# GEOLOGICA ULTRAIECTINA

Mededelingen van
het Mineralogisch-Geologisch Instituut der Rijksuniversiteit
te Utrecht

No. 10


## GEOLOGY OF THE SPANISH PART OF THE GAVARNIE NAPPE (Pyrenees) AND ITS UNDERLYING SEDIMENTS NEAR BIELSA (Province of Huesca)


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## GEOLOGY OF THE UPPER CRETACEOUS AND PART OF THE LOWER TERTIARY BETWEEN THE RIO ARAGON SUBORDAN AND THE RIO GALLEGO (SPANISH PYRENEES, PROVINCE OF HUESCA)


G.F.J. JEURISSEN

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CONTENTS

# SUMMARY

A description is given of the stratigraphy and the tectonics of the Spanish part of the Gavarnie nappe and and its underlying sediments near Bielsa. The outlines of the geological map coincide approximately with the frontier of the municipality of Bielsa. This study forms part of geological investigations carried out in the Spanish Western Pyrenees by members of the Geological Institute of the State University of Utrecht, Holland, under the direction of Prof. M.G. Rutten.

Not much attention could be paid to the geomorphology of the region.

The stratigraphy consists of the following series:

1. Ordovician. The Ordovician consists of schists, quartzites and migmatites. It was not studied in detail.

2. Silurian. The Silurian consists of black, carbonaceous shales with subordinate black limestones.

3. Devonian. In the Lower Devonian three formations are distinguished. The Lower Limestone Formation consists of partly recrystallized limestone and marble. Its thickness is variable because of tectonic action, but may have been some 300 m originally. The Detrital Formation consists of shales and graywackes with subordinate limestones. The thickness of this formation varies from 200 m to more than 600 m. The Upper Limestone Formation consists of thin bedded limestones. Its maximum thickness is 100 m. The Middle Devonian also comprises three formations. The lower and the upper formation are quite similar and consist of nodular or banded limestones and calcareous shales. The thickness of the lower formation is about 200 m. Of the upper formation a thickness of 100 m is exposed. The middle formation consists of limestones which are well bedded in its upper- and lower part. The central part is more massive. The thickness of the middle formation is about 100 m. The Upper Devonian occurs in a formation of griotte limestones which contains also part of the Lower Carboniferous. Conodonts give a strong indication for the occurrence of Famennian in this formation. The thickness of the Upper Devonian is unknown. The whole formation varies in thickness from 16 m to 38 m. The genesis of the griotte structures is ascribed mainly to tectonic action.

4. Carboniferous. The Lower Carboniferous consists of griotte limestones, lydites, limestones and shales. The occurrence of Upper Tournaisian in the griotte limestone formation, which also contains Upper Devonian, is probable because of its conodont fauna. The thickness of the lydite formation varies from 3 m to 24 m. On the lydite brown and red limestones and shales are found, which are 18 m to 70 m thick. Next follow black limestones with a thickness of about 60 m. The top of the Lower Carboniferous is formed predominantly by shales, which possibly represent a transition to the detrital Upper Carboniferous. The Upper Carboniferous consists of two formations. The lower formation is a monotonous alternation of graywackes and shales with a maximum thickness of about 500 m. The upper formation consists of green shales with intercalated sandstones and limestones. Its maximum thickness

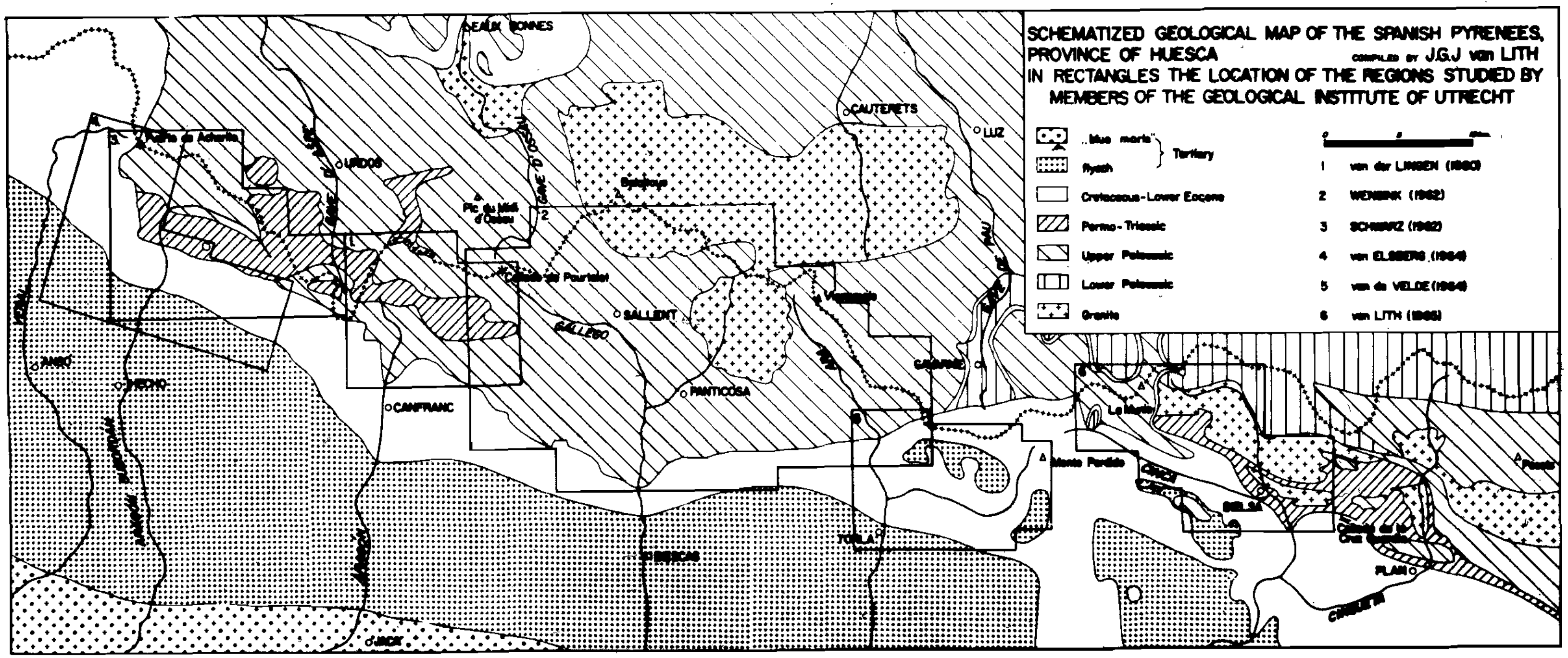


Fig. 1. Geological survey map of the Western Pyrenees with location of the regions studied by members of the Geological Institute of Utrecht.

amounts to about 400 m.

5. Permo-Triassic. At the base of the Permo-Triassic a conglomerate bearing zone with a maximum thickness of 3 m occurs in many places. It is followed by some 8 m of light grey sandstones. The higher parts of the series consist of an alternation of red sandstones and red shales. The capping bed is formed by green shales which grade into the Middle Triassic. The total thickness of the Permo-Triassic varies from 80 to 120 m.

6. Middle and Upper Triassic. The Muschelkalk is represented by dolomites, limestones and marls. Its thickness is less than 40 m. The Keuper is formed by gypsiferous marls and cellular dolomites. Its thickness cannot be established.

7. Jurassic. No Jurassic is found in the study region.

8. Cretaceous. The Cretaceous, as far as studied, consists of two series. The lower series, of Coniacian-Santonian age, is composed of medium bedded limestones and dolomites, which generally contain some quartz sand. The basal layer is formed by a conglomerate, which contains components of the immediately underlying rocks. The maximum thickness of this series is 150 m. The upper series has a Santonian-Campanian age and consists of limestones, sometimes argillaceous or cellular, and dolomites. Only a thickness of some 30 m was studied. No detailed study was made of the higher parts of the Cretaceous nor of the Tertiary.

The granite, which occurs in the study region, is covered by a thick waste mantle of pre-Permo-Triassic age. The granite itself was not studied.

From a structural point of view the region can be divided into three major units.

The first major unit consists of the crystalline basement of Ordovician and granite toghether with its autochthonous cover of Permo-Triassic, Middle and Upper Triassic and Coniacian-Santonian sediments. Generally the sedimentary cover is gently folded but zones of more intense folding do occur. The folding is supported to be the result of more or less vertical differential movements of the basement. Two structural trends can be discerned. One is about E - W, the other varies from N62°W to N74°W (Alpine direction in the Pyrenees). In so far as can be deduced from the faults in which the sedimentary cover is involved, the trend of the faults in the basement is generally about E - W. This agrees with the E -W trending folds in the sedimentary cover. The folds with Alpine direction correspond with an echelon arrangement and with rotational movements, of the E -W striking faults. Locally they correspond with faults which have an Alpine direction.

The second major unit is the Gavarnie nappe, which consists of Upper Paleozoic sediments that overlie major unit 1. Locally overthrust slices, consisting of Permo-Triassic and Middle and Upper Triassic occur at the base of the Gavarnie nappe. These were sheared off from the autochthonous sediments and probably moved over a shorter distance than the Paleozoic. In the Gavarnie nappe itself four tectonic units are distinguished. These units are thrust slices which originated probably during the movement of the Gavarnie nappe as a whole. The Silurian black shales formed the lubricating layer at the base of the Gavarnie nappe. As the age of the folding within each unit could not be ascertained the attitude of the folds cannot be used as a criterion for the direction of the thrust movement. It is the position of the third major unit which provides the evidence for a southerly movement of the Gavarnie nappe.

The third major unit consists of Cretaceous and Lower Tertiary rocks that overlie the major units 1 and 2. The top of the Cretaceous of major unit 1 has about the same age as the base of the Cretaceous of major unit 3. This suggests a normal order of succession from the Cretaceous of major unit 1 into that of major unit 3, where these units are in contact with each other. The third major unit moved, however, over a considerable distance from north to south to reach its actual position. This paradox is explained by an onlap of the Cretaceous on its basement. During the Alpine orogeny the Cretaceous and Lower Tertiary of the northern region, where sedimentation only began in the Santonian-

Campanian, moved southward on the back of the Gavarnie nappe. The basal thrust plane of the nappe happened to cut off the underlying autochthonous Cretaceous at about the same time-stratigraphic level as represented by the base of the autochthonous Cretaceous of major unit 3. The cover of Cretaceous to Eocene rocks moved farther southwards than the Paleozoic. In front of the Gavarnie nappe the base of this Cretaceous came into contact with synchronous autochthonous Cretaceous of major unit 1.

# 1. GENERAL INTRODUCTION AND GEOMORPHOLOGICAL ASPECTS

## 1.1. INTRODUCTION

### 1.1.1. Location of the region

The region investigated is situated in the western extremity of the Central Pyrenees, province of Huesca (fig.1). Its northern and northwestern boundary is formed by the state frontier with France. To the south and southwest the region is bordered by the scarps of Upper Cretaceous to Eocene sediments. The western boundary is a line just west of the watershed between the Rio Cinca and the Rio Cinqueta. The outlines of the map coincide approximately with the frontier of the municipality of Bielsa.

### 1.1.2. Purpose of the study

The study was set up to get an insight in the stratigraphy and tectonics of the Spanish part of the so called Gavarnie nappe and of the sediments which underlie it. The Gavernie nappe consists of Upper Paleozoic sediments which overlie Cretaceous sediments. It extends for the greater part on French territory.

This paper forms part of geological investigations carried out by members of the Geological Institute of the State University of Utrecht in the Spanish Western Pyrenees. These studies are directed by Prof. M.G.Rutten. Up to now 5 theses have appeared in the frame of this project (van der Lingen, 1960; Wensink, 1962; Schwarz, 1962; van Elsberg, 1964; van de Velde, 1964). The locations of the studied regions are indicated on the survey map (fig. 1.).

### 1.1.3. Methods of investigation

For the geological mapping use was made of the topographical maps sheet 146 (Bujaruelo), 147 (Liena), 178 (Broto) and 179 (Bielsa), scale 1 : 50000, published by the Direction General del Instituto Geografico y Catastral de Madrid. These maps were photographically enlarged to a scale 1:20000 to enable detailed mapping. The field work was performed during the summers of 1959 - 1962. The laboratory work was done in the Geological Institute of the State University at Utrecht, Holland, where also the collections are deposited. The determination of the foraminifera was carried out by Dr.J.E. van Hinte of this institute. The conodonts were determined by W.A. van Wamel.

### 1.1.4. Previous authors

In the previous century only occasionally some attention was payed to our region. The first detailed study was made by Bresson (1903) who investigated the Paleozoic of the French departments Hautes-Pyrénées and Basses-Pyrénées. In this study he described also the northernmost part of our region. The thesis of Dalloni (1910), concerning the geology of the Pyrenees of Aragón, has been of great value for preparing the present paper. He described the extension of the Gavarnie nappe on Spanish territory and the stratigraphy of this region. The study of Misch (1934) deals with the stratigraphy of the Mesozoic and with the Alpine tectonics of the part of the Pyrenees in which our region is situated.

A more detailed account of the history of the investigations on the Gavarnie nappe as a whole will be given in part 4.0.2.

### 1.1.5. General remarks

The municipality of Bielsa consists of five villages, viz. Bielsa (423 inhabitants), Javierre (35), Espierba (127), Parzán (110), Chisagües (44). The numbers of inhabitants are derived from the census of 1960. By road the region is only accessible from the south. A by-road bifurcating from the road connecting Ainsa with Plan, runs to Bielsa and farther north up to Hospital de Parzán. North of Hospital de Parzán a tunnel to France is under construction. Moreover, a by-road runs from Bielsa to Espierba. The chief means of subsistence are forestry (production of about 7000 $m^3$ p.a.) and cattle-breading (some 850 cows, 4500 sheep). Agriculture is of minor importance (mainly potatoes).

Hydroelectric power stations are found north of Parzán and at a dam in the Cinca valley above Bielsa (Pantano de Pineta). The power stations are connected by channels and pipelines. Formerly mining was of some importance. Up to the Civil War the lead mines of Liena were exploited by French companies. An abandoned aerial tramway leading into France testifies to those times. A recent evaluation led to the definite abandonment of these mines. An occurrence of fluorite is found southeast of Bielsa.

Although the Valle de Pineta is one of the most impressive valleys of the Pyrenees, only few tourists visited this region until recently.

### 1.1.6. Topographical names

For the same topographic objects different names are used by the French and the Spanish. In order to facilitate comparative studies of publications with French and Spanish topographical bases, a list is given of equivalent names (Table I).

## 1.2. GEOMORPHOLOGICAL ASPECTS

### 1.2.1. Relief

The maximal difference in elevation in our region is about 2200 m. The lowest point is found south of Bielsa at 940 m and the highest peak is formed by La Munia (3134 m). Two major mountain crests can be distinguished, The first is the Upper Cretaceous to Eocene range in the south: Monte Perdido, Sierra de las Sucas, Montinier, Maristá. It is only partly covered by our map. The second forms the state frontier in the north: la Capilla, Tormacal, Blancas de la Larri, La Munia. Other peaks reaching considerable heights are the Chinipro (2800 m), the Robiñera (3003) and the Barrosa (2763).

The slopes are moderate to very steep. In some parts of the region decreases of the slope angle occur which may represent planation levels, related to local or temporal base levels. However, gentler slopes may be caused also by lithology and cryoplanation. As we gave only little attention to the morphology we will not discuss the origin of these levels. A correlation with planation levels elsewhere in the Pyrenees would be absolutely hazardous. So we give only a review of the levels we observed.

| | |
|---|---|
| 1. 3000 m | the flat summit of the Robiñera |
| 2. 2700 - 2800 m | SW and S of La Munia, Chinipro, Barrosa |
| 3. 2500 - 2600 m | Lagos de La Munia, southern Chinipro Spur, Sierra de Liena, |
| 4. 2100 - 2200 m | La Estiva, Circo de Barrosa |
| 5. 1800 - 1900 m | Sierra de Espierba, Sierra Marqués |
| 6. 1500 - 1640 m | Larri valley, southern slope of the Pico del Cuezo, |
| 7. 1200 - 1350 m | Cinca valley. |

The valleys are partly longitudinal, partly transverse. Longitudinal valleys occur mainly in the sedimentary rocks where they follow structural trends (Fuen Santa, Upper Cinca, Real, Montillo), whereas the transverse valleys are mainly found in crystalline rocks (La Larri, Barrosa). The transverse valleys are consequent and in areas with sediments cataclinal.

### 1.2.2. Glaciation

Our area forms part of a region with impressive glacial erosion (Cirque de Gavarnie, Ordesa trough valley, Circo de Pineta, Cirque de Troumouse, Circo de Barrosa). However, most of the 'Cirques' are in fact trough-ends. A true cirque is found just south of the Collado de las Puertas. The Lago del Cao is situated in another cirque. Beautiful examples of trough valleys in our area are the valley of the Larri river and the upper Barrosa valley. The Cinca valley above Bielsa is an asymmetric glacial valley (fig.2). Its steep southern wall is formed by the scarp of a cuesta of Upper Cretaceous to Eocene sediments, its gentle northern side is a dip slope of Upper Cretaceous rocks. Hanging valleys and glacial steps are numerous, e.g. in the Fuen Santa, between the Fuen Santa and La Larri, between La Larri and the Cinca (fig.3), in the Cinca above Bielsa, in the Real near Chisagües, in the upper Barrosa and in the Barrosa near Hospital de Parzán.

Morainic deposits occur scarcely in our region. At the valley mouth of La Larri and in the Cinca valley above Bielsa morainic dams are found. Elsewhere the morainic deposits, in so far as they were deposited, have been largely washed away by later fluviatile erosion.

### 1.2.3. Stream erosion

After the retreat of the ice, the rivers attacked the glacial relief. The fluviatile erosion was concentrated on the glacial steps which formed local base levels. Upstream of these steps the rivers formed alluvial plains and developed meanders (Cinca and Larri). Gradually the glacial steps were deeply incised and gorges were formed. This involved a lowering of the local base levels and part of the alluvial deposits was cleared away. We now find remnants of alluvial terraces in the Cinca valley at 1600 m and in the Larri valley at 1240 m.

Gully erosion is mainly found in the shales of the Lower Devonian and Permo-Triassic. Sinkholes are situated on the foot of the cuesta scarp of Upper Cretaceous limestones south of the Comodoto and La Estiva (fig.4).

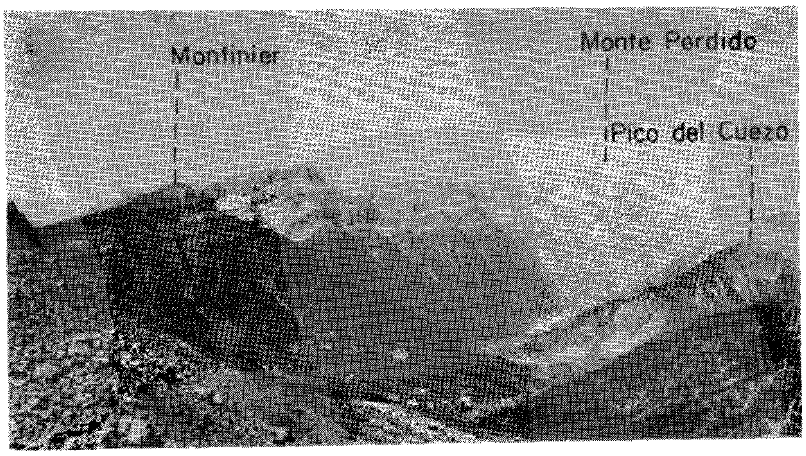


Fig. 2. The Cinca valley above Bielsa (Valle de Pineta). View from the Collado de la Cruz de Guardia in western direction.

TABLE I

EQUIVALENT TOPOGRAPHIC NAMES

| Topographical map sheets 146, 147, 179 | Carte géologique France feuille de Luz (251) | other equivalents |
|---|---|---|
| Barrosa | | Pic de Barroude |
| Blancas de la Larri (2707) | Pic de Pène-Blanque (2811) | |
| La Capilla (2830) | Pic blanc (2836) | |
| Chinipro (2800) | Pic de Lary (2805) | |
| Collado de las Puertas | Col de las Portes | |
| Comodoto (2362) | P. de Piedra Mucla (2369) | Piedra Nuela |
| Fuen Santa | Hount Sainte | |
| Hospital de Parzán | Hôpital de Bielsa | Hospital |
| La Mota (2580) | P. Chamenas (2641) | |
| La Munia (3134) | Pic de la Munia (3150) | |
| unnamed peak W. | Mont Arrouye (3039) | |
| of La Munia (2908) | | |
| unnamed peaks NE | Pic de Serre Mourène (3058) | |
| of La Munia | Pic de Troumouse (3086) | |
| Pico del Cuezo | el Queso | Quezo |
| Puerto Barrosa | Port de Barroude | |
| Puerto de Fenás | Port de la Canaou | |
| Puerto Viejo | Port-Bieil | |
| Punta Suelza (2973) | Pic de Suelza (2969) | |
| unnamed peak NW of | Punta Fulsa (2862) | |
| Punta Suelza (2867) | | |
| unnamed brook from | Bo. Ribarnera de la Petta | |
| Lgo. del Cao to Cinca | | |
| Rio Real | | Bco. de Chisagües |
| Robiñera (3003) | Pic de las Luseras (3007) | Pic de las Louseras |
| Sierra de Espierba | Sierra de Orabel | |
| Sierra Marqués | Sierra del Marqués | |
| (E of Bielsa) | (near Punta Suelza) | |
| Tormacal (2848) | Soum de Port-Bieil (2848) | |
| Urdiceto | Ourdissettou | Urdiseto |
| Valle de Pineta | Vallée de Pinède | |

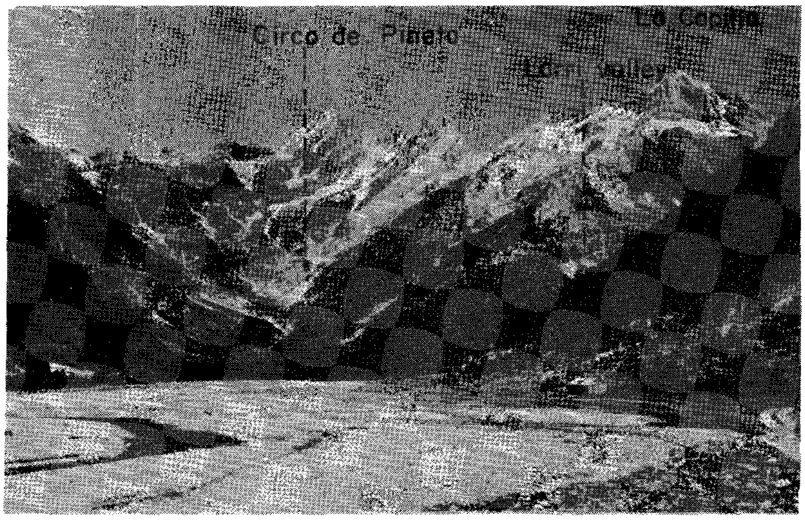


Fig. 3. Hanging valley of La Larri. View in western direction.

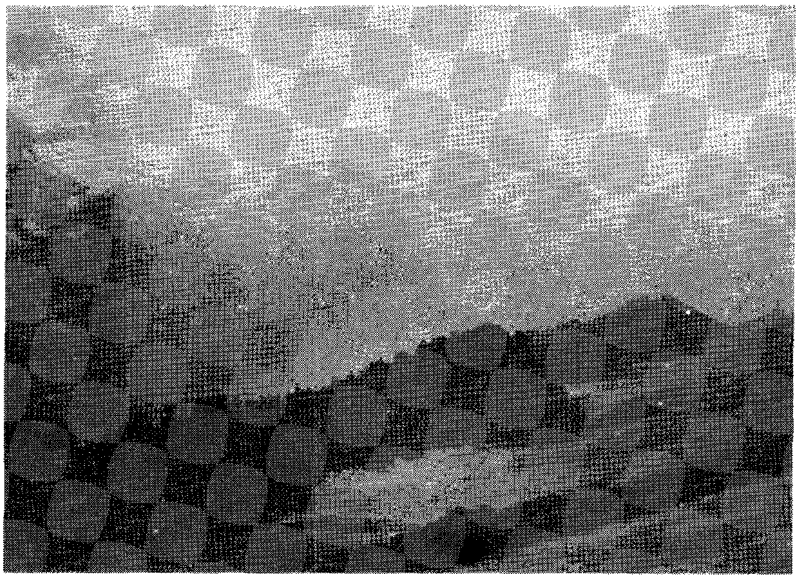

Fig. 4. Scarps of Upper Cretaceous south of la Estiva. View in western direction.

# 2. STRATIGRAPHY

## 2.0. INTRODUCTION

As fossils are extremely rare in the region of study, our stratigraphy is mainly based on lithology. Only exceptionally we found paleontological evidence for the age determination of a formation. In order to avoid confusion we prefer the chronostratigraphic terminilogy as usually applied in the Pyrenees and do not use formation names derived from geographic features as recommended by the International Subcommission on stratigraphic Terminology.

Therefore it should be kept in mind that we use the chronostratigraphic names of the various formations and groups as defined in our description of the local unit concerned. The assigned ages are based mainly on lithological correlation with formations of supposedly known age elsewhere in the Pyrenees. Accordingly the description of each system is preceded by an introduction in which a short review is given of the development and extension of that system in the Pyrenees.

## 2.1. CAMBRO-ORDOVICIAN

The oldest rocks of the Pyrenees consist principally of detrital sediments and are found mainly in the Eastern and Central Pyrenees but occur also in the Western Pyrenees near St.Jean Pied de Port. Age determination of these sediments is impossible, because of the absence of fossils. Consequently, the presence of Cambrian has been questioned for a long time. Recent authors however, (e.g. Cavet 1957) indicate the occurrence of a series comparable with the fossiliferous Cambrian of the Montagne Noire. In the Lower Ordovician also fossils are lacking. The Upper Ordovician however, yielded faunas in several places, mainly of Caradoc age.

On account of the difficulties met with in dating the Cambrian and Ordovician the whole of the oldest sediments is frequently indicated as Cambro-Ordovician. Its stratigraphy in some nearby regions has been summarized in Table II.

The Cambro-Ordovician series, possibly including even older sediments, are metamorphosed to a large areal extent and locally to a high degree. In low-grade metamorphic areas sericite-phyllites are found, grading into mica-schists. With increasing metamorphism one finds a sequence of mica-schists, andalusite-mica-schists, migmatites and quartzdiorites, and basal gneisses. The metamorphic rocks of the Central Pyrenees have been described by several authors, a.o. Guitard (1955, 1958), Raguin (1938), de Sitter and Zwart (1958, 1962). The reader is referred to these publications for further details and literature.

In our region rocks of supposed Ordovician age occur in the extreme north and further south of Bielsa in the Cinca valley and in the Larri valley. Though no detailed study was made of these formations, some general aspects will be mentioned.

In the Barrosa region we observed a monotonous formation of quartzites and quartz-biotite-schists, cut by many granitic dykes. These rocks are attributed to the Ordovician by Bresson (1903), Dalloni (1910) and Clin (1959). Barrow (1908) described two thin sections from this area in which he found decomposed andalusite. The boundary of the Ordovocian with the granite of Bielsa, which lies farther south than indicated by Clin, is not sharp in the Barrosa region. Towards the contact the granitic dykes in the Ordovician rocks become more and more numerous. The country rock passes into a mixture of granitic and wall rock material. Gradually the inclusions

TABLE II

Correlation of the Lower Paleozoic stratigraphies as given by: Destombes (1953), Clin (1959) and Waterlot (1961).

| | Destombes 1953 | Clin 1959 | Waterlot 1961 |
|---|---|---|---|
| Ashgill | | | |
| Caradoc | Calcaires à Echinosphaerites de Montauban-Luchon.<br>Schistes ardoisiers de Cazarilh et de Squiéry. | Calcaire de Salles et poudingues à billes<br>Schistes et quartzites bleus avec calcaires. | Alternance de calcaires noirs et de schistes bleu-noir (à faciès pseudo-griottes). |
| Llandeilo | Conglomérats à éléments de quartz et quartzite. | | |
| Arenig | Calcaires métallifères de Caralp.<br>Schistes et quartzites noirs (faciès pseudo-carburé). | Conglomérats.<br>Schistes et phyllades noirs. | Schistes noirs. |
| Upper Cambrian | Schistes et phyllades sombres de Squiéry. | Calcaires métallifères schistes satinés. | Alternance de calcaires bleus assez pyriteux et de schistes à allure carburée. |
| Middle Cambrian | Calcaires et Schistes de Squiéry et de Pouchergues en très petits lits. | Quartzites et quartzophyllades gris-verdâtre | Alternance de calcaires bleu-noir et noirs et de schistes gris-foncé ou noirs. A la base calcaires noirs en plaquettes. |
| Lower Cambrian | Schistes et quartzites claires de Squiéry avec quartzites dominant à la base.<br>Conglomérats à éléments quartzeux et quartzitiques de Squiéry reposant en concordance sur des phyllades sombres. | | Alternance de calcaires schisteux très clairs de marbres cipolins clairs et de calcschistes clairs. |

of country rock decrease and finally one finds a homogeneous granite. The width of this zone of migmatite-like rocks is some 200 m.

South of Bielsa we found a formation of grey-blue dense schists overlain discordantly by Permo-Triassic red beds. An X-ray investigation of these schists showed only mica's and some quartz without any trace minerals which could give an indication for the metamorphic facies. We suppose that these are the 'schistes bleus' of Clin (1959) and so attribute them to the Ordovician.

In the Larri valley we found migmatites. These rocks, however, are badly exposed because of deep weathering and the alluvial cover in the valley.

## 2.2. SILURIAN

### 2.2.0. Introduction

In the Pyrenees the presence of at least the greater part of the Silurian has been proved by fossils. Neither the lowermost nor the uppermost part are, however, known with certainty. Graptolites pointing to a Lower Landovery age have been reported only from one locality (Laruns, Basses Pyrénées, Dalloni 1952). However, a continous sedimentation from the Ordovician into the Lower Silurian is mentioned by several authors (a.o. Zandvliet, 1960; Cavet, 1957· de Sitter and Zwart, 1958). The same holds for the uppermost part of the Silurian which did not yield a distinctive fauna (Dalloni, 1913, 1930: Schmidt, 1931: Boissevain, 1934) but which passes gradually into the Lower Devonian in some places (Schmidt, 1931; Zandvliet, 1960· Cavet, 1957: Schulman, 1959 Mirouse, 1962). It must, however, be kept in mind that disharmonic folding of the Upper Paleozoic with regard to the Lower Paleozoic frequently obscures the original sedimentary relation between the Silurian and the Devonian.

The Silurian mainly consists of black shales, the characteristics of which have been described at some length by Kleinsmiede (1960). In the lower part quartzites are frequently found. A calcareous zone with *Orthoceras* and *Cardiola interrupta* usually occurs near the top and has been dated as Upper Wenlock (Dalloni 1913, 1930) or Ludlow (Schmidt, 1931: Zandvliet, 1960: Wensink, 1962). Pyrite, in the form of crystals, veins and nodules, is generally abundant.

The black colour of the Silurian has been the subject of several investigations. Capdecomme (1943) showed the colouring matter to be graphite. His analyses gave a C content of 3,8 - 6,1 % for the bulk composition, whereas very rich parts yielded 10,2 - 24,5 %. Analyses mentioned by de Sitter (1959), de Sitter and Zwart (1962) and Kleinsmiede (1960) show a carbon content of 0,3 - 8,8%. Some authors, however, attribute the black colour to finely divided iron sulfide (e.g. de Sitter, 1959: Fearnsides in Zandvliet, 1960).

The deposition of the Silurian rocks of the Pyrenees is generally regarded as to have occurred in a period of widespread quiet sedimentation in an euxinic environment.

### 2.2.1. Silurian of the region

The Silurian occurs in a strongly tectonized zone at the base of the Gavarnie nappe, not only in our region but also in France in the whole area occupied by the nappe (Bresson 1903). For graphic reasons the Silurian of the basal thrust zone could not be indicated on the map except in the Circo de Barrosa. The extension of the Silurian has been indicated in fig. 5. The zone of Silurian rocks varies greatly in thickness, attaining its maximum in the Circo de Barrosa with about 125 m, whereas in other places it may be reduced to a few metres or even to zero. However, this variation in thickness has nothing to do with the original sedimentary thickness but is the result of tectonic action.
In the Circo de Barrosa the Silurian is developed in its well known facies of black staining shales, which are often pyritic and weather rusty brown wherever they are in contact with water. We came across: *Monograptus* sp.

On the northern slope of the Sierra

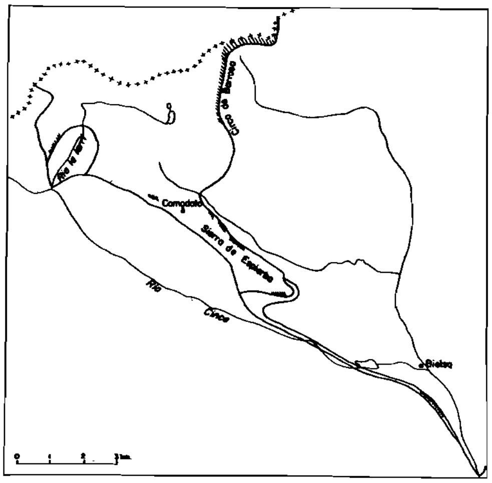


Fig.5. Observed localities of Silurian rocks (shaded).

de Espierba some limestone layers occur in the black shales. This may be the calcareous zone of the Upper Wenlock or Ludlow as mentioned in the introduction. Here we found: Orthoceras sp.

In the Larri valley we found the Silurian only at the western side of the valley where the black shales occur again at the base of the overthrust mass. The thickness is reduced to 3 m, whereas on the northern and eastern side of the valley the Silurian seems to be lacking altogether.

The easternmost occurrence of Silurian black shales at the base of the overthrust mass is found south of Bielsa.

Fig.6. Thin section of Silurian limestone with enigmatic structures of carbonaceous matter.

Between the Comodoto and La Estiva black shales and limestones occur in a short and narrow band. The shales contain calcareous nodules with a pyritic core, a phenomenon well known from the Silurian elsewhere in the Pyrenees. The black limestone contains only a few percent of opaque, probably carbonaceous, matter occurring between the small grained calcite crystals. We found also a grey limestone crowded by small black patches. It is not known whether these are organic remains of some kind or may have been formed by concentration of carbonaceous matter driven out during recrystallisation (fig. 6).

The Silurian age of the described rocks is based on the discovery of Monograptus, the lithology and the continuity to dated localities outside our region.

## 2.3. DEVONIAN

### 2.3.0. Introduction

In general the Devonian of the Pyrenees can be subdivided into Lower, Middle and Upper Devonian, either on a paleontological base or by lithological correlation with fossil-bearing sequences elsewhere. Each of these series may attain a maximum thickness of about 400 m.

In the Lower Devonian both the Gedinnian and the Coblentzian are represented. But the Gedinnian has been reported only from a few localities (e.g. Roussel, 1904; Dalloni, 1910, 1930; Laverdière, 1930; Schmidt, 1931). It shows a rather uniform development of shales, whereas the Coblentzian displays a more differentiated development. In the western Pyrenees the Coblentzian is composed of shales and graywackes and to a lesser degree of marls or limestones. Important lateral changes of facies occur (e.g. Wensink, 1962). Towards the east the graywackes disappear and in the Central Pyrenees the Lower Devonian has a more limy development. In some places it is even indistinguishable from the Middle Devonian (Dalloni, 1910; Boissevain, 1934; de Sitter and Zwart, 1958; Schulman, 1959).

The transition from the Lower Devonian to the Middle Devonian is generally formed by a calacareous zone, which contains *Spirifer cultrijugatus*. This zone is assigned to the Lower Devonian by some authors (a..o. Schmidt, 1931; Wensink 1962), to the Middle Devonian by others (a.o. Dalloni, 1910, 1930; Laverdière, 1930).

The Middle Devonian is mainly represented by the Couvinian. This series shows a widespread development and is generally fossiliferous. The Lower Couvinian occurs locally in the form of multicoloured *griottes*, shales and calcareous shales (Dalloni, 1910, 1910b, 1930: Wensink 1962). (For a discussion of the term *griotte*, see below). Generally, however, the whole of the Couvinian consists of grey limestones, mostly with weathered out silicified fossils and irregular dolomitic spots. The dolomitization has been ascribed either to an emergence following the Devonian or to magnesium expelled by regional metamorphism from deeper seated regions (de Sitter and Zwart, 1958). The basal part is well layered but the Couvinian becomes more massive, with transitions to reef limestone, higher in the series ( de Sitter and Zwart, 1958: van der Lingen, 1960; Wensink, 1962; Schwarz, 1962).

The Givetian is almost unknown in the Pyrenees though suspected in some places. Schmidt (1931) reported the Givetian from the Canfranc region. According to Wensink (1962) reef building may have continued into the Givetian but he found no index fossils. Dalloni (1930) assumes the whole of his '*calcaire massif à polypiers*' in the Catalan Pyrenees to be Givetian. Cavet (1957) supposes a Givetian age for the '*marbre flambé* in the Eastern Pyrenees.

Both the subdivisions of the Upper Devonian, the Frasnian and the Famennian, are present in the Eastern and Central Pyrenees, although generally not in complete sections. The Upper Devonian is well known for its *griotte* limestones since Barrois found the Upper Famennian fauna of the Ravin de Coularie in such limestones. Besides of *griotte* limestones, mainly red coloured, but also grey, green or many-coloured, the Upper Devonian of the Eastern and Central Pyrenees contains also gray or blue massive limestones, multicoloured shales and limestone conglomerates (Dalloni, 1930· Boissevain, 1934: Destombes, 1953 Cavet, 1957: de Sitter and Zwart, 1958, 1962· Zandvliet, 1960).

The genesis of the *griotte* limestones is still a matter of discussion. These limestones consist of an alternation of laminae or thin strata of limestone and shale. The bedding planes are undulated and the shale layers frequently come in touch with each other, thus forming a nodular limestone. Sometimes the limestone nodules are formed by cephalopods. Ovtracht and Fournié (1956) distinguished three facies of *griotte* limestone:

1. incontestable intraformational conglomerates.
2. *griottes sensu stricto*: alternation of clay and limestone laminae which underwent the effect of slumping nearly contemporaneous with their deposition. As a result the limestone zones are irregularly undulated, sometimes forming more or less elongated nodules.
3. massive *griotte limestones* = patched limestones: the clayey cover of the nodules is indistinct, the structure is recognizable only by colour shades. However, all transitions to *griottes s.s.* occur.

According to Zwart (1953) the destruction of the normal stratification is not caused by slumping but is the result of agitation of the sea. Some authors consider tectonic deformation as an important factor in forming the nodular habit. Zandvliet (1960) reports the occurrence of single *griotte* layers with the long axis of the nodules standing obliquely to the bedding and parallel to the cleavage. According to de Sitter and Zwart (1962), stretched nodules in the *griotte* limestones are formed by strong deformation accompanying cleavage folding.

Probably both slumping and later tectonic deformation may have been active in forming the structures of the *griotte* limestones.

As it is often thought that *griotte* type limestones are characteristic of the Upper Devonian in the Pyrenees, it should be kept in mind that similar rocks occur also lower in the Devonian and in the

Carboniferous.

The griotte limestones, which predominate in the Upper Devonian of the Eastern and Central Pyrenees, become less important towards the west. In the western Pyrenees impure limestones, shales and sandstones with Spirifer verneuilli form a considerable part of the Upper Devonian (Dalloni, 1910; Schmidt, 1931; van der Lingen, 1960· Mirouse, 1960a, b; Wensink, 1962: Schwarz, 1962).

### 2.3.1. Devonian of the region

#### 2.3.1.1 Lower Devonian

In the Lower Devonian we distinguish from the base to the top a Lower Limestone Formation, a Detrital Formation and an Upper Limestone Formation.

The Lower Limestone Formation is found in the Blancas de La Larri, in the Circo de Barrosa continuing to the south towards the Rio Real and in the Sierra de Espierba. Its thickness is quite variable because of tectonic action but originally it may have been some 300 m. The Detrital Formation is exposed in the northern part of the region in a zone all round the Blancas de La Larri from the Puerto de Fenâs to the Munia and northward into France. Further it is found in the southern slope of the Robiñera and all along the southern border of the Paleozoic, where it covers a considerable area northwest of the Comodoto, at La Estiva and in the Larri region. The minimum thickness of the Detrital Formation is 200 m but it may amount up to 600 m and more. However, the possibility of thickening by isoclinal folding and/or imbrication should be taken into account. The Upper Limestone Formation is found in a small zone from the Puerto de Fenas to the Robiñera and in the southern slope of the Robiñera. Its thickness varies from 0 m to 100 m. The Lower Limestone Formation consists of massive, mottled white, light grey, blue and pink, partly recrystallized limestone and marble. The limestones are conspicuous by their light weathering colour: in the Blancas de La Larri even a brilliant white (fig. 7). Bedding is only occasionally observable, mainly

Fig. 7. Strongly folded Middle Devonian on the wouthern slope of the western Chinipro spur. In the background the Lower Limestone Formation of the Blancas de La Larri.

on the margins of the massifs. Seen from a distance the eastern edge of the Blancas de La Larri faintly shows a thick bedding with layers of some 20 m thick. Lateral transitions into 'barrégienne' (ribbed limestone) occur locally in the Sierra de Espierba and in the Circa de Barrosa. Sometimes irregular brown-weathered dolomitic spots are found. Near the top the rock contains streaks of quartz sand, which become more numerous upwards and sometimes the top is formed by a quartzite layer. More lithological variations may occur but could not be observed because no good cross sections are exposed that are not inaccessible scarps. Because bedding is almost absent we consider the formation to consist largely of reef limestones. However, the strong recrystallization left no trace of reef building organisms. We did not find any fossils but Dalloni (1910) mentions Favosites and Cyathophyllum from our region.

Our Lower Limestone Formation forms part of a limestone band which is found over a large area in the southern part of the sheet Luz of the geological map of France 1 : 80000. This limestone, called 'dalle' by Jacquot (in Bresson, 1903), is attributed to the Middle Devonian ($d^{6-4}$) by Bresson. Its position immediately on the Silurian he ascribes to tectonic action.

According to Clin (1959) however, this limestone is Lower Devonian because of its occurrence in the same position in an undisturbed section near the Neste de Louron. This view is not shared by Mirouse (1962). Following Clin we attribute this limestone to the Lower Devonian.

The Detrital Formation in the northern part of the region consists of shales, graywackes and some limestones. The shales are darkgray, weathering grey or brown, and have a variable sand content of quartz with accessory zircon, tourmaline and ore. Sometimes they are somewhat limy. Occasionally impure limestones are intercalated. The graywackes are somewhat greenish dark grey and weather brownish grey. They are well developed near the Puerto de Fenás with layers up to 1,50 m thick. The graywackes are composed mainly by quartz and mica's. The content of rock fragments and felspar is very low. A zone of impure limestones with a thickness of some 120 m occurs in the Detrital Formation at a distance of about 100 m above the Lower Limestone Formation. This zone consists of an alternation of greenish smooth weathered bands and brown weathered bands with cavities. These cavities may partly represent fossil moulds. These limestones contain:

Fenestella sp.

and undeterminable coral remains.

Near the Puerto de Fenás Bresson (1903) found the following fauna:

Fenestella plebeia
Zaphrentis
Pleurodictyum
Atrypa reticularis

From the area of the Lago Superior de la Munia he reported:

Spirifer
Atrypa reticularis
Pleurodictyum
Fenestella

We only came across several Fenestella's.

In the southern slope of the Robiñera the formation is quite similar to that described above. Among the badly preserved fossils we identified : Fenestella sp.
Pleurodictyum sp.
Spirifer sp.
Stropheodonta sp.

In the southern outcrops of the Detrital Formation, from the Larri valley to the Sierra de Espierba, shales predominate. The graywackes and impure limestones are less developed here. The shales are blueish grey to black and weather grey to brown. They vary from pure argillaceous with paper thin lamination to arenaceous with irregular layers. Sometimes they grade into calcareous shales or impure limestones. The graywackes are well developed just east of La Estiva. The rock is medium grey and contains quartz (60 %), rock fragments (20 %), muscovite, biotite and chlorite (15 %) and felspar (5 %). Most of the felspar is plagioclase with about 30 %. An. In the western slope of the Larri valley limestones occur resembling those described from the Detrital Formation near the Blancas de La Larri. In these limestones we found:

Fenestella sp.
Atrypa sp.

On the Comodoto we came across:

Cyathophyllum sp.
Rhynchonella sp.
Zaphrentis sp.

Elsewhere hardly any fossils were found. But from the region west of Puerto Viejo which forms the continuation in France of the zone concerned, Bresson (1903) mentioned:

Pleurodictyum problematicum Goldf.
Fenestella aff. plebeia M'Coy
Atrypa reticularis Linn.
Zaphrentis
Spirifer
Phacops aff. potieri Bayle

Although our fossils could not be determined specifically, we attribute a Lower Devonian age to the Detrital Formation because Bresson dated this formation both in our region and in adjoining areas in France on account of a decisive fauna.

The Upper Limestone Formation consists of a grey dense limestone with layers from 0,5 cm to 5 cm thick and showing a platy habit. Sometimes the beds are 10 cm to 30 cm thick. The few corals we found are undeterminable. In general the formation is concordant with both the Lower Devonian and the Upper Carboniferous. In the last few metres up to the limestone the shales of the Detrital Formation have a gradual increasing number of intercalated limestones layers. The contact with the Upper Carboniferous is, however, more abrupt and frequently disturbed. Accordingly we suppose that this formation forms the upper part of the Lower Devonian. Bresson (1903) found these limestones over a large area in the French part to the Gavarnie nappe and assigned them to the Middle Devonian (d3b). We do not think, however, that it is equivalent with our Middle Devonian limestones (cf. part 2.3.1.2) because of its different lithology.

A similar formation is found along the southern boundary of the Detrital Formation in the southern slope of the Robiñera.

#### 2.3.1.2 Middle Devonian

The series we assign to the Middle Devonian is found in the Chinipro, in the Robiñera just east of the Collado de las Puertas, in the Comodoto, and in the western and northern side of the Larri valley. The series can be divided into three formations.

The lower and upper formations are quite similar and consist of impure limestones, generally nodular or banded, of calcareous shales and to a lesser degree of argillaceous shales. The thickness of the lower formation cannot be measured because of tight folding but it may amount to about 200 m. The upper formation is at least 100 m thick. The rocks of both formations are normally neutral light grey, but sometimes greenish or pink. They show reddish, brown, yellowish and greenish weathering colours. The shales contain some chloride and a little fine-grained quartz, but real arenaceous shales do not occur. No fossils were found in these formations. The lower one might be correlated with the multicoloured fossiliferous Lower Couvinian as reported by Wensink (1962) and Dalloni (1940, 1944, 1930). The upper formation probably represents a transition from the Middle Devonian to the Upper Devonian. It might be homologous with variegated shales and sandstones in the valley of the Gave de Brousset, described by Mirouse (1960) as corresponding with a part of the middle and upper series of the Devonian.

The middle formation consists of medium grey limestones, weathering light grey. The thickness of this formation is quite variable but averages 100 m. It forms a band which stands out clearly in the southern slopes of the Chinipro (fig. 7,8). In its lower and in its upper part the rock is well bedded in layers of 10 - 50 cm thick, whereas the central part is more massive with indistinct layers which have a thickness of 1 - 2 m. Silicified spots and streaks weather as protuberances with a brown colour. The limestone is generally small-grained crystalline.

The generally silicified and protruding fossils are few and it is difficult to detach them from the rock without damage, The Chinipro yielded:

- Alveolites sp.
- Favosites sp.
- Thamnopora sp.
- Philipsastrea sp.
- Zaphrentis sp.

In the exposure of Middle Devonian in the Robiñera we came across:

- Phacops sp.
- Chonetes sp.
- brachiopods undet.
- corals undet.

In the scree of the Comodoto we found

- Cyathophyllum sp.
- Zaphrentis sp.
- Favosites sp.
- Pleurodictyum sp.
- Philipsastrea sp.
- Tamnastrea ? sp.

In the Larri valley we found another Zaphrentis.

These fossils do not contradict a Middle Devonian age, which is probable because of the lithological similarities with Middle Devonian rocks outside our region.

#### 2.3.1.3 Upper Devonian

Upper Devonian occurs in our region in a griotte type limestone formation which contains also part of the Lower Carboniferous (cf. part 2.4.1.1). This for-

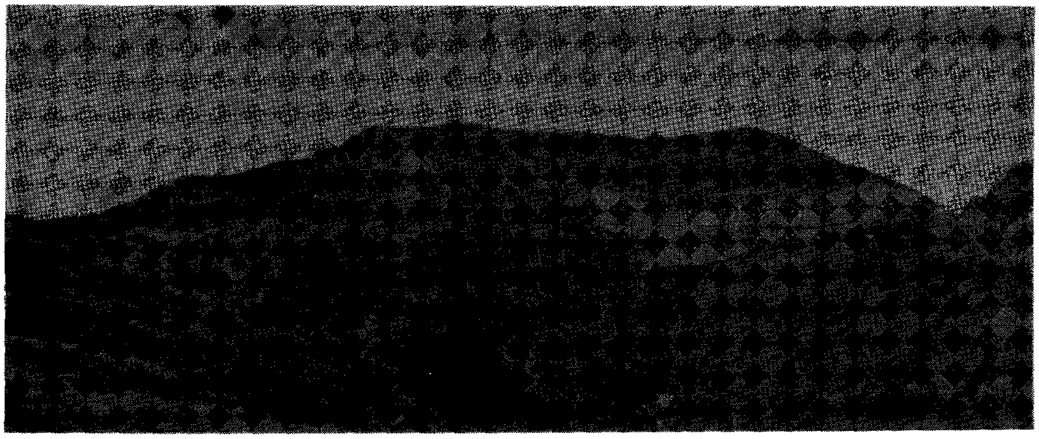

Fig. 8. View of the Chinipro from the south. LD : Lower Devonian (Detrital Formation), MD : Middle Devonian, UD - LC : Upper Devonian - Lower Carboniferous.

mation is found in the Estiva region and higher upwards to the summit of the Chinipro (fig.8). Smaller outcrops occur between La Estiva and the Comodoto near the point 2249 m (section VI). The thickness of the formation varies from 16 m to 38 m. (fig. 9). It consists of an alternation of thin strata and laminae of limestone and shale. In the lower part of the formation the limestone laminae are 0,5 - 1 cm thick, whereas they thicken upwards to strata with a thickness of several cm. The shaly intercalations likewise show a thickening from 2 cm at the base to 1 - 2 cm at the top of the formation. Several strata and laminae together form a bed. In the lower part the beds are 5 - 10 cm thick, whereas in the upper part thickness of 40 cm - 2 m are observed. The fresh rock has a medium gray colour with tendencies to blue, green or red, the shales being darker than the limestones. The weathering colour of the limestone generally is grey, but locally it may be pink. The shales weather brown and project by selective weathering. Seen at a distance the whole rock shows a yellowish grey colour.

In the griotte limestones the shaly layers are strongly undulated on cm to dm scale and frequently come in touch with each other. The limestone layers in between do not follow the undulating shales, but are adapted to the undulations by great variations in thickness. Often they are truncated by the shale layers. On a larger scale the connection between the limestone and shale layers is generally preserved and the sequence as a whole is not disturbed. On the other hand, the small scale undulations are not local, but occur throughout the whole formation. Locally the undulations of the shales are regular but in general the shale layers form a more intricate pattern.

When the undulations of the shales are regular, the isolated limestone elements are regurlarly shaped nodules which differ little in size. Cephalopods are often seen in these nodules but the fossils are very badly preserved.

In general, however, the shaly layers show irregular flow structures in which irregularly formed limestone elements of cm to dm size are embedded (fig.10). The larger limestone elements are somewhat coarser grained than the smaller ones.

These properties make it improbable that slumping played an important part in the origin of these structures. In our opinion they were mainly caused by tectonic deformation comparable with boudinage. During differential movements along the stratification planes the limestone layers reacted by plastic flow and recrystallization. Pre-existing irregularities in the stratification were accentuated and thus the limestone layers became divided into separate smaller and larger elements. These limestone elements were enveloped in the shaly material. Around the smaller limestone ele-

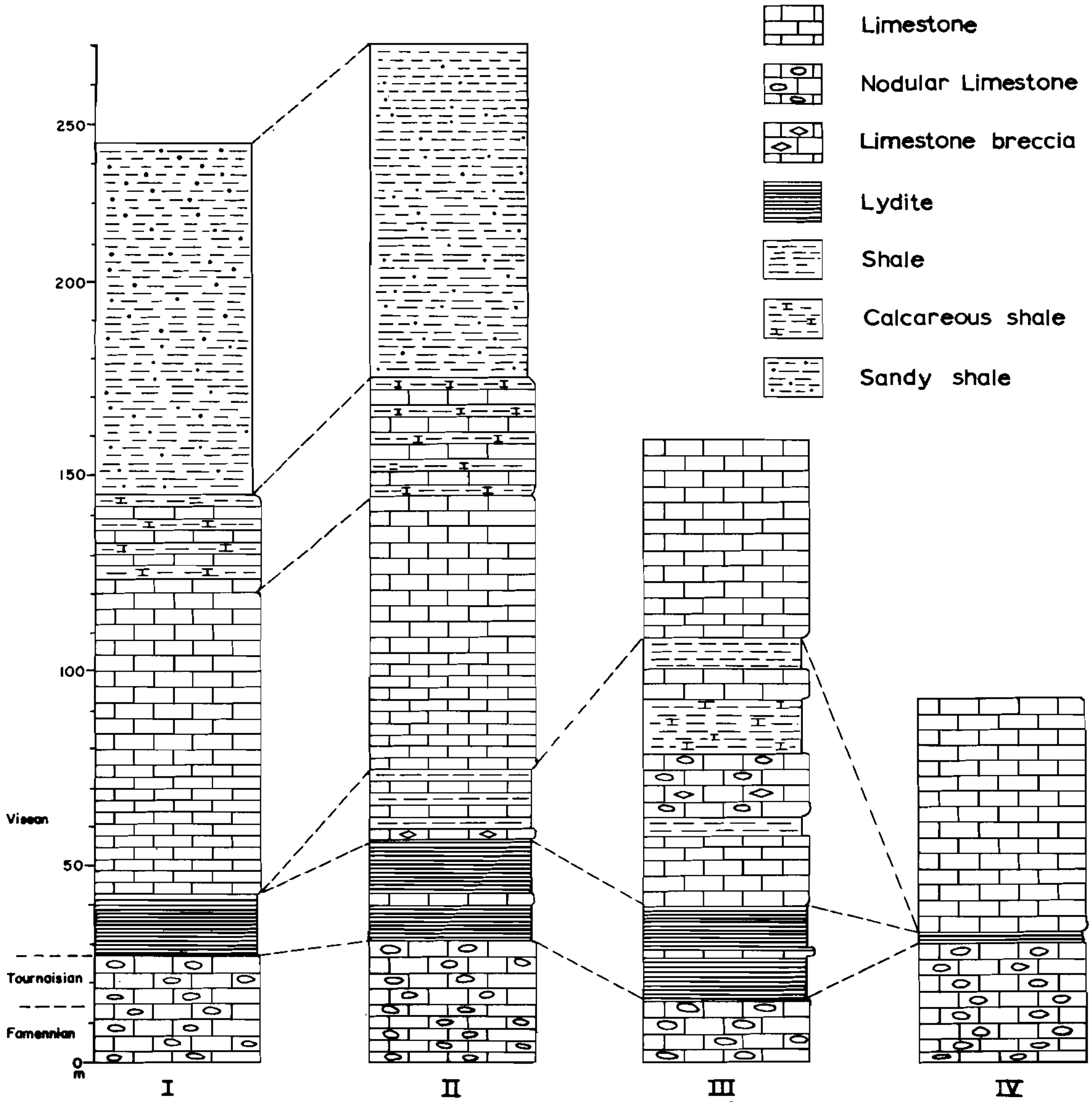


Fig. 9. Columnar sections of the Upper Devonian and Lower Carboniferous. I : Lower limb of Estiva syncline, II : Upper limb of Estiva syncline, III : Chinipro, IV : Eastern termination of Estiva syncline (point 2249 m).

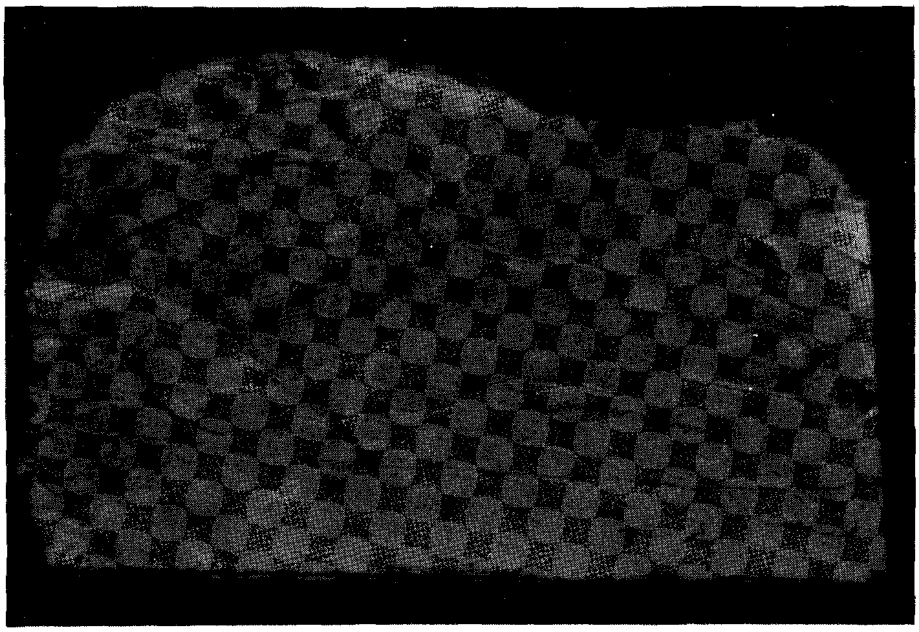

Fig. 10. Acetate peel showing griotte structure in Upper Devonian - Lower Carboniferous transition formation. Natural size,

ments the movement was largely taken up by the shaly material, in which these elements could move without internal deformation. The larger elements, however, encountered more resistance and suffered internal deformation, which is testified by stronger recrystallization. The fact that real slump structures are not apparent in our region does of course not exclude the possibility that slumping played a more important part in forming different structures in griotte type limestones elsewhere.

Apart from indeterminable cephalopods the only macrofossil we found is Rhynchonella sp. So the age determination of this formation could not be based on macrofossils. The occurrence of Upper Devonian in the described formation is based on conodonts. We found:

Scaphignathus velifera Ziegler
Palmatolepis glabra glabra Ulrich& Bassler 1926
Spathognathodus inornatus (Branson & Mehl)

This association can be correlated with the conodont fauna of the lower Platyclymenia--Stufe = lower Hemberg-Stufe in Germany (Ziegler 1962). This is a strong indication for the presence of the Famennian in this formation. We found no evidence for the occurrence of Frasnian in our region.

## 2.4. CARBONIFEROUS

### 2.4.0. Introduction

The Lower Carboniferous (Dinantian) in the Pyrenees is mainly represented by the Visean. Up to recently the occurrence of the Tournaisian in the Pyrenees was not known. However, gradual transitions from the Upper Devonian into the Lower Carboniferous were observed locally in the Central Pyrenees (Schulman, 1959; Zandvliet 1960 de Sitter and Zwart, 1962). Moreover a conodont fauna of Tournaisian age has been found by Ziegler (1959) in the valley of the Noguera Pallaresa. Also in our region the Tournaisian occurs (see part 2.4.1.1). So the assumption of a period of non-deposition during the Tournaisian cannot be maintained universally.

On top of the Tournaisian limestones, and locally perhaps on the Upper Devonian, a zone of lydites is found. It attains a thickness of some tens of metres at most. Locally the lydites contain nodules of calcium phosphate, particularly in their lower part. Sometimes a fine-grained breccia or a conglomerate of lydite fragments is found, covering the lydite or taking its place (Schmidt, 1931; de Sitter and Zwart 1958, 1962). A fauna from the phosphate nodules of Rimont (dep. Ariège) represents a zone corresponding with the Tournaisian-Visean boundary (Delépine, 1935 part 1).

The occurrence of the Visean is ascertained in several places by means of its fauna of goniatites, of which Goniatites crenistria is the most common. The Upper Visean of the French Pyrenees is characterized by the Productus fauna of Larbont and the Goniatites fauna of Mondette (Dubar, 1931 a,b; Delépine, 1935 part 2, 1937, 1953). In the Spanish Pyrenees goniatite faunas are known from several localities (a.o. Dalloni, 1910, 1930; Schmidt, 1931; Boissevain, 1934; Wensink, 1962). The Visean of the Western Pyrenees consists mainly of dark grey limestones which contain limestone breccias near the top. It attains a thickness of 50 - 100 m. Locally the series has a variegated colour or is developed as true griottes (Dalloni, 1910; Mirouse, 1960; van der Lingen, 1960; Wensink, 1962; Schwarz, 1962). Toward the e , the limestones decrease at the expense of shales and sandstones. East of the river Esera the limestones are no more but intercalations in a detrital series which comprises also the Upper Carboniferous (Dalloni, 1930; Boissevain, 1934; de Sitter and Zwart, 1958, 1962; Kleinsmiede, 1960; Zandvliet 1960).

The Upper Carboniferous shows a uniform development of alternating shales, sandstones and conglomerates over the whole Pyrenees.

The lower parts of the series, lying between the Visean and the fossiliferous Westphalian is considered to represent the Namurian, the occurrence of which has, however, not been proved by index fossils except in the Basses-Pyrénées. Here the Upper Visean grades into the Lower Namurian with Eumorphoceras bisulcatum (Delépine, Dubar, Laverdière, 1929; Delépine, 1953).

The Westphalian is well developed in the Western and Central Pyrenees, but it is missing in the Eastern Pyrenees. In many places it contains a rich flora (Roussel 1904; Dalloni, 1940; 1930; Schmidt, 1931; van der Lingen, 1960; Wensink, 1962; Schwarz, 1962). Locally coal measures are found. Limestone intercalations with marine fossils have been observed in the Upper Gállego and Canal de Izas regions (Schmidt, 1931; van der Lingen, 1960; Wensink, 1962).

The presence of the Stephanian in the Western Pyrenees is dubious (van der Lingen, 1960; Wensink, 1962; Schwarz, 1962) whereas in the Eastern Pyrenees it is well developed and contains coal beds and volcanic intercalations (Roussel, 1904; Dalloni 1930; Schmidt, 1931).

### 2.4.1. Carboniferous of the region

#### 2.4.1.1 Lower Carboniferous

The Lower Carboniferous in our region is found in the same areas as the Upper Devonian. The Lower Carboniferous on the Chinipro is shown in fig. 8. The Tournaisian occupies the upper part of the formation which has been described in part 2.3.1.3. This formation is overlain by a zone of lydites. The upper part of the Lower Carboniferous is formed by limestones and shales. Columnar sections of the Lower Carboniferous together with the Upper Devonian are given in fig. 9.

The presence of the Tournaisian in the Devonian-Carboniferous transition formation was ascertained on the basis of conodonts. A sample from the upper part of this formation contained:

Siphonodella duplicata (Branson & Mehl)
Siphonodella quadruplicata (Branson & Mehl)
Polygnathus inornata? E.R.Branson
Hindeodella fragilis? Hass 1959

The first three conodonts are characteristic for the Lower Pericyclus-Stufe in Germany (Bischoff, 1957). Hindeodella fragilis occurs in the Lower Mississipian in America (Lindström 1964). Thus the occurrence of Upper Tournaisian in this formation is highly probable.

On top of the Devonian-Carboniferous transition formation follows a zone of lydite which varies in thickness from 3 m to 24 m. The lydite is a well bedded chert with layers from 1 - 10 cm. It is generally black, but frequently it is banded light and dark grey, the dark bands being formed by black streaks, 0,02 - 0,2 mm thick, which sometimes have outgrowths perpendicular to the stratification. Thin layers of black shales are intercalacted. The rock contains a little calcite and calcite veinlets occur, some of which are cut by quartz veinlets. Locally a limestone bed with a thickness of a few metres occurs at approximately the middle of the formation. This limestone is light grey or pink and has 10 - 20 cm thick layers. In the upper part it is somewhat nodular.

The lydite is followed by a formation of brown and red limestones, which are often nodular and sometimes brecciated. They are associated with red, green and grey-brown shales. The thickness of this formation amounts to 70 m in the Chinipro, whereas in the higher part of La Estiva (upper limb of the Estiva syncline) it is only 18 m. In the lower part of La Estiva (lower limb of the Estiva syncline) these limestones and shales are lacking altogether. Far down the slope of the Chinipro we found a fossiliferous scree boulder of which the exact position in this formation is not known. It contained:

Goniatites cf. striatus (Sow.)
Goniatites cf. granosus (Portl.)
Encrinus?

indicating an Upper Visean age.

On top of this formation, but in the lower part of La Estiva immediately on top of the lydite, follows a formation of dark grey to black limestones, which weather light grey. Its thickness varies from 50 m to 77 m. In the lower part the layers are 5 cm to 15 cm thick and calcareous shales are intercalated, whereas upwards the shales disappear and the layers have a thickness of up to 1 m. Many white calcite veins in all directions clearly stand out against the dark rock. We came across:

Orthoceras sp.
Goniatites sp.

Dalloni (1910) mentions from the Chinipro:

Phillipsia Brongniarti, Fisch.
Glyphioceras crenistria, Phill.
Glyphioceras Malladae, Barrois
Prolecanites, sp.
Orthoceras giganteum, Sow.
Poteriocrinus minutus, Roem.
Encrines

The uppermost part of the Lower Carboniferous has been eroded on the Chinipro. Only at La Estiva the series continues with some 25 m of dark grey to black limestones in layers of 0,5 - 10 cm, sometimes 30 - 50 cm, with intercalated calcareous shales. The highest part exposed consists of dark grey somewhat arenaceous shales weathering brown-black to yellowish. This upper formation has a thickness of about 90 m. It possibly represents a transition to the detrital Upper Carboniferous.

#### 2.4.1.2 Upper Carboniferous

The series we ascribe to the Upper Carboniferus is found in a zone extending from the Tormacal in the extreme northwest of the region through the valley of the Fuen Santa, forming the Robiñera and ending in the Circo de Barrosa. The Munia mountain is also formed by this series. Another exposure is found in the area of the Barrancos Sobresplucas and Clot (southern slope of the Robiñera) and a small one on the Sierra de Espierba. The series can be divided into two formations. The lower formation consists of graywackes and shales (Graywacke Formation). It attains a maximum thickness of about 500 m. The upper formation consists of green shales with intercalated sandstones and limestones (Green Formation). Its thickness seems to vary quite considerably, attaining its maximum with about 400 m.

The Graywacke Formation is a monotonous alternation of dark grey graywackes and shales with an occasional cal-

careous intercalation.
The thicknesses of the layers vary from 5 cm to 50 cm, but sometimes they attain several metres. Laterally the beds are not very persistent. Sedimentary structures are extremely rare. The low dipping cleavage in the shales indicate that the generally steep north dipping formation is overturned. The main constituent of the graywackes, varying from 60 % to 80 %, is quartz. The groundmass consists mainly of sericite and chlorite and forms 35 - 15 %. The quartz grains, which are rounded and have a moderate sphericity vary in size from 0,1 to 0,2 mm. Their dimensions seem to be proportional with the quartz content of the rock. The grains are fairly to well sorted. Rock fragments form only a few percents. Muscovite flakes are rare. Accessory minerals are zircon, tourmaline and ore. The chlorite-sericite groundmass sometimes contains a little calcite.
A pure calcite matrix was observed once. The shales are generally dense and only show some larger mica flakes along the cleavage planes. Limestones are practically lacking altogether, but occasionally limy shales and sandstones with a calcareous matrix do occur.

The Green Formation consists mainly of green shales. They have a variable and sometimes high chlorite content and show numerous pyrite crystals ranging up to 1 cm in diameter. Locally the shales are calcareous. The intercalated limestone layers do not exceed a thickness of 75 cm in general, but locally they may attain thicknesses of 1,5 m. One finds mainly greenish grey nodular limestones with brown red and purple weathering colours. The sandstone intercalations are generally thin layered, contain up to 75 % quartz in a chlorite-sericite matrix whereas sometimes the groundmass is calcite. The grain size of the quartz is about 0,2 mm but may be as small as 0,04 mm. The smaller dimensions are generally found in rocks containing less than 60% quartz. So we probably deal with all intermediates between sandstones and shales.

The contact between the Graywacke Formation and the Green Formation is quite disturbed in some places, but elsewhere there seems to be no reason to doubt a true stratigraphic superposition. Moreover, the two formations accompany, each other to the west over a considerable distance beyond the French frontier (Bresson 1903). According to Bresson the series continues up to the summit of the Chinipro. This, however, is no normal order of succession, as the Chinipro cap forms a separate tectonic unit, consisting mainly of Lower Carboniferous (cf. part 2.4.1.1 and 4.1.1.2). The occurrence of nodular limestones with green and red colours is a feature in common to both the Lower Carboniferous and the Green Formation, but the latter is far more calcareous, with prevailing red and grey colours. So a correlation of the Green Formation with the Lower Carboniferous of the Chinipro by its lithology is not either warranted.

No fossils were found. We attribute the series to the Upper Carboniferous because of the lithological similarities of the Graywacke Formation with the fossil bearing Westphalian of the Western Pyrenees.

## 2.5. PERMO-TRIASSIC

### 2.5.0. Introduction

Continental, reddish, detrital sediments occur in many parts of the Pyrenees and are assigned by various authors to the Permian, the Triassic or, indefinitely, to the Permo-Triassic. Only in few cases an exact dating on a paleontological base is possible. In some regions a subdivision was made on a lithological base or on the occurrence of nonconformities.

The following is a short survey of some areas in the Pyrenees where the red beds have been dated and / or subdivided. In grey-green and red shales which concordantly cover the Carboniferous near St-Girons (Ariège), Caralp (1903) found a marine fauna which was determined by Haug as Lower Permian (Artinskian). This was confirmed by Schmidt (1931). Delépine (1931) however, re-examined the locality and assigned the series to the Carboniferous. North of the studied region, in the Neste d'Aure valley, a marine fauna was found by Dalloni (1957) within

TABLE III
Correlation of the subdivisions of the Permo-Triassic deposits in the Western Pyrenees according to Mirouse, van der Lingen and Schwarz.

| Mirouse (1958, 1959)<br>Gave d'Aspe and Aragón Subordán | Schwarz (1962)<br>Aragón Subordán | | van der Lingen (1960)<br>North of Canfranc |
|---|---|---|---|
| | West of Cravetas | East of Cravetas | |
| "Série de la Peña de Marcanton"<br>red pelites and greenish sandstones<br>500 m | "Marcanton series"<br>red sandstones and shales with a monogenic conglomerate at the base<br>400 - 1100 m | P3<br>polygenic conglomerates and red sandy shales<br>55 m | P3<br>red conglomerates, sandstones and shales<br>1300 m |
| "Série du Pic Baralet"<br>polygenic conglomerates red sand-stones and massive limestones<br>400 m | "Baralet series"<br>thick conglomerate layers, red pelites and sandstones<br>350 - 600 m | | |
| "Séries du Somport"<br>sandstones, pelites and multicoloured graywackes<br>400 m | "Somport series"<br>reddish sandy shales<br>30 - 120 m | P2<br>sandstones<br>3 m | P2<br>reddish sandstones with intercalations of conglomerates<br>250 m |
| | | P1<br>greyish or reddish shales with limestone intercalations<br>18 m | P1<br>limestone layers alternating with grey or red shales<br>100 m |

the red bed series. Though this fauna has strong affinities to the Carboniferous, Dalloni did not doubt its Permian age because of the stratigraphic position. A Rotliegendes flora was reported by Dalloni (1913) and Schmidt (1931) from the Noguera Pallaresa valley between Gerri and Sort at the base of the Permian, which here shows the same facies as the underlying Upper Carboniferous. In sandstones covering the Dinantian near Mauléon-Barousse (north of B. de Luchon) Dalloni (1938) found a Lower Permian flora. The occurrence of the Triassic in the red bed series was demonstrated by a Buntsandstein flora found by Dalloni (1911a, 1913) near Guils (between the Noguera Pallaresa and Segre rivers).

Only in few areas a nonconformity is found within the red bed series. The nonconformity at St. Girons between the fossiliferous shales and the overlying red series as mentioned by Caralp (1903) must be considered as the base of the Permian, when taking into account the determination by Delépine (1931). A nonconformity within the red series is reported by Dalloni (1913, 1930) and Misch (1934) from the area between the Noguera Ribagorzana and the Flamisell rivers and by Ashauer (1934) from the Segre valley. These nonconformities are considered as the boundary between Permian and Triassic by these authors. In the Western Pyrenees Permo-Triassic has been studied in the valleys of the Gave d'Aspe, Rio Aragón and Rio Aragón Subordán by Mirouse (1958, 1959), van der Lingen (1960) and Schwarz (1962). A correlation of their subdivisions is given in Table III. Locally Mirouse found a slight nonconformity between the 'Série du Somport' and the 'Série du Pic Baralet'. Because of the absence of monogenic conglomerates and of red to purple-coloured sandstones, both known from the Triassic of the Basque Country, Mirouse considered the whole series to have a Permian age. In the region studied by Schwarz the 'Somport series' overlies Carboniferous and Lower Devonian sediments with an angular unconformity. The nonconformity, locally observed by Mirouse between the 'Somport series' and the 'Series of the Pic Baralet' was not met with in this region. Generally the P1 series of van der Lingen conformably overlies the Upper Carboniferous. Neither van der Lingen nor Schwarz found evidence for an exact dating, so they called the whole series Permo-Triassic.

On account of differences in petrographical constitution Virgili (1961) distinguished a Permian and a Buntsandstein series in the valley of the Noguera Ribagorzana. The upper zone consists of red quartzitic sandstones, fine quartzose conglomerates and red claystones, with at its base a quartzose conglomerate resting on a clay level. It is this clay-conglomerate contact which she considers an important break in the sequence and which she assumes to be the boundary between Permian and Buntsandstein. The upper part of the lower zone is composed of red clays and felspathic sandstones with volcanic material and is assigned to the Saxonian. Its lower part is formed by calcareous graywackes and blackish shales with coal and graphitic particles. As distinct from the Buntsandstein-base with its quartzose conglomerate, the base of the Permian is formed by a polygenic conglomerate, less sorted and with lower roundness and sphericity. It overlies the Stephanian conformibly. The flora found by Dalloni (1913) and Schmidt (1931) confirms the Autunian age of this part.

Summarizing we can say that both the Permian and the Buntsandstein are represented in the Pyrenees. Exact datings however, are few and in most cases the series are assigned to their respective systems on a lithological base only.

### 2.5.1. Permo-Triassic of the region.

The Permo-Triassic occurs in the following areas. On the Barrosa mountain and in the northern part of the Circo de Barrosa it overlies the Ordovician. In the Larri valley again red beds cover the Ordovician. Farther to the west the Permo-Triassic is not again found in the Spanish Pyrenees as far as the region south of the Collado de Pourtalet (van der Lingen,

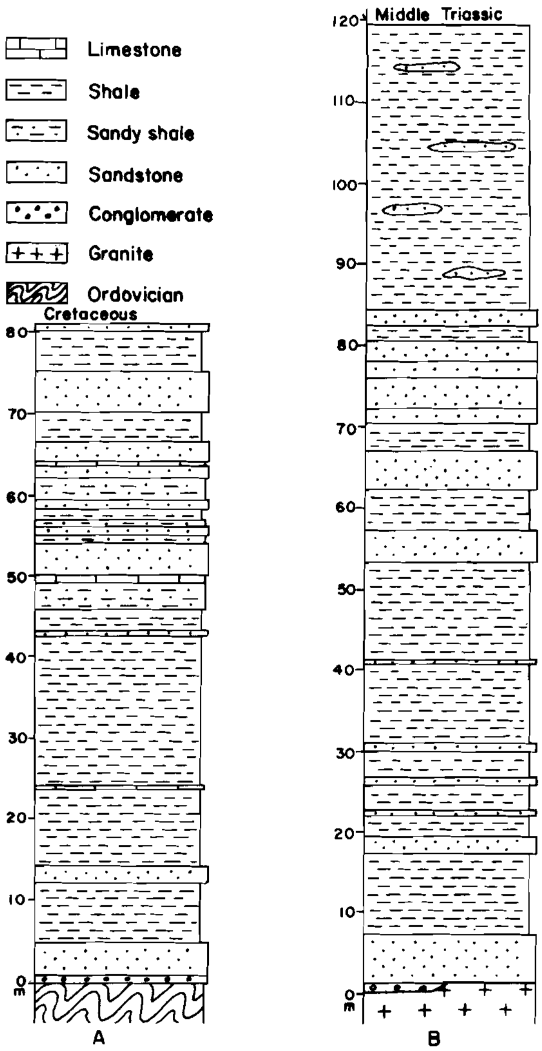


Fig. 11. Columnar sections of the Permo-Triassic in the Larri valley (A) and north-west of Bielsa (B).

1960). On the northern slope of the Real valley the series overlies the granite up to the Mota. It continues eastward to the Barrosa river where it is divided into two bands, one, bounded on both sides by granite, runs straight on, the other passes Bielsa and turns into the Montillo valley. In the extreme east of the region the two meet again. Near the Lago de Urdiceto two isolated outcorps occur. In the Larri valley the series has a thickness varying from about 80 m at the western side to 15 m at the northern side of the valley. Elsewhere the Permo-Triassic attains a thickness of about 120 m.

Of the two columnar sections (fig. 11), that taken near Bielsa is the most representative for the whole area. In many places a conglomerate bearing zone, varying in thickness from 5 cm to 3 m, occurs at the base. The conglomerates fill slight depressions in the basement surface. They split up into several layers, separated by red sandy shales when the total thickness of the zone exceeds about 15 cm. The components are generally well rounded. Their diameter does not exceed 5 cm. The pebbles we found near Bielsa are composed of quartz and quartzite, whereas Dalloni (1910) reported additional pebbles of limestone, schist and very rarely of granite from the Bielsa region. Only in the sandy groundmass we came across some rock fragments. In the Larri valley however, we did find pebbles of limestone and granite among the main mass of quartzose components.

On top of this conglomerate zone, or directly covering the basement, a formation occurs of light grey sandstones with a thickness of 6 m to 10 m. They are composed of sunangular to subrounded quartz grains (up to 75 %), grains of rock material, muscovite flakes and sericite. Accessory components are tourmaline, zircon and ore. Neither limestone fragments nor felspars are met with.

The higher parts of the series consist of an alternation of red sandstones and red shales. In the sandstones only quartz (some 40 %) and some rock fragments can be observed in thin sections. The rest cannot be determined as it is masked by the red colour. Some sandstones contain inclusions of red shale with a diameter up to 15 cm. An X-ray investigation of the shales showed quartz, muscovite, hematite and a little calcite. In the Larri valley some limy layers occur, but elsewhere no limestones are found. From about 50 to 80 m both columns show a predominant development of sandstones. The highest part consists of red shales with occasional sandy lenses.

At the top of the series in the Larri valley a grey sandstone layer is found, resembling those occurring at the base, but containing some felspar. On the Barrosa and in the extreme east of the area this sandstone appears again, associated with a conglomerate layer containing quartz components up to 30 cm in size (fig. 12). Northeast of Bielsa, in the isolated northern band of Permo-Triassic, a conglomerate, supposedly in the same stratigraphic position, is found, which contains cobbles of red shale and sandstone besides quartz compoents.

The capping bed of the Permo-Triassic is formed by green shales which grade into the Middle Triassic.

As fossils are lacking only a correlation with lithological similar series of known age would apply to ascertain the age of the red beds in our region. Taking into account the results of Virgili (1961), the occurrence of polygenic conglomerates

Fig. 12. Conglomerate in Upper Permo-Triassic NW of Barrosa mountain.

at the base argues for a Permian age of the Lower part of the series. The conglomerate layer on the Barrosa and in the eastern part of the region could represent the base of her Buntsandstein. In other parts of the area, however, the red series grade into the Middle Triassic without an intervening conglomerate layer. Moreover, because of lateral variations, a correlation between the two areas is reliable only when made in the field troughout the region between the Cinca and the Noguera Ribagorzana.

As there is no sound evidence for an exact dating we call the whole series Permo-Triassic.

## 2.6. MIDDLE AND UPPER TRIASSIC

### 2.6.0. Introduction

The Triassic of the Pyrenees occurs in the Germanic tripartite facies of Buntsandstein, Muschelkalk and Keuper. As the Buntsandstein facies has been treated together with the Permian in part 2.5, we deal here only with the Middle and Upper Triassic. The Muschelkalk and the Keuper occur rather discontinuously in the Northern and Southern Pyrenees along the border of the axial zone. In the Southern Pyrenees however, these series are not known west of our region as far as the Basque Country.

The Muschelkalk consists of limestones and dolomites with subordinate cellular dolomites and in the Eastern Pyrenees with some gypsum (Ashauer, 1934). Its thickness ranges from 20 to 80 m. In the Northern Pyrenees the Muschelkalk is absent between the rivers Salat and Saison (= Gave de Mauléon) (Viennot, 1927: Casteras, 1933).

The Keuper is represented by variegated clays and marls, by dolomite and limestone in thin layers and by cellular dolomites, rock salt and gypsum. The thickness of the Keuper is quite variable. Undisturbed sections are rare because of the highly plastic behaviour of this formation under tectonic stress.

In the Triassic of the Pyrenees ophites occur. This are intrusive rocks consisting of plagioclase and pyroxene with ophitic texture and accessory minerals. Ophites occur also in the Lower Cretaceous, associated with lherzolites (Zwart, 1953b).

Where the Triassic is overlain by the Jurassic the transition is gradual by means of dolomites and limestones (Misch,1934).

### 2.6.1. Middle and Upper Triassic of the region

The Middle and Upper Triassic of our region has been dealt with by Misch (1934) and we have little to add to his description. Fig. 13 represents a section through the Triassic south of Bielsa according to Misch.

The Permo-Triassic red beds are overlain by a couple of metres of green shales, which grade into dolomitic marls, cellular dolomites and limestones with bright yellow and brown colours. Their maximum thickness is about 25 m. These are the 'Obere Grenzschichte des Buntsandsteins' of Misch which he marked 'Röt' because of their analogy with the Upper Buntsandstein of middle Germany. This formation does not only occur along the southern margin of the Permo-Triassic red beds but also crops out in the Permo-Triassic zone northeast of Bielsa.

The formation, which is assigned to the Muschelkalk, occurs in the same areas as the 'Obere Grenzschichten des Buntsandsteins'. It consists of grey dense dolomitic limestones and dolomites, weathering of grey dense dolomitic limestones and dolomites, weathering light grey and yellowish grey. They often show a platy habit. The thickness of the Muschelkalk does not exceed 35 m. According to Misch yellow cellular dolomites and limestones occur both near its lower and upper boundary. In the Cuezo however, to the northwest of Bielsa, we found similar rocks in the basal part of the Cretaceous, which here overlies the Muschelkalk. Though no good exposure of the succession was found we doubt the existence of yellow cellular dolomites and limestones in the upper part of the Muschelkalk.

The formation which is attributed to the Keuper consists of variegated gypsiferous marls and cellular dolomites and limestones. It occurs near Espierba and locally along the Muschelkalk band. Near Espierba the Keuper was discovered first by de Charpentier (1823) who found it associated with ophite. Dalloni (1910) also mentions the ophite but Misch (1934) does not figure it on his map. We did not find again the locality. Not only Keuper, however, is found near Espierba but probably also lower formations of the Triassic. Tectonic action in this highly disturbed overthrust slice of plastic materials has mixed up strongly the various rock types.

## 2.7. JURASSIC

The Jurassic occurs all along the northern flank of the Pyrenees, whereas in the south it is lacking between the Basque Country and the river Esera (a.o. Viennot, 1927; Casteras, 1933· Dalloni, 1910; Misch, 1934). The Lower Jurassic is mainly formed by limestones and marls. The Middle Jurassic consists generally of dark fetid dolomites. A stratigraphical break comprising the Upper Jurassic and the lower part of the Lower Cretaceous was generally assumed. However, Casteras et al. (1957) showed that at least locally the series is complete from the Bajocian to the Neocomian.

No Jurassic is found in our region.

## 2.8. CRETACEOUS

### 2.8.0. Introduction

The Lower Cretaceous of the Pyrenees has about the same extension as the Jurassic. As mentioned in part 2.7, the stratigraphical break of the lower part of the Lower Cretaceous is not general. The upper part of the Lower Cretaceous, consisting of Aptian and Albian, is better developed. The Aptian consists mainly of limestones of the Urgonian facies. The Albian is formed predominantly by black marls and for the rest by limestones, sandstones and conglomerates. According

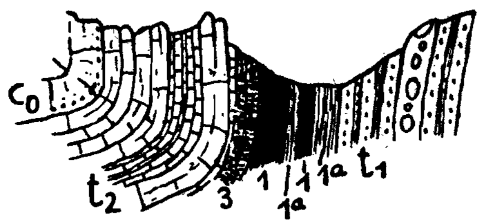


Fig.13. Section through the Tirassic south of Bielsa according to Misch (1934).

to Souquet (1964 d) the Albian is found also west of the Esera up to the Gavarnie district.

The Upper Cretaceous shows a different development in the Northern and in the Southern Pyrenees.

In the north the series starts with transgressive Cenomanian with breccias and conglomerates. This is followed by flysch type deposits up to the Upper Maastrichtian. The Upper Maastrichtian and the Danian consist of continental sediments.

In the Southern Pyrenees east of the Cinca the Cenomanian shows a predominantly marly development, but locally it is developed as limestone and then it forms part of the Upper Cretaceous limestone ('Oberkreidekalk') as defined by Misch (1934). In the region of the Segre river this Upper Cretaceous limestone comprises the Coniacian, farther west also the Santonian and in the Cotiella massif even part of the Campanian. The Upper Cretaceous limestone is covered by a marly formation which starts in the Santonian between the Segre and the Esera whereas farther west its lower boundary shifts to ever younger time-stratigraphic levels until in the Cotiella massif it is found in the Campanian. The upper part of this formation is formed by sandy marls and calcareous sandstones and represents the Maastrichtian.

The Cotiella massif ends west of the Cinca. North of it a band of rudistid-bearing limestones is found which is called the zone of Barbaruens by Misch (1934). This zone can be followed continuously westward along the southern boundary of the Paleozoic of the axial zone until it crosses the French frontier some 100 km farther west near the Puerto de Acherito The greater part of these rudistid bearing limestones are of Campanian age. Up to recently no paleontological evidence for older sediments had been found in the region between the Esera and Ara rivers (Dalloni, 1910; Misch, 1934; van de Velde, 1968). In the region of the Aragon Subordán the existence of Coniacian and Santonian was ascertained by Dalloni (1910) and van Elsberg (1968). According to Souquet (1964 a,b) also Cenomanian occurs in the Zone of Barbaruens. Moreover, Mirouse and Souquet (1964) showed the presence of Cenomanian limestones on the top of the Balaitous. The Maastrichtian of the easternmost part of the Zone of Barbaruens consists of marls. To the west their place is taken by brown calcareous sandstones of the Monte Perdido facies as defined by Misch (1934). According to van de Velde (1968), however, in the Ordesa region calcareous sandstones form only a minority in the series, which consists mainly of limestones, marly limestones and sandy limestones. West of the Ara river the Maastrichtian consists of limestones, calcareous marls and marly limestones. Usually a dolomitic level is intercalated in the middle part (Dalloni, 1910; Wensink, 1962; van Elsberg, 1968; Jeurissen, 1968).

The uppermost part of the Cretaceous in the Eastern and part of the Central Pyrenees is developed in the continental Garumnian facies (a.o. Dalloni, 1930; Ashauer, 1934; Misch, 1934). West of the Esera the Maastrichtian is covered by light grey marine limestones, which possibly are partly of Cretaceous, partly of Lower Tertiary age. However, there is little agreement in the dating of these sediments and in the names used for the various stages (Dalloni, 1910; Misch, 1934; Mengaud, 1939; Mangin, 1958, 1960; Hottinger and Schaub, 1960; von Hillebrandt, 1962, van Elsberg, 1968; van de Velde, 1968).

### 2.8.1. Cretaceous of the region

The Cretaceous of our region occurs in two separate tectonic units (major units 1 and 3, cf. part 4.1.1.0). The stratigraphy of the Cretaceous of each of these units will be treated separately after which an attempt will be made to establish the stratigraphic relations of the two series.

The Cretaceous of the first tectonic unit consists mainly of grey limestones. It is found in the Larri valley, in the Barrosa region, on the Sierra de Liena and from the upper Real to the region south of Bielsa. In the Larri valley the Cretaceous overlies the Permo-Triassic and is tectonically covered by the Paleozoic of the Gavarnie nappe. Its thickness increases from 30 m in the north to 90 m in the south. This however, is due to tectonics (cf. part 4.1.1.1.). In the Barrosa region the Cretaceous covers the Permo-Triaasic in the north, but farther south it directly overlies the granite, due to the wedging out of the Permo-Triassic. Its thickness does not exceed 20 m and generally is considerably smaller. On the Sierra de Liena isolated small and thin slabs of Cretaceous limestone occur, partly overlying the Permo-Triassic, partly immediately on the granite. Finally Cretaceous sediments overlie the Middle Triassic in a zone which extends from the upper Real region via the Pico del Cuezo to the region south of Bielsa. In this area the formation reaches a thickness of 150 m.

A columnar section of the Cretaceous of the western slope of the Larri valley is given in fig. 14. The basal layer is formed by a conglomerate with a thickness of 40 cm. It is composed of quartz components with sizes up to 1,5 cm in a calcite matrix. Fissures infilled by calcite occur in the components that are partly crushed. A similar conglomerate is found in the upper Real region where it attains a thickness of several metres. In general these conglomerates have a dark colour, caused by the dark grey to black groundmass. Farther east in the Real valley the basal layers are conglomerates or pudding stones with dolomite pebbles up to 2 cm in diameter in a groundmass of quartz sand and calcite. The pebbles consist of grey dense dolomite, weathering yellow and yellowish grey, and are obviously derived from the underlying Middle Triassic. Thus we find a correlation between the components in the basal conglomerate and the

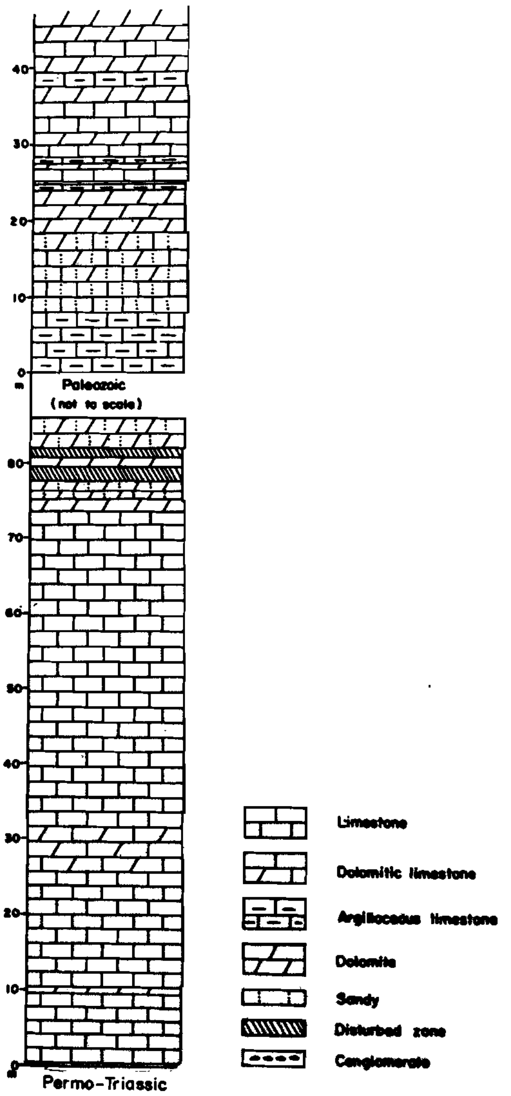


Fig.14. Columnar section of the Cretaceous of tectonic major unit 1 and part of tectonic major unit 3 in the Larri valley.

composition of the underlying rocks. The same phenomenon was observed by van Elsberg (1968) in the region of the Aragón Subordán.

In the Larri valley the basal conglomerate is overlain by some 10 m of dark grey limestones, weathering light grey. The thicknesses of the layers increase upwards from 0,25 to 1 m. On these limestones a streak of quartz pebbles with sizes from 0,5 - 1,5 cm is found. Next follow medium grey, often somewhat purplish or reddish limestones with a light grey weathering colour. Their thickness attains about 65 m. The individual thickness of the layers increases upwards to several metres. The limestones are fine grained and partly recrystallized. Generally they contain some quartz sand, but never in large amounts. Dolomitic and cellular layers are intercalated. The formation is concluded by some 12 m of medium grey dolomites, which weather yellowish grey. They abundantly contain irregularly formed concentrations of quartz sand. The dolomitic layers alternate with layers of black shales which are probably of Silurian age. This alternation is of tectonic origin and will be discussed further in parts 4.1.1.1. and 4.1.2.1.

In the other Cretaceous localities a rock sequence similar to that described above is found.

In several places rudistids occur in this formation but we found only badly preserved Radiolitidae. Dalloni (1910) mentions from our region;

Radiolites lumbricalis d'Orb
Sphaerulites
Hippurites

We found in thin sections:

| | |
|---|---|
| 'Textularia' | Polymorphinidae |
| Dictyopsella sp | Nummofallotia cretaceaus Schlumb. |
| Cuneolina pavonia d'Orb | Bolivina sp. |
| Cyclolina sp. | 'Rotalia' |
| Miliolidae | Pseudosiderolites sp. |
| Subalveolina sp. | Bryozoa |
| Lacazina sp. | Echinoidae |
| Vidalina hispanica Schlumb. | Rudistid and other mollusc fragments |
| Vidalina sp. | |

On account of this fauna a Coniacian-Santonian age is attributed to this formation.

Of the Cretaceous of the other tectonic unit (major unit 3, cf. part 4.1.1.0) only the lowermost part has been studied. It is represented in fig. 14.

In the Larri valley the series starts with dark grey argillaceous limestones followed by medium grey dolomitic limestones which upwards in the series become more and more dolomitic. The weathering colour is yellowish grey. The thickness amounts to about 18 m. The rock contains numerous concentrations of quartz sand with capricious outlines. So it is quite similar to the top part of the Cretaceous of the first tectonic unit. This formation is followed by limestones, sometimes argillaceous or cellular, and dolomites with grey and yellowish colours. The weathering colour is generally light grey. The thickness of this formation has not been established. In the columnar section only a thickness of 30 m is represented, the rest not being studied.

South of La Estiva and along the Sierra de Espierba a similar series is found. In the southeast of the region near the Maristá the lower part of the series consists of dark grey fine grained limestone with a little quartz sand. The weathering colour is yellowish to brownish grey. Dalloni (1910) mentions from this locality: Exogyra parvula Leym.

We came across: Alectryonia sp.
Exogyra sp.

In thin sections from the Cretaceous of the third tectonic unit we found:

| | |
|---|---|
| 'Textularia' | Robulus sp. |
| large agglutinants | Nummofallotia cretacea Schlumb. |
| Cuneolina sp. | Pseudosiderolites sp. |
| Miliolidae | Monolepidorbis sp. |
| Subalveolina sp. | Echinoidae |
| Lacazina sp. | Mollusc fragments |
| Vidalina sp. | |

This assemblage points to a Santonian-Campanian age.

It is not known whether the lithological differences between the rocks near the Maristá and those in the Larri valley are the result of lateral facies transitions or represent different time-stratigraphic units.

From the fossil assemblages of the two described series it appears that the first is older than the second. Moreover, in the Larri valley we found a clear resemblance between the upper part of the Cretaceous of the first tectonic unit and the lower part of the Cretaceous of tectonic unit 3. So it seems reasonable to suppose that originally the series of the two units formed a normal order of succession. The problem how this stratigraphic succession became divided into two units will be discussed in part 4.1.2.

The higher parts of the Cretaceous which have not been studied, are indicated on the map according to Misch (1934), slightly modificated after internal reports by Vreeken, Sanders and Mulder.

## 2.9. TERTIARY

As mentioned in part 2.8.0. the marine limestones and subordinate dolomites which cover the Maastrichtian are possibly partly of Cretaceous, partly of Tertiary age. Van Elsberg (1968), following Mangin (1958), and van de Velde (1968) place the Cretaceous-Tertiary boundary at the contact between the Maastrichtian and the higher formations.

Van de Velde attributes Dano-Montian and Ilerdian ages to limestones and dolomites covering the Maastrichtian in the Ordesa region. These formations are followed by flysch type deposits of Cuisian age.

On our map Dano-Montian and flysch type deposits are indicated according to Misch (1934) and internal reports by Vreeken and Sanders.

# 3. IGNEOUS ROCKS

## 3.1. GRANITE

Granitic and granodioritic rocks occur in the axial zone of the Pyrenees. A distinction can be made between leucocratic granites and biotite-granites to granodiorites.

The leucocratic granites occur only in the Cambro-Ordovician and are associated with migmatites. They are called 'granites hétérogènes', 'granites d'anatexie' or 'granites profonds' (e.g. Guitard, 1958). Raguin (1938) believed that the leucocratic granites were pre-Hercynian but Hupé (1947) pointed out that the metamorphism by these granites reached at least into the Devonian. Raguin later rallied to the idea of Hupé that the leucocratic granites are syntectonic in the Hercynian orogeny. Zwart (1953 a) and de Sitter (1956 a,b) were of the same opinion, but later Zwart (1958 a) pointed out that most of these granites have a postkinematic origin.

In contrast with the leucocratic granites, the biotite-granites and granodiorites are homogeneous and show sharp contacts. Locally they intruded into the Carboniferous ('granites en massifs circonscrits', 'granites supercrustaux'). They are generally regarded as late-tectonic. Zwart (1958 b) assumes that their consolidation took place after the post-kinematic phase of the regional metamorphism.

Intermediate types between the 'granites d'anatexie' and the 'granites en massifs circonscrits' are found also. The massif of Lys-Caillaouas is regarded as such (Raguin and Destombes, 1948; Clin, 1959· Clin et al., 1963).

According to Guitard (1958) all types of granitic rocks in the Pyrenees have been formed during the same granitisation episode. He explains the differences between the massifs by their position at different levels of the stratigraphic column, by the lithological varieties of the replaced material and by the consolidation in different phases of their evolution. Glangeaud (1958) assumes a genetic relation between the migmatites, the leucocratic granites and the granodiorites. According to him these are formed in this order by gradual homogenisation and ascension of accumulated mobile material.

In our region the so-called granite of Bielsa is found. As we made no detailed study of this massif we can only report some general features.

The rock is composed of quartz, orthoclase (frequently perthitic), plagioclase (An 25 - 35), biotite and/or hornblende. Muscovite is nearly absent. To the north it is bordered by the Ordovician. Generally the contacts are sharp but in the Barrosa region we found a zone of migmatite-like rocks (cf. part 2.1.). The contacts with the Permo-Triassic and with the Cretaceous are of sedimentary origin (cf. parts 2.5.1. and 2.8.1.). The granite is covered by a waste mantle with a thickness of some tens of metres. It consists of quartz sand in a claycy matrix. This weathering residue is particularly well preserved under the Permo-Triassic cover, which proves that it was formed before the deposition of the Permo-Triassic.

As no igneous contacts with youger rocks than Ordovician are found, we can say only that the granite is younger than Ordovician. It is older than Permo-Triassic because it has a sedimentary cover of that age. Probably the Bielsa granite has the same age that is assumed for the other biotite-granite and granodiorite massifs in the Pyrenees, *viz.* Upper Carboniferous.

3.2. HYPABYSSAL ROCKS

In our region many dykes are found which cut the Paleozoic sediments and the granite. The dykes we observed in the Upper Paleozoic were sampled but thin sections were not finished in time to describe these rocks in this paper. Most dykes are dark green or dark gray. They can be followed only over short distances.

In the Mesozoic sediments no igneous rocks were observed.

# 4. STRUCTURAL GEOLOGY

4.0. INTRODUCTION

4.0.1. General

In this section a short review will be given of the theories proposed to explain the genesis of the structure of the Pyrenees. For an account of the geographical and geological knowledge of the Pyrenees from antiquity up to 1893 the reader is referred to the thesis of Camena d'Almeida.

Neptunism was the first theory applied in the Pyrenees (Palassou, 1815; Ramond, 1801; de Charpentier, 1823). According to this theory the deposition of sediments took place in a retreating sea. Thus the oldest rocks should form the highest elevations of the mountain chains whereas at lower altitudes ever younger sediments should be found. Understandably, Ramond was quite surprised to find relatively young fossils at high altitudes in the Monte Perdido region. This was the beginning of a long-continued discussion about the structural situation of the Cretaceous and Eocene of that area.

Neptunism was followed by the theory of uplift (Dufrénoy and Elie de Beaumont, 1830; Leymerie, 1849). According to this theory mountain chains are formed by uplift as a result of volcanic activity. According to Dufrénoy and Elie de Beaumont the whole Pyrenees were originally covered by Cretaceous sediments which were separated in a northern and a southern band by the uplift of the central granite which split the sedimentary cover.

In the theory of continental sinking, on the other hand, the origin of mountain chains is attributed to the collapse of surrounding crustal parts by cooling of the earth. In this theory fault systems are predominating features (Magnan, 1874; Mallada 1878).

Meanwhile continued field work showed the importance of folds. These were explained by lateral compression as a result of shrinkage of the earth by cooling. This resulted in the idea of fold systems which stretched from one end to the mountain range to the other (de Margerie and Scharder, 1892; Roussel, 1893, 1904).

The theory of overthrust nappes was applied to the Pyrenees by Bertrand (1907, 1911, cf. Spitz, 1915) who designed a comprehensive nappe system for the whole chain. Although this theory found many adherents, there also were some fervent adversaries, a.o. Carez (1912) and Fournier (1911, 1925). Subsequent field work showed that many of the supposed facts on which the nappe theory was based were wrong and gradually the nappe system disappeared.

The '*chaine de fond*' theory of Argand was applied to the Pyrenees by Casteras (1933) and Jacob (1930). A *chaine de fond* is a mountain chain consisting of a core, folded and indurated at an earlier date, and of a sedimentary cover. The rigid core cannot be folded again and reacts on tectonic stress by faulting, whereas the mobile sedimentary mantle follows the movements of the core by folding ('plis de revêtement'). However, the cover can be detached from the core and develop folds on its own ('plis de couverture').

According to the 'chaine de fond' theory folding and hardening of the Pyrenean core occured during the Hercynian orogeny. This rigid basement now forms the axial zone. The exact age of the Hercynian orogeny is difficult to establish because of great age differences between the formations below and above the observed nonconformities. The most exact dating was made by Clin (1959) who observed a nonconformity between the Dinantian and the Middle Westphalian. Therefore he considers the Sudetic phase as the main Hercynian phase in the Pyrenees. Investigations of minor structures by geologists of Leiden University revealed the existence of four or five deformation phases, ranging from Lower Carboniferous to Stephanian (Zwart, 1960, 1963 Kleinsmiede, 1960 Zandvliet, 1960· de Sitter and Zwart, 1962 Boschma, 1963).

Folding of the post-Hercynian sediments occurred during the Alpine orogenetic period. Two phases can be distinguished. The first occurred between Lower and Upper Cretaceous and is often called the Austrian phase. In the Northern Pyrenees 'plis de revêtement' developed during this phase, whereas in the Southern Pyrenees it is represented only by a nonconformity. The main phase of the Alpine period is the Pyrenean phase, which, according to most authors, occurred in the Upper Eocene, ('plis de revêtement' and 'plis de couverture'). Mangin however, attributes an Oligo-Miocene age to the main deformation phase in the Western Pyrenees (Mangin, 1958, 1959; cf. de Sitter 1961). A similar view was put forward by Selzer (1934). Perhaps in different areas of the Pyrenees the main phase has been active at different times, or the formation of the Pyrenean structures might have been more or less continuous during the Tertiary.

Up to now most authors were of the opinion that the Hercynian core suffered no internal deformation during the Alpine period but was only affected by block faulting. So they applied the 'chaine de fond' theory, although frequently not explicity. Schwarz (1962), however, stated that the Alpine orogeny did affect the pre-Hercynian sediments by slip along bedding and cleavage planes and/or by internal plastic deformation. Van Elsberg (1968) observed that no displacement of any importance took place along the contact between the Cretaceous and older rocks, a fact which is in agreement with the statement of Schwarz. So the measure in which the pre-Hercynian sediments have been influenced by the Alpine orogeny, and consequently the applicability of the 'chaine de fond' theory, is still a matter of discussion.

The hypothesis of gravitational gliding was applied mainly to the southern border of the Pyrenees (de Sitter, 1954) 1956, b.c). Van Bemmelen (1955) designed a section of the Pyrenees according to this theory of gravity tectonics. In recent studies on the Pyrenees the idea of sliding tectonics becomes a more and more important factor in explaining the Alpine structures (Mangin, 1958, 1959· Wensink, 1962; van de Velde, 1968; van Elsberg, 1968).

### 4.0.2. Previous views on the structure of the Gavarnie nappe and the Ordesa region

In order to avoid confusion of thought we will give a definition of the regions under consideration (cf. schematized structural map).

The Gavarnie nappe is the mass of Upper Paleozoic sediments which overlie a thin sheet of Cretaceous sediments in the region of Gavarnie, Gèdre, Héas and farther to the east. Only a small part of the Gavarnie nappe extends on Spanish territory.

The Gavarnie district or Garvarnie region is formed by the valley of the Upper Gave de Pau near Gavarnie together with the Cirque de Gavarnie. Geologically the Gavarnie district is formed by part of the Gavarnie nappe with part of its basement. In the Cirque de Gavarnie occurs the northernmost part of the southern Cretaceous and Tertiary which overlie the Gavarnie nappe.

The Ordesa region is found south of the Gavarnie district. Geologically

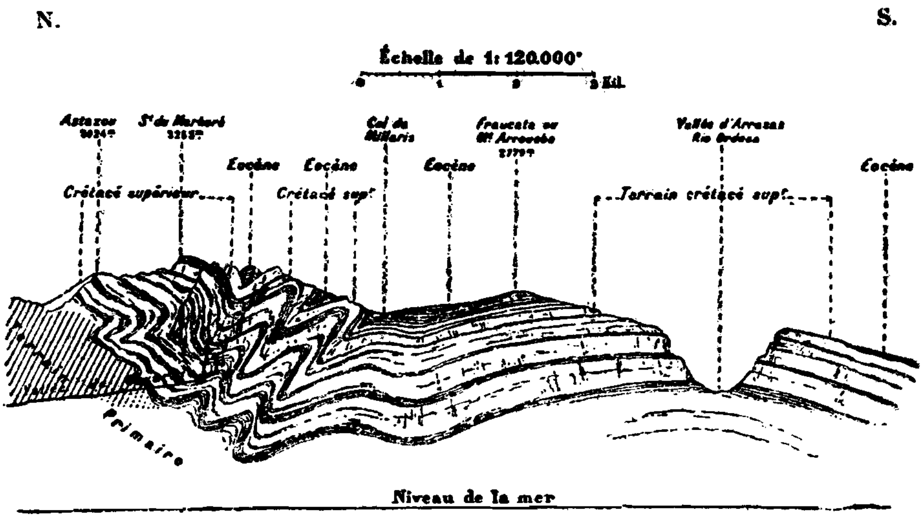


Fig. 15. Section through the Ordesa region after de Margerie (1886).

this region consists of Cretaceous and Tertiary sediments.

Sometimes a Monte Perdido region is distinguished. This region stretches from the Monte Perdido westward along the frontier crest. Thus it separates the Gavarnie district from the Ordesa region.

The investigations of the Gavarnie nappe date back to Ramond (1801) who observed Cretaceous beneath Paleozoic in the Cirque de Troumouse. He also found fossiliferous Cretaceous at high altitudes in the Monte Perdido region, which was difficult to reconcile to the idea of neptunism of that time.

According to Leymerie (1849) the fossiliferous sediments were taken to their actual position by uplift of the whole chain.

Stuart Menteath (1868) observed a south dipping fault in the Cirque de Gavarnie between the Paleozoic in the north and de Mesozoic and Tertiary in the south. From the position of the beds with relation to the fault and from the attitude of the fault he concluded that the Mesozoic and Tertairy sediments slid down from a higher position, instead of having been uplifted. Thus a century ago he put forward already the idea of downsliding which is generally accepted now.

For de Margerie (1886) the excessive altitude of the Cretaceous of the Monte Perdido region is caused by piled up recumbent folds (fig. 15). Carez (1900) admitted that this might be true but as the main reason he considered a pushing up from the south.

In 1903 appeared the study of Bresson which is of fundamental importance. He accurately described the succession of granitized Ordovician, Cretaceous and Upper Paleozoic in the valley of Gavarnie and Héas and farther east, up to the valley of Saux (fig. 16). From the position of the rudistids in the Cretaceous underneath the Paleozoic he concluded that the beds are right-side-up. Moreover he argued that the Cretaceous is autochthonous, because of the occurrence of a basal conglomerate which, apart form quartz pebbles, contains components of granitized sediments derived from the immediately underlying basement.

Stuart Menteath (1903), on the other hand, denied the existence of Cretaceous under the Paleozoic.

Carez (1904, 1903 - 1909) agreed with Bresson on the presence of Cretaceous beneath the Paleozoic. However, he considered the flat surface of the basement on which the Cretaceous rests, as caused by a thrust movement over this basement, and not by abrasion by the Cretaceous sea as supposed by Bresson. He remarked that at the entrance of the Cirque de Gavarnie the bedding of the Cretaceous is not parallel to the surface of the basement

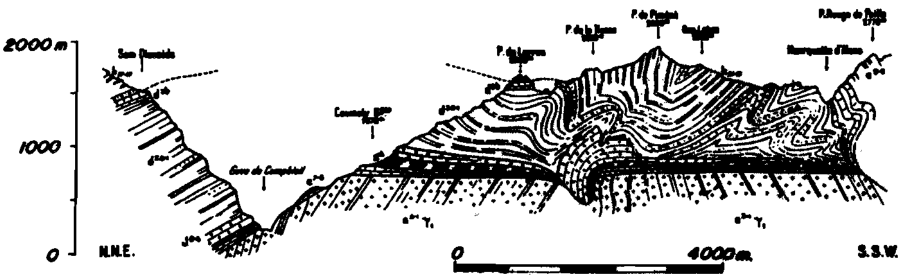


Fig. 16. Section through the Gavarnie nappe after Bresson (1903).

(fig. 17). The Upper Paleozoic of the Gavarnie region is overlain by Cretaceous of the Monte Perdido and Ordesa regions. The position of the Upper Paleozoic in between the two Cretaceous units was explained by Carez through thrusting of both Cretaceous sheets over a great distance from south to north whereas the Paleozoic moved from north to south over a short distance.

During the excursion of the Société géologique de France in the Pyrenees (Bresson 1906) it was noticed again that the lower unit of Cretaceous is autochthonous on the basement. Towards the end of the congress Carez joined this view. Moreover it was stated by the participants of the excursion that the folds of the Monte Perdido region are posterior to the thrust phenomenon and are due to subsequent epirogenetic uplift.

According to Dixon (1908) the thrust mass of Paleozoic has cleared away almost the whole of the Cretaceous series which originally covered the crystalline basement. It has left in place only the lowermost beds. The higher part of the series was crumpled up in front, in somewhat the same way a snowplough leaves a thin layer of snow on the ground and heaves up what lies immediately in front'. He considered the shear zones which occur in the Cretaceous limestone beneath the Paleozoic as caused by the influence of the thrust movement on top of it and not by thrusting of the limestone itself. Unlike Carez (1904, 1903-1909) he was of the opinion that the Cretaceous strata near Gavarnie are parallel to the basement. The upper part of this Cretaceous, however, thickens southwards, as it is cut off obliquely by the

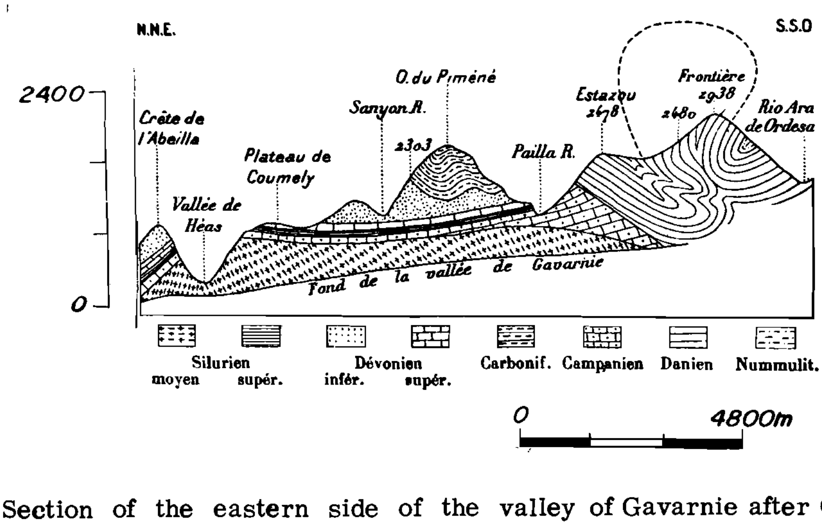


Fig. 17. Section of the eastern side of the valley of Gavarnie after Carez (1904).

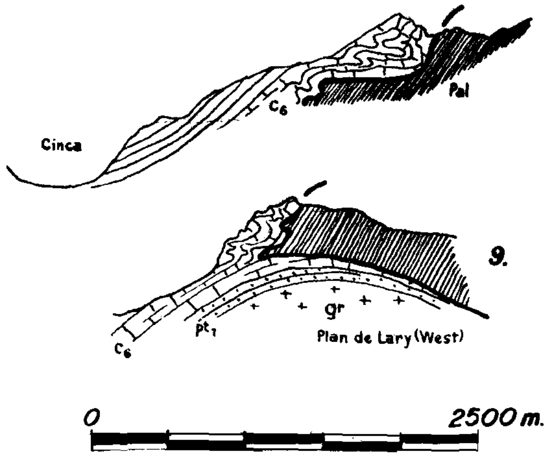


Fig. 18. Sections through the frontal part of the Gavarnie nappe in the western wall of the Larri valley according to Misch (1934).

overlying Paleozoic. He stated that 'it is difficult, if not impossible, to say where the overthurst begins or by how much its full width may exceed the 10 kilometres which is exposed'.

Dalloni (1910) established more accurately the outlines of the Gavarnie nappe on Spanish territory but did not give new views on its internal structure. His explanation of the structure of the Ordesa region as a recumbent fold is in contrast with the opinion of de Margerie (1886) who postulated an unfolded series. Selzer (1934) came to the same conclusion as Dalloni.

Misch (1934) observed no abnormal contact between the two Cretaceous units where these come into contact with each other in the Larri valley. Therefore he considered the structure as a wedge of Paleozoic, which was driven into a normal succession of Cretaceous limestones along a bedding plane (fig. 18).

According to Mengaud (1939) the basement of granitized Ordovician was pushed to the north where the granite massif of Néouvielle acted as a buttress so that its Paleozoic sedimentary cover was driven into the Cretaceous mass by counterpressure. He also figured the structure of the Ordesa region as a recumbent fold (fig. 19).

De Sitter (1954, 1956 b.c) explained the recumbent folds of the Monte Perdido and Ordesa regions by gravitational gliding. According to him the thrusting movement of the Paleozoic piled up the Cretaceous and Tertiary rocks in front of the thrust mass thus causing their downsliding.

Rutten (1955) observed the existence of an overthrust mass instead of a recumbent fold in the Ordesa region.

According to Clin (1956, 1959) the Gavarnie nappe extends farther east, than was observed earlier by Bresson. He established its eastern limit near the valley of the Neste de Louron (some 10 km east of the eastern limit of our schematized structural map).

In the Ara valley west of Gavarnie Wensink (1962) observed a dislocation ('dislocation of the Otal waterfalls'), which cuts off the Hercynian structures that lie to the northwest of it. The Paleozoic sediments between this location and the Cretaceous boundary in the south are strongly tectonized. In Visean limestones of this area Wensink observed a series of recumbent cascading folds which he thinks may be of Alpine origin. Possibly the dislocation of the Otal waterfalls represents the western boundary of the Gavarnie nappe. Its continuation beyond the French frontier is however, unknown. According to Wensink, two phases can be distinguished in the structural history of the Gavarnie district (fig. 20). 'As a result of the Alpine uplift of the axial zone the Upper Paleozoic near Gavarnie together with its cover of Mesozoic to Tertiary slid southward over the Cambro-Ordovician basement. By increasing upheaval the Upper Paleozoic sediments with their Upper Cretaceous to Eocene cover slid further southward. Thereupon other parts of the Mesozoic to Lower Tertiary rocks, which already were resting tectonically on the Cambro-Ordovician, were detached again except for the Campanian layers at their base'. Against the objection that the Lower Campanian might well be a normal sedimentary cover of Cretaceous deposits on the Cambro-Ordovician basement he remarks that there is no reason why the Cambro-Ordovician should not have been normally covered by Upper Paleozoic sediments.

According to van de Velde (1968) the Ordesa region is built up by four tec-

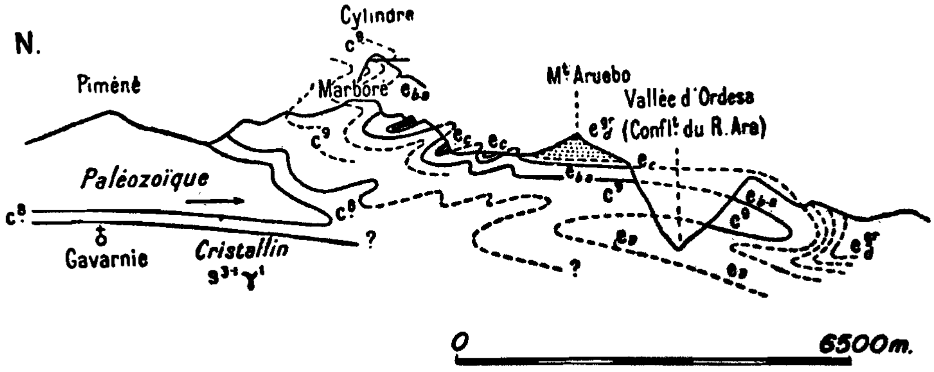


Fig. 19. Section through the Gavarnie and Ordesa regions after Mengaud (1939).

tonic units, which slid down from the rising axial zone during the Alpine orogeny (fig. 21 and schematized structural map).

Souquet (1964 c) still sticks to the idea of a recumbent fold in the Ordesa region.

In part 4.2 the facts and interpretations reported above will be combined with our own observations in order to describe the structural history of the Gavarnie nappe and related structures.

## 4.1. TECTONICS OF THE REGION

### 4.1.1. Description

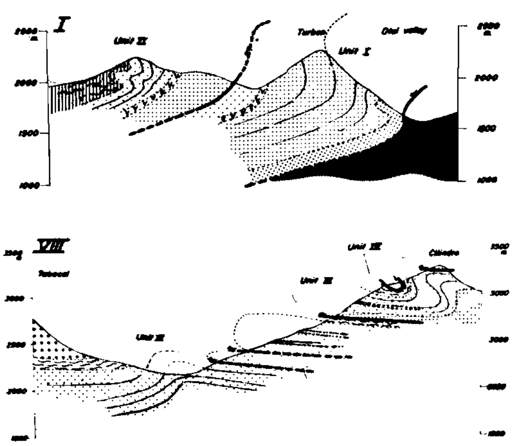

Fig. 21. Sections through the Ordesa region after van de Velde (1968).

#### 4.1.1.0 Introduction

The region can be divided into three major tectonic units: (for the situation of the tectonic elements see the inset of the geological map)

1. The crystalline basement, consisting of Ordovician and granite, together with its autochthonous cover of Permo-Triassic, Middle and Upper Triassic and Coniacian-Santonian sediments. This unit is found over a large area in the northern and eastern parts of the region and also in the west in the valley of the Larri river, where it crops out as a tectonic window.
2. The Gavarnie nappe, which consists of Upper Paleozoic sediments that overlie unit 1. Within this unit several overthrust slices can be distinguished. The Gavarnie nappe occurs in the northwestern part of the region, extending to the east, including part of the Sierra de Espierba.
3. An upper unit consisting of Cretaceous and Tertiary sediments covering units 1 and 2. It is found all along the southern margin of the map. According to van de Velde (1968) this major unit consists of four piled up minor thrust units in the Ordesa region. Only the lower part of the lowermost unit, as described by van de Velde, enters our region.

#### 4.1.1.1. Major unit 1. Crystalline basement with autochtho-

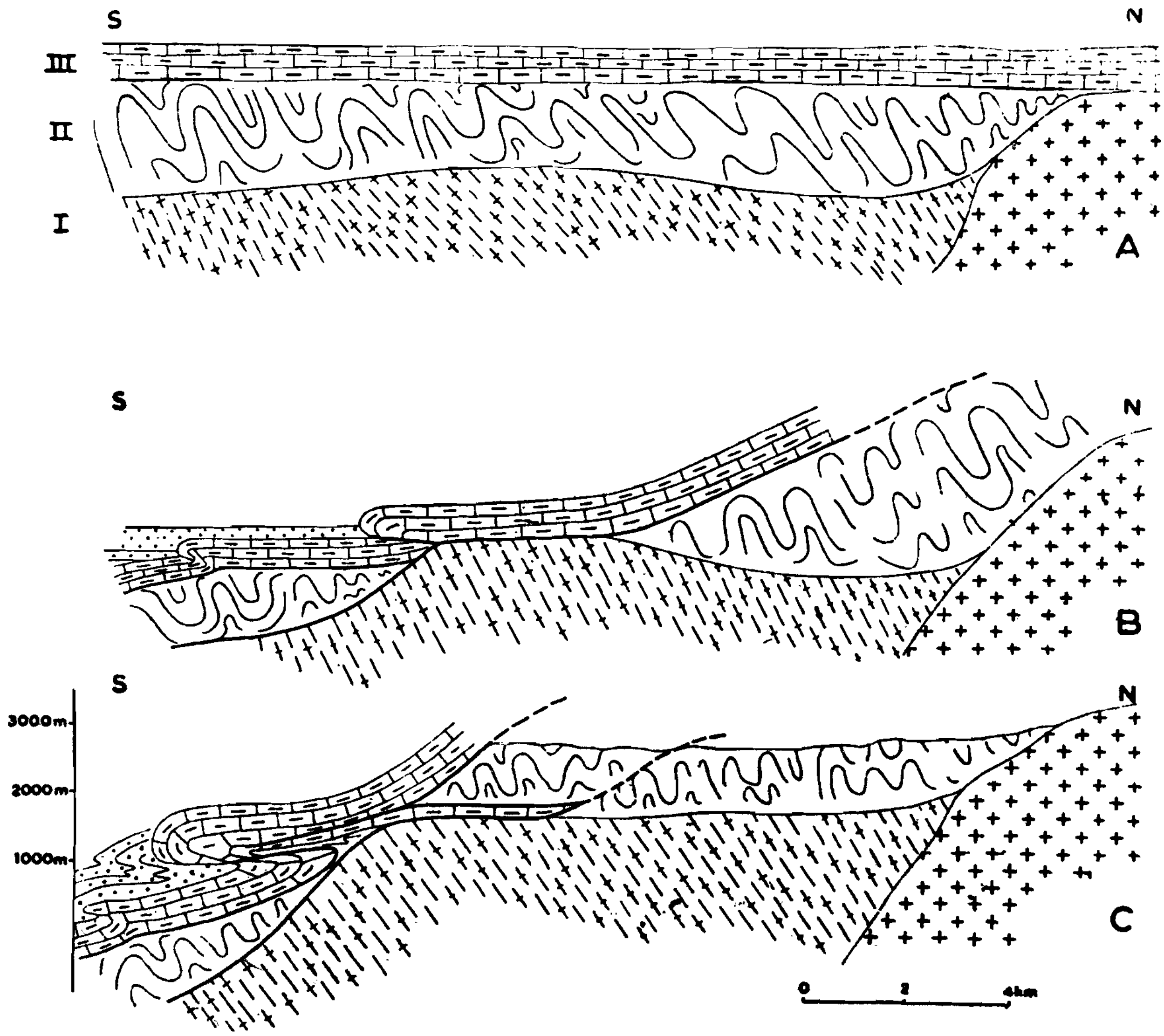


Fig. 20. Phases of movements during the Alpine orogeny near Gavarnie according to Wensink (1962).
A. I and II, Lower and Upper Paleozoic Stockwerke overlain by III, consisting of Upper Cretaceous and Eocene sediments (Upper Stockwerke): before the Alpine orogeny.
B. Uplift of the axial zone: southward sliding of III with locally attached II, III in contact with I near Gavarnie.
C. Continuing uplift; orogin of recumbent folds in III and also locally sliding of II.

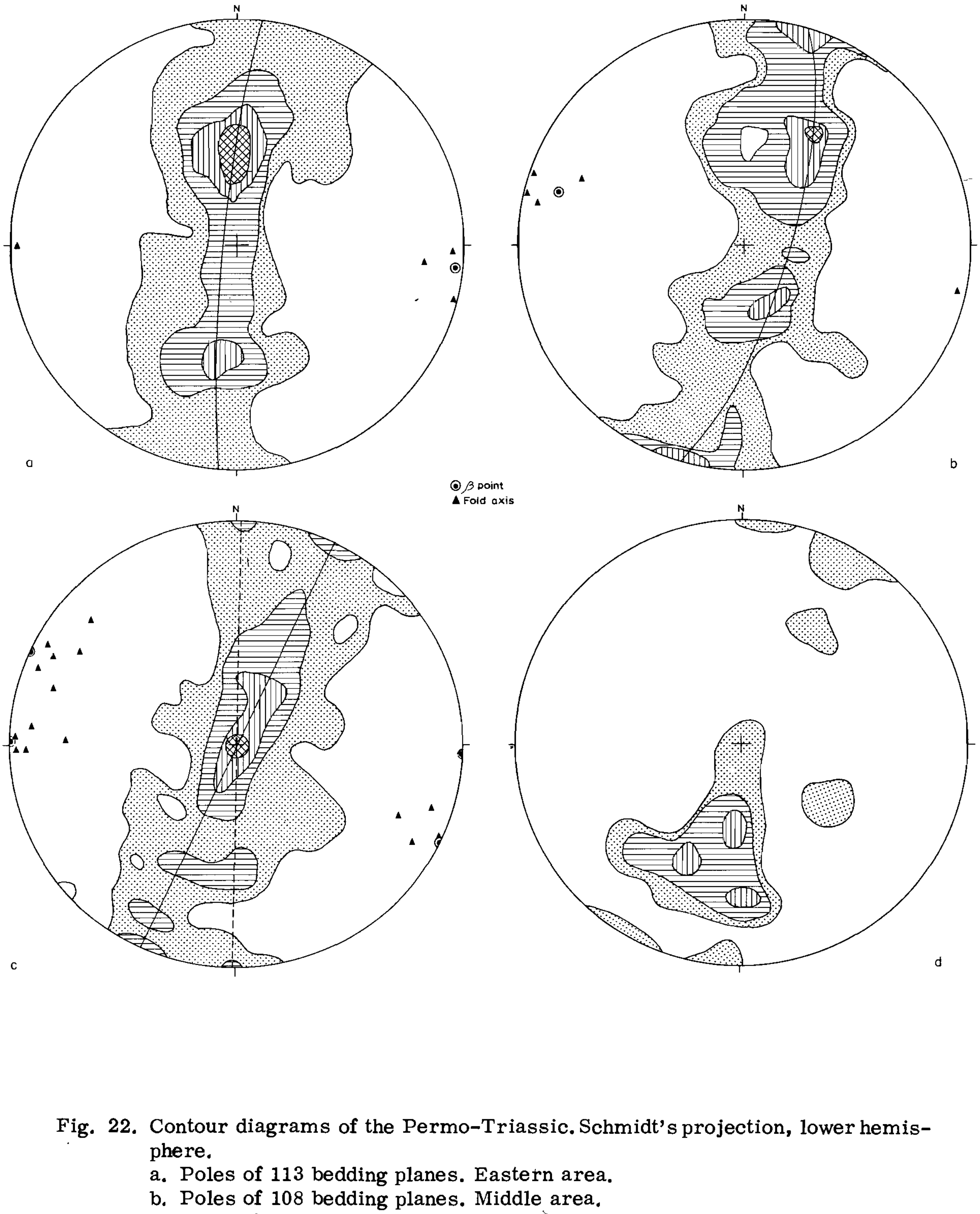


Fig. 22. Contour diagrams of the Permo-Triassic. Schmidt's projection, lower hemisphere.

a. Poles of 113 bedding planes. Eastern area.
b. Poles of 108 bedding planes. Middle area.
c. Poles of 144 bedding planes. Western area.
d. Poles of 136 cleavage planes. Whole region.

nous sedimentary cover

The crystalline basement has not been studied in detail. Only its post Hercynian tectonic features, in so far as they follow from the structures of its sedimentary cover, are described. Because of the differences between the structures of the different stratigraphic series of the autochthonous cover, the tectonics of each series will be described separately.

Permo-Triassic red beds cover the granite in the greater part of the region. In the northern part of the Circo de Barrosa and in the Larri valley the red beds overlie the Ordovocian with an angular unconformity. In the Cinca valley south of Bielsa, the Ordovician is covered by the Permo-Triassic with a nonconformity which was observed already by the Charpentier (1823).

As appears from the geological map the general trend of the Permo-Triassic changes from WNW - ESE in the west to E - W in the east. Contour diagrams illustrate these variations in structural trend. Poles of bedding planes of three areas have been plotted (fig. 22 a,b,c). In the Eastern area, covering the region of Urdiceto, Ibones and upper Bco. de Montillo, the main trend is N96°E with a plunge of 5° east. In the middle area, lying between the first area mentioned and the Cinca and Barrosa rivers, the principal trend is N 74°W with a plunge of 16°west. This western plunge can be deduced also from the map, as the base of the Permo-Triassic lies at 1000 m near Bielsa, whereas in the Upper Montillo it reaches heights between 1500 and 2000 m. The western area, west of the Cinca and Barrosa, shows a main trend of N64°W, without a plunge. A minor maximum can be discerned at N88°W, equally without plunge. The fold axes plotted in these diagrams approximate rather well the points derived from the bedding plane girdles. Poles of cleavage planes from the whole region have been plotted in fig. 22 d. Two maxima occur, one with a E - W trend, the other striking N 68°W. The interpretation of these data will be given in part 4.1.2.1.

Generally the red beds are gently folded, though zones of more intense folding do occur. The far-reaching denudation, which removed part of the Permo-Triassic left locally only synclinal structures. These find their expression on the map in narrow bands bounded on both sides by granite. Near the Lago de Urdiceto in the northeast corner of the map two such isolated synclinal structures are found. One is crossed by section XVI, the other by section XVII. The boundary of their southern limb with the granite is more or less disturbed in some places.

Another more or less comparable structure is formed by the long Permo-Triassic band, which runs from the Barrosa river north of Bielsa to the east (sections XII - XVI). We will call this the Marqués trench after the Sierra de Marqués which is situated south of this structure. Generally it is bounded on both sides by faults, but locally the contrast with the granite is not disturbed. The beds seem to form a steep monocline but the synclinal character of the structure is revealed by the occurrence of the basal light grey sandstones not only on the northern but locally also in the southern border. The synclinal structure appears also from the occurrence of Middle and Upper Triassic in the central part northeast of Bielsa (section XIII). North of Bielsa, near the western extremity of the Marqués trench, the Barrosa river cuts the botttom of the structure and the granite crops out (section XII). In the valley of the Rio Real a synclinorium is found which gradually dies out towards the west (sections IX-XI). Probably this synclinorium represents the western continuation of the Marqués trench.

Two faults occur in the area between the Rio Real and the Sierra de Liena. We will call the northern one the Liena. fault after the mines situated along it (Minas de Liena). The southern disturbance is called the Escorres fault after the Bco. de Escorres.

The Liena fault, which is ore-bearing, forms the boundary of the Permo-Triassic with the granite. In general it has a vertical attitude but locally its dip may flatten to about 50° south. Only in two places indications are found for the amount of slip. The occurrence of a slab of Cretaceous limestone against the fault on the southern

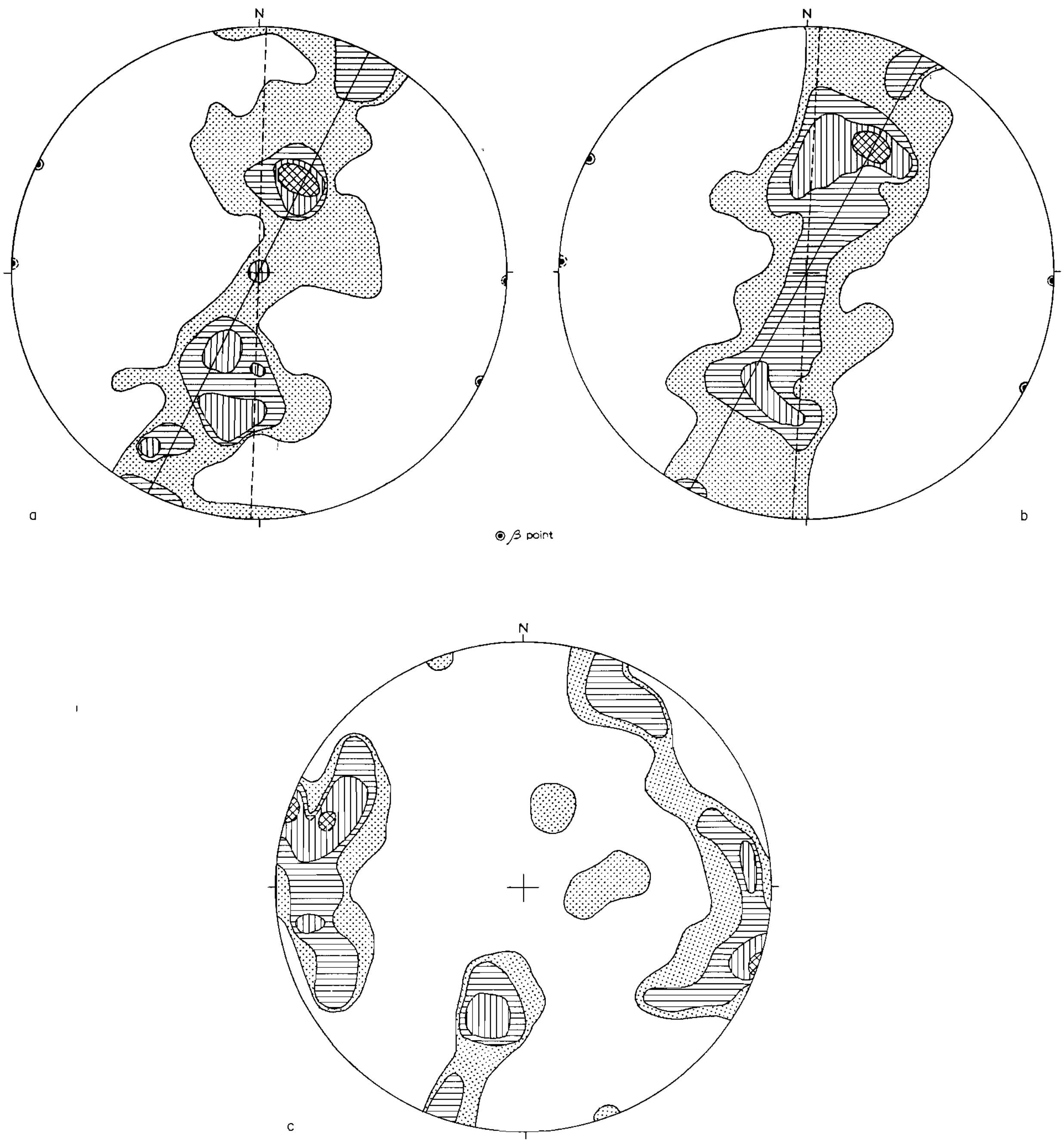


Fig. 23. Contour diagrams. Schmidt's projection, lower hemisphere
a. Poles of 51 bedding planes in Middle and Upper Triassic, Western area.
b. Poles of 93 bedding planes in Cretaceous. Western area.
c. Poles of 63 joints in Permo-Triassic and Cretaceous. Whole region.

side indicates a stratigraphic throw of at least some 100 m at this point, the southern block being downthrown (sectin IX). On the crest of the Sierra de Liena however, we find an opposite movement of 10 m at most (section X, X B). The opposite displacements at these two localities indicate a slight rotational movement with an angle of rotation of 5°. The pivotal axis of this rotation crosses an auxiliary fault, which occurs some 200 m south of the main fault. This fault might be regarded as a compensating structure, though an opposite rotation could not be established.

The Escorres fault is found about mid-way between the Liena fault and the Rio Real. In the Bco. de Ruego the Escorres fault is a thrust fault, the overthrust side consisting of granite. In the lower block a small overthrust slice consisting of Permo-Triassic and Cretaceous is caught between the main thrust plane and a drag fault (section VIII, VIII A). Towards the east the distance of thrust of the Escorres fault gradually diminishes (section IX). Finally, in the strip of Permo-Triassic sediments which extends from the Mota to the south, the movement along the fault, which here is nearly vertical, has reversed. The northern block is downthrown and thus a similar course of the direction of movement can be observed as is the case in the Liena fault. The point of zero slip in the Escorres fault lies just south of the corresponding point in the Liena fault.

Another zone where the gently undulating attitude of the Permo-Triassic is somewhat disturbed is observed southeast of Bielsa where the lower Bco. de Montillo runs through a tightly folded synclinorium (section XIV).

The Middle and Upper Triassic follows the undulations of the Permo-Triassic. Locally its incompetent members enabled the overlying Cretaceous to develop a more or less disharmonic folding with regard to the Permo-Triassic.

Poles of bedding planes of the Middle and Upper Triassic from the area west of the Cinca and Barrosa have been plotted in fig. 23. The main trend is N64°W with a minor maximum at N88°W, both without plunge. These directions are the same as those found in the Permo-Triassic of the western area.

The Cretaceous covers unconformibly Middle and Upper Triassic, Permo-Triassic and granite. There is no apparent nonconformity with the underlying sediments.

In the Circo de Barrosa and on the Sierra de Liena, where the Cretaceous has a thickness of less than 10 m, it shows a shaly structure with cleavage parallel to the bedding. This shaly structure is observed also in some places within the thicker parts of the series. In the northern and eastern side of the Larri valley we found a layer of laminated mylonite. This layer has a thickness of 15 - 50 cm and consists of alternating laminae of finer and coarser limestone. The limestone of the finer laminae is dense, whereas in the coarser laminae the grainsize varies from 0,03 to 0,1 mm. Laminated mylonite occurs also on the Sierra de Liena, but it is less developed there.

In the Larri valley the thickness of the Cretaceous increases from 30 m in the north to more than 90 m in the south (sections II - IV). This thickening is, at least locally, not gradual, but step by step other layers are added to the formation. The steps are coated by black shales, probably Silurian, and even between the layers these black shales are found. In fig. 24 a detailed section is given of the uppermost 13 m of the autochthonous Cretaceous of fig. 14. Between the Sierra de Liena and the Rio Real the transition cannot be observed, as no Cretaceous is found. In the Real valley we meet all at once with a thickness of some 150 m (sections VIII, IX).

In the Real region the Cretaceous follows the undulations of the underlying sediments, but locally disharmonic folding is observed, e.g. in the escarpment northwest of Bielsa. The incompetent members of the Middle and Upper Triassic act as a detachment zone (section XII, fig, 25). South of the Real synclinorium an anticline occurs, the steep northern limb of which forms part of the northern slope of the Pico del Cuezo (sections X, X A, XI). In the southern slope of the Cuezo the dip of the Cretaceous steepens towards the south from 20° to 45°. Along the part of the road from Bielsa to Espierba the limestone

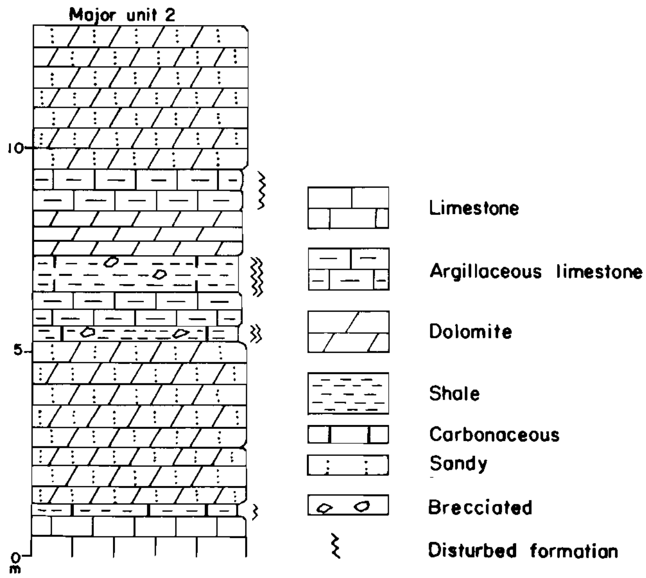


Fig. 24. Columnar section of the uppermost part of the Cretaceous of tectonic major unit 1 in the Larri valley. (Detail of fig. 14).

beds are subvertical in the southern limb of a minor anticline.

Farther east the Cretaceous is more disturbed. East of the Cinca we find the formation broken off in several small blocks which lie isolated in the Middle and Upper Triassic. Sometimes imbricated slices with anticlinal bends on their southern sides occur (sections XIV - XVI).

In fig. 23 b poles of bedding planes of the Cretaceous west of the Cinca and Barrosa rivers are represented. They show a main trend of N62°W with a minor maximum of N87°W, both without plunge. These directions fit rather well the bedding plane diagram of the Permo-Triassic of this area and that of the Middle and Upper Triassic (fig. 22 c and 23 a). In fig. 23 c poles of joints of both the Permo-Triassic and the Cretaceous from the whole region have been plotted. Three maxima can be discerned viz. N18°E, N6°W and N70°W.

#### 4.1.1.2 Major unit 2. Gavarnie nappe

In general the Gavarnie nappe overlies the autochthonous Cretaceous of major unit 1. Locally, where no Cretaceous is present, it is in contact with the Permo-Triassic or with the granite (section VIII).

Only in the Sierra de Espierba does a compliciation occur at the base of the overtrust mass (sections X, X A). Here we find overthrust slices, consisting of Permo-Triassic and Middle and Upper Triassic, which are caught between the autochthonous sediments and the main thrust plane of the Gavarnie nappe. This tectonic unit is indicated as unit A. No equivalent occurs elsewhere in our region nor has it been reported from the French part of the Gavarnie nappe. The continuation of unit A to the east is not exposed in the Cinca valley, but southwest of Bielsa Permo-Triassic and Middle and Upper Triassic are found again in anomalous position. Here the overthrust slice overlies autochthonous sediments as in the Sierra de Espierba, but is covered by Cretaceous of the third major unit, the Paleozoic being absent (sections XI, XII). South of Bielsa, along the western slope of the Cinca valley the structure disappears.

In the Gavarnie nappe we distinguish four minor units. The division into four units is based on the assumption that these units are thrust slices which originated during the Alpine movements. Their boundary planes are generally flat and cut sharply the structures on one or on both sides. Moreover, abnormal contacts occur locally within these units but they do not disturb the original stratigraphic successions to a considerable degree.

The first unit of the Gavarnie nappe is formed by Silurian and Lower Devonian. The Silurian black shales occur along the basal thrust plane of the Paleozoic mass. As a result of thrust movements, the thickness of this formation varies greatly, whilst locally it may have been squeezed out altogether. Flames of the black shales protuding into the overlying limestones can be observed in the Circo de Barrosa. Only in one place could we measure some fold directions, which, however, do not show a preferred orientation. Elsewhere the data were not reliable because of creep.

The Lower Devonian occupies the greater part of the Gavarnie nappe. In the Blancas de la Larri the Lower Devonian forms a large anticline overturned to the south which we call the Blancas anticline (sections II - VII). Within this Lower Devonian of the Blancas anticline the contact between the Lower Limestone Formation and the Detrital Formation is generally concordant on a larger scale. But in detail it may be disturbed by local discordances and friction breccias. The eastern end of the Blancas anticline is cut by a fault with a vertical separation of some 100 m (section VI, fig. 26). Along the boundary of the Blancas anticline with unit 2 the Upper Limestone Formation of the Lower Devonian occurs. It is well developed in the west near the Tormacal. In the valley of the Fuen Santa this formation is missing in the same area where in unit 2 the graywacke formation of the Upper Carboniferous is reduced or absent (sections III, IV). Near the Lagos de La Munia, where in unit 2 the whole series of the Upper Carboniferous is present again, the Upper Limestone Formation of the Lower Devonian also reappears in unit

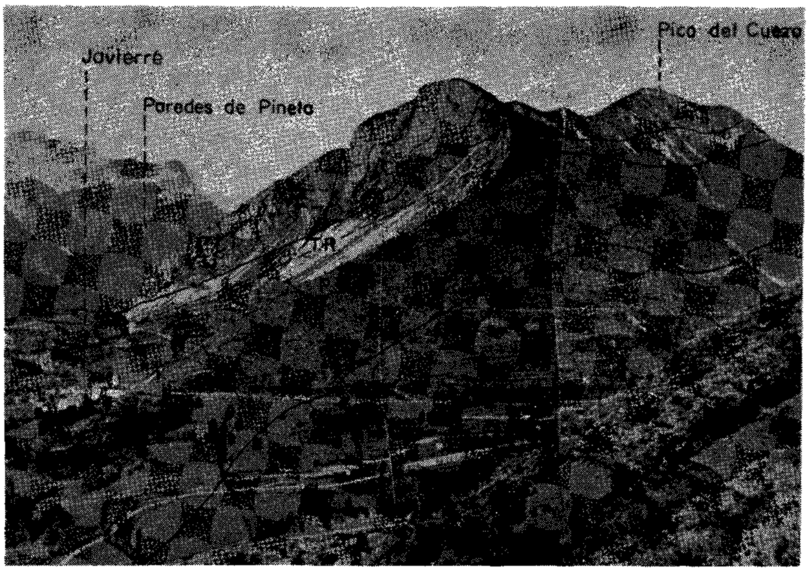


Fig. 25. The southwestern spur of the Pico del Cuezo. View towards the west. GR : granite, PT : Permo-Triassic, TR : Middle and Upper Triassic, CS : Coniacian-Santonian.


Fig.26. Fault cutting the eastern end of the Blancas anticline. View from the Lagos de la Munia. LLF: Lower Limestone Formation, DF: Detrital Formation, UC: Upper Carboniferous of unit 2.

1 and can be followed up to the Circo de Barrosa.

To the east of the Blancas anticline in the Circo de Barrosa, the Detrital Formation shows a series of smaller inclined folds (section VIII). The transition of the Blancas anticline into this quite different structure is not observable in this incompetent formation. The contact between the Detrital Formation and the underlying Lower Limestone Formation is rather flat in the Circo de Barrosa. Although it is obviously discordant, we took Detrital Formation and Lower Limestone Formation together in one tectonic unit as the original order of succession still exists. North of the French frontier, in the Géla valley and in the Cirque de Troumouse, the upper surface of the Lower Limestone Formation is sharply dentate by anticlines which pierce into the overlying Detrital Formation (fig. 27).

In following the Lower Limestone Formation from the Circo de Barrosa southwards, we find strong variations in its thickness as a result of tectonic action (section VIII, fig. 27). Near the French frontier it is about 150 m, whereas farther south it is squeezed out completely over some distance. In the eastern slope of the Robiñera there is a sudden increase in thickness to about 300 m, due to imbrication. Further southwards the Lower Limestone Formation, which up to here was subhorizontal, follows the base of the nappe in its descent towards the south. The formation is interrupted in the Real valley and appears again in the Sierra de Espierba with a dip of about 30° to the south (sections VIII - X). Here the Detrital Formation on top of the Lower Limestone Formation dips generally south with inclined folds as is the case in the Circo de Barrosa.

Farther west, however, this formation has a predominant dip to the north (sections I - IX). In the Larri valley we do not find again the Lower Limestone Formation. Here the Detrital Formation rests directly on a thin Silurian sheet or on the autochthonous Cretaceous. The Detrital Formation contains separate blocks of Middle Devonian limestones. One is found in the Comodoto with a nearly vertical attitude. It shows detailed folding manifested by minor peaks and troughs (section VII). Another slab of Middle Devonian is found in the western and northern side of the Larri valley (sections I, II).

The second tectonic unit of the Gavarnie nappe is formed by Upper Carboniferous, which extends from the Tormacal to the Robiñera. A related subunit, designated as 2a, is found south of the Robiñera in the Sobresplucas and Clot areas. The Upper Carboniferous of unit 2 forms a syncline ('Robiñera syncline', sections V, VI, VII) with in section VII the Green Formation in the core. In the Robiñera its flat lying southern flank is partly cut off by the contact with unit 1, whereas its steep northern flank is well developed here. Farther west in the valley of the Fuen Santa the synclinal axis is found farther

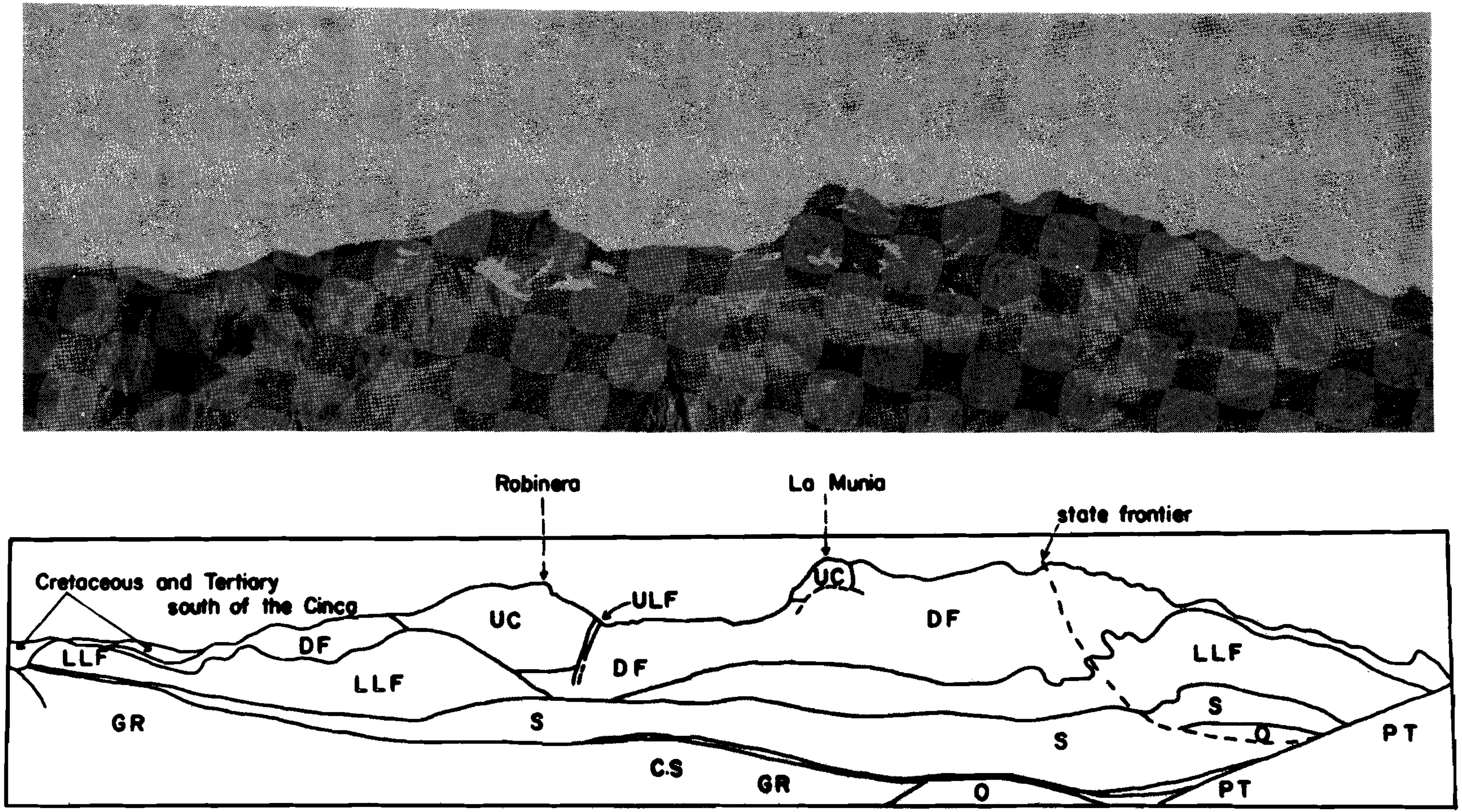


Fig. 27. Gavarnie nappe in the Circo de Barrosa. View from the western slope of the Barrosa mountain. Major unit 1 · Gr : granite, O : Ordovician, PT : Permo-Triassic, CS · Conician-Santonian. Major unit 2 ; unit 1: S : Silurian, LLF : Lower Limestone Formation, DF : Detrital Formation, ULF : Upper Limestone Formation. unit 2: UC : Upper Carboniferous.

north because of the lower topographic elevation and because of bending of the strike direction. An anticline ('Fuen Santa anticline') is found here south of the Robiñera syncline (sections V, VI). Further west the Robiñera syncline is even cut off completely by the overthrust between unit 2 and 1. Here the southern limb of the Fuen Santa anticline comes into contact with the Blancas anticline of unit 1 (section IV). In section III the Upper Carboniferous is reduced to a part of the Green Formation only. The folding of this unit appears again in the southeastern slope of the Tormacal where the Graywacke Formation reappears (sections I, II).

The top of the Munia is formed by a syncline in Upper Carboniferous graywackes, which probably can be connected with the Upper Carboniferous of the Robiñera (section VII). It then forms an erosional remnant of unit 2 floating on top of the Detrital Formation of unit 1.

The subunit 2a is formed by Upper Carboniferous with a vertical, or a steep northern dip (section VII). It is separated from the main unit by a part of the Detrital Formation of unit 1. To the west the northern boundary of unit 2a and the southern boundary of unit 2 approach each other and meet near the Collado de las Puertas. Here the two units come into contact and thus we find a doubling of the series (section VI).

The steep dipping Upper Carboniferous of the little outcrop on the Sierra de Espierba overlies the first unit and thus is in the same position as that of the Robiñera (section IX). It forms an outlier of unit 2.

The cleavage in the Upper Carboniferous is sensibly parallel with the axial planes and dips about 40° north.

The third tectonic unit of the Gavarnie nappe is formed by the greater part of the Chinipro and its spur to the west. It consists of Middle Devonian sediments which overlies Upper Carboniferous of unit 2 in the north and Lower Devonian of unit 1 in the south (sections IV, V, VI). The Middle Devonian is intensely folded, which can be observed very well in the western spur of the Chinipro (fig.7) The folds are overturned to the south and are overthrust locally.

The fourth and highest unit of the Gavarnie nappe is composed of two structures. The first, which we call the Chinipro cap, is a slice of Upper Devonian-Lower Carboniferous on the top of the Chinipro and on the spur towards La Estiva (section V, fig. 8). The second structure (unit 4a) is a syncline which occurs near La Estiva and which is also formed by Upper Devonian and Lower Carboniferous (sections V,VI).

The beds of the Chinipro cap dip gently south, but on the very top of the mountain they bend sharply to an overturned northerly dip.

The Estiva syncline is overturned to the south and lies with anomalous contacts all around in the Lower Devonian shales of the first unit. Apart from a minor fault, the western extremity of this syncline is regurlarly developed. To the east the lower limb dies out in the Lower Devonian shales, whereas th upper limb can be followed up to the Bco. Pietramula where the synclinal bend is still just exposed (section VI).

A subunit, designated as 4b, is found some 700 m to the south at the 2249 point, near the boundary with the Cretaceous of major unit 3. It consists of a little syncline, closed to the east and cut off toward the west, also built up by Upper Devonian and Lower Carboniferous. Underneath this structure another slice of the Upper Devonian-Lower Carboniferous series occurs, separated from the overlying syncline by a zone of Silurian black shales. Another isolated little unit of the same series, surrounded by Lower Devonian shales, occurs at La Estiva.

The variations in structural trend within the Gavarnie nappe as a whole follow better from the course of the strike lines on the map, than from contour diagrams of the bedding planes. From the geological map it appears that in the northwest the main trend is WNW - ESE, bending to E - W towards the east. In the southern part of the Gavarnie nappe a WNW - ESE trend predimonates. Poles of cleavage planes in the Paleozoic of the Gavarnie nappe have been plotted in fig. 28a. They show a strong E - W maximum.

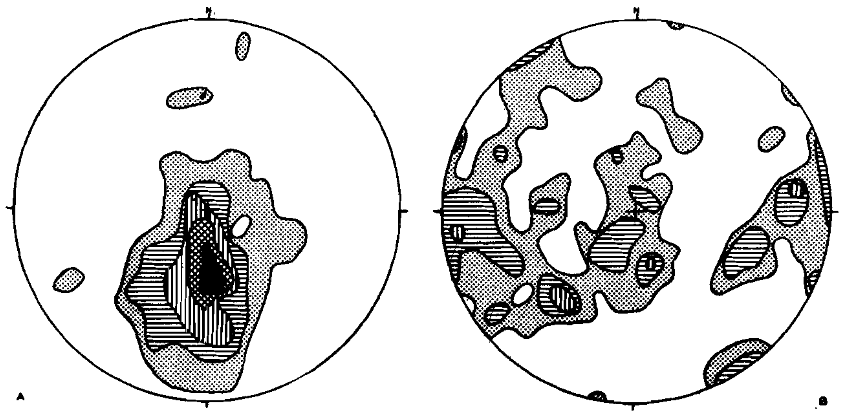


Fig. 28. Contour diagrams of the Devonian and Carboniferous of the Gavarnie nappe.
Schmidt's projection, lower hemiphere.
a. Poles of 279 cleavage planes.
b. Poles of 171 joints.

The joints plotted in fig. 28b show numerous minor maxima.

4.1.1.3 Major unit-3. Southern Cretaceous and Tertiary

The third major unit, which is composed of rocks of Cretaceous and Lower Tertiary age, covers both the autochthonous sediments and the Gavarnie nappe in the southern part of the region. It forms the continuation of the lowermost overthrust unit of the Ordesa region, as defined by van de Velde (1968). The average dip of the third major unit amounts to about 30°S north of the Cinca whereas south of the river dips of only some 10° are observed.

In the extreme west of the region the Cretaceous of the third major unit covers Paleozoic sediments belonging to unit 1 of the Gavarnie nappe. It forms several inclined folds in the western wall of the Larri valley (section I, fig. 29). The northernmost syncline is cut by a fault just north of the Capilla. The axial planes of the folds and the cleavage dip about 30° north.

At 1800 m in the western slope of the Larri valley the third major unit is in contact with the autochthonous Cretaceous, which forms the upper part of the first major unit and underlies the Gavarnie nappe. According to Misch (1934) a normal order of succession exists from the Cretaceous of our major unit 1 into the Cretaceous of major unit 3. The contact, however, is not well enough exposed to conclude with certainty to a concordance of stratification between the two units. This problem will be discussed in part 4.1.2.3.

The third major unit forms the southern boundary of the Larri tectonic window but the contact is not exposed. From the Larri valley to the east the Cretaceous of the third unit is again in contact with the Paleozoic of the Gavarnie nappe. The contact is exposed only locally, and nothing can be said about the nature of the contact plane. South of the 2249 m point west of the Comodoto the Cretaceous overlies unit 4b of the Gavernie nappe (section VI). At the contact, which dips 30° south, a conglomerate, brecciated by tectonic action, is found. It consists of components of grey limestone and lydite cemented by calcite. The next exposure of the contact is found on the Sierra de Espierba. South of the 1978 m point the Cretaceous is locally overturned, dipping 80° to the north. At its base a thin overthrust slice of Cretaceous limestone is folded in to the Lower Devonian, forming a synformal anticline (fig. 30). The higher part of the formation does, however, not join this structure and farther south it has again a southerly dip.

Farther east the lower boundary of the third unit is not exposed. From Es-

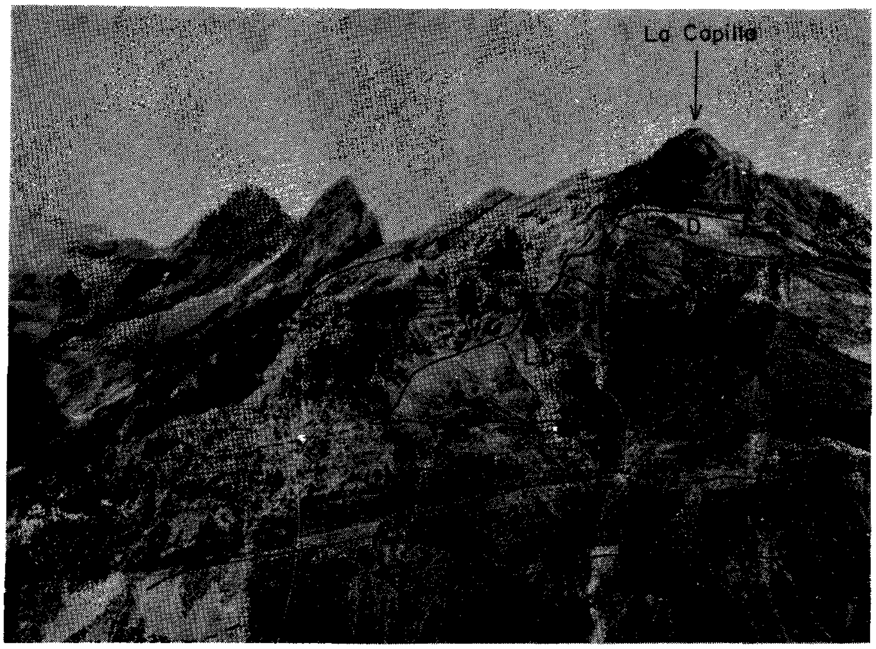


Fig. 29. Western side of the Larri valley. View from the eastern valley wall.
Major unit 1; O : Ordovician, PT : Permo-Triassic, CS : Coniacian-Santonian.
Major unit 2, unit 1; LD · Lower Devonian (Detrital Formation), MD :Middle Devonian.
Major unit 3· SC : Santonian-Campanian, M : Maastrichtian.

pierba to the Cinca south of Bielsa major unit 3 overlies the Permo-Triassic and Middle and Upper Triassic of unit A (sections X - XIII). The contact is, however, covered by alluvial and morainic depostis. On the slopes southeast of Bielsa and south of the Bco. de Montillo the contact of the third major unit with the autochthonous Middle and Upper Triassic is hidden by scree covering the foot of the scarps of the Cretaceous (sections XIV - XVII).

### 4.1.2. Interpretative

#### 4.1.2.0 Introduction

In this part we will try to give a dynamic interpretation of the structure described in part 4.1.1. We will discuss each major unit separately, after which a summary of the structural history of the region will be given in part 4.1.3.

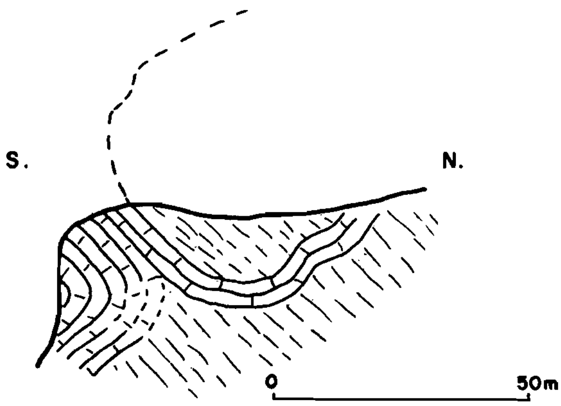


Fig. 30. Synformal anticline in Santonian-Campanian limestone in the Sierra de Espierba.

### 4.1.2.1 Major unit 1. Crystalline basement with autochthonous sedimentary cover

The first problem to be solved is the question whether the whole sedimentary cover is really autochthonous. The Permo-Triassic with its basal conglomerate filling up irregularities in the granite surface is certainly autochthonous. So is the Middle and Upper Triassic which overlies the Permo-Triassic in true stratigraphical superposition.

The Coniacian-Santonian, on the other hand, rests unconformably on Middle and Upper Triassic, Permo-Triassic and granite.
Discordant bedding between the Cretaceous and the underlying sediments was not observed, but a slight angular unconformity may exist all the same. However, the existence of a nonconformity can be explained by the transgression of the Upper Cretaceous sea over a tilted basement and so does not argue against the autochthonous position of the Cretaceous. According to Bresson (1903) the attitude of the rudistids proves that the Cretaceous beds are right-side-up and so cannot form the reversed limb of a recumbent fold. The indications that a correlation exists between the components of the basal conglomerate of the Coniacian-Santonian and the immediately underlying rocks are in favour of an autochthounous position (cf. part 2.8.1.). The same was observed by Bresson (cf. part 4.0.2.). Although bedding plane slips and disharmonic folding of the Cretaceous with regard to the Middle and Upper Triassic occurred locally, we did not observe important dislocations between the Cretaceous and the underlying rocks. We consider the shear zones in the Cretaceous limestones as been caused by the influence of the thrust movement of the Gavarnie nappe which acted as a 'traineau écraseur' (cf. Dixon 1908, part 4.0.2.).

From the above mentioned arguments it appears that the Cretaceous of the first major unit is autochthonous on the older rocks. Consequently the whole sedimentary cover of Permo-Triassic, Middle and Upper Triassic and Coniacian-Santonian is autochthonous on the basement of Ordovician and granite.

Another problem is how the folding in the sedimentary cover could develop on the rigid basement of granite and Ordovician. Of course the thick waste mantle which covers the granite could have enabled the overlying sediments to develop folds by transportation of the mobile weathered material into the anticlinal cores. From microstructures in the weathered granite it appears that internal movement did occur (fig. 31). But the amplitude of a series of folds developed in this way cannot exceed twice the thickness of the weathered layer. Now the amplitude of the folds observed attains more than 100 m locally, as against some tens of metres of the thickness of the weathered layer. So some additional mechanism must account for the folding of the sedimentary cover.

This mechanism can be found in more or less vertical differential movements of the basement. In the few places where we studied the granite we did not observe densely spaced fractures of other traces of intense internal deformation. The spacing of fractures along which differential

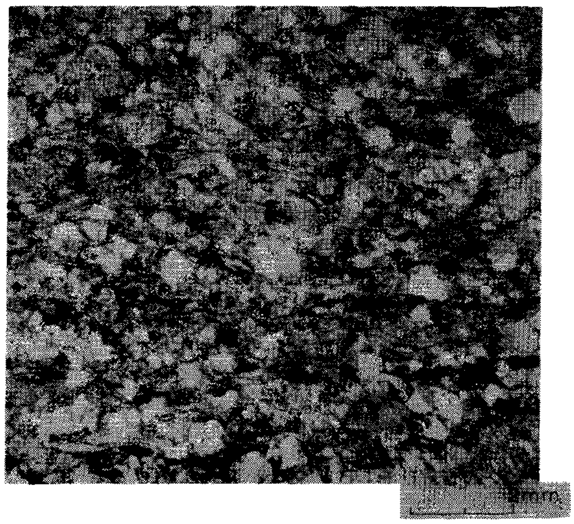


Fig.31. Thin section of weathered granite with deformation structures.

movements may have occurred must therefore be in the order of at least 10 metres. Even so the amount of slip along each fracture need be no more than a few metres to account for the fold amplitudes observed.

The consequence of this folding mechanism is that the fold pattern of the sedimentary cover must have been induced by the fracture and fault pattern of the basement ('plis de revêtement') In so far as can be deduced from the faults in which the sedimentary cover is involved, the trend of the faults in the basement is generally about E - W. Maxima in the same direction are found in the bedding and cleavage diagrams (cf. part 4.1.1.1., fig. 22, 23). We suppose that these maxima represent the folds which have been caused by the differential movements of the basement.

However, only in the eastern part of the region the E - W trend is predominant. The main maxima in the diagrams from other areas have a different direction, viz. N64°W - N74°W. This is the Alpine direction in the Pyrenees. In the middle and western parts of the region the predominant Alpine direction and the subordinate E - W direction cannot be separated regionally unless in very small subareas. As no cross-folding was observed, we think it improbable that they represent different phases. Therefore we suppose that the folds with Alpine direction were formed also by differential movements of the basement. The Alpine trend can be explained by an echelon arrangement and / or by rotational movements of the fractures and faults in the basement. Locally, however, the faults themselves have also an Alpine direction, e.g. the western extremity of the Marqués trench. The few joints measured (fig. 23c) fit the two fold trends rather well. The joints which form the main maximum can be considered as a transversal set in the folds of Alpine trend. The corresponding longitudinal set is represented by the set which strikes N70°W. The third maximum, which has a trend of N6°W, may represent the transversal set of the E - W folds. However, as only 63 measurements are available, the fit may be accidental.

#### 4.1.2.2 Major unit 2. Gavarnie nappe

The age of the thrust movement of the Gavarnie nappe cannot be established directly in our region. It must be post Coniacian-Santonian, as these sediments are overidden by the nappe. From indications furnished by the relation with major unit 3 follows that the movement occurred during the Pyrenean phase. These will be treated in part 4.1.2.3.

The direction of movement is generally assumed to be from north to south. The main argument for this view is found in the position of the third major unit and will be dealt with in the next section. As will appear from a consideration of the interrelations of the minor units, the attitude of the minor thrust planes also suggests a southerly movement. Moreover it has been put forward that the inclined folds of the Paleozoic point to a southerly movement of the Gavernie nappe (Bresson 1903, Dixon, 1908). It is possible, however, that the folding within the units of the Gavarnie nappe occurred already during the Hercynian orogeny and has nothing to do with the thrust movement. So a conclusion from these folds is warranted only in case their Alpine age is ascertained.

Therefore we must first try to establish the age of the folds in the Paleozoic of the Gavarnie nappe. In the structural trend of the Paleozoic two main directions can be discerned, viz. one about E - W, the other WNW - ESE. The first is a well known Hercynian direction. The WNW - ESE direction is parallel to the Alpine trend. This direction is, however, also known from regions in the Pyrenees where an Alpine influence is improbable. Moreover it has been demonstrated elsewhere that this trend represents a phase of the Hercynian orogeny (de Sitter and Zwart, 1962; Zwart, 1963; Boschma, 1963). The cleavage diagram (fig. 28a) shows a strong E - W orientation. Its maximum is found at a dip of 28° north. This corresponds to the E - W orientated folds with north dipping axial planes. However, the diagram gives no hold to distinguish between Hercynian and Alpine deformation. In the field we did not find strike-flexures or knick zones as for instance were observed by Schwarz (1962). Our joint diagram (fig. 28b) is too complex to be interpretable. So we are unable to ascertain the age of the folding in the Paleozoic of the Gavarnie nappe and consequently the attitude of the folds cannot be used as a criterion for the direction of the thrust movement.

The only evidence for an Alpine influence on the Paleozoic is found at its boundary with the third major unit. The folding of the Cretaceous in the western wall of the Larri valley (section I) was probably the result of differential movements of the Paleozoic along bedding and cleavage planes. Also in the Sierra de Espierba the Paleozoic is locally seen to be involved in the folding of the Cretaceous, forming the core of a synformal anticline (part 4.1.1.3, fig. 30).

The idea that the four units of the Gavarnie nappe are thrust slices which originated during the overthrusting of the nappe is suggested by the nature of their boundary planes. The plane which separates the first from the second unit extends over large parts of the Gavarnie nappe and can be followed up to the region where the nappe comes up from a deeper level in the north. To our knowledge thrust planes of a similar extent and equally clean-cut do not occur in the Pyrenees in regions which were only affected by the Hercynian orogeny. So we suppose that the basal thrust plane of the second unit developed contemporaneously with the movement of the Gavarnie nappe as a whole.

The continuation of our third and fourth units outside our region is not known. So we cannot ascertain whether their basal thrust planes extend as far as that of the second unit. The position of the third unit on both the second and first unit shows that the third unit reached this situation when unit 1 and 2 were already individualized. Consequently its basal thrust plane developed during the same phase or even later than the thrust plane between the first and second unit. The same holds good for the fourth unit with regard to unit 3 and 2. It follows that all four units of the Gavarnie nappe are thrust slices that originated during the movement of the Gavarnie nappe as a whole. In the following these units are discussed more in detail.

Along the basal thrust plane of the Gavarnie nappe the incompetent Silurian pelites formed the lubricating layer for the overlying thrust mass. In the Larri valley the Silurian shales were pressed in between the upper beds of the autochthonous Cretaceous where these are truncated by the thrust mass. The flames of Silurian which project into the overlying limestones in the Circo de Barrosa point to the South, thus showing a southerly movement of these limestones with regard to the Silurian.

The basal thrust slice (unit A), consisting of Permo-Triassic and Middle and Upper Triassic, was picked up from the autochthonous sediments during the thrust movement. It probably moved over a shorter distance than the Paleozoic.

Within the first unit of the Gavarnie nappe differential movements between the Lower Limestone Formation and the Detrital Formation of the Lower Devonian are shown by the occurrence of local nonconformities, by friction breccias and by disharmonic folding. These phenomena may be partly of Hercynian origin.

The second unit plunges into the Detrital Formation and butts against the Lower Limestone Formation of the first unit. Its southern boundary with the first unit is an underthrust which was overturned to northern dip in its lower part. The position of unit 2 with regard to the first unit points to a southern movement of the former over the latter. A slice of the Upper Carboniferous was detached from the main mass, to form minor unit 2a. Part of the detrital formation of the first unit was jammed in between the units 2 and 2a.

The third unit overlies the second unit, but does not follow its downward plunging. It slid farther south over the Detrital Formation of the first unit.

The two structures of the fourth unit cannot be connected directly because of the different development of the series in the Chinipro cap and in the Estiva syncline. Probably they are originally remote parts of the same minor nappe, of which the frontal part became entrapped in the Detrital Formation of unit 1 and formed the Estiva syncline. The rear part moved somewhat farther south. In this way the two parts, which were deposited at some distance from each other, moved closer together. The subunit 4b may be interpreted as the eastern termination of the Estiva syncline which has become detached and was transported southward.

#### 4.1.2.3 Major unit 3. Southern Cretaceous and Tertiary

Our third major unit forms part of the zone of Barbaruens as defined by Misch (1934) (cf. part 2.8.0). The structures of this zone are interpreted as gravitional by several recent authors (de Sitter, 1954, 1956; Wensink, 1962, Van Elsberg, 1968; van de Velde, 1968). According to van de Velde the overthrusting in the Ordesa region originated by downsliding of the Upper Cretaceous and Tertiary sediments from the rising axial zone during the Alpine orogeny. The inclined folds we observed in the western slope of the Larri valley fit in well with this picture of gravitational sliding.

According to Misch (1934) the autochthonous Cretaceous and the Cretaceous of our third major unit form a normal order of succession where they come in touch in the Larry valley. He is of the opinion that abnormal contact exists between these two units in the Larri valley. Therefore he considered all of the Cretaceous as autochthonous and interpreted the Gavarnie nappe as a wedge of Paleozoic, which was driven into this normal succession of Cretaceous sediments.

We found the contact hidden by scree. But during an excursion in the valley of Gavarnie we observed a fault contact between the same two units in the Cirque de Gavarnie. This fault was reported already by Bresson (1903, fig. 16) and by Carez (1904, fig. 17). Accordingly we assume that also in the Larri valley the Cretaceous of the third unit cuts off the autochthonous Cretaceous, but possibly under a very low angle.

The apparently normal stratigraphic relation between the two Cretaceous units, which led to the erroneous interpretation by Misch, can be explained as follows. As we saw in part 2.8.1. the authochthonous Cretaceous is of Coniacian-Santonian age, the Cretaceous of the third major unit of Santonian-Campanian age. Moreover the upper part of the autochthonous Cretaceous is lithologically similar to the lower part of the third unit. This suggests that the two units originally formed a continuous series. We assume that the Cretaceous of the third major unit was deposited at a considerable distance from its actual position. In this view the top layers of the authochthonous Cretaceous and the basal layers of the allochthonous formation might have been deposited either shortly after another, or contemporaneously. This can be explained by an onlap of the Cretaceous on its basement as a result of a gradual transgression of the Cretaceous sea from south to north. In this way the sedimentation in the south started in the Coniacian, but farther north the basal layers are of Santonian or even Campanian age.

During the Alpine orogeny the Upper Cretaceous and Lower Tertiary rocks of the northern region, where sedimentation only began in the Santonian-Campanian, moved southward on the back of the Gavarnie nappe. In our region the basal thrust plane of the nappe happened to cut off the underlying autochthonous represented by the base of the allochthonous Cretaceous of major unit 3. The Mesozoic to Tertiary sediments, which earlier had ridden on the back of the Paleozoic of the Gavarnie nappe, then slid down into the foreland. Their base and the top of the authochthonous Cretaceous, which are of about the same age, came into tectonic contact. This mechanism explains the tectonic interpretation of the delusive Cretaceous sequence in the Larri valley.

In this picture of gravitational downsliding of the Cretaceous from the Paleozoic into the southern foreland with truncation of the authochthonous sediments, the only possible direction of movement for the Gavarnie nappe is from north to south.

The Upper Cretaceous to Lower Tertiary cover moved farther southwards than the Gavarnie nappe. This 'Voraneilen der jüngeren Schichten' is typical for gravity tectonics. It is true that the basal thrust plane of the Gavarnie nappe has not a universally south dipping attitude. On the contrary it has a low dip to the north over the greater part of the Gavarnie nappe. This seems to contradict gravitational sliding of the Gavarnie nappe. However, it may be

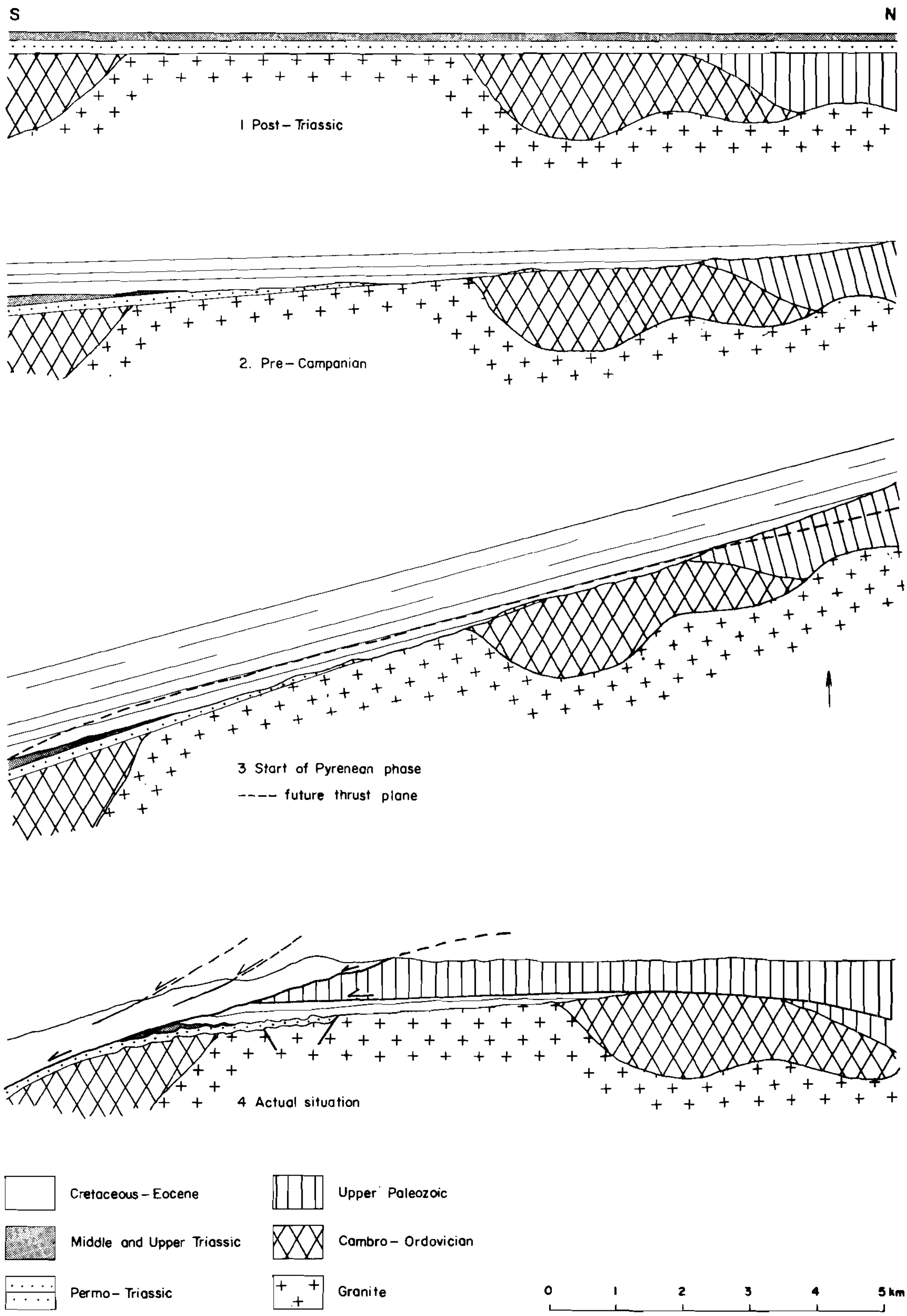


Fig. 32. Structural evolution of the Gavarnie nappe. For explanation see text.

assumed that the axis of uplift in the basement shifted from north to south. A shifting to the south of the tectonic activity during the Alpine orogeny agrees with the fact that the orogenetic phase between the Lower and Upper Cretaceous is of far greater importance in the Northern Pyrenees than it is in the south. The low northern dip of the basal thrust plane can be explained equally well by subsidence of the axial zone after the downsliding.

It follows that no conclusive evidence exists for a gravitational origin of the Gavarnie nappe alone. It can be explained as well by lateral thrust, depending on the geotectonic theory one adheres.

From the relations between our three major units we concluded that the overthrusting of the Gavarnie nappe occurred simultaneously with the movement of the third major unit. As the downsliding of the southern Cretaceous was dated as post-Cuisian in the Ordesa region (van de Velde, 1968), the same, or roughly the same, age can be assumed for the Gavarnie nappe.

## 4.2. CONCLUSION

Summarizing the facts and interpretations exposed in the previous parts, the structural history of our region can be described as follows (fig. 32.)

1. After the Hercynian orogeny and the intrusion of granitic batholiths the region was base-leveled. On the peneplain the Permo-Triassic and Middle and Upper Triassic sediments were deposited.
2. It is not known whether deposition took place in our region during the Jurassic and Lower Cretaceous. Anyhow, possible sediments were removed by subsequent erosion which also affected the Upper and Middle Triassic and Permo-Triassic. The basement was slightly tilted. This tilting was possibly related to the tectonic phase between the Lower and Upper Cretaceous. The Upper Cretaceous sea transgressed from south to north over the tilted basement. Thus the oldest sediments, of Coniacian age, were deposited only in the south whereas the younger beds reached ever farther to the north, until in the Santonian the sea covered also the Upper Paleozoic in the north.
3. After the remainder of the Upper Cretaceous and the Lower Tertiary were deposited, the axial zone started to rise.
4. In reaction part of the Upper Paleozoic together with its Mesozoic to Tertiary cover, slid down to the south. The cover of Cretaceous to Eocene rocks moved farther southwards than the Paleozoic, and slid into the marginal trough. In the Ordesa region these formed four piled up thrust slices. In front of the Gavarnie nappe the base of the detached mass of Cretaceous rocks came into contact with synchronous autochthonous Cretaceous. The part of the sedimentary cover that remained authochthonous reacted on the differential movements of the basement by forming 'plis de revêtement'.

In this account the movement of the Gavarnie nappe has been considered as gravitational. If lateral thrust is assumed for the origin of the thrust mass of Upper Paleozoic, the story is similar. In that case the movement of the cover of Mesozoic to Tertiary age only is ascribed to gravity (cf. de Sitter, 1954, 1956b,c).

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