shown by shoals. In the eastern part some quartz sand influxes took place, as shown by the quartzites intercalated in the limestones of the upper member of Sections 31 and 36 (Plate I and Fig. 4).

The marine limestones cover the entire area of quartz sand sedimentation. Bars and barrier islands, even if emerged, were subsequently covered. A relative subsidence of the area, probably strongest in the western part, must have been a condition for this transgressive sedimentation.

Limestone

The limestones of the upper member of the Basibé Formation originated mainly from lime muds, while some levels are composed of pellet muds. The latter are preferentially dolomitized. Quartz silt and sand are nearly absent (Figs. 4 and 23), but clay is present in the uppermost part of the member and increases in

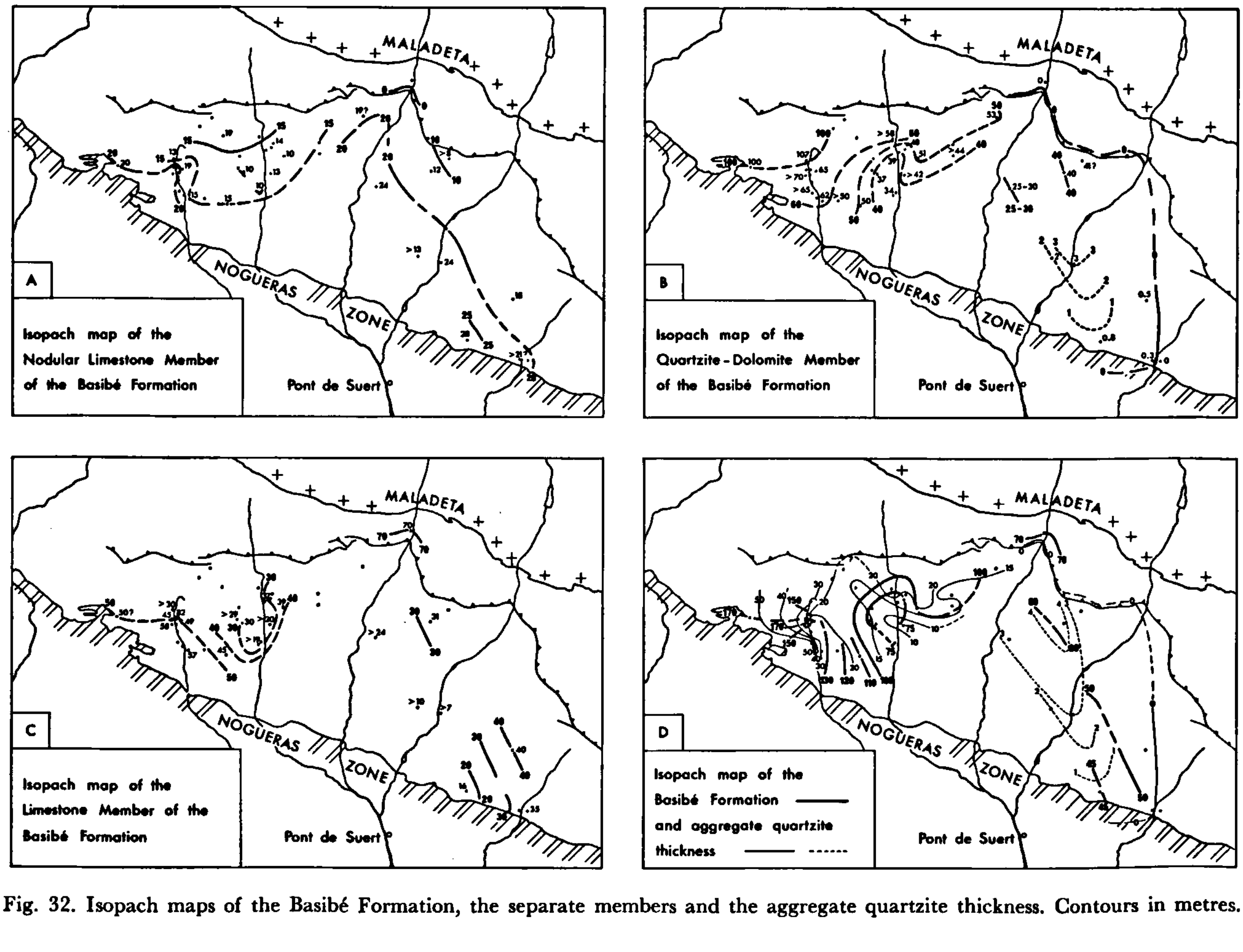


Fig. 32. Isopach maps of the Basibé Formation, the separate members and the aggregate quartzite thickness. Contours in metres.

content towards the transition into the Fonchanina Formation. This formation consists nearly entirely of argillaceous sediments.

In the limestones fossils are not frequent and generally occur as fragments. Many limestones do not contain any fossils at all, or only some sparse fragments. Some beds, however, have an abundant faunal content, though also mainly fragmental. Bioturbation was found in exceptional cases.

Although the fossil content is rather poor, it indicates an open shallow marine environment of deposition for these lime muds.

Energy conditions in the environment of deposition were usually quiet, as appears from the predominance of muddy sediments (Folk, 1959, 1962; Dunham, 1962; Plumley et al., 1962 and Leighton and Pendexter, 1962). This indicates deposition in water depths that were nearly free of wave and current action. Such quiet water areas are found at the lee side of barriers as well as in a deeper offshore environment below

In some levels, sedimentary structures and also high amounts of quartz silt and fossil fragments demonstrate local agitation. Considering the occurrence of small ripples, gullies and cross-bedding (all of centimetre scale), water movement was caused by currents. Occasionally even high energy conditions may have prevailed in the lime sedimentation environment, as shown by the intercalation of quartzites in some wave base.
sections east of the Ribagorzana River.

These influxes of quartz silt and sand occurred after relatively long intervals of non-deposition, or were sudden phenomena, since even in thin sections (Section 31, samples 946 and 947) the boundary between pure limestone and quartz silt is a straight line.

Microscopical analyses of samples has demonstrated that some limestone beds were deposited as pellet muds (Fig. 19). The occurrence of pellet muds is a common feature of recent shallow, quiet marine water environments where lime deposition occurs (Newell and Rigby, 1957; Cloud, 1962; Imbrie and Purdy, 1962; Purdy, 1963; Beales, 1965; Illing et al., 1965). The texture of the micrite pellets is explained by most authors by a process of transport, or inorganic precipitation, cementation and recrystallization, or by organic agglutination. The amount of pellets of faecal origin is thought to be important. The pellet limestones do not, however, show, burrows of the original pelleting organisms, which is in accordance with the findings of most authors. Besides, most of these structures would have disappeared due to the dolomitization that affected nearly all these limestones. A specific environment of deposition cannot be given, but rather shallow water is suggested by the findings in recent environments.

Another argument is probably the observation that some of the dolomitized pellet limestones, or pelmicrites (according to Folk, 1959, 1962), are overlain by layers with abundant fossil fragments (Fig. 20), so that shallow water is most probable. The accumulation of such fossil fragments in a muddy environment may be due to storm or current activity. Local irregularities in bottom topography, e.g. bars or shoals, may have had great influence on such processes, as is suggested by the isolated occurrence of these pellet mud deposits (Sections 12, 18 and 25).

The thicknesses of the individual limestone beds decrease considerably towards the top of the formation (Fig. 10). In the uppermost part, often only laminae are found. These laminae are of inorganic origin. Generally an alternation of limestone laminae and argillaceous limestone laminae occurs, indicating a low rate of sedimentation, repeatedly interrupted by influxes of argillaceous material.

Due to the introduction of increasing amounts of clay, the lime mud sedimentation had came to an end in the Fonchanina Formation.

The limestones of the upper member of the Basibé Formation are rather dark. This may indicate a relatively high content of organic material and hence deposition in an environment where low oxidizing or even reducing conditions prevailed during early burial. This is a common feature where fine-grained mud is being deposited.

Lime sedimentation in the area north of the Alpine thrust zone will have shown a great resemblance to the depositional environment of the limestones of the upper member of the central area. These sediments of the northern area are comparable to many parts of these upper limestones, also in faunal content. Uniform sedimentation conditions may therefore be expected, characterized by low energy and probably shallow water.

Since limestones comparable to the nodular weathering limestones of the Basibé Formation of the area south of the Alpine thrust zone are not found north of this zone, it is assumed that these were not deposited, or different conditions prevailed in that environment of lime deposition.

SUMMARY OF DEPOSITIONAL ENVIRONMENTS

> "*We may conclude that while littoral and neritic marine facies are consistent and reliable indicators of* relative *water-depth, they are unreliable as guides to* absolute *depths of deposition*".
>
> J. R. L. Allen, 1967, Depth indicators of clastic sequences. p. 441.

The four major lithologies distinguished which together compose the Basibé Formation, i.e. the nodular weathering limestones and their dolomitized counterparts, the silty to sandy argillaceous carbonates, the quartzites and the limestones, are each representative of a particular depositional environment. They are generally sediments deposited in shallow marine environments, where slight variations in water depth or bottom topography, turbulence and sediment supply caused differences in accumulation forms and lithological composition.

Within such a shallow marine environment many variations in sub-environments, each with their own characteristics can occur side by side. According to "Walther's Law of Facies", laterally adjacent sediments must also succeed each other vertically in a sedimentary sequence (Krumbein and Sloss, 1963; Visher, 1965). The vertical succession of sediments in the stratigraphic column of the Basibé Formation thus reflects a succession of lateral environmental distributions.

The sedimentary sequence of the Basibé Formation solely contains deposits of open shallow marine environments. This indicates that the depositional environment in general did not undergo fundamental changes in water depth.

The lower carbonates are interpreted as deposits of a shallow, well aerated, open marine environment according to their lithology, structures, textures and faunal content. Influxes of quartz sands led to the formation of submarine bars or barrier islands which may in their uppermost parts consist of subaerial deposits. Though they generally do not contain any fossils, these intercalated quartzites may be assumed to represent shallow marine or littoral deposits. The upper limestones are probably also shallow marine or somewhat deeper marine deposits. The term regressive can be used for the sequence from the lower carbonates to the quartzites, the term transgressive for the sequence from the quartzites to the upper limestones.

Differences in water depths of the environments were, however, small, which means that the supply of sediments and their deposition was more or less in equilibrium with the subsidence of the area. Slight differences (see Fig. 32) have, however, certainly occurred. A rather stationary depositional environment is inferred, probably a shallow shelf sea or epicontinental sea. A stable lowland occurred in the vicinity. The climate was warm.

The predominant sediment in this environment was lime mud. Clastic sediments that were introduced consist of very resistant minerals and represent a mature assemblage. This may be due to the composition of the source rock or to its weathering conditions. These clastic sediments, which consist mainly of quartz, probably became available by erosion, either in the hinterland or in the marine realm itself. Prolonged transport, probably by longshore currents, led to some sorting of the material and to the influx of these detritics in the shallow marine environment of carbonate deposition. The direction of these longshore currents remains uncertain, but may have been a NW-SE direction. Particular hydrodynamic conditions, related to water depth, submarine topography and wind directions, i.e. mainly wave action and to a lesser degree currents, sorted and winnowed the clastic sediment and created bars, barrier bars or barrier islands.

Leewards of these bars, somewhat restricted, shallow depositional environments existed, where silty to sandy argillaceous carbonates were deposited, in which burrowing was common. These open marine, though relatively, sheltered back-bar environments were characterized by the faunal content of the deposits, the absence of terrestrial sediments and evaporite minerals. The term "restricted" thus only applies to the lower energy level of this environment.

The quartz sand accumulation remained rather stationary in this slowly subsiding environment, although the bottom irregularities may have changed the courses of the transporting currents, so that the localities of sand deposition shifted repeatedly. A sudden stop in the sand supply, and more probably greater subsidence, again allowed all-over lime mud sedimentation during the deposition of the upper member. Due to continuing subsidence, the earlier deposits were buried, protected against erosion and preserved by the subsequent lime mud sedimentation which again occurred in a shallow marine environment. After a considerable period of lime deposition, the sedimentation no longer kept pace with the subsidence. Deepening of the environment occurred and influxes of clayey material took place. Another reason for ending the lime mud deposition, besides deepening, may have been an increased supply of argillaceous material, which became available as a result of changes in the hinterland or was brought in by currents from other parts of the sea bottom. The transition from the limestones of the Basibé Formation to the slates of the Fonchanina Formation is in most places rather gradual.

The diagenetic changes (Chapter V), which gave the deposits their present aspect, are strongly dependent on the original proporties of the various sediments and hence on the environment of their deposition.

SAMENVATTING

Dit proefschrift behandelt de primaire lithologie, de omstandigheden en gebieden van afzetting en de diagenese van de sedimenten der onderdevonische Basibé Formatie in het gebied tussen de rivieren Esera en Mañanet in de Centrale Spaanse Pyreneeën.

De Basibé Formatie bestaat in dit gebied uit de volgende gesteente typen: knollig verwerende kalksteen, dolomiet, siltige tot zandige kleihoudende dolomiet, kwartsiet en kalksteen.

De variaties in dikte van het onderste "member" (lichtgrijze knollige kalksteen) en van het bovenste "member" (vrij donkergrijze kalksteen) zijn niet groot. Het middelste "member" (een afwisseling van kwartsiet en dolomiet met enkele siltige tot zandige kleihoudende dolomiet-lagen) is wigvormig: de dikte neemt af van 100 m (waarvan 50 m kwartsiet) in het westen, tot 0 m in het oosten, binnen een afstand van ongeveer 35 km. De noord-zuid uitbreiding van

dit "member" is ongeveer 6 km. In de omringende gebieden bestaat de Basibé Formatie slechts uit kalksteen, in de oostelijke gebieden uit equivalenten van het onderste en bovenste "member".

De datering van de afzettingen is moeilijk, determineerbare fossielen zijn schaars, mede door de sterke tektonische deformatie. Conodonten uit kalksteenmonsters van het bovenste "member" geven een bovenste Onder-Devoon (Emsien) ouderdom aan. Een vergelijking met gesteenten van gelijke ouderdom uit de omringende gebieden in de Pyreneeën, toont aan dat toen in het oosten en noorden kalksedimentatie plaatsvond, terwijl in het westen meer detritische sedimenten werden afgezet.

De afzettingen van de Basibé Formatie in het bestudeerde gebied zijn zeer goed gelaagd; de laagdiktes variëren meestal van 10 tot 50 cm, in de kwartsieten soms tot 2,5 m. Sedimentaire strukturen zijn zeldzaam, slechts in de kwartsieten en kalkstenen komen parallelle laminatie en soms scheve gelaagdheid voor. In de dolomieten treden loodrecht op de gelaagdheid talrijke kwartsaders op.

Door petrografisch onderzoek is getracht de oorspronkelijke eigenschappen der verschillende sedimenten te onderscheiden van de fysische en chemische veranderingen (diagenese) die deze gesteenten na afzetting hebben ondergaan.

De knollige kalksteen is ontstaan uit fossielhoudend kalkslib, dat lithificatie onderging en waarin laatdiagenetisch door drukoplossing stylolieten en banden van stylolieten ontstonden.

Dolomieten uit het middelste "member" zijn ontstaan door dolomitisatie van knollige kalken. Stylolieten, fossielen en verschillen in kristalgrootte wijzen hierop. Uit het patroon van elkaar kruisende aders kan worden afgeleid dat kwarts in deze dolomieten indrong nà dolomitisatie welke volgde op stylolitisatie. De vervanging van de kalk door dolomiet wordt verondersteld te hebben plaatsgevonden door reaktie met Mg-rijk interstitieel water dat circuleerde in de kalken, welke in de direkte nabijheid van de kwarts-zandstenen voorkomen. Dit Mg-rijke water was tijdens de cementatie uit de kwarts-zandstenen verdreven.

Kalken van het bovenste "member" zijn ontstaan uit kalkslib en bestaan nu overwegend uit fijnkristallijne calciet. Enkele lagen van gedolomitiseerde pelmicrieten komen voor. Ook de onderste lagen van dit "member" zijn gedolomitiseerd. Stylolieten komen in deze kalken algemeen voor en zijn laat-diagenetisch van oorsprong.

De kalken van de Basibé Formatie noordelijk van de Alpiene overschuivings- en breukzone en nabij het kontakt met de Maladeta granodioriet-batholiet zijn gemetamorfoseerd.

Siltige en zandige kleihoudende dolomiet is een gedolomitiseerd kalkig sediment waarin ook mica-blaadjes en zware mineralen voorkomen.

Kwartsieten zijn ontstaan uit zeer zuivere, goed afgeronde en gesorteerde kwartszanden waarin een sterke kwartscementatie is opgetreden. Naast kwarts komen slechts de zware mineralen toermalijn, rutiel en zirkoon voor. Dit rijpe mineraalgezelschap moet afkomstig zijn van een laagland waar sterke verwering is opgetreden en/of van een ouder sediment. Na afzetting en bedekking ondergingen deze kwartszanden drukoplossing op de korrelkontakten terwijl daarnaast korrelaangroei en cementatie door kwarts plaats vond. De drukoplossing is mede beïnvloed door tektonische druk en door verhoging van temperatuur ten gevolge van de intrusie van de Maladeta batholiet, en leidde tot verschillende soorten korrelkontakten. Vervanging van kwartskorrels en secundaire kwarts door dolomiet is een belangrijk proces in deze kwartsieten.

Sedimentaire strukturen zijn schaars in de sedimenten van de Basibé Formatie en de interpretatie van de omstandigheden en gebieden van afzetting is daarom grotendeels op de petrografische kenmerken en de verspreiding gebaseerd. Een vergelijking met equivalente fossiele en recente sedimenten en afzettingspatronen droeg hiertoe bij.

Het kalkslib van het onderste "member" moet zijn afgezet in een gebied met rustig water, met weinig verstoring door golven en stromen. Mariene fossielen wijzen op een marien gebied van afzetting. Het sediment was goed doorlucht. Het klimaat was warm. In dit gebied werden kwartszanden door kuststromingen aangevoerd en door golf- en stroomwerking als banken of strandwallen afgezet. Deze kwartszand-accumulaties bleven vrij stationair in het bestudeerde gebied, gezien de opeenstapeling van kwartsietlagen in het centrale en westelijke deel. Migratie binnen het gebied zal zeker hebben plaatsgevonden. Hierdoor ontstonden de plaatvormige zandlichamen.

Uit de zelden optredende scheve gelaagdheid kan geen aanvoerrichting noch een zeewaartse of landwaartse kant worden afgeleid. De meeste kwartsieten hebben een homogeen uiterlijk, vermoedelijk een gevolg van omwoeling door organismen. Enkele lagen met graafsporen zijn waargenomen. Eolische afzettingen zijn waarschijnlijk niet aanwezig.

Aan de lijzijde van de banken of kustwallen bestond een energetisch rustiger gebied. Hier werden de siltige tot zandige, kleihoudende kalksedimenten afgezet. Fossielen en het ontbreken van continentale en evaporitische afzettingen wijzen erop dat dit een open marien afzettingsgebied was. De kwartszanden van de banken of strandwallen zijn vaak over deze afzettingen gemigreerd en daarop neergelegd.

Voortgaande daling van het gebied had bedekking van deze sedimenten door de kalk van het bovenste "member" tot gevolg. Deze sedimentatie van kalkslib vond ook weer in een ondiep marien milieu plaats en werd mogelijk door beëindiging van de kwartszandaanvoer.

Kalkafzetting kwam tot een eind toen meer kleiig materiaal in het gebied werd aangevoerd.

REFERENCES

Adams, W. L., 1964. Diagenetic aspects of Lower Morrowan, Pennsylvanian, Sandstones, northwestern Oklahoma. Am. Assoc. Petroleum Geologists Bull., 48, p. 1568—1580.

Adams, J. E. & Rhodes, M. L., 1960. Dolomitization by seepage refluxion. Am. Assoc. Petroleum Geologists Bull., 44, p. 1912—1920.

Allen, J. R. L., 1964. Sedimentation in the modern delta of the river Niger, West Africa. In: Deltaic and shallow marine deposits (Straaten, L. M. J. U. van, editor). Developments in Sedim., 1, Elsevier Publ. Co., Amsterdam, 464 p.; p. 26—34.

—, 1965 a. Coastal geomorphology of eastern Nigeria: beach-ridge barrier islands and vegetated tidal flats. Geol. Mijnb., 44, p. 1—21.

—, 1965 b. Late Quaternary Niger delta, and adjacent areas: sedimentary environments and lithofacies. Am. Assoc. Petroleum Geologists Bull., 49, p. 547—600.

—, 1967. Depth indicators of clastic sequences. Marine Geol., 5, p. 429—446.

Almela, A. & Rios, J. M., 1947. Explicacion al mapa geológico de la provincia de Lérida 1 : 200.000. Madrid, Rev. Inst. geol. minero España, 193 p.

American Commission on Stratigraphic Nomenclature, 1961. Code of stratigraphic nomenclature. Am. Assoc. Petroleum Geologists Bull., 45, p. 645—660.

Andrews, P. B. & van der Lingen, G. J., 1969. Environmentally significant sedimentologic characteristics of beach sands. N. Z. Jl. Geol. Geophys., 12, p. 119—137.

Atwood, D. K. & Bubb, J. N., 1970. Distribution of dolomite in a tidal flat environment: Sugarloaf Key, Florida. Journ. Geol., 78, p. 499—505.

Avias, J., Joseph, C. & Dir. Expl. ERAP., 1967. Le Dévonien de la Montagne Noire (France). Int. Symp. Devon. Syst. Calgary, 1967, 1, p. 21—26.

Ball, M. M., 1967. Carbonate sand bodies of Florida and the Bahamas. Journ. Sed. Petrol., 37, p. 556—591.

Bathurst, R. G. C., 1964. Diagenesis and paleoecology: A survey. In: Approaches to paleoecology (Imbrie, J. & Newell, N., editors), John Wiley and Sons, New York, p. 319—344.

Beales, F. W., 1965. Diagenesis in pelletted limestones. In: Dolomitization and limestone diagenesis (Pray, L. C. & Murray, R. C., editors). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 13, p. 49—70.

Bernard, H. A. & LeBlanc, R. J., 1965. Résumé of the Quaternary geology of the Northwestern Gulf of Mexico province. In: The Quaternary of the United States (Wright, H. E. jr. & Frey, D. G., editors). Princeton University Press, Princeton, p. 137—185.

Bird, E. C. F., 1967. Coastal lagoons of southeastern Australia. In: Landform studies from Australia and New Guinea. (Jennings, J. N. & Mabbutt, J. A., editors). Cambridge University Press, 434 p.; p. 365—385.

Bissell, H. J. & Chilingar, G. V., 1967. Classification of sedimentary carbonate rocks. In: Carbonate rocks, Origin, occurrence and classification (Chilingar, G. V., Bissell, H. J. & Fairbridge, R. W., editors). Developments in Sedim., 9A, Elsevier Publ. Co., Amsterdam, 471 p.; p. 87—168.

Blatt, H., 1967. Original characteristics of clastic quartz grains. Journ. Sed. Petrol., 37, p. 401—424.

Blatt, H. & Christie, J. M., 1963. Undulatory extinction in quartz of igneous and metamorphic rocks and its significance in provenance studies of sedimentary rocks. Journ. Sed. Petrol., 33, p. 559—579.

Borg, I., Friedman, M., Handin, J. & Higgs, D. V., 1960. Experimental deformation of St. Peter sand: A study of cataclastic flow. In: Rock deformation (Griggs, D. & Handin, H., editors). Geol. Soc. Am., Mem. 79, p. 133—191.

Boschma, D. 1963. Successive Hercynian structures in some areas of the Central Pyrenees. Leidse Geol. Med., 28, p. 103—176.

Boyd, D. R. & Dyer, B. F., 1966. Frio barrier bar system of south Texas. Am. Assoc. Petroleum Geologists Bull., 50, p. 170—178.

Bredehoeft, J. D., Blyth, C. R., White, W. A. & Maxey, G. B., 1963. Possible mechanism for concentration of brines in subsurface formations. Am. Assoc. Petroleum Geologists Bull., 47, p. 257—269.

Bresson, A., 1903. Etudes sur les formations anciennes des Hautes et Basses Pyrénées (Haute Chaîne). Bull. Serv. Carte géol. Fr., 14/93, 278 p.

Carozzi, A. V., 1960. Microscopic sedimentary petrography. John Wiley & Sons, Inc., New York, London, 485 p.

Cavet, P. 1957. Le Paléozoïque de la zone axiale des Pyrénées orientales françaises entre le Roussillon et l'Andorre. Bull. Serv. Carte géol. Fr., 254/55, 216 p.

Cayeux, L., 1935. Les roches sédimentaires de France. Roches carbonatées (calcaire et dolomies). Masson et Cie., Paris, 447 p.

Charlet, J. M., 1968. Etude préliminaire du massif granitique de la Maladeta (Pyrénées centrales espagnoles). Ann. Soc. géol. Nord, 88, p. 65—75.

Chilingar, G. V., Bissell, H. J. & Wolf, K. H., 1967. Diagenesis of carbonate rocks. In: Diagenesis in sediments (Larsen, G. & Chilingar, G. V., editors). Developments in Sedim., 8, Elsevier Publ. Co., Amsterdam, 551 p.; p. 179—322.

Clin, M., 1964. Etude géologique de la haute chaîne des Pyrénées centrales entre le cirque de Troumouse et le cirque du Lys. Thesis, Nancy, 1959. Mém. B. R. G. M. 1964, 27, 379 p.

Cloud, P. E., 1962. Environment of calcium carbonate deposition west of Andros Island, Bahamas. U.S. Geol. Surv. Prof. Paper 350, 138 p.

Conolly, J. R., 1965. The occurrence of polycrystallinity and undulatory extinction in quartz in sandstones. Journ. Sed. Petr., 35, p. 116—135.

Curray, J. R., 1960. Sediments and history of Holocene transgression, continental shelf, northwest Gulf of Mexico. In: Recent sediments, northwest Gulf of Mexico (Shepard, F. P., Phleger, F. B. & Andel, T. H. van, editors). Am. Assoc. Petroleum Geologists, Tulsa, 394 p.; p. 221—266.

Curray, J. R. & Moore, D. G., 1964. Holocene regressive littoral sand, Costa de Nayarit, Mexico. In: Deltaic and shallow marine deposits (Straaten, L. M. J. U. van, editor). Developments in Sedim., 1, Elsevier Publ. Co., Amsterdam, 464 p.; p. 76—82.

Dalloni, M., 1910. Etude géologique des Pyrénées de l'Aragon. Thesis Univ. Paris, 444 p. Marseille.

—, 1913. Stratigraphie et tectonique de la région des Nogueras (Pyrénées centrales). Bull. Soc. géol. France, 4/13, p. 243—263.

—, 1930. Etude géologique des Pyrénées catalanes, Ann. Fac. Sc. Marseille, 26/3, 373 p.

Damestoy, G., 1961. Etude de la série dévonienne de la vallée de Quintua (Basses-Pyrénées). C. R. Soc. géol. Fr., 1961/3, p. 75—77.

Dapples, E. C., 1959. The behavior of silica in diagenesis. In: Silica in sediments (Ireland, H. A. editor). Soc. Econ.

Paleontologists Mineralogists, Spec. Publ., 7, p. 36—54.
—, 1962. Stages of diagenesis in the development of sandstones. Geol. Soc. Am. Bull., 73, p. 913—933.
—, 1967a. Diagenesis in sandstones. In: Diagenesis in sediments (Larsen, G. & Chilingar, G. V., editors). Developments in Sedim., 8. Elsevier Publ. Co., Amsterdam, 551 p.; p. 91—125.
—, 1967b. Silica as an agent in diagenesis. In: Diagenesis in sediments (Larsen, G. & Chilingar, G. V., editors), Developments in Sedim., 8, Elsevier Publ. Co., Amsterdam, 551 p.; p. 323—342.
Davies, D. K., 1969. Shelf sedimentation: An example from the Jurassic of Britain. Journ. Sed. Petrol., 39, p. 1344—1370.
Deffeyes, K. S., Lucia, F. J., & Weyl, P. K., 1965. Dolomitization of Recent and Plio-Pleistocene sediments by marine evaporite waters on Bonaire, Netherlands Antilles. In: Dolomitization and limestone diagenesis (Pray, L. C. & Murray, R. C., editors). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 13, p. 71—88.
De Groot, K., 1969. The chemistry of submarine cement formation at Dohat Hussain in the Persian Gulf. Sedimentology, 12, p. 63—68.
Destombes, J. P., 1953. Stratigraphie des terrains primaires de la Haute-Garonne. C. R. 19e Sess. Congr. géol. intern., Alger 1952, Sect. 2, fasc. 2, p. 107—129.
—, 1959. Position stratigraphique des schistes à Trilobites de Cathervielle (Haute-Garonne). Bull. Soc. Et. Sc. Angers, N. S., 2/89, p. 111—113.
Dickson, J. A. D., 1966. Carbonate identification and genesis as revealed by staining. Journ. Sed. Petrol., 36, p. 491—505.
Dillon, W. P., 1970. Submergence effects on a Rhode Island barrier and lagoon and interferences on migration of barriers. Journ. Geol., 78, p. 94—106.
Dooge, J., 1966. The stratigraphy of an Upper Devonian carbonate-shale transition between the north and south Ram Rivers of the Canadian Rocky Mountains. Leidse Geol. Med., 39, p. 1—53.
Dunham, R. J., 1962. Classification of carbonate rocks according to depositional texture. In: Classification of carbonate rocks (Ham, W. E., editor). Am. Assoc. Petroleum Geologists, Mem. 1, p. 108—121.
Eden, J. G. van, 1970. A reconnaissance of deltaic environment in the Middle Eocene of the south-Central Pyrenees. Geol. Mijnb., 49, p. 145—157.
Evamy, B. D., 1969. The precipitational environment and correlation of some calcite cements deduced from artificial staining. Journ. Sed. Petrol., 39, p. 787—793.
Fairbridge, R. W., 1967. Phases of diagenesis and authigenesis. In: Diagenesis in sediments (Larsen, G. & Chilingar, G. V., editors). Developments in Sedim., 8, Elsevier Publ. Co., Amsterdam, 551 p.; p. 19—89.
Fisher, W. L. & Rodda, P. U., 1969. Edwards Formation (Lower Cretaceous), Texas: Dolomitization in a carbonate platform system. Am. Assoc. Petroleum Geologists Bull., 53, p. 55—72.
Fisk, H. N., 1959. Padre Island and the Laguna Madre flats, coastal south Texas. In: Second coastal geography conference (Russell, R. J., editor). Coastal Stud. Inst. Louisiana State Univ., 472 p.; p. 103—152.
Folk, R. L., 1959. Practical petrographic classification of limestones. Am. Assoc. Petroleum Geologists Bull., 43, p. 1—38.
—, 1962. Spectral subdivision of limestone types. In: Classification of carbonate. rocks (Ham, W. E., editor). Am. Assoc. Petroleum Geologists, Mem. 1, p. 62—84.
—, 1965. Some aspects of recrystallization in ancient limestones. In: Dolomitization and limestone diagenesis (Pray, L. C. & Murray, R. C., editors). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 13, p. 14—48.
Friedman, G. M. & Sanders, J. E., 1967. Origin and occurrence of dolostones. In: Carbonate rocks, Origin, occurrence and classification. (Chilingar, G. V., Bissell, H. J. & Fairbridge, R. W., editors). Developments in Sedim., 9A, Elsevier Publ. Co., Amsterdam, 551 p.; p. 169—265.
Füchtbauer, H. & Müller, G., 1970. Sedimente und Sedimentgesteine. Teil II Sediment-Petrologie (Von Engelhardt, W., Füchtbauer, H. & Müller, G.). E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, 726 p.
García Saínz, L., 1940 a. Las superficies de erosión que preceden a los glaciares cuaternarios del Pirineo Central y sus recíprocas influencias. Est. Geogr. Madrid, 1, p. 45—73.
—, 1940 b. Morphologische Skizze des Zentralabschnittes der spanischen Pyrenäen in der Tertiär- und Quartärzeit. Peterm. Geogr. Mitt., 86, p. 366—371.
Garrison, R. E. & Fischer, A. G., 1969. Deep-water limestones and radiolarites of the Alpine Jurassic. In: Depositional environments in carbonate rocks (Friedman, G. M., editor). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 14, p. 20—56.
Germann, K., 1968. Diagenetic patterns in the Wettersteinkalk (Ladinian, Middle Trias), Northern Limestone Alps, Bavaria and Tyrol. Journ. Sed. Petrol., 38, p. 490—500.
Gilbert, C. M., 1954. Sedimentary rocks. In: Petrography Williams, H., Turner, F. J. & Gilbert, C. M.). W. H. Freeman and Co., San Francisco, 406 p.; p. 249—406.
Ginsburg, R. N., 1957. Early diagenesis and lithification of shallow-water carbonate sediments in South Florida. In: Regional aspects of carbonate deposition (LeBlanc, R. F. & Breeding, J. G., editors). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 5, p. 80—100.
Hartevelt, J. J. A., 1970. Geology of the Upper Segre and Valira valleys, Central Pyrenees, Andorra/Spain. Sheet 10, 1 : 50,000. Leidse Geol. Med., 45, p. 167—236.
Heald, M. T., 1955. Stylolites in sandstones. Journ. Geol., 63, p. 101—114.
—, 1956. Cementation of Simpson and St. Peter Sandstones in parts of Oklahoma, Arkansas and Missouri. Journ. Geol., 64, p. 16—30.
—, 1959. Significance of stylolites in permeable sandstones. Journ. Sed. Petrol., 29, p. 251—253.
Heddebaut, C., 1965. Recherches stratigraphiques et paléontologiques dans le massif des Aldudes (Basses-Pyrénées). Bull. Soc. géol. Fr., 7/7, p. 631—639.
—, 1966. Données nouvelles sur le Silurien et le Dévonien des Pyrénées Basques. Ann. Soc. géol. Nord, 86, p. 197—200.
Hollenshead, C. T. & Pritchard, R. L., 1961. Geometry of producing Mesaverde sandstones, San Juan basin. In: Geometry of sandstone bodies. (Peterson, J. A. & Osmond, J. C., editors). Am. Assoc. Petroleum Geologists, Tulsa, 240 p.; p. 98—118.
Höllermann, P., 1968. Die rezenten Gletscher der Pyrenäen. Geograph. Helv., 23, p. 157—168.
Hollmann, R., 1962 Über Subsolution und die "Knollenkalken" des Calcare Ammonitico Rosso Superiore in Monte Baldo (Malm; Norditalien). N. Jb. Geol. Paläont., Mh., 1962/4, p. 163—179.
Hoorn, B. van, 1970. Sedimentology and paleogeography of an Upper Cretaceous turbidite basin in the south-Central Pyrenees, Spain. Leidse Geol. Med., 45, p. 73—154.

Hoyt, J. H., 1969. Chenier versus barrier, genetic and stratigraphic distinction. Am. Assoc. Petroleum Geologists Bull., 53, p. 299—306.

Hubert, J. F., 1962. A zircon-tourmaline-rutile maturity index and the interdependence of the composition of heavy mineral assemblages with the gross composition and texture of sandstones. Journ. Sed. Petrol., 32, p. 440—450.

Hsü, K. J. & Siegenthaler, C., 1969. Preliminary experiments on hydrodynamic movement induced by evaporation and their bearing on the dolomite problem. Sedimentology, 12, p. 11—25.

Illing, L. V., Wells, A. J. & Taylor, J. C. M., 1965. Penecontemporary dolomite in the Persian Gulf. In: Dolomitization and limestone diagenesis (Pray, L. C. & Murray, R. C., editors). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 13, p. 89—111.

Imbrie, J. & Buchanan, H., 1965. Sedimentary structures in modern carbonate sands of the Bahamas. In: Primary sedimentary structures and their hydrodynamic interpretation (Middleton, G. V., editor). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 12, p. 149—172.

Imbrie, J. & Purdy, E. G., 1962. Classification of modern Bahamian carbonate sediments. In: Classification of carbonate rocks (Ham, W. E., editor). Am. Assoc. Petroleum Geologists, Mem. 1, 279 p.; p. 253—272.

International Subcommission on Stratigraphy and Terminology, 1961. Stratigraphic classification and terminology. 21st Intern. Geol. Congr. Copenhagen, 1960, Rept. Session, Norden, 25, p. 1—38.

Jacob, Ch., 1935. A propos du versant méridional des Pyrénées centrales. Bull. Soc. géol. Fr., 6, p. 78—80.

Jacob, Ch., Fallot, P., Astre, G. & Ciry, R., 1927. Observations tectoniques sur le versant méridional des Pyrénées centrales et orientales. C. R. 14e Congr. géol. intern., Madrid 1926, 2, p. 335—411.

Johnson, K. G. & Friedman, G. M., 1969. The Tully clastic correlatives (Upper Devonian) of New York State: A model for recognition of alluvial, dune (?), tidal, nearshore (bar and lagoon), and offshore sedimentary environments in a tectonic delta complex. Journ. Sed. Petrol., 39, p. 451—485.

Jurgan, H. 1969. Sedimentologie des Lias der Berchtesgadener Kalkalpen. Geol. Rundschau, 58, p. 464—501.

Keizer, J., 1953. La géologie de la couverture sédimentaire du Massif de l'Arize. Leidse Geol. Med., 18, p. 229—253.

Kendall, C. G. St. C. & Skipwith, Sir Patrick A. d'E., 1968. Recent algal mats of a Persian Gulf lagoon. Journ. Sed. Petrol., 38, p. 1040—1058.

King, C. A. M., 1959. Beaches and coasts. Edward Arnold Publ. Ltd., London, 403 p.

Klein, G. deVries, 1967. Paleocurrent analysis in relation to modern marine sediment dispersal patterns. Am. Assoc. Petroleum Geologists Bull., 51, p. 366—382.

Kleinsmiede, W. F. J., 1960. Geology of the Valle de Arán (Central Pyrenees). Leidse Geol. Med., 25, p. 129—245.

Krauskopf, K. B., 1959. The geochemistry of silica in sedimentary environments. In: Silica in sediments. (Ireland, H. A., editor). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 7, p. 4—19.

Krumbein, W. C. & Garrels, R. M., 1952. Origin and classification of chemical sediments in terms of pH and Eh potentials. Journ. Geol., 60, p. 1—33.

Krumbein, W. C. & Sloss, L. L., 1963. Stratigraphy and sedimentation. Freeman and Comp., San Francisco, London, 660 p.

Krynine, P. D., 1946. The tourmaline group in sediments. Journ. Geol., 54, p. 65—87.

Kuenen, Ph. H., 1959. Experimental abrasion: 3. Fluviatile action on sand. Am. Journ. Sci., 257, p. 172—190.

Kwon, H. J., 1969. Barrier islands of the northern Gulf of Mexico coast: sediment source and development. Report 75, Coastal Stud. Inst. Louisiana State Univ., 51 p.

Lagny, P., 1964. Précisions sur l'âge d'une série schisto-gréseuse dévonienne de la vallée du Laxia (Basses-Pyrénées). C. R. Soc. géol. Fr., 1964/4, p. 144—145.

Lane, D. W., 1963. Sedimentary environments in Cretaceous Dakota sandstone in northwestern Colorado. Am. Assoc. Petroleum Geologists Bull., 47, p. 229—256.

Laporte, L. F., 1967. Carbonate deposition near mean sea-level and resultant facies mosaic: Manlius Formation (Lower Devonian) of New York State. Am. Assoc. Petroleum Geologists Bull., 51, p. 73—101.

Laverdière, J. W., 1930. Contribution à l'étude des terrains paléozoïques dans les Pyrénées occidentales. Mém. Soc. Géol. Nord, 10/2, 131 p.

LeBlanc, R. J. & Hodgson, W. D., 1959. Origin and development of the Texas shoreline. In: Second coastal geography conference (Russell, R. J., editor). Coastal Stud. Inst. Louisiana State Univ., 472 p.; p. 57—102.

Leighton, M. W. & Pendexter, C., 1962. Carbonate rock types. In: Classification of carbonate rocks (Ham, W. E., editor). Am. Assoc. Petroleum Geologists, Mem. 1, 279 p.; p. 33—61.

Lingen, G. J. van der, 1960. Geology of the Spanish Pyrenees, North of Canfranc, Huesca province. Estud. geol., 16, p. 205—242.

Lith, J. G. J. van, 1965. Geology of the Spanish part of the Gavarnie nappe (Pyrenees) and its underlying sediments near Bielsa (province of Huesca). Geol. Ultraiectina, 10/1, 64 p.

Logvinenko, N. V. & Remizov, I. N., 1964. Sedimentology of beaches on the north coast of the Sea of Azov. In: Deltaic and shallow marine deposits (Straaten, L. M. J. U. van, editor). Developments in Sedim., 1. Elsevier Publ. Co., Amsterdam, 464 p.; p. 245—252.

Lotze, F., 1964. The distribution of evaporites in space and time. In: Problems in paleoclimatology (Nairn, A. E. M., editor). Interscience Publ., John Wiley & Sons, London, p. 491—509.

Masters, C. D., 1967. Use of sedimentary structures in determination of depositional environments, Mesaverde Formation, Williams Fork Mountains, Colorado. Am. Assoc. Petroleum Geologists Bull., 51, p. 2033—2043.

Maurel, M., 1965. Stratigraphie du Dévonien inferieur et moyen de la Montagne Noire. (Résumé). Coll. Dévon. inf., Rennes, 1964. Mém. B. R. G. M., 33, p. 63.

Maxwell, J. C., 1960. Experiments on compaction and cementation of sand. In: Rock deformation (Griggs, D. & Handin, J., editors). Geol. Soc. Am., Memoir 79, p. 105—132.

McKee, E. D., 1957. Primary structures in some recent sediments. Am. Assoc. Petroleum Geologists Bull., 41, p. 1704—1747.

—, 1964. Inorganic sedimentary structures. In: Approaches to paleoecology (Imbrie, J. & Newell, N., editors). John Wiley and Sons, New York, p. 275—295.

McKee, E. D. & Sterrett, T. S., 1961. Laboratory experiments on form and structure of longshore bars and beaches. In: Geometry of sandstone bodies (Peterson, J. A. & Osmond, J. C., editors). Am. Assoc. Petroleum Geologists, Tulsa, 240 p.; p. 13—28.

Mey, P. H. W., 1967 a. Evolution of the Pyrenean basins during the Late Paleozoic. Int. Symp. Devon. Syst. Calgary, 1967, 2, p. 1157—1166.

—, 1967 b. The geology of the Upper Ribagorzana and Baliera valleys, Central Pyrenees, Spain. Leidse Geol. Med., 41, p. 153—220.

—, 1968. Geology of the Upper Ribagorzana and Tor valleys, Central Pyrenees, Spain. Sheet 8, 1 : 50,000. Leidse Geol. Med., 41, p. 229—292.

Mey, P. H. W., Nagtegaal, P. J. C., Roberti, K. J. & Hartevelt, J. J. A., 1968. Lithostratigraphic subdivision of post-Hercynian deposits in the South-Central Pyrenees. Leidse Geol. Med., 41, p. 221—228.

Mirouse, R., 1962. Recherches géologiques dans la partie occidentale de la Zone primaire axiale des Pyrénées. Thesis Toulouse, 672 p. Mém. Carte géol. Fr. Paris, 1966, 451 p.

—, 1965. A propos des couches de passage du Dévonien inférieur au Dévonien moyen dans la haute vallée d'Ossau (Basses Pyrénées). (Résumé). Coll. Dévon. inf., Rennes, 1964. Mém. B. R. G. M., 33, p. 64—65.

—, 1967. Le Dévonien des Pyrénées occidentales et centrales (France). Int. Symp. Devon. Calgary, 1967, 1, p. 153—170.

Misch, P., 1934. Der Bau der Mittleren Südpyrenäen. Abh. Ges. Wiss. Göttingen, math.-phys. Kl. 3, Folge, H. 12, 168 p.

Mohr, C. G. van der Meer, 1969. The stratigraphy of the Cambrium Lancara Formation between the Luna River and the Esla River in the Cantabrian Mountains, Spain. Leidse Geol. Med., 43, p. 233—316.

Moore, D. G. & Scruton, P. C., 1957. Minor internal structures of some recent unconsolidated sediments. Am. Assoc. Petroleum Geologists Bull., 41, p. 2723—2751.

Müller, G. M., 1964. Methoden der Sediment-Untersuchung. Sediment-Petrologie (Von Engelhardt, W., Füchtbauer, H. & Müller, G.) Teil I. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, 303 p.

Murray, R. C., 1960. Origin of porosity in carbonate rocks. Journ. Sed. Petrol., 30, p. 59—84.

—, 1969. Hydrology of South Bonaire, N. A.-A rock selective dolomitization model. Journ. Sed. Petrol., 39, p. 1007—1013.

Murray, R. C. & Lucia, F. J., 1967. Cause and control of dolomite distribution by rock selectivity. Geol. Soc. Am. Bull., 78, p. 21—35.

Nagtegaal, P. J. C., 1962. Geologie van de Nogueras Zone. Internal Report, Geol. Min. Inst., Univ., Leiden.

—, 1966. Scour-and-fill structures from a fluvial piedmont environment. Geol. Mijnb., 45, p. 342—354.

—, 1969. Sedimentology, paleoclimatology, and diagenesis of post-Hercynian continental deposits in the south-Central Pyrenees, Spain. Leidse Geol. Med., 42, p. 143—238.

—, 1971. On the origin of griotte limestones. Leidse Geol. Med. in preparation.

Newell, N. D. & Rigby, J. K., 1957. Geological studies on the Great Bahama Bank. In: Regional aspects of carbonate deposition (LeBlanc, R. J. & Breeding, J. G., editors). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 5, p. 15—79.

Nicolaî, A., 1963. Contribution à l'étude géologique du bassin du Gave de Pau entre le granite de Cauterets et la Zone nord-pyrénéenne. Thesis, Bordeaux, 122 p.

Nussbaum, F., 1935. Die Seeen der Pyrenäen. Eine geographische Studie. Mitt. Naturf. Ges. Bern, 1934, p. 1—184.

—, 1938. Die eiszeitliche Schneegrenze in den Pyrenäen. Verh. 3. intern. Quartärkonf., Wien, 1936, p. 198—200.

Oele, E., 1964. Sedimentological aspects of four Lower-Paleozoic Formations in the northern part of the province of Léon (Spain). Leidse Geol. Med., 30, p. 1—99.

Oomkens, E., 1967. Depositional sequences and sand distribution in a deltaic complex. Geol. Mijnb., 46, p. 265—278.

Osgood, R. G., 1970. Trace fossils of the Cincinnati area. Palaeontographica Americana, 6 (41), p. 281—444.

Ovtracht, A., 1967. Le Dévonien du domaine Nord-Pyrénéen Oriental. Int. Symp. Devon. Syst. Calgary, 1967, 1, p. 27—35.

Park, W. C. & Schot, E. H., 1968. Stylolites: Their nature and origin. Journ. Sed. Petrol., 38, p. 175—191.

Pélissonnier, H., 1958. Observations géologiques sur la partie NW de la feuille de Luchon au 50.000e. Bull. Carte géol. Fr., 255/56, 1958, p. 181—187.

—, 1959. Dolomitisation de la série dévono-dinantienne dans les Pyrénées de la Haute-Garonne et de l'Ariège. Bull. Soc. géol. Fr., 7/1, p. 625—630.

Perseil, E.-A., & Tourenq, C., 1963. Nouvelles données sur le Paléozoïque du massif de l'Arize (Ariège). Bull. Soc. géol. Fr., 7/5, p. 835—837.

Pettijohn, F. J., 1957. Sedimentary rocks. 2nd ed. Harper & Brothers, New York, 718 p.

Pettijohn, F. J., Potter, P. E. & Siever, R., 1965. Geology of sand and sandstone. 2nd pr. 1966, Indiana Geol. Surv. & Dept. Geol., Indiana Univ., Bloomington, 208 p.

Phleger, F. B. & Ewing, G. C., 1962. Sedimentology and oceanography of coastal lagoons in Baja California, Mexico. Geol. Soc. Am. Bull., 73, p. 145—181.

Plumley, W. J., Risley, R. W., Graves, R. W. & Kaley, M. E., 1962. Energy index for limestone interpretation and classification. In: Classification of carbonate rocks (Ham, W. E., editor). Am Assoc. Petroleum Geologists., Mem. 1, p. 85—107.

Potter, P. E., 1967. Sand bodies and sedimentary environments: a review. Am. Assoc. Petroleum Geologists Bull., 51, p. 337—365.

Purdy, E. G., 1963. Recent calcium carbonate facies of the Great Bahama Bank. 2. Sedimentary facies. Journ. Geol., 71, p. 472—497.

Reineck, H. E., 1967. Layered sediments of tidal flats, beaches, and shelf bottoms of the North Sea. In: Estuaries (Lauff, G. H., editor). Am. Assoc. Adv. Sci., Publ. 83, p. 191—206.

Renton, J. J., Heald, M. T. & Cecil, C. B., 1969. Experimental investigation of pressure solution of quartz. Journ. Sed. Petrol., 39, p. 1107—1117.

Rijckborst, H., 1967. Hydrology of the Upper-Garonne basin (Valle de Arán, Spain). Leidse Geol. Med., 40, p. 1—74.

Roberti, K. J., 1970. Hercynian gravity folding in the area of the Flamisell and Mañanet valleys, south Central Pyrenees, province of Lérida, Spain. Leidse Geol. Med., in press.

—, 1971. Geological map of the Pallaresa and Flamisell valleys, south Central Pyrenees, province of Lérida, Spain. Sheet 9, 1 : 50,000. Leidse Geol. Med., in preparation.

Rock-color chart, 1951. 2nd. ed. Geol. Soc. of America, New York.

Roehl, P. O., 1967. Stony Mountain (Ordovician) and Interlake (Silurian) facies analogs of recent low-energy marine and subaerial carbonates, Bahamas. Am. Assoc. Petroleum Geologists Bull., 51, p. 1979—2032.

Roger, Ph., 1965. Etude stratigraphique et structurale de la zone de Nogueras entre l'Esera et l'Isabena (Huesca-Espagne). Act. Soc. Linnéene de Bordeaux, 102, p. 1—25.

Rusnak, G. A., 1960. Sediments of Laguna Madre. In: Recent sediments, northwest Gulf of Mexico. (Shepard, F. P., Phleger, F. B. & Andel, T. H. van, editors). Am. Assoc. Petroleum Geologists, Tulsa, 394 p.; p. 153—196.

Sabins, F. F., 1963. Anatomy of stratigraphic trap, Bisti field, New Mexico. Am. Assoc. Petroleum Geologists Bull., 47, p. 193—228.

Sanders, J. E. & Friedman, G. M., 1967. Origin and occurrence of limestones. In: Carbonate rocks, Origin, occurrence and classification. (Chilingar, G. V., Bissell, H. J. & Fairbridge, R. W., editors). Developments in Sedim., 9A. Elsevier Publ. Co., Amsterdam, 471 p.; p. 169—265.

Scherp, A., 1963. Die Petrographie der paläozoischen Sandsteine in der Bohrung Münsterland 1 und ihre Diagenese in Abhängigkeit von der Teufe. Fortschr. Geol. Rheinl. Westf., 11, p. 251—282.

Schmidt, H., 1931. Das Paläozoikum der spanischen Pyrenäen. Abh. Ges. Wiss. Göttingen, math.-phys. Kl. 3. Folge, H. 5, 85 p.

Schmidt, V., 1965. Facies, diagenesis, and related reservoir properties in the Gigas Beds (Upper Jurassic), Northwestern Germany. In: Dolomitization and limestone diagenesis (Pray, L. C. & Murray, R. C., editors). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 13, p. 124—168.

Schwarz, E. J., 1962. Geology and paleomagnetism of the valley of the Rio Aragón Subordan north and east of Oza. Spanish Pyrenees, province of Huesca. Estud. geol., 18, p. 193—240.

Sears, J. D., Hunt, C. B., & Hendricks, T. A., 1941. Transgressive and regressive Cretaceous deposits in southern San Juan basin, New Mexico. U. S. Geol. Surv. Prof. Paper 193-F, p. 101—121.

Seguret, M., 1964. Sur le style en tête plongéante des structures pyrénéennes de la zone de Nogueras (versant sud des Pyrénées centrales). C. R. Ac. Sc., 259, p. 2895—2898.

—, 1966. Sur le charriages de la zone des Nogueras (versant sud des Pyrénées). C. R. Soc. géol. Fr., 1966/1, p. 17—18.

Sharma, G. D., 1965. Formation of silica cement and its replacement by carbonates. Journ. Sed. Petrol., 35, p. 733—745.

Shelton, J. W., 1967. Stratigraphic models and general criteria for recognition of alluvial, barrier-bar, and turbidity-current sand deposits. Am. Assoc. Petroleum Geologists Bull., 51, p. 2441—2461.

Shepard, F. P., 1960. Gulf coast barriers. In: Recent sediments, northwest Gulf of Mexico. (Shepard, F. P., Phleger, F. B. & Andel, T. H. van, editors). Am. Assoc. Petroleum Geologists, Tulsa, 394 p.; p. 197—220.

—, 1963. Submarine geology. 2nd. ed. Harper & Row, Publ., New York, 557 p.

Shepard, F. P. & Moore, D. G., 1955. Central Texas coast sedimentation: characteristics of sedimentary environment, recent history, and diagenesis. Am. Assoc. Petroleum Geologists Bull., 39, p. 1463—1593.

Shinn, E. A., 1968. Practical significance of birdseye structures in carbonate rocks. Journ. Sed. petrol., 38, p. 215—223.

—, 1969. Submarine lithification of Holocene carbonate sediments in the Persian Gulf. Sedimentology, 12, p. 109—144.

Shinn, E. A., Ginsburg, R. N. & Lloyd, R. M., 1965. Recent supratidal dolomite from Andros Island, Bahamas. In: Dolomitization and limestone diagenesis (Pray, L. C. & Murray, R. C., editors). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 13, p. 112—123.

Shinn, E. A., Lloyd, R. M. & Ginsburg, R. N., 1969. Anatomy of a modern carbonate tidal-flat, Andros Island, Bahamas. Journ. Sed. Petrol., 39, p. 1202—1228.

Shirley, J., 1964. The distribution of Lower Devonian faunas. In: Problems in palaeoclimatology (Nairn, A. E. M., editor). Interscience Publ., John Wiley & Sons Ltd., London, p. 255—261.

Siever, R., 1959. Petrology and geochemistry of silica cementation in some Pennsylvanian sandstones. In: Silica in sediments (Ireland, H. A., editor). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 7, p. 55—79.

—, 1962. Silica solubility, 0°—200°C, and the diagenesis of siliceous sediments. Journ. Geol., 70, p. 127—150.

Siever, R., Beck, K. C. & Berner, R. A., 1965. Composition of interstitial waters of modern sediments. Journ. Geol., 73, p. 39—73.

Simpson, S., 1957. On the trace-fossil Chondrites. Quart. Journ. Geol. Soc. London, 112, p. 475—499.

Sippel, R. F., 1968. Sandstone petrology, evidence from luminescence petrography. Journ. Sed. Petrol., 38, p. 530—554.

Sitter, L. U. de, 1964. Structural geology. McGraw Hill, London, New York, 551 p.

—, 1965. Hercynian and alpine orogenies in northern Spain. Geol. Mijnb. 44, p. 373—383.

Sitter, L. U. de & Zwart, H. J., 1958. Geological map of the Paleozoic of the Central Pyrenees, sheet 3, Ariège, France. 1 : 50,000. Leidse Geol. Med., 22, p. 351—418.

— & —, 1962. Geological map of the Paleozoic of the Central Pyrenees, sheets 1 Garonne, 2 Salat France. 1 : 50,000. Leidse Geol. Med., 27, p. 191—236.

Skinner, H. C. W., 1963. Precipitation of calcian dolomites and magnesian calcites in the southeast of South Australia. Am. Journ. Sci., 261, p. 449—472.

Soliman, S. M., 1964. Primary structures in a part of the Nile delta sand beach. In: Deltaic and shallow marine deposits. (Straaten, L. M. J. U. van, editor), Developments in Sedim., 1. Elsevier Publ. Co., Amsterdam, 464 p.; p. 379—387.

Souquet, P., 1967. Le Crétacé supérieur sud-pyrénéen en Catalogne, Aragon et Navarre. Thesis, Fac. Sci., Toulouse, 529 p.

Straaten, L. M. J. U. van, 1959 a. Minor structures of some recent littoral and neritic sediments. Geol. Mijnb., 21, p. 197—216.

—, 1959 b. Littoral and submarine morphology of the Rhône delta. In: Second coastal geography conference. (Russell, R. J., editor). Coastal Stud. Inst. Louisiana State Univ., 472 p.; p. 233—264.

—, 1965. Coastal barrier deposits in South- and North-Holland, in particular in the areas around Scheveningen and IJmuiden. Med. Geol. Sticht. Nw. Serie 17, p. 41—76.

Taylor, J. M., 1950. Pore-space reduction in sandstones. Am. Assoc. Petroleum Geologists Bull., 34, p. 701—716.

Taylor, J. C. M. & Illing, L. V., 1969. Holocene intertidal calcium carbonate cementation, Qatar, Persian Gulf. Sedimentology, 12, p. 69—107.

Thomas, W. A. & Mann, C. J., 1966. Late Jurassic depositional environments, Louisiana and Arkansas. Am. Assoc. Petroleum Geologists Bull., 50, p. 178—182.

Thompson, W. O., 1937. Original structures of beaches, bars, and dunes. Geol. Soc. Am. Bull., 48, p. 723—752.

—, 1949. Lyons sandstone of Colorado Front Range. Am. Assoc. Petroleum Geologists Bull., 33, p. 52—72.

Thomson, A., 1959. Pressure solution and porosity. In: Silica in sediments (Ireland, H. A., editor). Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 7, p. 92—110.

Trurnit, P., 1968. Pressure solution phenomena in detrital rocks. Sedimentary Geol., 2, p. 89—114.

Usdowski, H. E., 1967. Die Genese von Dolomit in Sedimenten. Mineralogie und Petrographie in Einzeldarstellungen, 4, Springer-Verlag, Berlin, 95 p.

—, 1968. The formation of dolomite in sediments. In: Recent developments in carbonate sedimentology in Central Europe. A. Processes of carbonate formation and diagenesis. (Müller, G. & Friedman, G. M., editors). Springer Verlag, Berlin, 255 p.; p. 21—32.

Visher, G. S., 1965. Use of vertical profile in environmental reconstruction. Am. Assoc. Petroleum Geologists Bull., 49, p. 41—61.

—, 1969 a. How to distinguish barrier bar and channel sands. World Oil, 168/6, p. 106—113.

—, 1969 b. Grain size distributions and depositional processes. Journ. Sed. Petrol., 39, p. 1074—1106.

Von Engelhardt, W., 1967. Interstitial solutions and diagenesis in sediments. In: Diagenesis in sediments (Larsen, G. & Chilingar, G. V., editors). Developments in Sedim., 8. Elsevier Publ. Co., Amsterdam, 551 p.; p. 503—521.

Waldschmidt, W. A., 1941. Cementing materials in sandstones and their influence on migration of oil. Am. Assoc. Petroleum Geologists Bull., 25, p. 1839—1879.

Walker, T. R., 1960. Carbonate replacement of detrital crystalline silicate minerals as a source of authigenic silica in sedimentary rocks. Geol. Soc. Am. Bull., 71, p. 145—151.

—, 1962. Reversible nature of chert-carbonate replacement in sedimentary rocks. Geol. Soc. Am. Bull., 73, p. 237—241.

Waterlot, M., 1965. Sur quelques caractères du Dévonien de la région de Benasque (Province de Huesca), Espagne. Ann. Soc. géol. Nord, 85, p. 247—250.

—, 1969. Contribution à l'étude géologique du Carbonifère Ant-Stéphanien des Pyrénées centrales espagnoles. Mem. Inst. Geol. Min. España, 70, 259 p. Thesis, Lille, 1967, 380 p.

Weber, P., 1965. Bildung und Regelung von Kalkknollengefügen. Untersuchungen im Oberdevon des Rheinischen Schiefergebirges. Decheniana, 118, p. 55—84.

Weimer, R. J., 1961. Spatial dimensions of Upper Cretaceous sandstones, Rocky Mountain area. In: Geometry of sandstone bodies (Peterson, J. A. & Osmond, J. C., editors). Am. Assoc. Petroleum Geologists, Tulsa, 240 p.; p. 82—97.

Wennekers, J. H. N., 1968. The geology of the Esera valley and the Lys-Caillauas massif, Central Pyrenees. Sheet 7, 1 : 50,000. Thesis, Univ., Leiden, 46 p.

Wensink, H., 1962. Paleozoic of the Upper Gállego and Ara valleys, Huesca province, Spanish Pyrenees. Estud. geol., 18, p. 1—74.

Werner, F., 1963. Über den inneren Aufbau von Strandwällen an einem Küstenabschnitt der Eckernförder Bucht. Meyniana, 13, p. 108—121.

Zandvliet, J., 1960. The geology of the Upper Salat and Pallaresa valleys, Central Pyrenees, France/Spain. Leidse Geol. Med. 25, p. 1—127.

Zenkovich, V. P., 1967. Processes of coastal development. (Steers, J. A., editor). Oliver & Boyd, Edinburgh and London, 738 p.

Ziegler, W., 1959. Conodonten aus Devon and Karbon Südwesteuropas und Bemerkungen zur bretonischen Faltung. N. Jahr. Geol. Paläont.,Mh., p. 289—309.

Zwart, H. J., 1953. La géologie du massif du Saint-Barthélemy (Pyrénées, France). Leidse Geol. Med., 18, p. 1—228.

—, 1960. Relations between folding and metamorphism in the Central Pyrenees, and their chronological succession. Geol. Mijnb., 22/39, p. 163—180.

—, 1964. The structural evolution of the Paleozoic of the Pyrenees. Geol. Rundschau, 53, p. 170—205.

—, 1965. Geological map of the Paleozoic of the Central Pyrenees, sheet 6, Aston, France, Andorra, Spain. 1 : 50,000. Leidse Geol. Med., 33, p. 191—254.

Zwart, H. J. & Mey, P. H. W., 1965. Quelques remarques sur la schistosité et la zone des Nogueras dans les Pyrénées. C. R. Soc. géol. Fr., 1965/3, p. 107—108.