THE GEOLOGY OF THE UPPER SALAT AND PALLARESA VALLEYS, CENTRAL PYRENEES, FRANCE/SPAIN

J. ZANDVLIET

(Pre-issued 24-2-'60)

CONTENTS

ABSTRACT

A sequence of more than 4000 m of marine sediments, mainly unfossiliferous and apparently without any unconformities, range in age from probable Cambrian to pre-Hercynian Carboniferous. The lower formations are of neritic facies and there is no indication of a Pyrenean basin before the Devonian, the deposits of which are much thicker in the centre of the present axial zone than on the margins.

A relatively thin band of black shales of Silurian age acted as a tectonic lubricant and thus its presence resulted in a marked disharmony between the infra- and supra-structures. The infra-structure is very complicated and consists of multiple composite anticlinoria and synclinoria in which the tectonic shortening is mainly accounted for by the smallest fold unit — the tightly isoclinal micro-folding. Fold axes and b-lineations of this cleavage micro-folding plunge consistently in the same direction over sharply delimited areas of up to hundreds of square km. In the supra-structure the micro-folding plays a much smaller role than in the infra-structure; the folding is less composite and high-amplitude folds of some 1000 times larger dimensions provide a real shortening of about 40—50 %. A thinning of roughly 20 % of the Devonian sediments by compression has been calculated from fracture phenomena in thin slate intercalations in limestone beds. This thinning thus gives an apparent shortening which is greater than is actually the case.

The northern boundary of the main dome of Lower Palaeozoic is formed by a steep flexured zone with a throw of at least 2 km. Adjacent to this flexure on the northern side is a zone of steep isoclinally folded Upper Palaeozoic rocks cut by an E—W branch of the North-Pyrenean fault system, resulting in a tilting of both blocks towards the north. The main dome is flanked to the south by a deep Upper Palaeozoic syncline of which the southern flank in the Monseny area passes into recumbent folds directed towards the south.

After the main folding arching caused a fanning out of the originally vertical structure elements. Genetically related to this fanning is a late fracture cleavage (knick-zones) which displaces the syn-tectonic cleavage in such a way as to indicate a dilatation in a N—S direction.

A subsequent, yet pre-Triassic vertical jointing, visible on aerial photographs, shows a complicated picture with many strike maxima of poor regional consistency. These major lineaments greatly influence the drainage.

Important remnants of pre-glacial denudation surfaces have been preserved and lie at 2400—2600 m and 1850—2350 m altitude. The lower altitudes of these ranges are found towards the west of the area. The snow line of the last glaciation — derived from the lowest level of nivation cirque excavation — lay at 1500—1600 m in the north rising to 2100—2200 m in the south.

A purely petrographical description is given of granodiorite batholiths, dykes, sills and basic rock intrusions. The talc of Fonta probably originated from dolomite by metasomatic addition of large quantities of hydrothermal quartz which penetrated from the granodiorite intrusion along a fault plane. The galena and sphalerite occurrences of Carbauère are also connected with a fault.

INTRODUCTION

AVANT PROPOS

The following thesis forms an independent contribution to a detailed geological survey of a more than 100 km wide section through the Palaeozoic core of the Central Pyrenees (axial zone + satellite massifs), directed by Dr L. U. de Sitter, professor in structural geology and Dr H. J. Zwart, both of the Geological-Mineralogical Institute of the State University of Leiden, Holland.

In this survey, which will contain 9 sheets, the Salat-Pallaresa area (sheet 5) is the central part (see index map II to the geological map 1:50.000). It is situated in its entirety in the axial zone west of Andorra between northern latitudes 42°30′36″ and 42°48′09″ and eastern longitudes 4°40′ and 5°08′15″ and covers an area of about 1260 km^2, of which 4/5 is in Spain and the rest in France. This area was largely mapped by the author himself. In the marginal parts use was made of unpublished internal reports of geological students of the above-mentioned geological institute, often supplemented with new data (see index map I to the geological map and p. 126).

The purpose of the mapping was to obtain as complete a geological map as possible, with special reference to some special tectonic problems. Use was made of the following topographical maps and aerial photographs:

French part: Carte de la France 1:20.000, published by the Institut Géographique National of the Ministère des Travaux Publics et des Transports. Sheets: Pic de Maubermé (XX-47) no. 4, Aulus les Bains (XX-48) no. 1 to 8 and L'Hospitalet (XXI-49) no. 1. These maps were found to be very accurate. In the field the author also had at his disposal aerial photographs (13×18 cm) to a scale of 1:20.000 or 30.000 of the above-mentioned institute (run Aulus-Vicdessos-Ax).

Spanish part: Mapa Militar de España 1:50.000, published by the Instituto Geográfico y Catastral. Sheets: Isil (149), Noarre (150), Esterri de Aneo (181), Tírvia (182), Andorra (183), Sort (214), Seo de Urgel (215) and Bellver (216). The editions after 1945 are on the whole accurate enough for the purposes of geological mapping to that scale. The author had the opportunity to consult in the field for a short time aerial photographs of the eastern part of the Spanish area in connection with a geological reconnaissance for a hydro-electric company. From this material the topographic base of the geological maps was composed. The maps in the text and the glacial and physiographic map 1:100.000 are reductions of the 50.000 map.

In order to facilitate map-reading a list of topographic names — used in this paper — is given together with their grid references or geographical coordinates on p. 120.

N.B. All measurements of cleavage, stratification, jointing, lineation etc. are in degrees (360° system).

The Salat-Pallaresa area is divided into two areas with entirely different climates by an almost inaccessable high crest (2600—3140 m), which runs

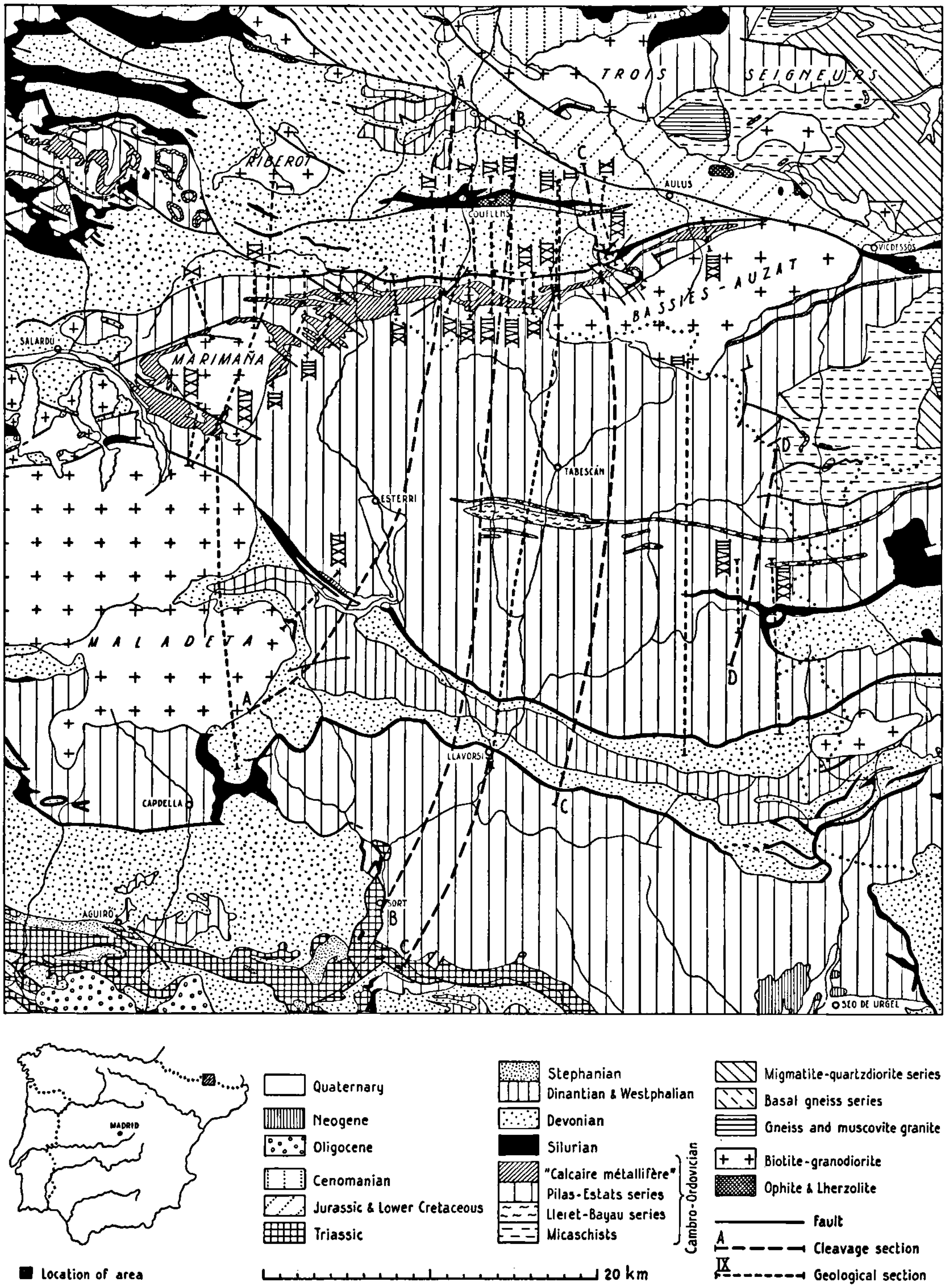


Fig. 1. Geological key map of the sheet 5 and adjacent areas.

from north-west to south-east and which also forms the boundary between Spain and France. The northern slope is rather cool and extremely humid, owing to low hanging clouds frequently filling up the valleys from the north and to the high precipitation. In summer the landscape has a fresh green appearance. The woods consist mainly of beeches. The southern slope on the contrary is dry and warm and the dominant colour is yellow. Here birch-woods predominate. Pine-forests are mostly confined to the granite massifs.

The steep northern slope is drained for the greater part by tributaries of the Salat and in the extreme north-east also by the Ariège. The run off flows via the Garonne into the Atlantic Ocean. The southern slope is the drainage-area of the Pallaresa and its tributaries, in the extreme east also of the Sègre. The two rivers join and debouch into the Ebro, which drain into the Mediterranean. The south-western part of the Marimaña granite drains partly underground into the Garonne.

Owing to the great range of altitude (600—3141 m) and the rugged relief, the area is well exposed, in the higher woodless parts sometimes almost completely.

The accessibility in the valleys, especially on the Spanish side has been improved in recent years by the laying out of a number of motor-roads. Accessible now from France are: Salau on Salat (860 m), Estillon on Alet (820 m) and Aulus on Garbet (700 m) and from Spain: Puerto de la Bonaigua (2072 m), Alós de Isil on Pallaresa (1270 m), Servi and Gabas on Unarre (1420 m), Tabescán on Cardós (ca 1120 m), and Areo on Vall-farrera (ca 1420 m).

The area is thinly populated, the villages are small. On the Spanish side the population lives on horse- and cattle-breeding (cows, sheep and goats) and a little agriculture (grain and potatoes). Esterri de Aneo is the centre of the district with prosperous shopkeepers. The French Pyrenean population, decimated by the migration to Toulouse, is poor, since a few cows, some sheep and the hay-making are here the only means of subsistence. The watering-place of Aulus with thermal springs has some bigger hotels.

PREVIOUS AUTHORS

Since Palassou in 1784 wrote his "Essai sur la minéralogie des monts Pyrénées", many hundreds of articles have appeared on the geology of the Pyrenees, particularly on its stratigraphy.

The Palaeozoic, called "terrain de transition" up to 1860/70, also aroused great interest in this initial phase. Important authors at this time who also worked in or around the present area were, among others: Charpentier (1812—1823) Dufrénoy (1830/34), François (1841/43), Durocher (1844), De Verneuil & Keyserling (1861), Zirkel (1867), Mussy (1868/69/70), Magnan (1871/74), De Lacvivier (1882/84/86) and Jacquot (1887/90).

After Caralp's important thesis (1888), Roussel's works (1893/1904) and Bertrand's geological surveys in the first decade of the 20th century, resulting in the 1:80.000 map (sheets Bagnères-de-Luchon 1911 and Foix 1912) the interest in the Palaeozoic in general and in that of the axial zone of the Central Pyrenees in particular, slackened. It is true, however, that there appeared some important publications on the Palaeozoic of the Catalan Pyrenees about 1930 (Jacob, Dalloni, Schmidt, Boissevain), of which Dalloni's (1930) and Schmidt's (1931) partly or wholly cover the area at present under consideration.

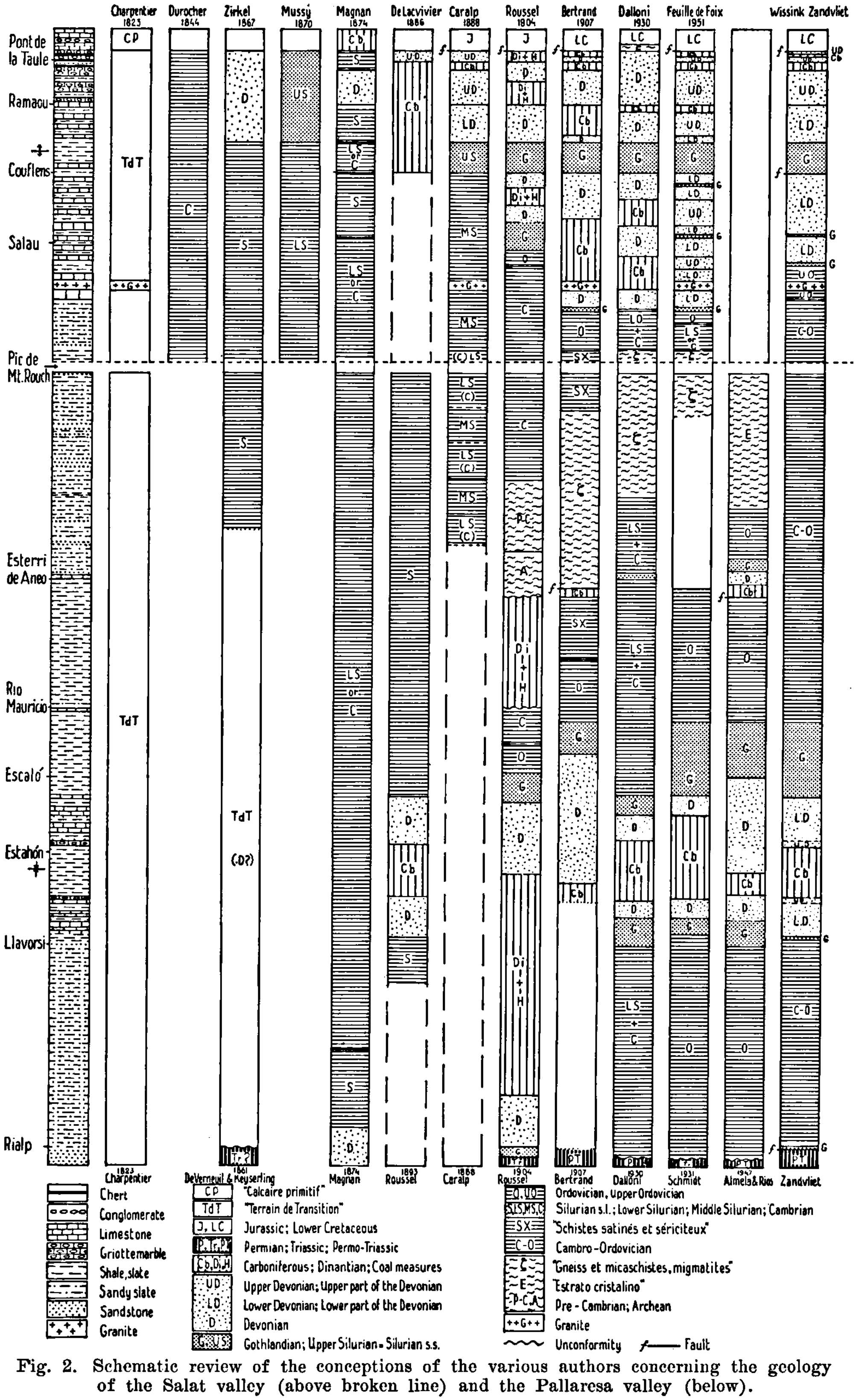


Fig. 2. Schematic review of the conceptions of the various authors concerning the geology of the Salat valley (above broken line) and the Pallaresa valley (below).

After the second world war the interest in the Palaeozoic has shown a strong revival and not only by French geologists. Also Spanish and Dutch geologists have exerted themselves to elucidate the remotest history of the Pyrenean chain, which had proved so difficult to unravel.

Of the French workers we should mention: Raguin, Durand (already before 1945), Hupé, Destombes, Thiébaut, Cavet, Autran and Guitard. The two last mentioned authors investigated the geology of a part of our area. Of the Spanish geologists it is especially Llopis Lladó, Fontboté, Solé Sabarís, who have dealt with the Palaeozoic, whereas, from Holland De Sitter, Zwart, Keizer, Wissink, Snoep, van Alphen, Francken and Kleinsmiede worked in or round our area and have already published data.

In order to summarize the views of these authors as far as they concern our region we present here a tabular summary showing their conception of one general north to south section, fig. 2.

It is obvious that the views of the older authors are less well founded, and the table, particularly the Salat valley section, clearly illustrates the ever-changing interpretation of the Palaeozoic succession. As early as 1888 Caralp gave a section of this valley, with the greater part of which we can agree. Later investigations did not always achieve further progress. Roussel (1893) also gives an excellent description of some sections of the syncline of Tírvia-Espot, and it remains incomprehensible why he changed his view in 1904, without any field evidence.

As far as the older literature is concerned, we are well-served by Caralp's "analyse critique des travaux antérieurs" in his thesis of 1888. Carez (1903/09) in his great work gives an outline of more than 1900 geological publications bearing upon the Pyrenees. Casteras (1933) in his work on the structure of the northern flank of the Central and Eastern Pyrenees adds a great number of publications that have appeared since that of Carez.

CHAPTER I

THE RELIEF

Introduction

The Pyrenean (late-Eocene) orogenic phase, the last important tectonic activity of a long succession starting in the Upper Palaeozoic, was immediately followed by active erosion and denudation of the axial zone. The coarse Oligocene conglomerates of Pobla de Segur represent the sedimentary result of this denudation phase.

The resulting aplanation has been preserved in some remnants of high plateaus in the axial zone.

The morphogenetic uplift which brought these surfaces to their present altitude started probably towards the end of the Miocene and was intermittently active till far into the Pleistocene and perhaps even later; short periods of uplift were followed by long periods of rest with intensive destruction, in which large parts of the mountains were denuded.

In this period of uplift the relief was rejuvenated by deeply incising Tertiary rivers and in their high drainage basins the snow accumulated several times during the Pleistocene, forming glaciers, which in their turn put their stamp on the relief of the highest parts of the chain. During the inter-glacial periods, when the ice partly or entirely withdrew, glacial erosion was replaced by a strong fluviatile erosion deepening the valleys and breaking down the glacial erosion-forms, a process which continues to the present day.

Naturally the Hercynian structures and the nature of the rocks have also played their part in the physiography. We shall discuss these factors successively in connection with their influence on the relief in the mapped area.

A. THE PRE-GLACIAL RELIEF

The pre-glacial morphology is characterized by a series of comparatively horizontal denudation surfaces lying at different altitudes and originated during the periods of morphogenetic rest; they are separated from each other by rather steep slopes. The oldest levels naturally lie highest and in each subsequent erosion cycle they were further broken down so that only small remnants of these surfaces are left. The dating, especially of the oldest and highest levels meets with great difficulties, because a direct correlation with synchronous sedimentary deposits is generally not possible. Indeed the ages given to any one level by various authors, differ greatly.

On the whole there is agreement, however, that the uplifts of large areas in the Central Pyrenees have been very regular and consequently denudation surfaces of similar altitude over rather large distances along the axis of the chain are considered as belonging to the same erosion cycle. Later dislocations, however, may have disturbed this schema in places (cf. Llopis Lladó, 1946 and 1947, p. 124; Pannekoek, 1935/37).

A good coordination of the separately investigated areas is still lacking; many valleys are known only very fragmentarily and this holds good in particular for our area, although a few authors mention it briefly (García Sáinz, 1933, 1940 a: Flamisell; García Sáinz, 1935, 1940 a: Pla de Beret, Basiero, Marimaña; Sermet, 1949/50: upper Salat).

In the Central Pyrenees we can distinguish on both sides of the chain some characteristic and constant denudation levels, of which the one situated between 1900 and 2300 m is most widely known (Faucher, 1935; Sermet, 1950; García Sáinz, 1935; Boissevain, 1934; Solé Sabarís, 1951).

Goron, 1937, calls this level "le niveau des hautes platformes", Boissevain (p. 126) speaks of "le niveau des fonds des cirques". A higher level can be distinguished lying in general between 2400 and 2500 m (Faucher, 1938; Sermet, 1950; García Sáinz, 1940 a).

A still higher denudation level is sometimes assumed to be represented in the more or less constant altitude of the high summits, the so-called summit level (complete peneplanation) mostly at 2700 to 2900 m, the age of which varies greatly according to the various authors, from Hercynian (Faucher, 1938, p. 36) to Younger Tertiary. According to García Sáinz (1940 a, p. 48) the highest tops of Maladeta (3300 to 3400 m), Panticosa (2900 to 3100 m), Riús-Saburedo (2900 to 3000 m) and Montcalm (3080 m) are remains of an even older erosion cycle of Cretaceous age. Llopis Lladó (1947, p. 131 ff) specifies this as pre-Senonian. Below 1800 m there also occur some levels which are less regular, but their correlation over large areas is doubtful.

In the Spanish part of our area occur a rather large number of more or less horizontal plateaus, truncated mountaintops and crests and some high erosion terraces, which are always characterized by weathered or decalcified rocks. Some of these surfaces truncate indifferently formations of various ages and/or lithology and there is no reason to suppose that these levels have a structural origin. See the physiographic map 1:100.000.

Some of the largest plateaus are:

	altitude	*surface*	
1. Coma Romadera	1850—2055	2,9 km²	on Devonian limestones and slates, Silurian shales and Ordovician slates
2. Pleta Anterrius	2200—2350	2,9 „	on Ordovician slates and sandstones (fig. 3)
3. Montareño	2400—2550	2,6 „	on Ordovician slates and sandstones
4. Perefita	1920—2200	2,3 „	on Ordovician slates and limestones
5. Pico de la Plana ...	2320—2472	1,8 „	on hornfels of the Ordovician

The plateau of Coma Romadera continues over the Bco de Caregue as an erosion terrace of the same altitude.

In all some 55 practically horizontal surface remains were recognized as such, as well as more than 20 sloping weathered surfaces. They are summarized in the following table. The figures indicate the number of erosion remains of one level and not their size.

French side

West	sloping	horizontal	Central	sloping	horizontal	East	sloping	horizontal
——	—	—	——	—	—	2960—3077,4	1	1
2700—2736	1	—	2600—2850	4	—	——	—	—
2440—2580	1	1	2440—2576	1	1	2460—2650	3	3
1933—2260	—	2	1800—2220	5	2	1850—2273	3	10
1800	—	—	1800	—	4	1800	—	6
	2	3		10	7		7	20

Spanish side

West	sloping	horizontal	Central	sloping	horizontal	East	sloping	horizontal
——	—	—	——	—	—	——	—	—
2600—2880	4	1	——	—	—	2600—2800	1	1
2320 à 2400 —2650	2	3	2400—2600	—	1	2400—2600	—	3
1800—2072	—	5	1900—2290 à 2350	—	6	2220—2350	—	3
1800	—	—	1800	—	2	1800	—	—
	6	9		—	9		1	7

From the table it appears that on both sides of the mountain chain two important denudation levels are present which are also known at the same altitude elsewhere in the Central Pyrenees viz. the level of ca 2400 to 2600 m and the one of ca 1850 to 2350 m. This is borne out by the presence of surfaces extending uninterruptedly from one side to the other side of the central watershed at about the same altitude (Port d'Aula 1933 to 2260 m and more or less also the Renacha plateau 2450 to 2580 m).

a. The denudation level of 2400 to 2600 m and the remains of pre-glacial slopes occurring above it

To this level, preserved practically only on the Spanish side, belong some of the largest plateaus, as those of Montareño (2400 to 2550 m), Campirme (2450 to 2600 m), Renacha (2450 to 2470 m). Locally a plateau may culminate in a somewhat higher rounded top, jutting out above the real denudation surface (Monadnock) e. g. Montareño (2593 m), Campirme (2633 m).

As a rule, no more horizontal denudation remains are found above 2600 m. There occur numerous remains of worn slopes, giving the impression of being very old. Such slopes always dip away from the highest zones, on the French side of the frontier-crest they incline 20 to 35° to the north and on the Spanish side 15 to 25° to the south. The obliquely truncated top of Montcalm, situated there where the frontier-crest turns to the south-east, inclines 18 to 19° to the north and north-east. On the eastern border of the Encantats granite, finally, old slopes are found dipping towards the east. It is clear that these sloping surfaces cannot be correlated with the so-called "Gipfelflur", for the recorded slopes of 18° to 35° (mostly calculated from the very accurate French topographical maps 1:20.000), do not correspond with that of an aplanation surface. They cut the highest tops obliquely and are certainly younger than the "Gipfelflur". In the opinion of the writer it is probable that they are synchronous with the denudation level of 2400 to 2600 m when the mountains did not extend more than 500 m above this level and formed the hill sides of this mature landscape. If these slopes had originated after the next uplift, they would have been steeper because the new river systems then were cutting deeply into the original landscape.

In 1940 a García Sáinz gave a map of the denudation surfaces of the southern Pyrenees between Gállego and Pallaresa. On this map the author considers all denudation remains between 2400 and 2700 m as having originated during the Eocene-Pontian cycle, those between 2800 and 3400 m as features of a Cretaceous cycle. The maps available in 1940 were so inaccurate that many features of García Sáinz's map are completely erroneous, some plateaus are in reality peaks and other distinctive ones are not marked at all (Niña 2700 to 2869 m, Pico de Fonguera 2720 to 2880 m). Moreover, in this part of the Maladeta massif no distinction can be made between the two erosion cycles of García Sáinz. The surface east of Estanque Negro (2655 to 2689 m) (according to García: Eocene-Pontian) inclines to the south-east, which leads to the supposition that this surface was higher in north-western direction and in fact we find there a very pronounced denudation surface (not recorded by García) on the Pico de Fonguera (2820 to 2880 m), inclining somewhat to the east. It seems probable that both denudation remains have been formed during the same cycle of degradation. Pico de Fonguera is as high as Pala Pedregosa and many other tops which García considers to be remains of the Cretaceous surface.

b. The denudation level of 1850 to 2350 m alt.

This wide-spread level, also present in many places on the French side, consists of the plateaus of Coma Romadera (1850 to 2055 m), Pleta de Anterrius (2200 to 2350 m, fig. 3), Perefita (1920 to 2200 m), Pleta del Salau (1850 to 1985 m) and the plateau round Port d'Aula (1930 to 2260 m) and some other less important plateaus. That of Port d'Aula, especially north of the pass, contains rather large circular sinkholes of which the Courret des Etangs has a diameter of 200 m and a depth of 30 m. Some four small lakes of which the Etang d'Aréau is the best known lie also on this plateau, always on slates between limestones. Their origin is probably glacial.

The most accessible passes in the high watersheds lie at equal altitudes with this 1850 to 2350 m level; sometimes they form part of it, as e. g. Port d'Aula 2260 m and Puerto de la Bonaigua 2072 m. They also date from the same erosion cycle. These passes lie often in north—south divides (Puerto de Bonaigua 2072 m, Puerto de Ovella 2330 m and Collado de Cabrís 2310 m) or in places where a divide generally running east—west has locally a north—south deviation (Port d'Aula 2260 m and Port de Salau 2087 m). This makes one suppose that the drainage at that time had a more east—west direction than nowadays.

The east—west valley of the Ribera de Boldís, which is incised in the 2400 to 2600 m level of Montareño and hangs some 800 or 900 m above the present Lladorre river (fig. 3), may also date from this time, as also the very gently worn south-west slope of Montareño to ca 2000 m and many similar slopes in the upper Salat area between 1850 and 2200 m (round Pic de la Tèse).

During the formation of this denudation surface the base-level of erosion

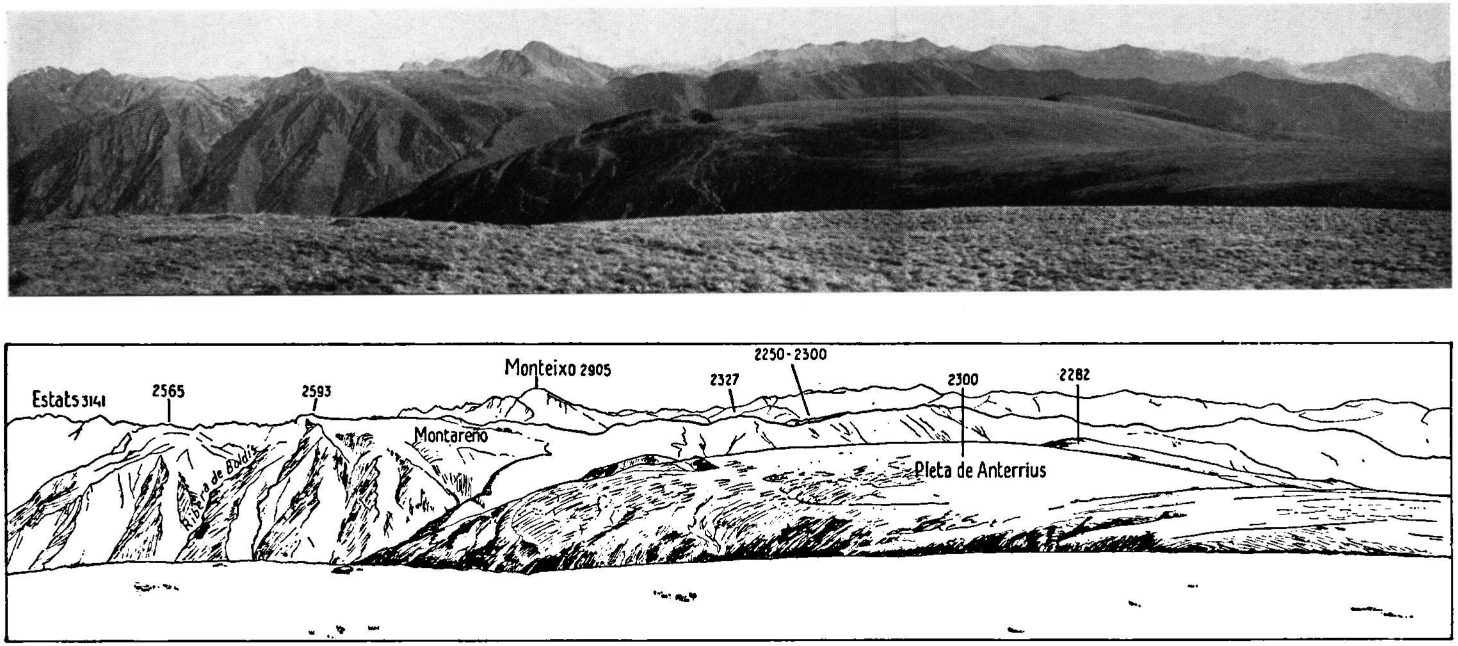


Fig. 3. Denudation surfaces of Pleta de Anterrius and Montareño looking cast (left) to south-east (right).

has probably not remained continuously at the same level; some comparatively small uplifts must have taken place in the meantime. Boissevain (1934, p. 126—131) distinguishes in the upper Sègre some four erosion stages at 1950, 2020 to 2050, 2150 and 2300 m alt. Small areas in the Salat-Pallaresa area sometimes make the impression that also here some erosion stages (of one cycle) can be distinguished. When such a survey is extended over a larger area the height intervals of the various denudation surfaces gradually overlap each other. Therefore we are of opinion that these erosion stages only have a local significance.

In the Spanish part of the mapped area there is a clear decrease of altitude of this surface towards the west; west of the Pallaresa this denudation level is situated between 1850 and 2100 m, at least so far as it is preserved (see also the table on p. 11). We must add to this that also the 2400 to 2600 m level of Pico de la Plana (2320 to 2472 m) is here rather low. Even the very regular summit level of the Marimaña massif is considerably lower than elsewhere in this area, viz 2600 to 2660 m rising eastwards to 2760 m alt. The Pla de Beret, finally, which has entirely kept its pre-glacial relief and which is of the same age as the 1900 to 2100 m surface of the Val d'Aran extends between 1800 and 1900 m.

García Sáinz (1935, p. 77—79) assumes for the area of the Pla de Beret a late-Quaternary tilting towards the south-south-east on the ground of the fact that during the Riss glaciation the drainage of the Pla de Beret still took place towards the Val d'Aran in contrast with the present drainage towards the Pallaresa. At any rate García's contention is not in contradiction with the data given in the preceding paragraph.

This surface is also well-developed in the Valle de Arán. Between the Aguamoix and the Artiès valleys there is an incision in this surface filled up by a 200 m thick sequence of conglomerates, sands and clays with numerous lignite layers in between (De Sitter 1954 a, p. 276, 1956 a, p. 217). Jelgersma, 1957, considers these sediments, on the ground of their pollen content as very probably Upper Miocene. Hence this surface is either Upper Miocene or a little older. Kleinsmiede (1960), however, doubts this dating and thinks that the sequence was down-faulted before the denudation surface was formed and thus that the surface is post-late-Miocene.

c. *The denudation levels below 1800 m alt.*

Besides the two very widely spread aplanation levels described above, there appear especially in the French part numerous incidental remains of denudation surfaces below 1800 m. With the exception of the 1040 to 1170 m level they have little influence on the relief of this area. This particular level, however, reveals itself in the important Col de la Trape (1111 m), in an erosion terrace near Cruzous (1170 m), both of them on Upper Devonian sediments, and in the large plateau de Géou (1040 to 1165 m), dipping somewhat to the north-east. The latter truncates the crest between the Garbet and the Trape river and lies on Mesozoic limestones with numerous sinkholes. The Garbet river which cuts through the Mesozoic synclinal zone downstream from Aulus has flowed westwards during the formation of this level along the southern border of this zone, draining the Alet and the Salat as tributaries. It is not impossible that the cutting of the Garbet through the Mesozoic fault zone is closely linked with a rejuvenation of the North-Pyrenean fault zone resulting in a tilt towards the north-north-east of the large plateau de Géou.

This caused the outlet via the present Col de la Trape to be obstructed and the course through the Mesozoic zone to be cut. Directly north of this zone the new Garbet joined the Estagette which already flowed along the northern border of the zone in question.

During the latest glacial period the Garbet glacier partly used the former course via Col de la Trape to the Alet of which witness the large quantities of glacial deposits on the slopes of the Trape valley and on the pass itself.

B. GLACIAL RELIEF FEATURES

Although the Pleistocene glaciations have not drastically changed the relief of the Pyrenees, its influence in detail is very clearly visible, as is witnessed by the frequent occurrences in the higher parts of the Mountain chain of U-shaped valleys, hanging tributary valleys, cirques, glacial tarns, "roches moutonnées", glacial striae and grooves, moraines, etc.

These are undoubtedly traces of more than one glaciation, the maximum number suggested by some authors being four. There is considerable doubt, moreover, as to the specific influence on the relief of each glaciation. The latest glaciation has, according to García Sáinz (1940 b, p. 368), entirely modified or intensified the erosional forms of older glaciations.

The extension of the latest glaciation is comparatively easy to determine (see physiographic map 1:100.000) and for the purpose of determining its limits use was made of polished zones, "roches moutonnées", glacial striae and unweathered erratic boulders. All of them are considered as features of the latest glacial period.

Of the few authors who have dealt with glaciations in the present area, in particular may be mentioned Nussbaum (1935, 1956: the whole mapped area), Chevalier (1954: French part) and García Sáinz (1935, p. 75—88: Pallaresa with Mauricio and Flamisell). In addition to these, Llopis Lladó (1946, p. 195) gave some information on the glaciation of the Vallfarrera, and Solé Sabarís (1957, p. 67—70 in Livret guide) described glacial deposits near Tírvia.

In the area of sheet 5 the following valleys once contained glaciers: Riberot, Estours, Angouls, Salat, Alet and Garbet, on the French side, Pallaresa with Bonaigua, Mauricio and Unarre, Estahón, Cardós, Vallfarrera and Tór, on the Spanish side.

With the exception of the Riberot and the Estours valleys, which for the greater part lie outside our area, these glaciated valleys will be discussed briefly.

a. The glaciated valleys

The four glaciated valleys of the northern side of the chain in the mapped area all belong to the drainage basin of the Salat. The GLACIER of the ANGOULS valley is the least important; judging from the shape of the valley the glacier never reached as far as the village of Angouls. The névé basin lies in the rather soft, little resistant Devonian limestones and slates, and a terminal moraine, therefore, is absent.

A distinct glacier basin lies at the foot of the Pic de Montaud, and lower down it runs into a cirque-like depression at ca 1170 m alt. with steep sides and waterfalls, called the Founiérous.

Only a little more important was the SALAT GLACIER, which hardly reached as far as the village of Salau, i. e. a distance of only 6½ km from the watershed. The old papermill of Salau has partly been built on the terminal moraine. Downstream from Salau the Salat valley is narrowly V-shaped. The glacial erosion upstream is insignificant. This is due to the small extension of the surface above the Würm snowline, and the very steep gradient of the supply valleys, causing the ice to flow down quickly and preventing thick ice accumulation (see the table on p. 21 and fig. 5).

Cirque d'Anglade (ca 1500 m alt.) is a deep basin, 800 by 400 m in size with almost perpendicular sides, closed by a threshold which juts out some 20 m above the cirque floor sloping down only 3°. The Anglade river reaches the bottom of the cirque by means of steep falls the last of which is 150 m high. Below the waterfall and at the sides of the cirque there lie big boulders which towards the centre and outlet pass into finer grained material. Just before the outlet the water disappears underground on the boundary between the "calcaire métallifère" and the Cambro-Ordovician sandstones, only to reappear some 300 m lower in the valley. The writer considers this cirque to be a lake filled up in post-glacial times. Glacial striations on the cirque side of the threshold inclining 20° to the south, confirm this point of view.

The ALET GLACIER formed by the Escorce glacier and the somewhat less important Ossèse glacier, advanced considerably further north than the Salat glacier. Although their two drainage basins are about equally large, the surface above the Würm snowline is much larger in the Alet area than in the Salat area (fig. 5). This must be the reason that these areas, which otherwise are very similar have undergone glaciation to so different a degree. The Alet glacier just reached La Bincarède, where a rather pronounced terminal moraine is present at 17 km from its source at an altitude of 630 m.

Some retreat stages can be distinguished in the Alet valley by the terminal moraines just downstream from Trein d'Ustou and near Estillon. Between Estillon and La Bincarède the valley is clearly U-shaped, down to Trein d'Ustou even very broadly so (fig. 4) in sharp contrast with the Salat valley which at this altitude is narrowly V-shaped.

Via Col de la Trape and the valley of the same name, part of the glacier ice flowed from the Garbet to the Alet glacier.

According to Chevalier (1954, p. 134) there also existed between the two glaciers a connection via the Plateau de Géou just north of the mapped area. The Guzet valley hangs about 300 m above the present Alet near Sérac.

The Cirque de Cagateille (ca 1150 m) can be compared with the Cirque d'Anglade, but is less imposing. The beautifully developed glacier-basins ("vans" as Chevalier (1954, p. 104) calls them) mostly end in a cirque lake or tarn with steep walls, e. g. that of Etang d'Alet (1910 m) and that of Etang de la Hillette (1797 m). Some 13 lakes occur here, all of glacial origin.

Both the névé basins of the Alet and of the Garbet glaciers are for the greater part situated in the Auzat-Bassiès granite.

The GARBET GLACIER has penetrated furthest north. According to Nussbaum (1935, p. 68) the terminal moraine lies near Plech, 1 km south-east of Oust, that is to say at a distance of 22 km from the watershed. Almost down to the terminal moraine the valley is U-shaped, above Aulus very broadly so. The upper course has four rock steps of which the two topmost ones hold lakes viz. Etang Bleu (1989 m) and Etang du Garbet (1683 m), the next step with a waterfall of 50 to 60 m high leads to the Cirque de Garbettou (ca 1400 m) with a threshold closing the outlet, probably a filled up lake. The lowest rock step, finally, bars the broad, oblong, flat depression of Agneserre (at ca 1100 m alt.), at the foot of the Auzat-Bassiès granite. In the drainage basin of the Garbet, with its glaciated tributary valleys of the Fouillet, Ars and Lauze, some 10 glacial lakes occur. The valleys all originate in well-developed glacier-basins.

A small part of the glacier flowed down over the Col de la Trape to the Alet glacier.

Of the three south-Pyrenean glaciers in our area that of the PALLARESA valley was by far the largest, although some part of its original upper valley must have been captured by the Garonne before the latest glaciation (Kleinsmiede 1960). The top of the Ruda glacier reaches so high that part of its

Fig. 4. Glaciated valley of the Alet looking north-west. From left to right the villages: St. Lizier, Bielle and Trein d'Ustou.

ice flowed into the Pallaresa valley over the low Puerto de Beret (1870 m). Glacial striae and "roches moutonnées" on this col point that way.

García Sáinz (1935, p. 75—88) holds that the upper course of the Pallaresa had been very little glaciated during the Würm and that as a result of the small extension and isolation of the Marimaña massif its glaciers would hardly or not at all have reached the Pallaresa valley. We cannot share this view.

It is true that the shape of the main valley and that of the northern Marimaña tributaries (Marimaña, Lausanas and Cireres) show but little glacial influence. The pre-glacial relief is here still clearly preserved. It is, however, a fact that fresh glacial striations and unweathered granite boulders occur everywhere to a great height on the northern and eastern slopes of the massif in question and, to a lesser degree, also on the left bank of the

Pallaresa; they have deposited where the two glaciers joined. The moraine in the valley of the Cireres for instance contains a great many granite boulders which can only originate from the main valley glacier, because no granite outcrops in the tributary valley.

The fact that the glacial influence on the upper course of the Pallaresa has been so slight, must be due in our opinion to the stagnant character of the Pallaresa glacier which, although piled up high, carried down little ice, because the large tributary glaciers of Areu and Bonaigua prevented its natural flow. In this connection the small gradient of the valley and the glacier is also significant. It amounts to about 0.022 (included the post-glacial fluviatile incision) for the valley floor between Montgarri and Alós and for the surface of the glacier (resp. at 2000 m near Montgarri and almost 1900 m near Alós) it must have been about 0.007. The glaciation of the Marimaña massif itself has been very extensive, especially on its eastern and western sides, witness the large glacial basins with tarns and the U-shaped drainage valleys. Undoubtedly the surface above the Würm snowline was considerable (compare fig. 5).

The Pallaresa valley is rather wide to 5 km upstream from Alós. Although the post-glacial erosion has been rather strong, parts of an old valley floor have been preserved as erosion terraces downstream from Montgarri. They lie about 100 m above the present river bed. Also between Alós and Isil a similar terrace is present 80 m above the steeply incised river.

Where the Bonaigua glacier joined the Pallaresa glacier we find a large comparatively flat depression, the Barbaña, between 1100 and 1200 m alt. The Pallaresa, and to a less extent also the Bonaigua, are here incised to a depth of 150 m. An equally high and steep rock step forms the transition of the Barbaña to the depression of Esterri which we consider as a former lake, filled up post-glacially; it is one of the few large finger lakes of the Pyrenees, occurring at an altitude of ca 900 m, where most South-Pyrenean Würm glaciers had already melted off (see table on p. 21). In 1954 the marshy southern part of the depression was dammed to form a reservoir as part of a hydro-electric scheme.

The massive Bonaigua glacier had its main origin in the Basiero granite. Part of the ice of the Ruda glacier flowed over the Puerto de Bonaigua and benefited the Bonaigua glacier.

The hanging valley of the Son also had a glacier, as is shown by the conspicuous terminal moraine about 1 km west of Son del Pino. The left tributary valley of the Son and also the moraine contain many granite boulders, but no granite outcrops in this valley. Probably the Cabaña-Bonaigua glacier flowed temporarily over the Collado del Paso del Coro (1929 m) and carried these granite boulders into the Son valley.

It seems improbable that the Unarre glacier should have contributed much to the Pallaresa glacier. It has been more of a filling-up of the space against the Pallaresa glacier, whereby the glacier-ice of the Pallaresa even penetrated far into the Unarre valley. The moraine south of Unarre contains many granite boulders which must have been supplied by the Pallaresa glacier, because no granite outcrops in the drainage basin of the Unarre. The moraine above Servi is probably a terminal moraine of a retreat stage. Post-glacial erosion reached as far as Servi, but upstream of this village the U-shape has been preserved in both valleys. According to García Sáinz, op. cit. the Pallaresa glacier stopped at Guingueta at about 920 m alt. However, fresh erratic boulders were found on the left slope at more than 700 m height above Esterri.

The Mauricio glacier has also added ice to the Pallaresa glacier, as appears from fresh erratic boulders 2½ km upstream of Espot on the right bank, at a height of over 250 m above the present valley floor. Moreover, the tributary valley of Estahís hangs about 250 m above the Mauricio. The delta-like gravel deposits, described by Nussbaum (1935, p. 87 and 1956) are no proof that the Mauricio glacier has never joined the Pallaresa glacier. In our opinion these gravels have been deposited as a delta in a lake occupying the growing space between the retreating Mauricio and Pallaresa glacier. To about 2 km downstream of Espot the valley is broadly U-shaped and it hangs about 150 m above the Pallaresa.

After the junction of this glacier with that of the Pallaresa valley, the thickness of the latter rapidly decreased. Downstream of the Mauricio the Pallaresa valley is rather narrow, but widens again near Escaló until about 3 km further south a 50 to 100 m high erosion terrace narrows the valley again. It consists of Devonian limestones and not of hard schists with quartzite bands, as Nussbaum (1956, p. 80) states. The terminal moraine is not clear. It is probably correct to consider as such (Nussbaum, 1935, p. 85) the thick debris deposits with angular granite boulders 2 or 3 km upstream of Llavorsi on the right bank of the Pallaresa. The total length of this glacier must then have amounted to about 51 km.

The CARDÓS GLACIER has a considerably smaller drainage basin than the Pallaresa (see fig. 5). After the junction of the Tabescán and the Lladorre glacier near the village of Tabescán the Cardós glacier arising from this junction is no longer been fed by tributary valleys. The Tabescán glacier was somewhat less potent than the Lladorre one, as appears from the fact that the Tabescán valley has a slightly hanging outlet.

For the rest the two valleys correspond in outline. Thus, in both valleys there occur rock steps when two or more glaciers came together, with at their foot a round or oblong basin with steep sides and a flat floor. The rock step is often steeply incised by later headward erosion. A characteristic instance of this is the Pleta de Bohavi at ca 1500 m alt., where the valleys of the Certescáns, Rumedo and Brohate are united with the Lladorre. The basin is closed on the downstream side by a distinct threshold of solid rock. Also at the junction of the Bedó and Noarre streams and that of the Tabescán and Noarre streams, rock steps with basins have been formed. The latter rock step has been steeply incised over one km length in the Tabescán valley.

In some higher tributary valleys typical pater noster cirques have been formed mostly with a series of lakes. The Bedó valley for instance has five rock steps each with its own lake; the Noarre valley has four rock steps of which three contain a lake and the Rumedo valley has four rock steps each with its own lake.

In all there occur some 52 lakes, in the Cardós area all of glacial origin, among which the Lago de Certescáns is over 70 ha (= 175 acres) large and 96 m deep (at 2233 m alt.).

The outlets of the smaller tributary valleys are often hanging rather high above the main valley or the larger tributary valleys.

The erosion terraces occurring in the Cardós area are all characterized by "roches moutonnées" with glacial striae. That of Noarre certainly forms an old glacial valley floor, but the origin of the other erosion terraces is not clear, although they are certainly glacially affected.

Near Tabescán there is a distinct terminal moraine of a retreating stage. Between Tabescán and Lladorre the Cardós valley is V-shaped, downstream from Lladorre valley-narrows alternate with round or broad basins which sometimes have the character of broad flat thresholds with glacial striae. The threshold south of Lladrós is more than 50 m higher than its basin upstream. Before this step had been completely cut, it certainly enclosed a lake. As far as Casibrós some six basins occur each closed by a threshold. Downstream from Casibrós the valley hangs above the widened depression of Ribera de Cardós.

Although glacial deposits are found in many places in the Cardós area, among other places near Surri, 160 to 200 m above the present valley floor, a distinct terminal moraine is absent, but the glacier probably did not reach the Vallfarrera. At least, in the narrowing downstream of the depression of Ribera no convincing proof that the glacier has passed through this canyon was found. If this is correct the length of the Cardós glacier amounted to some 28 to 29 km. Nussbaum (1956, p. 70) records moraine material near La Fábrica (halfway Tírvia). In the opinion of the writer this occurrence of debris on the right bank near La Fábrica forms part of a large landslide which has slipped down along the top of the Devonian on the outcrop of the Silurian shales. The 30 to 40 m thick deposits of the terrace of Tírvia are ascribed by Solé Sabarís in the Livret guide, 1957, p. 67—70, to older glaciations and inter-glacial (Riss to Mindel) deposits. This assumption seems hardly justified as we know that the bulk of this material consists of Carboniferous slates derived from the left bank of the Cardós river.

The end of the VALLFARRERA GLACIER is also difficult to determine, a clear terminal moraine being absent. The shape of the valley as far as Arahós has been reworked more or less glacially, but from there on the valley is narrowly V-shaped over some distance. Nussbaum (1935, p. 92) considers the considerable quantity of rather fine debris with erratic material on the left bank between Alins and Arahós as a terminal moraine. In our opinion the greater part of this debris came from the mountain-slope above it, and only a very small part, and then only the outside of this thick debris deposit, is material transported by the glacier (see Accumulation terraces, p. 27). As far as Arahós the length of the glacier amounted to some 25 or 26 km.

The glacier valley itself is occasionally clearly U-shaped, e.g. near Areo and south of Alins. The upper course, which has a more east-west direction, was certainly U-shaped, as appears from some parts of the old valley floor which are left as erosion terraces. The river has incised a 100 m deep and several km long gorge in this Würm valley floor. The formation of this gorge was very much facilitated both by macro-joints running parallel to the valley, and by cleavage trending in about the same direction.

The numerous glaciated tributary valleys nearly all hang conspicuously high above the main valley floor. The B^{co} de la Llaguña, for instance, is some 300 m above the Würm valley floor, and more than 400 m above the present river level.

In the drainage basin of the Vallfarrera occur some 25 glacial lakes, most of which are, however, very small. Above the Lago Fondo (2480 m alt., B^{cos} Sottlo) occurs the only lake on a pass in our area. This 100 m long unnamed lake lies at 2875 m alt. and is situated on the frontier between France and Spain.

The smaller TÓR GLACIER probably did not reach the Vallfarrera glacier. The Tór valley is only slightly U-shaped and does not carry any lakes.

b. Glaciation of the mapped area. General remarks

The higher the glacial erosion forms occur, the fresher they are. There has been little modification in the granite massifs for example since the last glacier ice melted. The long duration of the glaciation naturally exerted a very intensive influence on the relief, fluviatile erosion having hardly penetrated to these high areas. In contrast with the high altitude glaciation the glacial forms in the lower areas are often for the greater part broken down by fluviatile erosion, and moreover the time during which a glacier extended so far was too short to imprint a clear glacial character on these areas.

As early as 1883 Penck — after him many other authors — pointed out that the glaciation of the French side of the chain descended to a lower level than that on the Spanish side. Significative factors are both the exposure of the main slope to sunshine and the difference in precipitation. In general the large French Pyrenean glaciers of the Central Pyrenees descended to a level of ca 400 to 500 m alt. and the Spanish ones to 800 to 1000 m. As a result the French glaciers were mostly longer than the Spanish ones.

The glaciers of Western Ariège prove to be exceptional. They appear to have been considerably shorter and their terminations lie generally some hundreds of metres higher than normal.

Compare the table:

		average gradient between source and 1000 m	length of glacier	altitude of terminal moraine	
French side	Garonne	0,067	70 km	± 430 m	Nussbaum 1935
	Riberot	0,246	13 „	± 620 „	„ „
	Angouls	0,411	4½ „	850 „	
	Salat	0,385	6—6½ „	860 „	
	Alet	0,289	17 „	630 „	
	Garbet	0,230	22 „	520 „	Nussbaum 1935
	Ariège	0,079	63 „	400—420 „	„ „
Spanish side	Esera	0,063	35 km	850 m	García Sáinz 1940b
	Ribagorzana	0,098	22 „	980 „	„ „ „
	Tort	0,078	22 à 23 „	890 „	„ „ „
	Flamisell	0,092	18½ „	990 „	Nussbaum 1935 García Sáinz 1940b
	Pallaresa	0,027	51 „	820 „	Nussbaum 1935
	Cardós	0,090	28 à 29 „	850 „	„ „
	Vallfarrera	0,080	25 à 26 „	900 „	„ „
	Valira	0,064	28 „	960 „	„ „

Obviously only the area above the snowline has any practical significance as supply-area for the glacier. Nussbaum (1938) broadly fixes the Würm snowline on the French side of the frontier zone in the mapped area at 1800 to 1900 m. This figure given by Nussbaum corresponds with the small nivation cirques which are found occasionally and which are considered as approximately indicating the snowline (cf. Daly 1912, p. 593; Flint 1947, p. 95—96; Llopis Lladó 1951, p. 87). On the ground of this level of lowest cirque excavation. supplemented with Nussbaum's map, we have drawn the map of fig. 5, representing the névé areas of the glaciers as the surface above the Würm snowline.

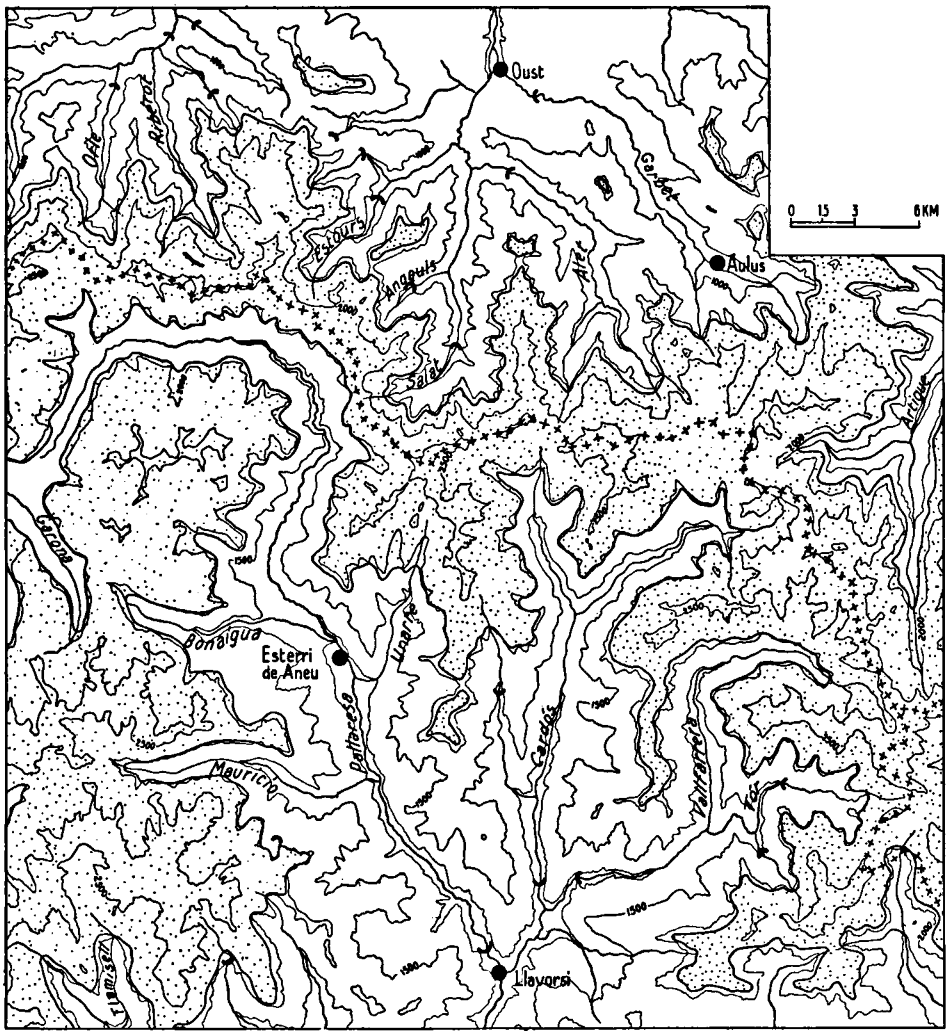


Fig. 5. Snowline map of the last glaciation in the Salat-Pallaresa area. Originally snow-covered areas are stippled. Arcs: lowest points reached by glaciers.

The snowline rose going north to south from about 1500 to 1600 m alt. in the north to 2100 to 2200 m alt. in the south of the mapped area. Local factors influenced this situation to some degree: on slopes with an exposition towards the north the permanent snow descended further down than on slopes exposed towards the south. The main crest, running from east to west especially showed great differences in this respect, the snowline on the northern flank certainly lying some 300 to 400 m lower than on the southern flank. The position of the snowline was also somewhat influenced by the differences in size of the horizontal extension of the ice cap. Small isolated high areas had a slightly higher snowline than similar less-isolated areas. Thus the snowline in the Garbet area (part of the large and high Estats massif) lies at ca 1600 m alt., or about 100 m lower than in the Angouls-Estours area (1700 m). Under otherwise equal circumstances the lengths of the glaciers seem to be proportional to the extent of the névé area, cf e. g. the Angouls, Salat, Alet and Garbet areas.

The small supply-area of our French glaciers corresponds also with the very steep gradient of their valleys (table on p. 21) due to the very small breadth of the Couserans Pyrenees, amounting only to some 20 to 30 km. The main watershed there is little, if any, lower than elsewhere.

Many glacial valleys in the mapped area find their origin in one or more large glacier basins or "Karplatten" (Nussbaum) or "vans" (Chevalier): broad, rather shallow and gently sloping excavations with flat bottoms and steep walls. These basins are separated from each other by sharp crests which downstream gradually increase in height with respect to the basin bottom. They end rather suddenly in case the basins on either side belong to the same névé area of a single glacier. The basin bottom is always covered with "roches moutonnées" and often also with lakes (fig. 6).

A glacier basin rarely passes with one step into a glacial valley, in general over-deepening took place in several successive steps, each time resulting either in a deep cirque with steep sides, or a cirque lake or just in a rock step.

Sometimes a number of glacier basins, each with its own lakes opens into one large cirque. The Escorce (= Alet) basin in which four large "vans" empty themselves via steep rock steps into the Cirque de Cagateille (fig. 7) illustrates this. Other complex basins are less regular, such as that of the Peguera (Saburedo massif) and of the Cabaña (Basiero massif). Over-deepening also very often took place where two or more glaciers joined e. g. Pleta de Bohavi (fig. 8).

Sometimes a large wide basin-like depression slightly sloping upwards contains a number of mostly small lakes separated from each other by numerous "roches moutonnées", e. g. the glacier basin of the Basibé (Marimaña massif) with six lakes between 2310 and 2350 m and that of the Baborte (Vallfarrera), where there are seven lakes between 2335 and 2375 m, among which is the rather large Lago Baborte; at a somewhat higher level four lakes occur between 2425 and 2455 m.

Thus the slope of the crests of the divides between the glacial basins is generally less than that of the basin profiles. Hence it is often concluded (Solé Sabarís 1951, p. 89, and others), that the broad flat glacier basins are gauged in the gently sloping pre-glacial relief, which has been only preserved on the crests of the divides. Another argument in favour of this view is the flat top remnants on these divides which are found occasionally, e. g. between the Hillette and the Turguilla in the Alet area, where the flat top slopes 25° towards the north. This old surface lies 50 to 80 m above the glacial bottom directly beside the crest and some 100 to 300 m above the centre of the glacier basins on either side. A similar remnant of an old surface is present between the Quer-Ner and the Fontaret in the Salat area, it lies at most 100 to 150 m above the glacial valleys on either side. There are also indications of this phenomenon in the Marimaña granite.

Fig. 6. *Roches moutonnées* controlled by joint pattern. Escorce valley above Cirque de Cagateille, Auzat-Bassiès granite.

Fig. 7. Glaciated Turguilla (left) and Hillette (right) valleys above Cirque de Cagateille. Post-glacial incision has taken place along major joints. Auzat-Bassiès granite.

The glacial excavation of the round and weathered surfaces of pre-glacial age of the Cap de la Galèche and the Pic de la Tèse (Salat valley) reaches to a depth of more than 150 m. It seems likely that at least a part of the very broad glacier basins are excavations in a pre-glacial relief.

It is remarkable that the influence of glaciation on slopes directed towards the east reaches a lower level than on slopes directed towards the west. Compare for instance the north-south divide between the Salat and the Alet, the one between the Fouillet and the Escorce north of Pic de Cerdá and the one between the Tabescán and the Lladorre. Often the top of these crests is somewhat asymmetrical in so far as the upper 100 m slope facing east is

Fig. 8. Pleta de Bohavi, probably a late-glacial filled up lake, as seen from the eastern end.

considerably steeper than that facing west. Lower down this difference in slope is not maintained. The asymmetrical north-south crests with gently sloping western slopes and steep eastern slopes described by Sermet (1950, p. 396 ff) from areas further to the west (e. g. between Pique and Garonne) do not occur here. These asymmetrical crests are according to Gaussen and Schrader (in Sermet 1950, p. 397) due to the prevalence of western winds laden with snow. They would more or less polish the west-slopes and prevent eroding but lead to snow accumulation on the leeward (= east)side of the crest.

In fact it has often been noted in spring that on the east-slopes against the steep upper part of the crest a thick mass of snow was present up to the top, whereas the west-slopes were entirely free from snow. It is not thought that the same factor has operated in the case of the Col de la Trape which falls steeply towards the east, as Sermet states. The valley of the la Trape is clearly a captured valley with a pre-glacial relief (see p. 14 ff).

Lakes occur only in those areas which were glaciated and lie above the Würm snowline. This points to a glacial origin, although the pre-Würm and even pre-glacial relief has no doubt played a part in the formation of many lakes, certainly at least with regard to its localisation. In this respect the reader is referred to p. 31, where the influence of macro-joints on the relief has been described.

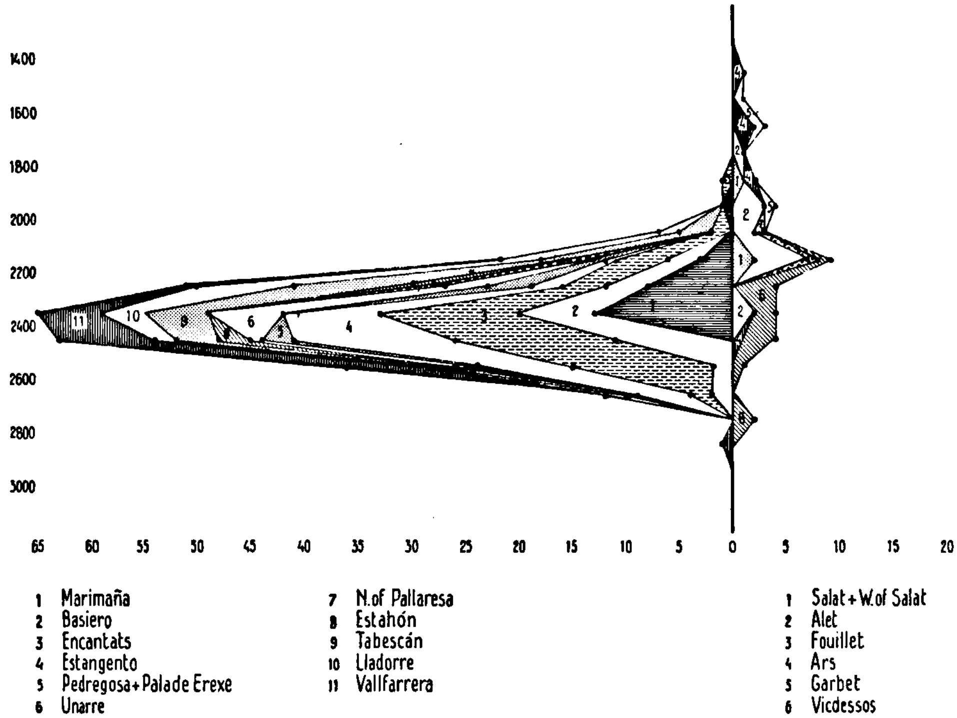


Fig. 9. Altitude-frequency graph of lakes occurring in the various drainage and other units. Left of vertical line — Spain, to the right — France. On the Spanish side 1—5 represent granite areas and 6—11 drainage units while on the French side 1—6 are drainage units. The plots for successive units are cumulative.

Most of the lakes lie in the granite massifs but where the glacial conditions and altitude are the same in other rocks, as for instance in the upper course of the Vallfarrera, the frequency of lakes is nearly the same. In all 298 lakes were counted, of which 259 lie on the Spanish side of the frontier.

In fig. 9 the vertical and areal distribution of these lakes has been graphically represented. From this it appears that the lakes in the French part descend to a lower level (1400 m) than in the Spanish part (1800 m). This too, is the result of the glaciation descending further on the French than on the Spanish side. However, owing to the very steep gradient of the French valleys there is hardly any place for lakes. The steep Salat valley, for instance, has only one very small glacial lake (Laquet du Mail).

c. Late-glacial stages

Above 2100 m alt. there occasionally occur semi-circular to V-shaped walls with the opening facing towards the mountain either at the foot of high crests or in cirques or barring cirques. They are 2 to 10 m high and consist of angular loose debris, mostly mixed with argillaceous material. The radius of these curved walls varies from 50 to 300 m. Sometimes several curves have been stringed together like garlands, e. g. at the foot of the crest east of the Lagunas de Basibé (Marimaña massif). The wall always encloses a depression of which the soil is often marshy or which contains a small lake. The walls generally carry no vegetation and have a very young appearance. They strongly suggest small moraines formed in a very late stage of the glaciation.

García Sáinz, who in 1941 wrote a lengthy article on the epi-glacial phases of the Spanish Pyrenees has suggested this origin of the walls and ascribes them to the very latest epi-glacial phase, the Daun. The altitude at which they occur varies from 1950 to 2650 m with a maximum between 2200 and 2350 m, both on the French and Spanish side.

García Sáinz described two other epi-glacial stages older than the Daun, viz the Bühl, the oldest, which pushed farthest ahead, characterized by rather thick moraine walls at the back of the U-shaped Würm valleys, and the Gschnitz, characterized by detrital material, which spread like a mantle in front of the glacial cirques. This difference in aspect of the moraines is supposed to be due to certain climatic factors. For further detail we refer the reader to García Sáinz's article (p. 259 ff).

Regarding altitude, the Gschnitz moraines lie between the Bühl and the Daun. There is a Bühl moraine, according to García Sáinz, where the B^{cos} de Estangento and Sallente join and a Gschnitz moraine east of the Lago Estangento on the Coma de Espós.

d. Accumulation terraces

The present relief is, at least in the Spanish part, also determined by a great number of sloping terraces, in the main valleys hanging up to a maximum of 600 or 700 m above the present valley floor. They occur along the valley sides or in small, non-glaciated tributary valleys. Often they continue part of the way up these small valleys. Downwards the terrace slope gradually decreases; the broad terraces above Areo and Espot become almost horizontal. Finally they are terminated by an erosional scarp. In many cases these sloping terraces have been incised by a brook, the incision showing that the terraces consist of loose material thickening towards the valley. Some of them are not incised (e. g. near Areo) and then the water of the brook percolates through the loose terrace debris.

The terraces consist of angular and badly sorted flat slate fragments, in general smaller than 10 cm embedded in argillaceous material. This debris almost certainly came from directly above. It is rather well-stratified, the slate fragments always lie with their flat sides practically parallel to the upper or lower side of the terrace, depending on their place in the terrace. Consequently the terraces strongly give the impression of being truncated alluvial fans. There are, however, some facts which do not correspond with this view:

1) a similar terrace east of Espuy (Flamisell) contains rounded, erratic granite blocks mixed with material from above. This is also the case with the large terrace along the Vallfarrera between Alins and Arahós.
2) it is an established fact that the height of the terraces above the valley floor is not arbitrary; on either side of the valley they are about equally high. One has the impression, moreover, that the terraces can be arranged in one or more series each of which descends gradually until the present valley floor is reached.
3) in some places it has been recorded that the upper terraces and the upper scouring zone of the Würm glacier lie at an almost equal height. Nevertheless some small terraces are found higher than the Würm glacier has ever reached.

These facts seem to suggest that these terraces are not individual alluvial fans. It is moreover improbable that fans should have filled up the valleys in post-glacial times to some 600 to 700 m and then be eroded away for the greater part.

If we assume that the fans were formed during the glaciation, accumulating partly against and on the sides of the glacier like kame terraces, the contradicting evidence seems to fall into place. The small incising tributary rivers started their work already during the glacial stage. The large terrace near Espot has a shape strongly suggesting this. Below the confluence of the Mauricio (= Escrita) and Peguera valleys, the U-shape of the main valley is formed on one side from the terrace wall.

In this connection the granite boulders in the terrace of Espuy have some significance. Although no glacial striae can be found on the granite boulders, owing to weathering, these erratics have probably been deposited by the Würm glacier as lateral moraine against the terrace.

It seems probable that these accumulation terraces are kame terraces, the greater bulk of their material, however, being fluvio-glacial.

These terraces will have to be studied more closely before a definite explanation can be given of their origin. The above interpretation is based on comparatively few data. As far as we know, none of the authors writing about the Pyrenees, has described similar terraces.

C. POST-GLACIAL RELIEF FEATURES

After the melting of the glaciers the rivers re-incised, terminal moraines were broken off, lakes were filled up, scree and alluvial fans were formed, and in places where brooks and rivers flowed more quietly, sand and gravel were deposited.

One of the most typical post-glacial phenomena is the formation of deep gorges by headward erosion. In our area these gorges do not occur higher up than at an altitude of 2200 m and exclusively in former glaciated areas, where they incise irregularities of the glacial valley profile.

They occur for instance in:

a) *hanging valleys,* where they discharge in the deepened valley, e. g. the hanging Tabescán valley above the Río de Lladorre with a gorge of about 800 m length.

b) *thresholds,* e. g. in the Cardós valley between Ribera and Lladorre with some 6 dissected ridges (p. 20).

c) *rock steps in a valley,* e. g. in two rock steps (880 m and 940 m) in the Ossèse valley with gorges of 200 to 300 m length or in the back face of cirque-like depressions, where mostly two or more valleys discharge. Instances of this are the gorges of the rivers Certescáns (400 m) and Brohate (1600 m) above the Pleta de Bohavi; the gorges of the rivers Hillette (600 m) and Turguilla (750 m) above the Cirque de Cagateille; the gorges of the Noguera Pallaresa (4000 m) and Río Bonaigua (1000 m), above the depression of Esterri de Aneo. The Vallfarrera valley above 1400 m is steeply incised over more than 5000 m.

The gorges are often as deep as the rock steps are high, upstream the depth of the gorge gradually decreases, so that in the case of long gorges a regular gradient of the river is mostly effected, as is the case with the gorge of the Pallaresa, which near Esterri is some 150 m deep and disappears between Isabarre and Borren. Shorter gorges are often closed upstream by a waterfall. Sometimes a gorge ends downstream in a waterfall, e. g. in the Ossèse at ca 880 m and at ca 1300 m, in the latter case a waterfall of over 50 m high falls over a very resistant porphyrite sill. The gorge, some tens of metres deep, breaks upstream through a high and broad dolomite bank and ends in the depression of Maillat, which must have contained a lake before the headward erosion had advanced so far.

As appears from the above examples, the length of the gorges varies strongly; not all the rock steps are cut by gorges. Generally this is a question of the resistance of the rock. Along the steep backface of the cirque d'Anglade where the Anglade river falls down over ca 160 m, no sign of headward erosion can however be perceived. The rock is here a hard and compact quartzitic sandstone. Elsewhere, in places where this same rock and roughly the same amount of water occurs, gorges are present hundreds of metres long. The explanation of this must be sought in the presence of macro-joints (p. 31, 94), along which the rivers have incised. This is very clearly the case on the northern border of the Auzat-Bassiès granite (fig. 7). Most valleys which carry off water towards the north have a rock step at the contact with the adjacent rock as a result of differential erosion. All these rock steps have been cut by a long gorge along a macro-joint. Also the gorges of the Cardós between Lladorre and Ribera and of the Tabescán have been formed partly or entirely along joints.

This headward erosion has nowhere given occasion to streamcapture.

Some glacial lakes, especially in the lower zones have been filled up after the retreat of the ice, as the Cirque d'Anglade, the Pleta de Bohavi, the depression of Esterri, etc.

Above 2000 m the fluviatile erosion is very slight. The glacial erosion forms are still very fresh there, especially in the granite massifs, scree and alluvial fans, however, often hide the original U-shape of the valleys, e. g. in the Unarre valley above Servi.

D. THE INFLUENCE OF THE HERCYNIAN STRUCTURE ON THE PRESENT RELIEF

Hercynian features — due to lithologic and structural factors — can be seen everywhere in the present relief. The resistance of rocks to erosion is governed by two factors, their lithology and their tectonization. Thus a quartzite — which is never strongly cleaved — will turn out to be very resistant. As the rock contains more clay matrix and is consequently softer, the tectonization (cleavage) increases, and the rock becomes still less resistant.

In places where the porphyrite sills are not cleaved (e. g. in the Salat-Alet area), they often jut out like sharp crests above the surrounding weaker rocks (fig. 10). Directly north of the Marimaña granite, however, these porphyrites are strongly cleaved and behave with respect to the erosion like the surrounding soft slates.

Fig. 10. Serrated ridge formed by hard porphyrite sill in upper part of the Cambro-Ordovician. Left bank of Ossèse valley. Back-ground: Devonian of the Pic de la Tèse mass.

The very incompetent Silurian pelites show in general a particularly low resistance to erosion. Consequently the Silurian is characterized by smooth and negative denudation forms, in particular where their outcrops have a considerable extension as in the anticline of Couflens (Col de la Serre du Cot 1546 and Col de Pause 1527) and round Col d'Escots 1618; the drainage too, often follows these soft shales, e. g. between Col de Mánega and Os de Civís.

As a rule the homogeneous granite massifs form the highest parts of the chain. In the Fouillet-Ars-Garbet area the granite forms a kind of wall jutting out above the adjacent rocks, though there is no question of a recent movement along a fault contact. Sometimes, however, the contact-metamorphic country rock — especially hornfels and limestone marble — seems to form the highest points, e. g. the crest of Teso at the eastern border of the Basiero granite and the crest of Cuenca-Moredo, bordering the Marimaña granite in the east. Also the curve of the upper Pallaresa round the Marimaña granite must be due to differential erosion as a result of lithologic differences and not to structural factors. The Mauricio valley in the core of the syncline of Tírvia-

Espot between two granite massifs seems to be structurally controlled at first sight, but may also be explained by lithologic differences. The fault valleys of the Liesca and the upper Riberot on both sides of the central crest are not purely structural either because the fault runs in the little-resistant Silurian.

One of the most striking Hercynian features in the present relief of the axial zone is the presence of a pattern of large lineaments. In this respect we refer to lineaments of fault character but not showing a measurable offset or throw (see p. 94). The course of many small and of some large valleys has been determined partly or entirely by such macro-joints. The Hillette, Turguilla and Ars rivers, for instance, which drain the northern part of the Auzat-Bassiès granite are for a large part joint-valleys. Also outside the granite massifs impressive examples can be given: the Cardós valley, for instance, is determined between Lladorre and Casibrós over a distance of more than 5 km by a prominent lineament which continues north of Lladorre in the left slope of the valley. In places where hard rock is present (e. g. in the rock steps) the river runs exactly in the joint-plane, but in the rounded basins filled with Quaternary sediments it does not, but the valley remains very straight. A clear zone of parallel joints is present here and the Cardós river shifts sometimes from one joint to another.

Other large joint-valleys are the Montalto valley (tributary of the Lladorre), the Tabescán valley and, somewhat less clearly, also the upper course of the Vallfarrera where it runs westward. In such joint-valleys a rapid destruction of the glacial valley bottom is the rule and erosion will easily lead to the formation of very long gorges, as is shown by the upper course of the Vallfarrera, where a 100 m deep gorge occurs over nearly the whole east-west course. This appears also to be the case in many other valleys if one compares the joints on the geological map with the morphologic map.

The writer is under the impression that the joint pattern is limited to the centre of the axial zone, and chiefly to the oldest rocks and the granite massifs. It is not very likely that the general north-south drainage was caused by this joint pattern, but nevertheless on a comparatively small scale the valley formation in the mapped area has made use of the Hercynian pattern of large and small joints, a factor which hitherto had not been realized. Many cols are dependent on macro-joints, e. g. Collado de Sellente, Puerto de Boet and many other unnamed passes (see geological map).

The levels of some lakes have fallen appreciably in post-glacial times. The cause of this is headward erosion along a joint, incising the outlet. Thus the level of the lake Gerbel (Basiero granite) has fallen about 20 m owing to erosion along the joint of the B^{co} de Gerbel.

Also the shape of the lakes, especially in the granite, is often determined by the local joint pattern; Lago Estats, for instance, is bordered on all sides by joints. Other lakes have one or two sides which are formed by joints (Bassibié and Rosario in the Marimaña granite (fig. 43), Lago Baborte and Lago Fondo in the Vallfarrera area.

CHAPTER II

STRATIGRAPHY

Introduction

Although the Palaeozoic sediments can at once be differentiated as such from younger deposits by the presence of a distinct cleavage, the division and determination of the age of these strata present great difficulties, especially in the axial zone. Only very few fossils are found, at any rate too few to give a detailed stratigraphy and certainly too few to establish time-stratigraphic units and to trace these laterally. There are, fortunately, a few lithologically characteristic horizons with a comparatively wide lateral extension which assist in establishing a time-stratigraphic framework. The usefulness of these horizons are limited however, by the facts that they are not everywhere present (e. g. chert horizon with phosphatic nodules dated as Upper Tournaisian or Lower Visean *) and that similar lithologies sometimes occur in other parts of the stratigraphic column. An example of the latter is the carbonaceous black shales, which apart from the Silurian are also occasionally found in the pre-Stephanian Carboniferous and deep in the Cambro-Ordovician, but never in such continuous bands as the Silurian black shales. Another instance is the griotte facies which mainly exists in the Famennian but is also known from the older Devonian and from the Lower Carboniferous.

Although in our region these lithologic markers are generally unfossiliferous and thus cannot be equated with time-stratigraphic units, they clearly represent the best available basis for relating the rocks to the ideal stratigraphic time-scale.

In the present area special use is made of the very typical black shale horizon, containing Silurian graptolites in some places. This characteristic horizon can easily be traced over the whole area and is used as the key for the entire mapping. Everything stratigraphically below this marker — by superposition or where this was not observable, by making use of tectonic data (plunges of fold axis, structures, etc.) — is called Cambro-Ordovician (Lower Palaeozoic), everything above it Upper Palaeozoic. The Devon and Carboniferous sequences are arbitrarily divided by the above-mentioned chert and nodule bed. When this is absent, the Carboniferous is taken as the sediments following the series of multi-coloured slates and griotte marbles, characterized elsewhere in the Pyrenees by Upper Devonian fossils.

The stratigraphic division and mapping is thus founded on one well-characterized marker, on relative age differences, on descriptions of dated strata in literature (the griotte and many-coloured slate series of the Upper Devonian) and on our own observations of dated strata outside the mapped

*) Durand Delga & Lardeux (1958) and Cavet (1958) have shown that cherts in the Mouthoumet massif and in the Eastern Pyrenees already occur locally in the very upper part of the Devonian.

area. Consequently the formation boundaries recorded on the map and discussed in this chapter cannot be expected to coincide with the ideal-time-stratigraphic names which have been applied to them.

The disadvantages and dangers of the application of ideal-time-stratigraphic names are recognized. For instance, further discoveries of fossils might necessitate constant revision of the so-called System-boundaries. However, at the present stage of mapping in the axial Pyrenees this method can be justified, pending the creation of formations in specified type areas, as the most practical way of broadly outlining the stratigraphic sequence with the available fossil evidence.

A. THE LOWER PART OF THE CAMBRO-ORDOVICIAN

The description of Palaeozoic strata will be made in the original order of deposition. Below the Silurian a thick series of slates, sandstones, quartzites and micro-conglomerates occupy the greater part of the Upper Pallaresa-Cardós basin with some coarse conglomerates, marbles and calcareous slates at the top. These topmost horizons are regarded as the upper part of the Cambro-Ordovician, and everything below it is referred to as the lower part of the Cambro-Ordovician. This latter succession can be subdivided into the Pilás-Estats series, the name derived from two mountain groups in the west and east, forming the greater bulk of the Cambro-Ordovician, and the Lleret-Bayau series, cropping out between Lleret and the lake of Bayau as limestones and black ferrugineous slates in the core of an anticline.

a. Lithologic description

1. *The Pilás-Estats series*

This thick series which takes up the greater part of the Upper Pallaresa-Cardós basin, consists of slates, sandstones, quartzites and micro-conglomerates, which laterally as well as vertically pass into each other and thus form a very variable alternation on the map.

The pure slates outcrop especially in the south against the northern flank of the syncline of Tírvia-Espot. They are mostly light grey, greyish-green, dark grey, black, often with purplish-brown cleavage planes, or with a shine of sericite on these planes ("schistes argentés", especially south of Esterri de Aneu). The slates are often provided with some pyrite, weathered or not. The slates always split according to the cleavage, which is finer in proportion as the slate has less sand admixture (paper slate). They laterally pass into hard, sandier and less schistose slate varieties and fine-grained sandstones, which are mostly of a lighter colour than the pure soft slates. The two slate types form units which can be mapped.

The reader is referred to the coloured geological map 1:50.000 which reflects the lithologic composition of this part of the Cambro-Ordovician. In these slates ripplemarks were found between Alós and Isil, but nowhere current bedding.

The sandstones, quartzites and micro-conglomerates are predominantly light grey to greenish-grey, rarely dark purplish-grey (Lago Lagola), usually with spots of weathered pyrite, varying from yellowish to brown, often also with streaks or cubes of unweathered pyrite, mostly rather compact, little schistose. Sometimes alternating sandy slates and sandstones occur in thin layers (1 mm to a few cm thick), which are variable in colour: the slate layers are dark grey to black, the sandstone whiter, sometimes somewhat reddish (fig. 11). Often these so-called "schistes rubanés" are closely related to sandy varieties of slates, having a higher content of sand.

The grain size is generally small so that it is difficult or impossible to observe individual grains with the naked eye. Locally the sandstones are coarser, when they gradually pass into micro-conglomerates which have a fine-grained, sometimes schistose matrix of fine-grained quartzite and slate, in which there are pebbles with a diameter of up to 8 mm. These pebbles consist mostly of crystal-clear rounded quartz grains and quartz aggregates, which stand out darkly against the predominantly light grey matrix. It is seldom that (for instance S. of Lago Mayor de Sens) light grey quartzite and dark slate pebbles occur in addition to quartz grains.

Fig. 11. Micro-folded and banded sandy slates ("schistes rubanés") of the Pilás-Estats series (Cambro-Ordovician). Lago Lagola.

The micro-conglomerates are especially found in zones with massive conglomerate beds, ½ m to several metres thick, alternating with fine-grained quartzitic slates. In the Brohate area these micro-conglomerate beds are generally much thinner (2 to 30 cm), and often have sharp top margins without any grading into finer grained sediments. In other cases, however, graded bedding is present in these rocks as it also is in the dirty sandstones. These rocks rarely contain carbonate. 1 km south of Pic de Brohate a normal micro-conglomerate is present, in which occasionally there occur larger pebbles of quartz and quartzite with a diameter of 2 to 10 cm. It strikingly resembles the coarse conglomerates occurring in the upper part of the Cambro-Ordovician. It is not impossible that it is in fact the same conglomerate, although the intervening slate series towards the Cambro-Ordovician limestones of the Artigue seems thicker than usual. The micro-conglomerate zone of which the above-mentioned conglomerate is a part, continues, however, along the southern side of the Bassiès granite to the west-north-west, where this zone in the Ossèse and Salat areas runs adjacent and parallel to the limestone of the upper part of the Cambro-Ordovician.

The lithologic changes in a lateral sense often take place over short distances so that the two flanks of even a small structure consist for the most part of quite different rocks. Folds except from the micro-folds, are rarely distinguishable in outcrop (cf. the quartzite in anticlinal structure north of Areo). Thus the northern flank of the great Old Palaeozoic arching appears to have a much sandier facies than the southern flank.

2. *The Lleret-Bayau series*

Between the rivers Estahón and Cardós a thick limestone zone outcrops in an anticlinal structure, south of which some less important limestone bands are present. The former consists of an alternation of thin, rather compact grey to whitish, rarely (near Boldís) darker limestone layers and pure slates. The limestones occasionally show some thin sandstone and slate layers, which locally predominate in thickness and pass into slate lenses, which contain little or no lime.

The limestone series south of the broad limestone band consists chiefly of grey calcareous slates.

As a rule these limestones are accompanied by a dark, indurated, highly pyritous band, which can be traced further east in irregular outcrops down to Lago Bayau.

b. *Stratigraphic aspects*

1. *The Lleret-Bayau series*

The Lleret-Bayau series is certainly older than the neritic slate sandstone series, described above, as they occupy the core of an anticlinal fold. The latter is marked in the field by outcrops, which are broader in the river valley than on the crest.

The view of Dalloni (1930), who regards the black zone as Silurian and of Almela and Ríos (1947), who place these limestones and dark slates in the Silurian, Devonian and Carboniferous, bordered in the south by a fault, are lithologically, stratigraphically and structurally incorrect.

It is much more probable that Cavet's "série de Canaveilles" (1951) from the Eastern Pyrenees outcrops here.

This series is defined by Cavet (1951, 1958) as situated conformably under the "schistes de Jujols" (see below) and running entirely parallel with the periphery of the gneiss massifs. It consists of a 1500 to 2000 m thick series of sericitic slates in which intercalations occur of carbonaceous slates and of grey or white often banded limestones and limestone-dolomite marbles, which are characterized by an abundance of iron-ore like that of the "ceinture ferrifère" of the Canigou massif *); it also sometimes contains siliceous lime-

*) Cf. Guitard (1954) the iron ore results chiefly from a late-Hercynian metasomatic replacement by siderite in the limestone.

stones, limestone sheets and sericitic calcareous slates. A characteristic feature is that the limestones and the slates constitute a "complexe schisto-calcaire" with a rapid alternation of the two components. The author describes this series in the north-east, south and west of the Canigou massif and further west on the Col de Puymorens, east of Andorra. Cavet correlates this series in tentative terms with the Cambrian series of the Montagne Noire, especially on account of the identical facies of the slate-limestone complex.

According to Zwart (in De Sitter & Zwart 1959) the zone of the Col de Puymorens can be followed towards the west for some distance in the southern flank of the eastern continuation of the Tór syncline, which here plunges towards the west.

In the northern flank of this syncline the series can be followed further west, although interrupted and cut off in some places by the important Soldeu-Hospitalet dislocation. At Lago Bayau this zone of black slates with an occasional limestone, flanked by quartzites reaches the Vallfarrera area.

A secand band was found by Zwart somewhat to the north in the fault zone of Mérens; this zone can be followed uninterruptedly westwards to where it crosses the French-Spanish frontier, south of the Port de Boet.

This series cannot everywhere be directly recognized as such in the mapped area. The black slates, which are still very characteristic near Lago Bayau, become less black, harder, more siliceous and strongly ferrugineous towards the west and north of Areo. The first limestone appears here in this series as an anticlinal core. In the Cardós and Estahón area, finally these limestones — still in an anticlinal structure — achieve a great development. A highly ferrugineous, locally dark and indurated band accompanies the northern border of the limestone of Lleret-Boldís. The ferrugineous dark zones otherwise occur only in places and cannot be followed over great distances.

West of the Sierra de Campirme this series is no longer present or at least can no longer be recognized as such. Whether the thin limestones of Escalarre-Esterri belong to the same stratigraphic group it is difficult to say.

Although the lithology in a westerly direction does not correspond too well with the development of the "série de Canaveilles" in Andorra and in the Canigou massif, it is still thought that the complex of ferrugineous hard slates, non-ferrugineous limestones and intermediate grey slates are identical with Cavet's "série de Canaveilles". They lie wholly in the trend of this series in Andorra and the Col de Puymorens.

2. *The Pilás-Estats series*

The "Lleret-Bayau series", of which no exact limits can be given, is followed by the Pilás-Estats series as described above. This series is known in a large part of the Pyrenees, without any great facies differences.

(Compare e. g. for the Northern Pyrenees: Laverdière (1930), Lamare (1939): western part; Destombes (1949—1955), Raguin (1938), Allaart (1954): central part; Roussel (1904), Cavet (1951/58): eastern part and Dalloni (1910/30), Schmidt (1931) for the Southern Pyrenees.)

Most workers consider this series as Lower Palaeozoic. A subdivision, however, is difficult to make through the lack of fossils and of characteristic and constant horizons. Only in some widely separated places there have fossils been found; *Calymenia Tristani* (guide fossil for the Llandeilo) and *Asaphus* spec. have been recorded from an identical series, among other places at a comparatively short distance from the mapped area near Bencarrech and Sentein (western Ariège, cf. Roussel 1893, p. 6) and near Seo de Urgel (south of Andorra, cf. Schmidt 1931).

Hence it is probable that the Llandeilo is also represented in our Pilás-Estats series.

From strata lying between this Llandeilo and the "Série de Canaveilles" no fossils are known in the Pyrenees.

The Pilás-Estats series never passes in the present area beyond the sericite epi-metamorphic stage; only east of the frontier approaching the Aston massif can the first biotite be found. Why all geological maps since 1907 (L. Bertrand "carte géologique provisoire de la partie orientale des Pyrénées") carry a broad zone of gneisses etc. reaching from the Aston to the Marimaña granite remains a mystery.

It is difficult to estimate the stratigraphic thickness of this Pilás-Estats series, especially as the intensive micro-folding, well-shown in the "schistes rubanés" (see fig. 11), has strongly thickened the series. The importance of this influence can only be guessed but certainly can not be neglected. Hinges of larger structures are mostly camouflaged by the intensive cleavage and the highly varying lithology prevents correlations from one flank to the other. On the whole the influence of these factors is too little realized by the Pyrenean geologists. Thus Destombes evidently does not take any account of it in his lithologic stratigraphic studies of the Htes-Pyrénées, Hte-Garonne and Hte-Ariège (1949/50/53/55 etc.). His tunnel sections, otherwise carefully surveyed (1947, Destombes & Vaysse) are invalidated, because they lack the support of extensive surface field work. All this makes us hesitate to attempt an estimate of the thickness of the Pilás-Estats series.

Cavet (1958) gives a thickness of 1500 to 2000 m for the equivalent of the Pilás-Estats series ("schistes du Jujols") situated between the fossiliferous Asghillian ("grauwacke à Orthis") and the "série de Canaveilles". In our region the interval between the black shales of the northern flank of the Tór syncline and the Lleret-Bayau series in the core of the next anticline would indicate a thickness a little over 2000 m, but it may get thicker westwards.

B. THE UPPER PART OF THE CAMBRO-ORDOVICIAN

a. Lithologic description

In many places, especially in the south and the west of the mapped area, the sandstone-slate succession of the Pilás-Estats series is terminated by a conformable, badly sorted, oligomict conglomerate. The latter consists of sub-rounded to sub-angular pebbles of vein quartz, dark, whitish or greenish sandy slate (near Burch), lying in a greywacke matrix. Pebbles of limestone and micro-conglomerates were nowhere found, but a single pebble of a metamorphic rock was noted in the conglomerate of the Puerto de la Bonaigua. The size of the pebbles varies greatly as also does the thickness of the conglomerate itself (see table). The pebbles are always very much larger (1½ cm) than the largest component parts of the matrix (2 mm). The quartzite pebbles are sometimes highly recrystallized (e. g. near Os de Civís) and then the conglomerate cannot easily be recognized as such. The pebbles are frequently elongated and boudinaged and often resemble exudation quartz nodules more than conglomerate pebbles.

	Puerto de la Bonaigua	Moredo (west from Alós)	near Tór	above Burch (east from Tírvia)	Son valley
Maximum pebble diameter	75 cm	15 cm	40 cm	10 cm	?
Approximate mean pebble diameter	2—15 cm	2—5 cm	2—10 cm	2—3 cm	2—5 cm
Total thickness of the conglomerate ...	100—200 m	20—30 m	20—40 m	½—80 m	a few m
N.B.	with thick slate-sand-intercalations, very schistose in places	very schistose pebbles cleaved with the matrix	with slate intercalations	matrix often greenish, many slate pebbles	

Fig. 12. Sheared conglomerate of the upper part of the Cambro-Ordovician. South-western slope of Moredo.

When the matrix is very schistose, the pebbles are often strongly flattened in the cleavage. The flattening of the pebbles of the conglomerate on the slope of Moredo is very strong here and the cleavage of the matrix runs through the pebbles (fig. 12). When the pebbles are less distorted, e. g. along the road to the Puerto de la Bonaigua, pebbles are also found showing a cleavage deviating from the general cleavage direction. It could not be decided whether a real cleavage was concerned here or only a jointing. In general, however, the cleavage affects only the matrix. The matrix of the conglomerate north-east of Estang Pudó, against the Marimaña granite has been completely granitized in contrast with its quartzite pebbles, which at least macroscopically do not show any thermal influence.

Apart from the Puerto de la Bonaigua and Tírvia, Schmidt (1931) records this same conglomerate at Escalarre (near Esterri). The occurrence of the latter is not known to the author, but the small extension and the strong wedging out, which is characteristic of these conglomerates make recognition often dependent on one single good outcrop.

These conglomerates are in turn followed by another slate series of varying thickness, which towards the top has a somewhat different character from the sandstone and slate series below the conglomerates. Near the Puerto de la Bonaigua and Alós the conglomerate is immediately followed by sandy slates. Near the above mentioned Col some thin white and light grey slaty limestone lenses are intercalated in these slates. Otherwise, this slate series is uniformly bluish-black and black and in the south and south-east dark grey. The cleavage is good, the sand content generally low and the purple-brown cleavage planes often found in older slates are absent. Towards the top these slates are transitional into the Silurian pelites.

Round the Marimaña granite and along the northern border of the main dome a thick, massive or coarsely crystalline limestone marble or dolomite outcrops, the "calcaire de Bentaillou" or the "calcaire métallifère" of Mussy (1869/70) (so named because of the abundance of ore units). This limestone horizon is lithologically characterized by its coarseness, conspicuous white colour and massive habit. Confusion with other Palaeozoic limestones is impossible in the mapped area if only on lithologic grounds.

The limestone marble is generally massive, predominantly white and whitish-grey in colour, rarely darker bluish-grey or black, with irregular flow structures. In many places these fine-grained marbles pass into coarse-grained varieties, which are poorly stratified; near Moredo for instance it consists of calcite crystals which are occasionally 15 mm long, possibly due to thermal metamorphism by the Marimaña granite. In general, however, the marmorization and crystallization seem to be due to regional and perhaps even to dislocation metamorphism. Locally the limestone is rather porous and consists of detrital grey calcite grains of 1 to 2 mm in diameter, which gives it a saccharoid aspect. This limestone is rapidly disintegrating in the Pouil area and then forms a pure lime (calcite) "sand". Owing to the absence of stratification the outcrops are always rounded and smooth.

In the area of Cuenca and Moredo, west of Alós and locally east of the Puerto de la Bonaigua the metalliferous limestone begins with a thin, white, massive limestone layer, which is generally no thicker than 10 m. This is followed by a sequence of rhythmically banded limestones varying from 40 to 150 m in thickness (called by Caralp (1888) "barrégiennes"); limestone beds of 1 to 50 cm thick alternate with sandy to clayey beds, of the same thickness or somewhat thinner which have a baked or silicified appearance. These indurated layers weather out on the outcrop surface and often contain some carbonate; in the thermal contact aureole of the Marimaña granite they are converted into lime-silicate rock, whereas the limestone layers do not macroscopically show new minerals. Cleavage is commonly well developed in these limestones and then the competent indurated beds are connected with each other by numerous, very thin layers of the same material parallel to the cleavage of the limestone beds. A similar but coarser feature has been described by Van Alphen (1956, p. 488) and it remains doubtful whether this process is purely mechanical or whether recrystallization has also played a part in it (see fig. 13). These "barrégiennes" are also very clearly developed east of the Puerto de la Bonaigua and in the Cougnets valley also at the base of the thick limestone series. For the rest "barrégiennes" are not restricted to the base and are not even restricted to this Cambro-Ordovician limestone, as this kind of limestone is also found in the limestones of the Lleret-Boldís series and in the limestones of the lower part of the Devonian.

In the area of the rivers Cougnets, Ossèse and Escorce the limestone is for the most part replaced by a massive, coarsely crystalline, non-stratified dolomite of a whitish, light brown or bluish-grey colour. Often whitish beds of a few mm alternate with bluish-grey ones; sometimes these are again obliquely intersected by another alternating system of similar layers

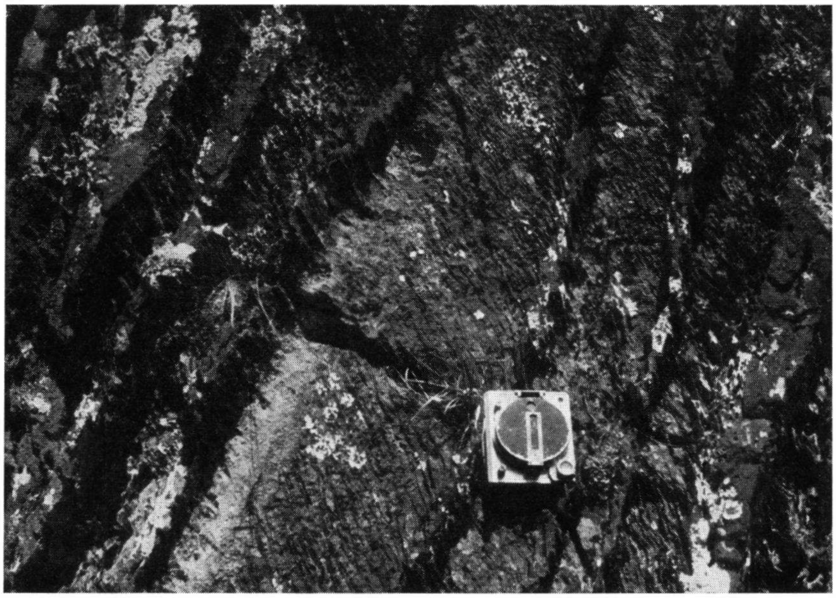


Fig. 13. Metalliferous limestone with hard sandy intercalations ("barrégiennes") of the upper part of the Cambro-Ordovicien. Cirque d'Anglade.

("structure en zigzag", Zwart 1954, fig. 1). One of the two systems probably represents the stratification, the other system possibly the original cleavage, partly obliterated by recrystallization. Along these cleavage planes replacement must have taken place as in the "barrégiennes" (see above).

North of the Puerto de la Bonaigua there are numerous yellowish-brown dolomite "patches" in the limestone which contrast strongly with the white marble. These "patches" have a diameter varying from 10 cm to some tens of m, and the dolomitization has no relation to the stratification. North of the Marimaña granite the limestone has nowhere been dolomitized.

The dolomites of the Cougnets, Escorce and especially of the Ossèse are characterized by many quartz veins and nodules of stem-quartz, with idiomorphic, clear quartz crystals up to a length of 8 cm. They contain also many pyrite crystals in pentagonal-dodecahaedral form with a thin limonite coating. Also outside the dolomites quartz and pyrite are sometimes present in the "calcaire métallifère", although not in such great quantities.

This "calcaire métallifère" is altogether absent in the synclines of Espot-Tírvia and Tór. Instead of this there occur in some places here and rather constantly along the syncline of Tór, grey or greyish-blue, sometimes pink (near Tór) slaty limestones and calcareous slates, which are well stratified, rather compact and are never crystalline. They often have somewhat satiny cleavage and stratification planes. The Devonian limestones of both synclines show this same characteristic. They can, however, be easily distinguished from the thin (30 m) slaty limestones in the Upper Cambro-Ordovician. In the north these calcareous slates are more important, especially north of the Marimaña granite and between the Cougnets and the Escorce, where they can be distinguished only with great difficulty from the Devonian limestones, when the typical Silurian pelites are absent or when they do not occur in their characteristic form. In places where the "calcaire métallifère" is present, these calcareous slates always occur as a younger series.

Between the two limestone series of this region and sometimes intercalated in the limestones we find black slates, poor in sand which were described above and in which a conglomerate band occurs below the waterfall of the Ossèse. This conglomerate has a dark grey to black, fine-grained matrix with somewhat lighter, quartzite pebbles, which are mostly not bigger than 2 cm and which are rather well-sorted and well-rounded, but tectonically more or less flattened (// cleavage). The conglomerate is only of limited extension.

b. Stratigraphic, geographic and facies aspects of the upper part of the Cambro-Ordovician

The upper part of the Cambro-Ordovician, containing as it does, two conglomerate and two limestone horizons, is of great stratigraphic significance because it is the only, reasonably accurately dated section of the Lower Palaeozoic in the Pyrenees.

1. *Distribution and nature of the limestones*

In the Central and Eastern Pyrenees between the Pau valley in the west and the Aude and Fresser valleys in the east, a limestone-slate horizon generally occurs near the top of a slate-greywacke series a short way stratigraphically below the black shales of the Silurian (cf. Bresson 1903, Destombes 1953, Roussel 1893, Dalloni 1930, Schmidt 1931, Boissevain 1934, Fontboté 1949, etc.). This horizon is also present in the southern flank of the Montagne Noire, cf. Gèze (1949).

The "calcaire métallifère", on the contrary is only known in a comparatively small area which extends from the Garonne in the west up to and including the St. Barthélemy massif in the east (see fig. 14 and the lithologic columns of fig. 15).

If we draw up an isopach map for the "calcaire métallifère", making use of data given by Kleinsmiede (1960, Valle de Arán), Allaart (1954, Aston massif) and Zwart (1954, St. Barthélemy massif) and the author (fig. 14), then it appears that in its present folded form it covers a surface of some 100 by 20 km, situated chiefly on the northern border of the present-day axial zone and in the St. Barthélemy massif. The limestone has its thickest development between the Bonaigua pass and the Ossèse valley. Even if the

Hercynian north—south compression is taken into account, the sedimentation area appears to be more or less elongated and lens-shaped.

Let us sum up some typical qualities of this "calcaire métallifère" occurrence:

1. comparatively small elongated area of deposition,
2. high content of primary Mg,
3. massive often unstratified habit,
4. great local differences in thickness over short distances and a somewhat unconformable character at the top (e. g. between the Cougnets and Ossèse valleys),
5. occurrence of coarse-grained, porous calcitic limestone of detrital origin,
6. occurrence of a 450 m thick dolomite in the Ossèse valley, which has a very small lateral extension (in an east—west direction) and which is here not regarded as an anticline of the normal "calcaire métallifère" (see also fig. 19).

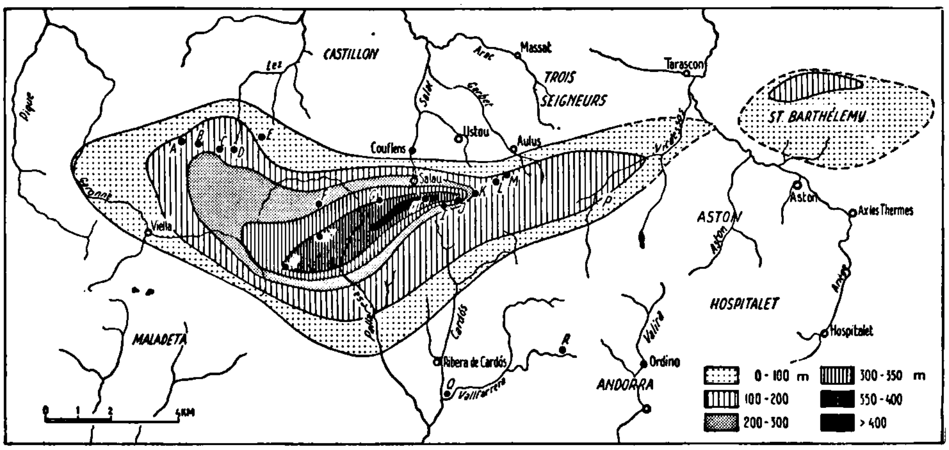


Fig. 14. Isopach map of the metalliferous limestone. Thicknesses in m. Locations of the lithologic columns of fig. 15 are also indicated (A—R).

Although the complete absence of fossils and the enormous thickness in the centre of the limestone area are not characteristic for a real bioherm a reef origin for the "calcaire métallifère" must, à priori, not be considered out of the question.

2. *Distribution and nature of the conglomerates*

Compared with the extension of the calcareous slates, conglomerates occur in the top layers of the Cambro-Ordovician distributed over a much longer distance. The deposition area covers practically the whole of the Pyrenees, the Mouthoumet massif and the Montagne Noire; the conglomerates are, however, less continuous. They are always situated below the calcareous slates, sometimes above the "calcaire métallifère", sometimes below it, and sometimes both above and below (see fig. 15).

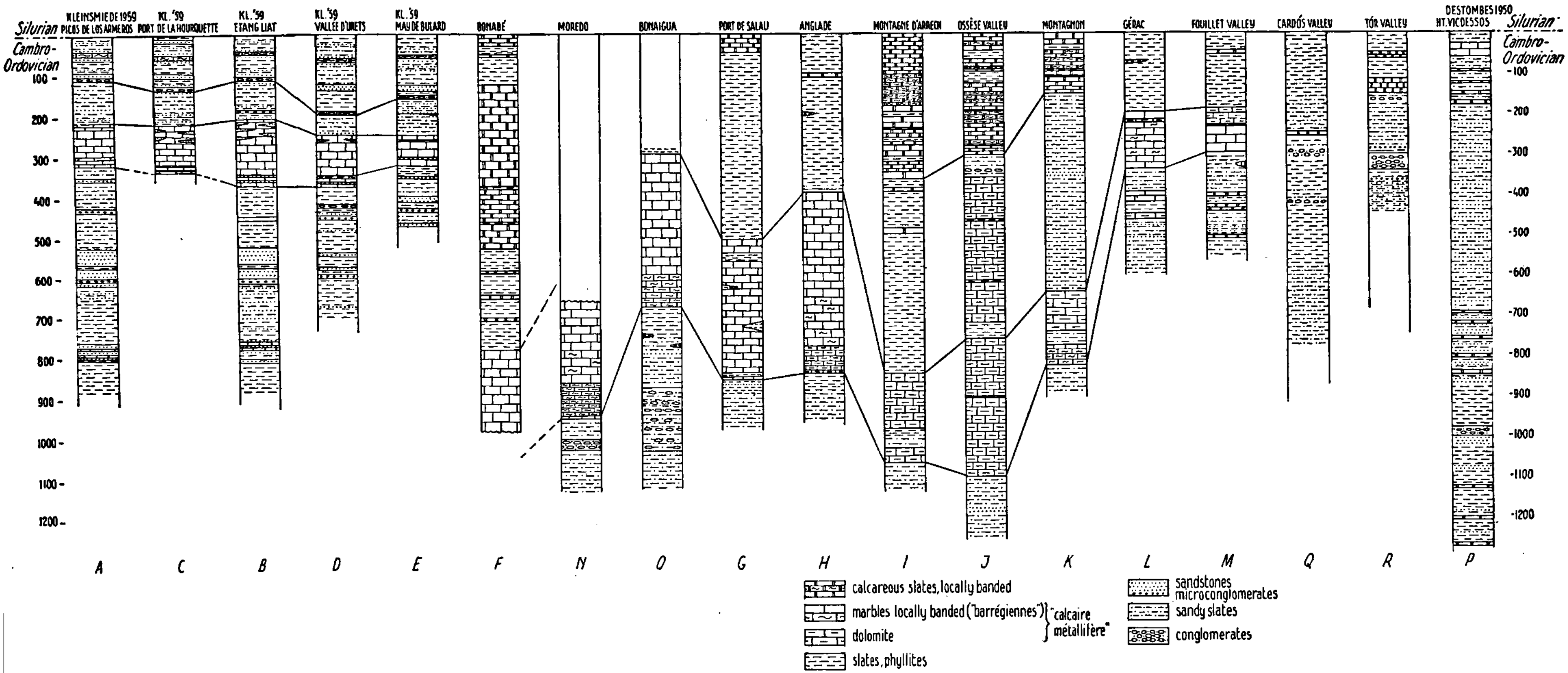


Fig. 15. Lithologic columns of the upper part of Cambro-Ordovician in the Garonne, Pallaresa, Salat et Vicdessos valleys. Locations of the columns are indicated in fig. 14. N.B. Kleinsmiede 1959 should be read Kleinsmiede 1960.

On the whole the conglomerate consists of pebbles of quartz and quartzite, while indurated slates in a sandy clay matrix are rare; occurrences of limestone pebbles (Roussel 1893: north of the Quérigut massif), gneiss pebbles (Roussel 1904, Fontboté 1949: Ter valley) and extrusive rocks (Fontboté op. cit., Raguin 1946 a: western Ariège) are very rare. The pebbles must have come from far, as micro-conglomerate pebbles have not been recorded, which one would not expect when the source area was nearby because the latter are as hard and resistent as the quartzites and are very frequent in the Cambro-Ordovician of the Pyrenees.

Schmidt (1931) is of the opinion that the origin of these pebbles lies to the south and supposes that the pebbles are Arenig sandstones and presumably Cambrian quartzites.

An unconformity has nowhere been recorded below any of the conglomerates, except by Schmidt (1931), who describes an angular unconformity of 60° near Vilamur (Sort). However, this outcrop represents the unconformable base of the Trias.

3. *The age of the limestones, slate-greywackes and conglomerates*

The upper calcareous slates and the slate-greywacke series beneath it down to the conglomerate contain in many places, distributed over a large part of the Pyrenees and the Montagne Noire a fauna of Brachiopoda, Cystidea, Polyzoa and Coelenterata: the classic "faune à *Orthis*". Two *Orthis* species are frequently found in it: viz. *O. calligramma* Dalm. and *O. Actoniae* Sow., of which the former species — only occurring in the greywackes and slates — can be considered as characteristic of the lower part of this section: the Caradocian s.s. (Schmidt 1931). *Orthis Actoniae* — occurring in the calcareous slates — would, on the contrary, belong to the upper part of the Caradocian s.l., viz. the Ashgillian (cf. Schmidt 1931 and Dreyfuss 1948). The greywackes of the Catalan Pyrenees especially have yielded an abundant Caradocian fauna, among other places near Seo de Urgel (Dalloni 1930, Schmidt 1931).

Although in the mapped area no fossils were ever found in this calcareous slate series, there is no reason to doubt their identity with the Caradocian s.l. in which case the calcareous slates might represent the Ashgillian.

Autran & Guitard (1955) mention a "faune à Orthis, Bryozoaires, Polypiers" of the Pic des Aubières (north-east of Et. d'Aréau), of the Fraychet river (tributary of the Alet) and near Col du Picou de la Mire, which they consider Caradocian. On stratigraphic and structural grounds, however, the writer believes that these strata belong to the Devonian (see p. 53).

The Caradocian greywackes themselves often directly overly a conglomerate, sometimes (Sierra de Tosas) separated from it by black slates with Fucoids, the latter being correlated by Dalloni (1930) with the slates with *Asaphus* and *Calymene* cf. *Tristani* of Bencarrech in western Ariège, which would be Llandeilo. Hence Dalloni considers the conglomerate to be Llandeilo. On the whole, however, the conglomerates are described as Caradocian (cf. Destombes 1953). According to Gèze (1949) the Llandeilo is absent in the Montagne Noire and the Caradocian begins with a transgressive conglomerate.

The non-fossiliferous "calcaire métallifère" is considered by Dalloni as belonging to the Devonian, as also by Bertrand (1907/11). Durand & Raguin (1943), De Sitter & Zwart (1950) and others argue for a probable Caradocian age. Caralp (1888) and Destombes (1953), however, place these marbles below the Caradocian. The last-mentioned author considers the conglomerate as a constant horizon; from the Montagne Noire to far west in the Pyrenees this conglomerate always separates the fossiliferous Caradocian from the "calcaire métallifère", the latter is consequently considered to be older: Arenig or Upper Cambrian.

This extreme view of Destombes is certainly not correct: the conglomerates of Alós de Isil and of the Puerto de la Bonaigua are clearly situated below the limestone. Near Bentaillou occur two conglomerates, one below and one above the limestone (see fig. 15). Outside the area of the "calcaire métallifère" only one conglomerate is known. But it is doubtful that this conglomerate is everywhere synchronous and even if that were the case we are still confronted with the problem of with which of the two conglomerates in the "calcaire métallifère" area it ought to be correlated. As long as no fossils are found it cannot be ascertained whether the marble is of Caradocian age or older, and for the present we are even doubtful whether the conglomerates and the "calcaire métallifère" are everywhere synchronous horizons.

Both the intraformational conglomerates and the epi-continental limestone indicate a period of emergence, of which the origin like that of the Llandeilo hiatus in the Montagne Noire should be found in epeirogenetic movements connected with Caledonian movements.

C. THE SILURIAN

The sequence of conglomerates, limestone-dolomites, calcareous slates and dark slates of the upper part of the Cambro-Ordovician is followed by a comparatively thin, but very characteristic sequence of black shales. These black shales outcrop in the core or in the flanks of some of the larger structures of the mapped area, such as the great Lower Palaeozoic dome, the anticlines of Couflens and Maubermé and the syncline of Tór and Tírvia-Espot with the continuation in the flanks of Monseny, and also in some smaller, partly diapiric structures in the centre of the Devonian of the upper Pallaresa.

a. Lithologic description, geo-chemical aspects

The Silurian deposits consist in their typical development of very fine-grained, fissile, black shales which often stain the fingers; they are poor in quartz, not very rich in iron but usually rich in carbonaceous matter. In places the shales are rich in Silurian graptolites. In an undisturbed position these extremely fissile shales easily split, without breaking, into very thin and remarkable large, somewhat flexible sheets. On the other hand when they are strongly affected tectonically the Silurian pelites have a greasy shine and an irregular undulating fissility which, as a rule, does not show parallelism with any tectonic direction. The Silurian pelites are characterized more than any other Palaeozoic sediment by their pronouncedly incompetent character which has been of the greatest tectonic significance (see p. 76).

A consequence of this feature is that it is not possible to determine the original thickness of this formation with precision. The differential movements which have taken place everywhere between roof and wall have as a rule left little of its original thickness, they are either thickened or thinned or even absolutely squeezed out.

Compare for instance the thickness of the Silurian of Mount Monseny with that in the syncline of Tírvia-Espot, south-east of Monseny or with that on the northern border of the Lower Palaeozoic dome where the black shales are in some places completely squeezed out. In the first case we are inclined to take a thickness of several hundred metres for granted; in the second case a thickness of 20 to 30 m is more in keeping with the field-data which also show that the Silurian here is about equally thick over a rather long distance and in both flanks.

In addition to this the disharmony between Lower and Upper Palaeozoic structures is less evident in this syncline than for example on the northern side of the central dome. The conclusion that the Silurian in the syncline of Tírvia-Espot was developed very thinly indeed seems to be justified. It is, nowadays, held that the Silurian in the Central Pyrenees does not exceed a thickness of 200 m (De Sitter 1954c, p. 296, Destombes 1953, p. 112, etc.). which is still considerably more than in the Tírvia-Espot syncline. The black shales show very little resistence to erosion. Consequently the Silurian is characterized by soft and smooth erosion forms.

The Silurian pelites contain much iron in the form of pyrite crystals and veins, on the surface mostly in a strongly weathered condition. The bedding planes are often of a rusty-brown. The iron-content of a very black sample derived from a tunnel freshly made for a hydro-electric plant not far south of the mouth of the river Mauricio on the Noguera Pallaresa is however not particularly high (according to De Sitter 1954c, p. 295: $FeO + Fe_2O_3 = 7{,}7\ \%$) but it occurs in easily soluble form, so that wherever water trickles over the rocks a thick ferrugineous cake of iron-ochre is left on the surface. Hence Silurian springs taste strongly of iron, and sometimes even slightly of H_2S; stagnant water often has a bluish iron-film. The very high alumina-content of the same sample is remarkable, viz. 33,5 %, as also is the small quantity of free quartz expressed by the low SiO_2-content (40 %). Kleinsmiede (1960, Valle de Arán) gives some analyses which show however a less extreme picture, viz. 60—75 % SiO_2 and 19—23 % Al_2O_3, ± 20 %.

Capdecomme (1943, p. 189) analysed Silurian pelites of Marignac (Hte Garonne) especially on their carbon-content reaching 3,8—24,5 %. These results are in good accordance with data given by Kleinsmiede (1960) who found a staining black shale of the Valle de Arán to contain 4,4—8,8 % carbon. However, a not staining sample of Silurian pelites shows about the same carbon-ratio as a black sample of the upper part of the Cambro-Ordovician derived from the Bonabé area, viz. $< 0{,}5\ \%$. The high C-content is a typical property of black shales (cf. e. g. also Pettijohn 1949, table 84 analysis "F" of a black shale of Dry Gap (Georgia, U. S. A.) with 13,11 % carbon. According to Capdecomme the carbon is present in the form of graphite. The shales often have a greasy shine as of black lead, especially on the undulating slickensided surfaces. The occurrence of chiastolite in contact-metamorphic Silurian pelites all over the Pyrenees also points to a high carbon-content. In the contact-aureole of the Riberot granite on the northern bank of the upper Pallaresa for instance this mineral is very prominent. Chiastolite is found occasionally also in metamorphic Carboniferous and Cambro-Ordovician sediments of the Pyrenees, but not in the mapped area.

In contrast with the above-mentioned analyses Mrs De Sitter—Koomans (in De Sitter 1954c, p. 295) found a carbon-content of only 0,3 % in a very black, staining sample, whereas most dark shales have a carbon-content of 0,5—1,5 % (the average carbon-content of shales is according to Leith & Mead (1915, p. 76) and Clarke (1924, p. 30) 0,81 %). Nevertheless the black colour of black shales is not always due to carbon. Fearnsides (1905, p. 613) ascribes the black colour of the Upper Cambrian Dolgelly beds in north-Wales, which contain very little carbonaceous matter, to very finely divided FeS_2, an opinion which is shared by many.

The sulphur-content of the analyses given by Kleinsmiede (1960) and De Sitter (1954c) of samples without macroscopic pyrite varies from 0,011 to 0,39 %. However when combined with iron this proportion of sulphur would not give sufficient pyrite to give the rock black staining properties. Apparently the Pallaresa sample contained only a few thin staining layers along which the rock splits very easily giving the rock the appearance of being entirely a well staining rock.

Occasionally (e. g. Rouze, east of Couflens) the rocks are weathered to a whitish mass probably very rich in aluminium (alumearth?). Both Chaubet (1937, p. 50) and Caralp (1888, p. 185 and p. 339) state the occurrence of alum in weathered Silurian pelites. Gypsum crystals, especially on the warm, dry, southern slope of the Pyrenees occur as needles in this weathered matter.

It is clear that, if the sediment owes its black colour to iron, the black colour should not occur if alum is formed but on the other hand, it is plausible that in such case the graphite would disappear from the rock as CO_2.

In their typical development the black shales cannot be confused with any other shale or slate in the mapped area. In the eastern Pyrenees black shales, however, are known from the much older "Série de Canaveilles" (Cavet 1951a + b; Cavet 1958). The black slates of the equivalent of the "Série de Canaveilles" in the mapped area cannot be confused.

However the Silurian is not developed everywhere in its characteristic habitus. The black shales may pass into a more greyish-black, non-staining, less fine-grained, less soft and less fissile variety, which also contains less iron. This facies can hardly be distinguished from the dark slates of the upper part of the Cambro-Ordovician.

Such is the case for the whole of the Silurian between the Plá de Beret and the Salat on the northern border of the Lower Palaeozoic dome. It is only between the Bcos de las Lausanas and the Circres that it is present in its characteristic habitus. On the other hand the Silurian in anticlinal and diapiric structures, outcropping in the Devonian of the upper Pallaresa does on the whole show the development in the black shale facies. The fact that similar less characteristic pelites also belong to the Silurian is indicated by the graptolites, mostly *Monograptus,* which are found in it near Escaló.

The occurrence of limestones in the black shale series is of secondary importance. The only important outcrop is formed by light bluish-grey calcareous slates, alternating with numerous thin black staining layers in the core of the anticline of Couflens. It extends over more than one km on either side of the Salat and attains a thickness of at least 70 m.

Similar, but less important limestones, outcrop on the Renacha plateau, east of the Alet, between the mouth of the Mauricio in the Pallaresa and Escaló, round Mount Monseny and near Tór. Near Etang d'Aréau, finally, there occur a few thin greyish-blue, yellow-brown-weathering, hard fossiliferous limestone plates in grey calcareous slates. A lens of white, massive limestone outcrops in the Col de la Serre du Cot.

The horizon with black elliptical, often dolomitized, fossiliferous limestone-nodules, so well-known to many authors, was not found in the mapped area; we have come across a black, more or less schistose limestone, into which the horizon of limestone-nodules locally passes, according to Chaubet (1937, p. 98), Destombes (1953, p. 112) and others. This black, sometimes rather granular limestone occurs in the Ossèse valley, about 2 km south of Estillon, in the northern anticline of the Silurian of the Col d'Escots, near the mines of Carbauère on the western bank of the Escorce and occasionally in the Silurian which outcrops in the Devonian of the upper Pallaresa. South of Monseny, just outside the mapped area these black limestones contain *Cardiola interrupta* Sow.

East of the Ossèse there occurs locally at the base of the black shales a dark, fine-grained, ferriferous, quartzite of 1 to 1½ m thick. It was found near the mines of Carbauère and in the valleys of the Fouillet and the Ars.

Near Arigail-Portet on both banks of the Alet the Silurian is very much indurated; so much so that here and there a rock is formed resembling a black chert; the stratification is greatly disturbed and many quartz veins occur in it. This silification of the Silurian is connected with a large east—west fault (see p. 78).

The transition to the Devonian limestone is generally not abrupt, but occupies a zone several metres thick, consisting of an alternation of thin layers of staining shales and grey calcareous slates, the thickness of which varies from several mm to a few cm. Round Mount Monseny the upper limit of the Silurian cannot be fixed with precision, for the transitional zone has a thickness here of many tens of metres. In this transitional zone the limestone and the black shale layers are somewhat thicker.

The transition to the black slates of the upper part of the Cambro-Ordovician is often very gradual, so that here, too, it is difficult to define the exact boundary.

It is clear that on account of the scarcity of fossils the lower and upper limit of the Silurian have been fixed only on lithologic grounds.

b. The environment of the Silurian pelites

The Silurian pelites must have been deposited in an euxinic environment as only graptolites have been found and no benthonic forms although the latter would have been more easily preserved than the delicate graptolites. The graptolite-concentrations point to a sudden mass mortality, e. g. because of toxic bottom water (containing H_2S)mingling with surface water as a result of convection currents.

The original thickness of the mud is strongly reduced by compaction with the result that graptolites which are almost equally thick as broad in limestone, are extremely thinly compressed in the black shales. Debyser (1955, p. 332) found a water-content of 80 % in recent lagoon sediments of the Ivory Coast which may be compared to our Silurian pelites.

The fine grain, the very even lamination, the low free quartz-content, the uniform character of the sediment and the absence of a benthonic fauna, point to a deposition in quiet stagnant depths below wave base, either remote from land or near land without much relief.

c. Stratigraphic aspects

On the whole the Silurian of the Pyrenees is rather distinctive. The uniform development of the black shale facies makes it an easily mapped rock-stratigraphic unit, but also faunally the Silurian is well-known: rich faunas, especially of graptolites, but also of molluscs, brachiopods and trilobites are found in many places. These faunas point to the presence of at least the greater part of the Silurian, as only the Lower Llandovery (s.s.) and the Upper Ludlow and Downtonian are not characterized by fossils and consequently not known as such, cf. Laverdière (1930, Lez and Orle), Dalloni (1930, p. 72, Catalan Pyrenees), Schmidt (1931 = 1944, p. 140, Southern Pyrenees), Chaubet (1937, p. 96, Southern flank of Montagne Noire).

Dalloni (1913, p. 244) however supposes that the limestones with *Pterygotus* and *Ceratiocaris* which he found between the Pallaresa and the Sègre near Castellás represent the youngest stage of the Silurian. The same author (1952) describes some graptolites of Laruns (Bse—Pyr.) which seem to point to a Lower Llandovery (s.s.) age.

Also in the present area the writer found a number of fossils, mainly in previously unknown localities. Unfortunately the graptolites sent up for determination have as yet not been returned. The localities are the following:

1. Along the road from Rieu to Faup, just after the first hair-pin bend (about one km north-west of Couflens) numerous graptolites.
2. Neighbourhood of Col de la Serre du Cot (east of Couflens) one specimen of the genus *Monograptus.*
3. In the northern Silurian anticline beneath Col d'Escots at about 1150 m alt. graptolites and crinoids. Mr. G. Waterlot (Lille) was so kind as to determine these graptolites:
   Monograptus Halli Barr. (zone 21—22 of Elles and Wood)
   M. cf. *Sandersoni* Lapw. (zone 18—19 of Elles and Wood) specimen to 30 cm
   M. cf. *jaculum* Lapw. (zone 19—21 of Elles and Wood)
   Hence the Upper Llandovery (s.s.) is certainly present here.
4. In the hair-pin bends of the motor-road from the Pallaresa valley to Espot. Numerous graptolites.
5. Pallaresa valley 1 km north-west of Estarón. Numerous graptolites.
6. Pallaresa valley, near Escaló. Roussel (1893, p. 82) mentions here *Monograptus priodon* Bronn. (zones 22—29, Tarannon and a part of the Wenlock) and *M. proteus* Barr.
   In the black shales of the southern flank of the Tírvia-Espot syncline, just south of our area numerous graptolites have been found, Dalloni 1930, p. 57) mentions some ten species of graptolites near Llesuy, representing the Wenlock. The author, too, found a great many graptolites here.
7. On the southern bank of Etang d'Aréau in calcareous slates alternating with some thin compact limestone-bands occur many Orthoceras, trilobites, *Cardiola interrupta* Sow. and *Dayia navicula* (determination Dr Shirley, New Castle).
   These lime slates have consequently to be placed in the Ludlow. Without many lithologic changes they pass into calcareous slates and limestones which are usually reckoned to be Devonian. If these Ludlow limestones form a continuous rock-horizon, we shall have to take into account that the boundary Silurian-Devonian, mapped as the level above which no black shales occur, should be placed higher. This makes the Silurian of the map merely a rock-stratigraphic unit.

The Silurian limestones, often characterized by a rich fauna of molluscs and brachiopods are always reckoned to the upper part of the Silurian. Destombes & Vaysse (1947, p. 405) and Destombes (1953, p. 112) describe in the Vallée de la Pique a black thinly stratified limestone with calcareous nodules in which there occur many Orthoceras and *Cardiola interrupta* Sow. This limestone was already considered to be Lower Wenlock by Barrois (1887). Dalloni (1930, p. 72, Catalan Pyrenees) describes limestones, belonging to the Upper Wenlock and Lower Ludlow on the ground of faunas rich in molluscs and brachiopods. Also Boissevain (1934, p. 48, upper Sègre) mentions black and blue limestones from the upper part of the Silurian. Schmidt (1931 = 1944, p. 143) attributed limestones with Orthoceras in which he found some characteristic trilobites near Montardit, to the Lower Ludlow. Finally Chaubet (1937, p. 98, southern flank of Montagne Noire) considers black shales with black marly nodules as Wenlock and grey dolomitic limestones as belonging to the Upper Wenlock and Lower and Middle Ludlow.

Owing to tectonic complications it is, as a rule, difficult to determine the stratigraphic place of the non-fossiliferous Silurian limestones in the present area, the author has, however, the impression that in this area the limestones also occur exclusively at the top of the black shale sequence.

Quartzites are also known elsewhere from the base of the Silurian. They are always placed in the Llandovery because they lie amidst black shales with a Llandovery fauna; cf. Laverdière (1930, western Pyrenees), Schmidt (1931 = 1944, p. 140, 144) and Boissevain (1934, p. 44 ff) upper Pallaresa and upper Sègre. Chaubet (1937, p. 97) also considers "schistes quartzeux" as Llandovery.

Summarizing we may say that the Silurian in the Pyrenees and Montagne Noire has developed very uniformly and that the black shale facies seems confined to this series. It is, however, not true that the whole Silurian has developed only in black shale facies. At the base (Llandovery) there locally occurs a quartzite, the upper part (Wenlock and Ludlow) contains many lithologically-different limestones. It is not impossible that a part of the limestones which we reckoned to the Devonian also belong to the Silurian and represent a part of the Ludlow and the Downtonian.

D. THE DEVONIAN

The Silurian is followed by a limestone slate series, which in the north consists mainly of an alternation of the two components; in the south-east (synclines of Tór and Tírvia-Espot) however, it consists predominantly of limestone. The upper part of the sequence is generally variegated and the limestones, moreover, nodular. These nodular limestones or "griotte" are usually considered to be Upper Devonian, everything below it will be described as the lower part of the Devonian. In view of the somewhat different succession in the lower part of the Devonian in the north and in the south of the mapped area we shall discuss these Devonian areas separately.

a. Lithologic description

1. *The lower part of the Devonian on the northern border of the axial zone*

This Devonian is characterized by a coarsely rhythmical alternation of limestone and slate layers, each of which has a thickness of some tens of metres and can be followed over long distances.

The calcareous rocks. — In a fresh condition the limestones are dark blue to dark bluish-grey, rarely light grey, with a weathered surface layer of 2 cm maximum thickness of brownish-grey to light grey colour. The limestone is either compact and badly stratified with rounded weathered surfaces or a rather thinly stratified platey limestone. The colour of the bedded limestones is somewhat lighter, more greyish-blue to greyish-brown, inside with light coloured bedding planes which, owing to finely divided sericite, shine like silk. They pass into limestone slates of the same colour, also with a sericite gloss. The limestones are rarely crystalline. Marbles very seldom occur east of the Port de Salau, more often north of the upper Pallaresa, especially as a result of thermal metamorphism round the Riberot granite. Chiefly west of the Salat the limestones contain practically everywhere, dark, thin slate layers (varying from 1 to 5 mm, rarely up to 1 cm), mostly broken rectangularly and drawn apart. These layers occur at a distance of ½ to 30 cm from each other. Towards the west these thin layers gradually become more siliceous. In the western region these limestones, banded in a finely rhythmical way, resemble the banded sandy limestones from the upper part of the Cambro-Ordovician, where the layers are on the whole much sandier than in the Devonian. South of the Col de la Trape adjacent to the griotte marbles occurs a very massive, hard, blue limestone with much calcite. In the valley of the Salat, south of Salau, the basal limestone occasionally shows a peculiar weathering with numerous amygdaloidal cavities, of ½ to 2 cm length arranged in rows parallel to the stratification. This selective weathering is probably caused by the inhomogeneous composition of the limestone.

In only a few places (Spioulou and Rabé, Salau area) the limestones are dolomitized on a small scale, the colour of the dolomite being whitish-grey. The dolomitic parts always contain some quartz in veins or patches.

The non-calcareous rocks. — Usually these non-calcareous rocks are pure slates, which vary in colour from dark grey to bluish-black and are practically always very fine-grained and of homogeneous composition. It is only west of Bonabé that the grain is coarser and the composition sandier. The slates are very fissile, but the fissility decreases with increasing sand content. The fissility plane (probably the cleavage) is mostly finely wrinkled. Towards the north the slates become lighter grey and east of the Alet even greenish-grey to light green, as also locally in the Devonian of the upper Pallaresa.

2. *The lower part of the Devonian of the Tór and Tírvia-Espot synclines*

This southern facies of the Devonian is characterized, in particular east of the Pallaresa, by a much more pure calcareous development. The rare slates can be traced only for a short distance. Towards the west and the north-west pure slates occur more frequently, though nowhere so frequently and so regularly as in the northern zone.

East of the Pallaresa this lower part of the Devonian consists of compact limestones and calcareous slates. The colour of these rocks varies from whitish to light brownish-grey, whereas the colour of the slates sometimes inclines to sea-green. Locally some light greyish-green slates are intercalated. The limestones in the syncline of Tór are on the whole very schistose.

Towards the west the limestones become more marbly, probably as a result of the thermal contact metamorphism of the Maladeta granite, and in the same direction the number of very hard, siliceous layers increases. These layers are 1 to 100 mm thick and always protrude by weathering. Round the Teso and the Monseny these limestones look very much like the "barrégiennes" of the upper part of the Cambro-Ordovician. In this region massive blue limestones alternate with greyish-green and brown calcareous slates. North of Espot the slates, here varying from blue to greenish-grey, are sometimes very sandy. In general this Devonian as it extends in a north-western direction bears a greater resemblance to the Devonian of the upper Pallaresa.

In the above-mentioned syncline greyish and whitish limestones are found directly on top of the black shales of the Silurian.

The Devonian on the crest between the rivers Pallaresa and Cardós consists from top to bottom of:

Upper part:	
— whitish to yellowish-white limestones and calcareous slates with traces of griottes 1 to 2 dm ø, alternating with yellow-brown, grey or sea-green slates, sometimes grey with a purple shine 1 to 4 dm ø	30—40 m
Lower part:	
— grey, mostly calcareous paper-slates, alternating with some thin (1 dm) compact grey limestones, towards the bottom with lighter coloured, calcareous slates and limestones	120—140 m
— rather compact, greyish-blue limestone with white calcite veins	4 m
— grey to dark grey calcareous paper-slates, towards the bottom with some limestones (5 to 50 cm ø)	60—70 m
— compact grey limestone with white calcite veins or layers, somewhat more schistose in the middle	50—60 m
— yellow-brown to grey calcareous slates alternating with a few thin greyish-brown limestones	15 m
— paper-slates, grey, inclining to green	3 m
— greyish-blue, rather compact limestone	½ m

3. *The upper part of the Devonian*

The series of the rather monotonously coloured limestones and slates, as described above passes towards the top into a series of many-coloured slates and nodular limestones, the latter predominating.

This multi-coloured complex outcrops in the north-east against the North-Pyrenean fault, in the fault zone of Couflens-Arigail and in the syncline of Tírvia-Espot, east of the Pallaresa.

In their typical development the rather massive marble-limestones are whitish, yellow-brown, reddish-brown, red, sometimes violet-red, light violet-grey or sea-green, as a rule with a more or less clear indication of lime nodules, which are ½ to 2 cm in mean diameter, mostly almond-shaped, and arranged in rows parallel to the stratification. These so-called "griottes" lie in a "matrix", containing somewhat more clay than the griottes themselves. This matrix is mostly of the same colour as the griottes and is generally very subordinate, just a thin layer separating the nodules (Farrera). Similar very typical griotte marbles are, in fact, only found in the mapped area in the syncline of Tírvia-Espot east of Burch.

The griotte marbles generally occur in layers, which are some dm to some metres thick, alternating with slates, which are mostly calcareous and have the same colour variation as the griottes, viz. red, red-violet, bluish-green, sometimes with light violet and green spots, black, very often greyish inclining to green or red.

The two components alternate without being sharply separated. On the whole taking them separately the limestones or slate beds are each of one constant colour (unless they are mottled); the alternation of griotte marbles and slates, however, produces a variegated effect.

With the less typical griotte marbles of the northern Devonian — in so far as they outcrop in the mapped area — the matrix often occupies a greater volume; occasionally there occurs a single griotte layer, where the long axis of the griotte stands obliquely to the stratification parallel to the cleavage (among other places near Sérac).

In the Couflens fault zone griotte marbles occur in this series only by way of exception (e. g. 500 m south-east of the Col de la Serre du Cot). There the limestones are rather mottled in green and reddish or greyish-green with light lilac, or light ochre-coloured to light reddish-brown with somewhat darker spots.

In the syncline of Tírvia-Espot this variegated series is only present east of the Pallaresa. West of Burch the griottes gradually disappear, the colours are still present approximately as far as the Pallaresa, but further west and north-west no distinction can be made between a lower and an upper part of the Devonian.

b. Stratigraphic aspects

The series, described above is certainly younger than the Silurian on which it rests. Few fossils are found in it, at least in the mapped area. In the lower part crinoid columnals occasionally occur and more rarely straight nautiloids. Near Salau a deformed simple coral was found.

De Lacvivier (1882) mentions the discovery of goniatites near Pont d'Ustou.

Autran and Guitard (1955) describe a "faune à Orthis, Bryozoaires, Polypiers", which they found near Pic des Aubières, in the river Fraychet and near the Col du Picou de la Mire. According to the authors this fauna is characteristic of the Caradocian. We cannot agree with this view because the "schistes jaunâtres troués et à nodules marneux ou gréseux" of the river Fraychet, in which these fossils were found, are situated between an anticline in the north, with a graptolitic Silurian core and in the south a succession of very characteristic Silurian between the Devonian and the Cambro-Ordovician. Detailed mapping has revealed two synclines and an anticline in this intermediate zone. In the core of the latter structure some Silurian outcrops on the right bank of the Alet a few metres above the bridge at 945 m, only 100 to 150 m south of the locality, where the fossiliferous layer crosses the Alet.

For other localities of this fossiliferous layer similar structural arguments refute the hypothesis that Ordovician occurs tectonically intercalated with the Devonian-Silurian series. The explanation would seem to be that the basal limestone is fossiliferous in places and the fauna bears a superficial resemblance to that of the Caradocian. The griotte marbles and variously-coloured slates can certainly be correlated with the dated strata of the same lithology in the country adjacent to the mapped area.

Although the comparatively great abundance of goniatites and the peculiar texture of the griotte marbles have made the latter familiar to the geologists of the Pyrenees at a very early date — in 1841 Françoix mentions the ocurrence of goniatites in similar rocks in Ariège —, it was not until 1898 that the age of the Upper Devonian of the Pyrenean griotte series, although often conjectured, was confirmed. In that year Barrois found the first determinable ammonoids, among which *Clymenia*, in the griotte marble of the ravine of the Coulari situated between Cierp and Cathervielle (Hte-Garonne). Since then rather abundant Upper Devonian faunas of this series have been described, among others by Dalloni (1930) and Schmidt (1931). According to Dalloni (op. cit.) griottes represent only the Upper Frasnian and Famennian.

The griotte texture, however, is not entirely restricted to the Upper Devonian; even in the older Devonian (Monseny: base of the Devonian) and in the Carboniferous (Vilarubla: on the basal cherts) griotte-like limestones have been found (cf. also Bresson 1903), nowhere, however, on such a large scale and so variegated as in the upper part of the Devonian and certainly not alternating with multi-coloured slates.

We therefore consider the griotte and many-coloured slate series as belonging to the Upper Devonian; the lower limit of the griotte series does not of course correspond, however, with the base of the Upper Devonian (see also Dalloni's table). Hence we prefer to speak in this connection of the upper part of the Devonian.

Less well-known is the series, occurring below the griottes. This series of limestones, calcareous slates and slates, is seldom fossiliferous, mostly of uniform appearance in the axial zone and has been placed by various authors in almost every section of the Palaeozoic (see e. g. fig. 2 giving the views of the authors who have worked in our area).

Its position between the Silurian pelites and the griotte marbles defines its age as the lower part of the Devonian.

Outside the axial zone the Devonian is faunally and partly also lithologically better characterized.

Casteras (1933, central and eastern Pyrenees), Keizer (1954 Arize), Zwart (1954, St. Barthélemy) gave divisions of the Devonian of the northern marginal zone, based on fossils and lithology, Dalloni (1930) of the southern marginal zone, on faunal grounds and Destombes (1953, Vallée de la Pique) of the northern border of the axial zone, partly founded on fossils (see table between p. 56 and 57).

It appears from the table that the Devonian in the central Pyrenees is present in two more or less different facies, passing into each other, viz. a non-differentiated monotonous Devonian, which consists of limestones, calcareous slates and slates, practically always unfossiliferous, occurring in the centre of the axial zone and a sequence in which a lithologic distinction can be made between Lower, Middle and Upper Devonian, each subdivision often being characterized by fossils and known in the northern as well as in the southern marginal zone.

The axial Devonian of the Valle de Arán shows towards the top a much sandier, deeper marine facies with well-developed graded bedding than in the present area. This facies is restricted to the area between Las Bordas and Montgarri (see Kleinsmiede 1960). This sandy facies wedges out towards the east in the Devonian of the upper Pallaresa and perhaps also in the Devonian north of Espot; strata, which are both reckoned by us to the lower part of the Devonian. The Aranese Devonian would then likewise belong to the lower part of the Devonian.

c. *The thickness of the Devonian*

In order to get a better insight into the stratigraphic development and the eventual tectonic deformation of the lower part of the northern Devonian the Salat-Alet area was mapped in great detail (see the 20.000 scale map of

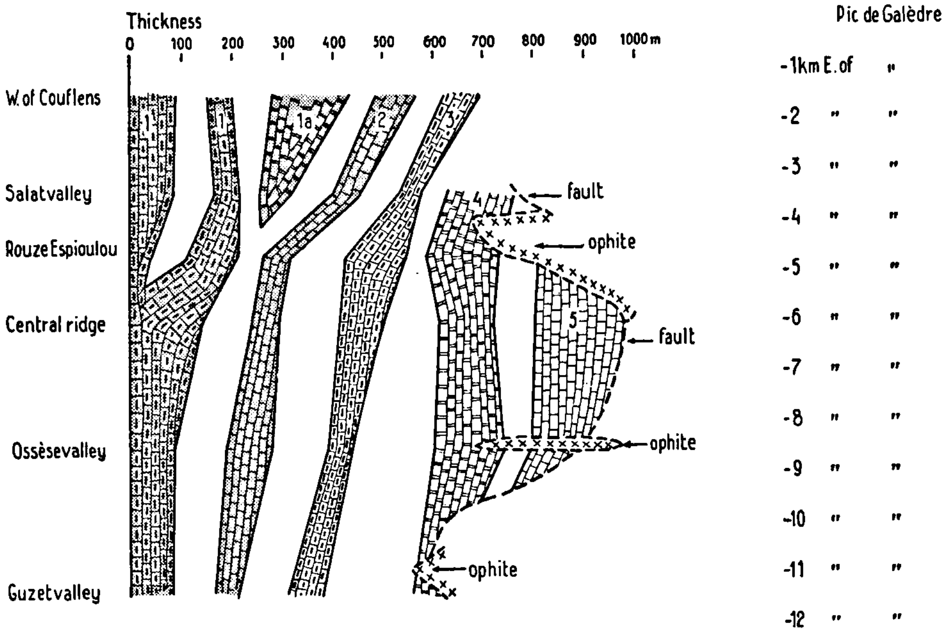


Fig. 16. Variations in thickness of the various Devonian limestone layers (1—5, see text) and intercalated slate units in the Salat-Alet area.

this area; also below p. 68). This mapping demonstrates that the Devonian south of the fault zone of Couflens-Portet contains 5 limestone horizons — further west even 6 —, alternating with slates. These horizons appear to be constantly present in the whole Salat-Alet area, although the thickness of the separate horizons varies (see fig. 16).

Thus the basal limestone thickens in a westerly direction and in the Salat area it becomes two layers through the intercalation of a slate layer. Another limestone (1a) appears here between limestone (1) and (2), which in its turn wedges out towards the east and the north. Consequently the Devonian in the west is considerably thicker than in the east, and the difference in thickness is mainly due to the thickening of the very lowest part of the Devonian. This basal Devonian also thins towards the north, while limestone (1a) also wedges out in this direction.

Taking these variations in thickness into consideration, it is clear that no generally applicable measure of thickness can be given. Moreover the outcropping Devonian becomes only gradually younger towards the north in spite of the many fold axes. In other words no single complete vertical section is available for measuring the thickness and we can only proceed by adding the thicknesses of the separate limestone and slate horizons obtained from areas which were originally situated far apart (10 to 15 km between limestone (1) and (5). In view of the differences in thickness in the direction from east to west varying from 160 to 485 m over a distance of 10 km for the Devonian older than limestone (2), an even greater variation can be expected perpendicular to the basin axis. Moreover, it is not certain that these rock units are also time units.

Neglecting these objections for the moment, we arrive at a minimum thickness of about 1000 metres for the outcropping lower part of the Devonian.

Along the whole breadth of the area mapped in detail the Devonian series is cut off by the Couflens fault zone with ophites, and thus the total thickness of the lower part of the Devonian still remains unknown. North of the fault zone, however, the lower part of the Devonian is complete, but unfortunately another objection presents itself viz. the low proportion of outcrops which makes it impossible to observe the presence, if any, of folds, with any degree of certainty. Without these folds the thickness would amount to some 1300 m, but it is very probable that at least one anticline and one syncline are present, through which the true thickness would be reduced to about 800 m. Hence it would follow — taking into account the thinning in northerly direction — that the Devonian older than the "griotte" would be approximately equivalent with the five known limestone and slate layers.

North of the Silurian of Couflens the basal Devonian limestone is practically absent and there are no indications that the Silurian-Devonian contact is not a normal one. Probably the basal limestone was not deposited so far north. The same tendency has been noted further west in the Vallée de la Pique.

It is impossible to give characteristics of each of the five limestones; their lateral variations are of the same order as the differences between the separate limestones. In fact even the lateral differences appear to be much greater than the vertical ones. From west to east the five limestones change more or less in the same sens, viz. in the east the limestones are somewhat more massive, whereas slate beds of a few mm thickness occur in the limestone in the west (see p. 71). The limestones (4) and (5) make the impression to be somewhat more massive than the lower ones. Possibly it is the influence of the marginal facies zone which makes itself felt here. Perhaps one could correlate the limestones (4) and (5) with the massive limestone of the Middle Devonian of the marginal facies (Arize and St. Barthélemy massifs). The compact blue spar limestone of the Col de la Trape possibly represents this marginal facies just penetrating into the axial zone facies.

The slates are rather uniform although towards the top of the series they become slightly greener.

Less is known concerning the Devonian of the upper Pallaresa and round the Riberot granite. Whether the increase of thickness in a western direction continues beyond the Salat valley could not be ascertained. In the north of the area the upper part of the Devonian is little seen. At least in

STRATIGRAPHIC TABLE OF THE DEVONIAN OF THE CENTRAL PYRENEES

	North-Pyrenean zone (Arize) (Keizer 1954)	Northern border of the axial zone (Hte-Garonne) (Destombes 1953)	Northern border of the axial zone (upper-Pallaresa-Salat-Alet) (Zandvliet)	Northern flank of the Tírvia-Espot syncline (Zandvliet)		South-Pyrenean zone (Central and Eastern Pyrenees) (Dalloni 1930)
Dévonian supérieur	— Calcaire compact à grain fin au sommet silicieux, à la base mieux stratifié et noduleux avec intercalations de schistes. *Oxyclymenia* sp. — Calcaire gris foncé, tacheté de rouge, (griotte). *Cheiloceras* sp. — Calcaire noduleux, bien lité, rouge (ou vert), les nodules sont séparés par des calcschistes. Fréquemment dans les nodules des goniatites ou corraliens (griotte); loc. des poudingues. Brachiopodes, Corraliens ou Spongiaires. (100 à 120 m) — Schistes et calcschistes, violets et verts, transition vers les griottes; banc de dolomie localement (près de Riverenert). (50 à 70 m)	f Calcaires griottes à faune de Clyménies du Famennien supérieur, du ravin de Coulari. (40 m) e Schistes et calcschistes clairs et versicolores, quelques lits calcaires et dolomitiques peu épais. (100 à 120 m)	Grey-white, green, red or light lilac marmorized griotte-limestones alternating with some green-grey to purple usually-calcareous slates. North of the Pic de Fonta max. 400 m.	Massive, somewhat griotte-like, dirty-white and yellow-white limestone beds of a thickness of 1—2 dm, alternating with grey-green to sea-green (sometimes with a purple shine) slate beds, 1—4 dm in thickness. (30—40 m) Towards the west (Espot) not distinguishable from the lower part. Towards the east (Burch) characteristic green and red colours and distinct griotte nodules.	upper part of the Dévonian	VI *Famennien:* griottes à Goniatites, Clyménies, Posidonomies, etc., supportant les lydiennes du Carboniférien. V *Frasnien:* Base des marbres griottes à *Géphyroceras retrorsum.* Calcaires et schistes noirs, à *Phacops cryptophthalmus, Gephyroceras retrorsum, Buchiola retrostriata, Rhynchonella cuboides, Spirifer Verneuili, Productella productoides.*
Dévonian moyen	— Dolomie à grain fin, compacte, gris, loc. à grain grossier. (100 à 150 m) — Calcaire massif bleu, à grain grossier, spathique; loc. dolomitisé et loc. fossilifère, transition vers le bas en calcschiste. Crinoïdes, Brachiopodes, *Strophonema?, Chonetes, Atrypa, Pentamerus* sp. (30 à 120 m)	d Schistes bleus en dalles, à trilobites de Cathervielle. (20 à 30 m) c Calcaires clairs à encrines, en bancs plus ou moins massifs. (30 à 35 m)	Massive to very schistose, dark grey-blue to light brown-grey limestone beds, 30—160 m thick which are interbedded with numerous thin dark slate layers west of the Salat river; in the Salat-Alet area there are 5—6 limestone beds present. These limestones alternate with: — ca equally thick, dark grey to blue-black usually non-calcareous slate beds, which upwards become lighter green-grey in colour and are arenaceous locally west of the Bonabé area. Limestone and slate beds are altogether some 1000 m thick in the Salat-Alet area.	— Rather dark paper slates and slates with a few thin grey limestone layers which increase in proportion downwards until the whole rock has been changed in a calcareous slate. (200 m) — Massive grey-blue limestone with a few calcareous slate beds of a thickness of several m. (65 m) — Grey paper slates with some thin limestone and calcareous slate beds. (30 m) Towards the west alternating limestone and slate beds, north of Espot more sandy slates and "barrégiennes" occur. Towards the east the slates are lighter in colour.	lower part of Dévonian	IV *Givétien:* Calcaire massif à polypiers. III. *Eifélien:* Calcaires colorés, calcschistes et schistes rouges et verts à *Phacops, Anarcestes, Agoniatites,* représentant le niveau à *Spirifer cultrijugatus.*
Dévonian inférieur	— Calcschiste, gris foncé, schistes avec intercalations de calcaire fossilifère. Près de Peybagué: grès rouge ferrugineux et très fossilifère, avec intercalations de calcaire rouge. *Rhynchonella,* Crinoïdes, *Spirifer* sp. Trilobites, Brachiopodes (*Atrypa*). (80 m)	b Calcschistes bleu mat et calcaires en petits bancs (à faune coblencienne en Htes-Pyrénées). (80 à 150 m) a Schistes noirs, subardoisiers luisants. (15 à 20 m)				II. *Coblentzien:* 2. Calcschistes, schistes micacés et calcaires de Gerri et de Bellver avec *Phacops occitanicus, Bronteus* et petits brachiopodes. 1. Schistes grauwackeux à Fenestelles et grands brachiopodes, de plus en plus réduits vers l'Est. I. *Gédinnien:* schistes ardoisiers ou en dalles de Llesp et de Monros à *Tentaculites* (horizon de Cerler en Aragon).
	Total thickness of the Devonian 360—540 m	Total thickness of the Devonian 285—395 m	Total thickness of the Devonian of the Salat-Alet area ca 1400 m. — towards the north and east thinner.	Total thickness of the Devonian 330 m. — towards the north-west thicker.		

the mapped area it is not terminated by the Carboniferous. Wissink (1956, p. 517) estimates a maximum thickness of 400 m for this Devonian round Pic de Fonta (north-west of Couflens). The thickness of the lower part would here amount to 500—800 m, which on average is less than we determined for equivalent beds north of St. Lizier and considerably less than for the Devonian south of this area (a minimum of 1000 m for the lower part).

Still further north in the Arize massif the Devonian as a whole is 360 to 540 m thick (Upper Devonian 150 to 190 m and Lower and Middle Devonian: 210 to 350 m, Keizer 1954). So from this it appears that there is a strong decrease of thickness of the Devonian in a northerly direction.

The southern Devonian, on the contrary, is appreciably thinner. On the crest between Pallaresa and Cardós the thickness amounts to some 300 m (upper part: 30 to 40 m and lower part: 250 to 300 m); near Espot about 350 m altogether. Further west the Maladeta granite cuts off the syncline, so that it can no longer be decided whether the thickness of the Devonian really increases in a north-westerly direction, as may be expected.

Summarizing it can be said that the Devonian of this section of the Pyrenees attains its maximal thickness in the axial zone, which rapidly decreases towards the marginal zones and that the facies gets a more differentiated character in the same direction.

E. THE CARBONIFEROUS

Notwithstanding the absence of fossils and of any characteristic Lower Carboniferous marker such as the chert horizon with phosphate nodules, the non-calcareous dark slates and micaceous sandstones, situated on top of the griotte marbles and variegated slates of the upper part of the Devonian are considered to belong to the Carboniferous.

In the mapped area they only outcrop in the core of the Tírvia-Espot syncline.

a. Lithologic description

The rocks consist for the most part of dark grey to black, often somewhat sandy, very fissile micaceous slates which occasionally show finely-rippled cleavage- or stratification-planes. Varieties poor in sand are especially very fissile (paper-slate). For the rest the Carboniferous east of the Pallaresa is comparatively little cleaved and the rocks mostly tend to split along the stratification.

Sandstone to greywacke bands 0.1 to 4 m thick occur, chiefly west of the Pallaresa. The sandstones vary from light yellow-brown to dark grey, while the greywackes are more dark greyish-green in colour. The mean grain size is very often less than 1 mm. The light-coloured sandstones, often with brown or black spots are very similar to some sandstones from the Stephanian of the Sort area.

All these rocks, but especially the more sandy varieties contain small (max. 2 mm) clastic muscovite flakes.

The Carboniferous is recognizable even from afar by the dark grey to black colour of the weathered rocks which clearly stand out against the much lighter colours (yellow-white to light brown) of the bordering Devonian. Moreover, the Carboniferous landscape has a rather rough appearance in contrast with the smoother forms peculiar to the Devonian. It is also significant that these formations are characterized by an entirely different vegetation, which very often strikingly delimits the Devonian-Carboniferous contact. The Devonian is on the whole rather bare, with dry pastures or with high, thinly planted trees, while the Carboniferous especially between the Pallaresa and the Cardós is covered with an almost impenetrable brushwood of *Quercus Ilex*. To the east this is replaced by a broom vegetation.

Owing to its uniformity this series can hardly be subdivided and only the Carboniferous of the Mauricio is rather varied. A section north-west of Espot of a series of more than 500 m thickness, showing practically no secondarily folding consists from top to bottom of:

— bluish-black slates with rather numerous light coloured sandstone to darker coloured greywacke beds which vary in thickness from 0.1 to 0.8 m .... ca 200 m
— bluish-black slates with numerous thin, grey sandstone layers and lenses of a few mm thick; towards the top these sand layers greatly increase in thickness .... ca 200 m
— grey to greenish-blue slates and paper slates, practically without sand layers, with locally a brownish-grey calcareous slate band of 20 to 30 m thick in the lower part .... ca 150 m

The Carboniferous east of the Pallaresa consists almost entirely of dark slates, which are less sandy than west of this river. Against the Devonian the slates are somewhat greener, they are also sometimes calcareous, but limestone beds are no longer found here. Sandstone beds are rare, but many thin, grey sandstone layers do occur in the slates, although not to the same degree as further west.

East of Farrera the slates are strongly desintegrated, earth-like, exceedingly black and often resemble the Silurian pelites.

East of the Pallaresa there occur also numerous quartz veins in which siderite, chlorite and talc are found, among other places very frequently along the new road from Tírvia to Glorieta de Montesclado and also 800 m south-south-east of Aydi and 1 km south-west of Farrera.

b. Stratigraphic aspects

The upper part of the Devonian, in particular west of the Cardós, passes rather gradually and fully conformably into the Carboniferous: from greenish-grey slates with limestone, which we consider to be Devonian, over grey to greenish-blue slates with only an occasional thin limestone to dark grey or black slates with some sandstone or greywacke layers: the sedimentation seems to have been continuous without a trace of a hiatus.

The complete absence in this Carboniferous of the chert horizon at its base, which is found almost everywhere in the Central and Eastern Pyrenees, in the Montagne Noire and in the Feixa-Vilarubla region south-east of Sort, at a distance less than 20 km from the Carboniferous of the present area is of particular interest. This typical horizon has been dated as Lower-Visean with perhaps the Upper Tournaisian included (Delépine 1929, 1937), indicating that the whole or at least the greater part of the Tournaisian had not been deposited in the Pyrenees, the same condition prevails in the Cantabric Mountains and in the Montagne Noire. Cherts have recently been recorded in the Upper Devonian of the Mouthoumet-massif (Durand Delga & Lardeux, 1958) and in the Eastern Pyrenees (Prades: Cavet, 1957 in Durand Delga et al), from which it appears that locally at least there need not be a hiatus in the base of the Carboniferous.

In general the chert horizon is followed by a thick series of mostly micaceous sandstones, conglomerates and slates in which there occasionally occur limestones of Visean age. In the basal series these often resemble the griotte marble of the Upper Devonian. This Culm series is in places characterized by a Dinantian (Upper Visean) fauna, for instance near Mondette (Delépine 1931), Larbont and St. Antoine (Lartet 1884, Roussel 1885) in the Arize massif. This series closely resembles the Carboniferous of the Tírvia-Espot syncline, except for the absence of conglomerates in the latter.

This syncline most probably continues towards the west in the syncline of the Plá dels Estañys north of the Maladeta massif, carrying a practically identical Carboniferous sequence, although somewhat more sandy and coarse-grained and showing some conglomerate bands. According to Kleinsmiede (1960) the shale content increases towards the east. This Carboniferous from Plá dels Estañys is characterized by frequent plant remains: the "grauwacke à Calamites". Several determinations have been made: Zeiller (1886) does not argue for a definite age, Faura y Sans (1928) decides on a Dinantian (Tournaisian and Visean)-Westphalian age, based on a very limited flora. Dalloni (1910), however, correlates this Carboniferous on the ground of identical facies with the Middle Westphalian of Aragón (río Escarre) which is determined by a better flora.

Summarizing the evidence, we can say that the Tírvia-Espot Carboniferous is certainly older than the Hercynian folding and therefore older than the Upper Westphalian of Aguiró (Roussel 1904, Dalloni 1910) which unconformably covers the Hercynian folds of the border of the axial zone.

CHAPTER III

STRUCTURE

Introduction

The general structure on sheet 5 is relatively simple. It consists in the first place of a large Cambro-Ordovician arched dome, plunging and narrowing towards the west and occupying the central and eastern part of the map area. The northern border of this dome, containing three granite plutons and a thick reef limestone, dips steeply down in a flexure. North of this flexure there occurs a set of isoclinal folds in the Devonian, containing consecutively younger rocks in a northern direction. A partly double branch of the North-Pyrenean fault zone, accompanied by ophite intrusions cuts with an E—W trend into these Devonian folds and caused a tilting towards the north on both sides of the fault.

In the north-west we find the easterly plunging Cambro-Ordovician Maubermé anticline disappearing beneath the Devonian. Near the northern limit of the map the axial zone is truncated by a major branch of the North-Pyrenean fault and by a Lower Mesozoic syncline both trending WNW.

The southern boundary of the main dome is formed by the north dipping, steep, isoclinal syncline of Tírvia-Espot filled with Devonian and Carboniferous rocks. The Tór syncline in the eastern portion of the dome, and north of the Tírvia syncline is less deep but also isoclinal.

The trend is rather uniformly east—west to WNW—ESE. The intrusive granite batholiths interrupt this consistent strike somewhat, the strikes curving clearly round these massifs as if they have been pushed aside.

These structures are of Hercynian origin, except for those formed by the North-Pyrenean fault movements. These main structures will be treated successively in the following pages and afterwards some particular structural details and later transformations will be described. At the end an attempt will be made to reconstruct the structural phases in their chronological order and in their mutual relation.

N.B. The position of sections I to XXVIII is indicated in fig. 1.

A. MAIN STRUCTURES

a. The main dome

The Cambro-Ordovician rocks in the arched uplift of the upper Pallaresa-Cardós region consist mainly of a monotonous sequence of pelitic and psammitic sediments with frequent lateral and vertical alternations in which key beds or characteristic units are lacking. Moreover, the intense cleavage and the frequency of micro-folding (half-wavelength usually less than 10 cm) would greatly impede the tracing of a distinct horizon. For the same reasons it is often impossible to find out whether a wedging out of a lithologic unit has a structural or sedimentary origin. Thus the unravelling of the minor-

Fig. 17. N—S cross-sections through sheet 5. For location see fig. 1.

and intermediate-scale folding of this Lower Palaeozoic dome, which is certainly more complex than a simple anticlinorium, is particularly difficult. The folds have a wide range of wavelength and amplitude. Only some of the larger ones can at once be recognized as such, because rocks of a different lithologic nature from the base of this Lower Palaeozoic (the Lleret-Bayau series) or from the top-most and higher strata (metalliferous limestone, Silurian black shales, Devonian limestones) outcrop in the core of the structure in question.

For instance the Tór syncline, characterized by Silurian and Devonian rocks (sections XXVII and XXVIII, fig. 26) may be compared with the Lladorre anticline and the anticline north of La Forsa on the Vallferrera, both with the Lleret-Bayau series (sections II and III, fig. 17). The latter structure is also marked by a quartzite layer which makes a closure.

Other evidence points to a comparatively gentle secondary folding which is far from isoclinal, for in some places sub-horizontal stratification occurs over areas of several sq km. The metalliferous limestone of the Cuenca-Moredo region for instance, slopes rather uniformly at 20° or 30° towards the west. Also to the east in the area of the Estats, where the cleavage is less intense, indications can be found in the field of rather flat, gently undulating layers. This tendency is also very clearly present in the metamorphic Aston massif, which forms the eastern continuation of the upper Pallaresa-Cardós dome (De Sitter & Zwart 1959, p. 6 and 11).

Finally the area round Mount Monseny in the south-east of the map area, shows a clearly folded sequence (Cambro-Ordovician, Silurian and Devonian), in which the limits of the Silurian black shales lie rather horizontal, although the lower limit of the black shales on the western slope is quite strongly undulating (section I, fig. 17).

These flat structures often seem to exist only when viewed as a whole especially when observed from some distance. Looked at more closely the layers are strongly undulating, thus thickening the sequence structurally. The stratification always remains steep. The limestone of the Cuenca-Moredo is cut off towards the north by a vertical fault with dextral movement: the Moredo fault. The limestone then continues towards the north in a series of rather sharp folds, plunging somewhat towards the west. On an average the limestone remains at the same altitude in a northern direction. The sharp structures are consequently undulations in a sub-horizontal plane. In a western direction they gradually form a westward-plunging synclinorium of which the southern flank is on the east end cut out by the longitudinal Moredo-fault (sections X to XIII, fig. 18).

The quartzitic zone south of Lago Lagola wedges out laterally in the slopes of the Campirme and the Pilás in a number of tongues. In the case of some of these tongues it has been ascertained that they form the core of small anticlines. A peculiar anticlinal structure, cut by late-tectonic vertical cross-faults of small throw outcrops west of Pic de Cerda in the Gérac valley (section XXII, fig. 20). This somewhat coffer-shaped Gérac anticline can be recognized by a comparatively thin non-stratified dolomite layer, part of the metalliferous limestone. The contact with the slates at the base of the dolomite occasionally stands out clearly in the flat closure and this is characterized by some small, pointed anticlines (the synclines are never seen to outcrop curiously enough). The presence of these small structures causes the contact dolomite-slate to be everywhere rather steep.

Hence we must picture this Lower Palaeozoic arching as a multiple

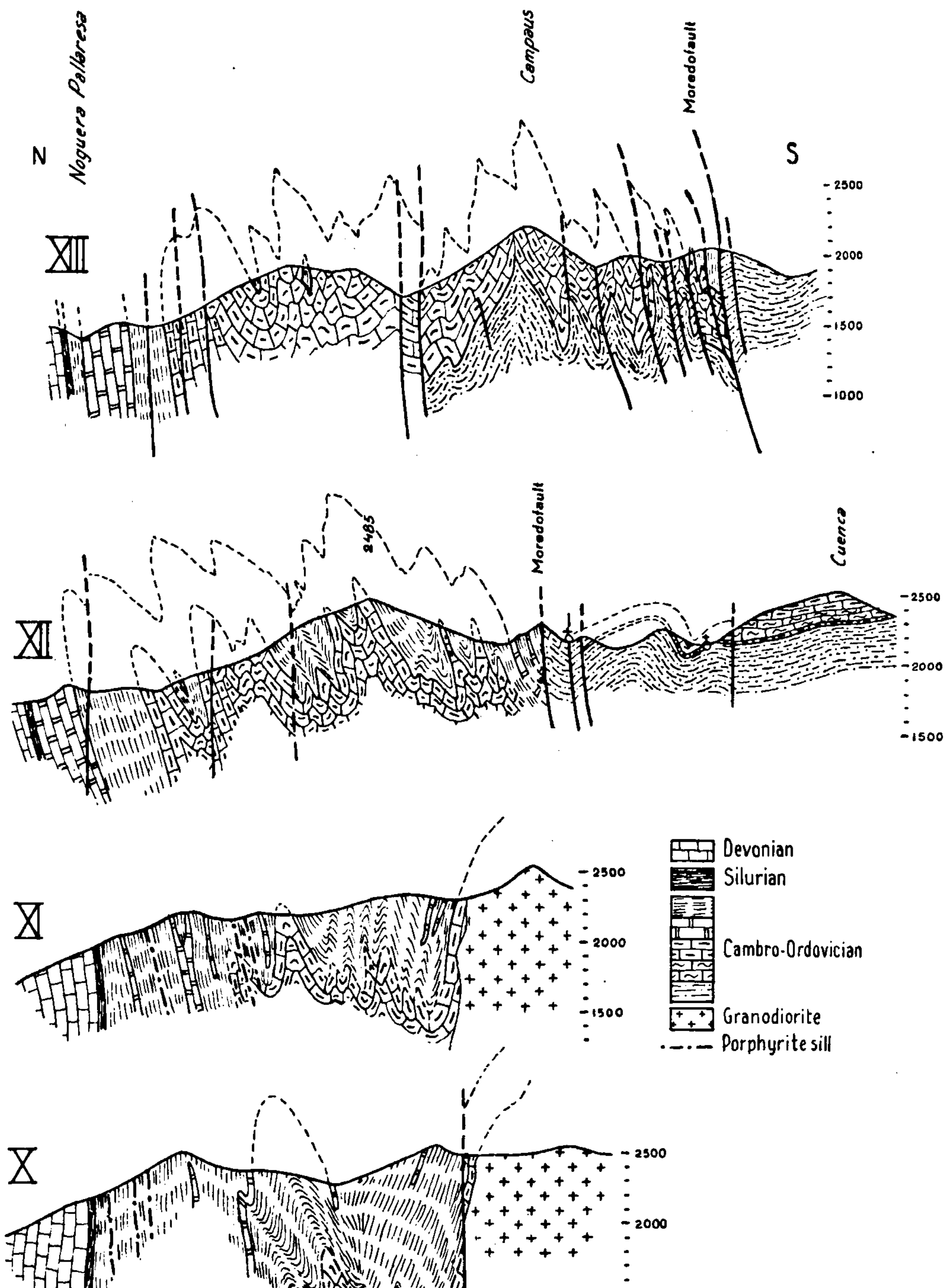


Fig. 18. N—S cross-sections through the northern border-zone of the main dome, N and NE of the Marimaña granite.

composite anticlinorium, since a structure of a higher order may always consist of a number of structures of a lower order. The folds, observed in this Lower Palaeozoic can be divided roughly into size orders, and since the folds, observed in this Lower Palaeozoic are roughly symmetrical the half-wavelength has been taken as representative of the wavelengths although a precise relationship is not implied. These size orders are:

1. structures with a half-wavelength of the order of
   n. 10^4m (central dome, Devonian synclinorium of the Valle de Arán)
2. structures with a half-wavelength of the order of
   n. 10^3m (Tór syncline, Bonabé syncline, Bonaigua anticline. Lleret-Bayau anticline)
3. structures with a half-wavelength of the order of
   n. 10^2m (Gérac anticline, Campaus structures, structures in the quartzite of Lago Lagola)
4. structures with a half-wavelength of the order of
   n. — n. 10 m (observed in numerous places)
5. structures with a half-wavelength of the order of
   n. 10^{-1} — 10^{-2} m (microfolding)

The lower orders 4 and 5 and to a lesser extent also 3 form a contrast with the higher orders since they are directly controlled by the nature of the sedimentary sequence, principally the thickness of the competent layers, if it is small ("schistes rubanés") then the lower limit of the folding is formed by the micro-folding; if it is greater, then the lower limit of the folding lies at a higher level (in the case of the "calcaire métallifère" for instance in the order of n. 10^2 m). The deformation which resulted from the variable rock-facies in the Cambro-Ordovician is not everywhere the same, neither in a lateral nor in a vertical sense. It is obvious that such a complicated state of folding cannot possibly be achieved by pure concentric shear, especially in the case of the lower structure orders and thus planar (cleavage) shear provides the only possibility of movement in the stress field. The effect of a similar composite type of folding is that the lower orders of folding 5, 4 and 3 account for the bulk of the deformation while the structures of higher order (1 and 2) remain rather gently undulating, little compressed, with proportionally little amplitude. This idea is expressed in the sections I, II and III of fig. 17.

The northern border of the main dome has a slightly different character in the following points:

1. The average folding is a little more compressed than elsewhere in the dome. The metalliferous limestone, outcropping practically exclusively in this border zone, is on the whole strongly isoclinal with individual structures of the order of n. 10^2 m. Folds of a lower order are only sporadically found (sections X to XXIII, fig. 18—20). The character of the folding of this border zone resembles the likewise isoclinal and isodimensional Devonian folding (sections IV to X, fig. 22) more than the internal folding in the dome outside the border zone as described above.
2. The Lower-Palaeozoic dips steeply under the Upper Palaeozoic and the latter, owing to successive folds shows steadily younger beds towards the north, from which can be deduced, that the regional dip of the Upper Palaeozoic rocks is only 15° or 20° towards the north. The same effect

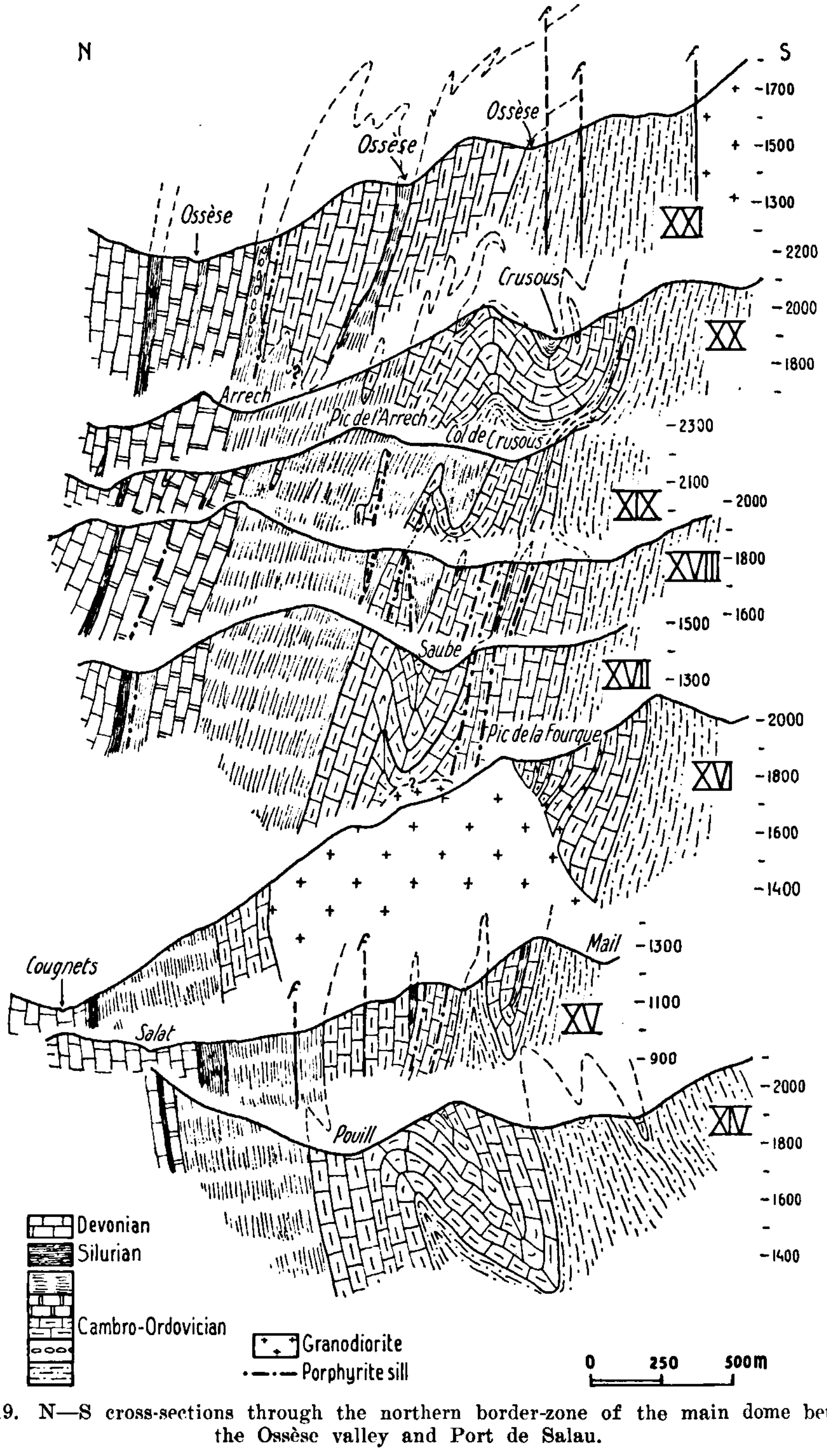


Fig. 19. N—S cross-sections through the northern border-zone of the main dome between the Ossèse valley and Port de Salau.

is also seen to some extent in the Lower Palaeozoic, in the "calcaire métallifère" north of the Marimaña granite and near Port de Salau. Between the two zones with a northward regional dip there is a comparatively narrow zone, in which no repetition occurs, this being a long steep flank. The incompetent Silurian usually forms part of this flank but in some places it is lacking, as a result of structural complications. In the Ossèse valley it has locally even become a fault zone with fault breccias and mineralisations at Carbauère. This steep flank has the character of a flexure and the local absence of the Silurian points to differential movements.

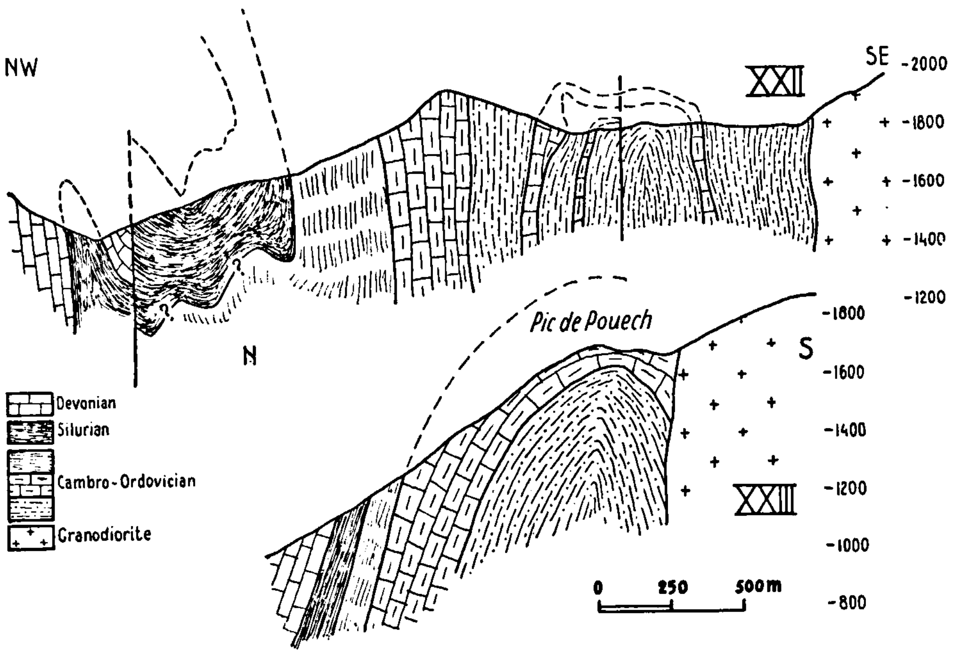


Fig. 20. N—S cross-sections through the northern border zone of the main dome east of the Escorce valley.

3. The northern border of the dome is consolidated by three granites (from west to east the Marimaña, Salau and Auzat-Bassiès granites), which lie practically in the same trend and give the impression of being the outcropping parts of a large granite ridge. The flexure border and granite ridge may be genetically connected. The flexure border is older than the fracture faults in the granite, for the flexure border is moved in the Escorce area by these faults. There are, however, indications that the flexure continued to form during or after this late fault movement. The situation on either side of the fracture faults is not the same (see the geological map).

From the sections it is concluded that the net throw of the flexure zone amounts to some 2½ km (Ossèse valley). This large throw must be present along the whole northern border, as repeated high-amplitude folding cores of Silurian are seen neither in the Cambro-Ordovician synclines to the south or Devonian anticlines to the north. Exceptions occur below the Col d'Escots, where Silurian cores in the Devonian folds clearly indicate the lower side of the flexure and in some other places as a result of the very elongated isoclinal or even diapiric Devonian folding. East of the Col the Silurian is thicker and less tectonized than west of the Col. The character of the flexure of the border zone remains clear however.

A similar steep border zone is also present south of Capdella on Flamisell just outside the sheet 5 area, likewise on the border of Infra- and Supra-Palaeozoic. The Silurian, practically horizontal in the Monseny area, suddenly plunges away steeply towards the south under the recumbently-folded Devonian of the Flamisell.

The contrast between the strongly folded metalliferous limestone north of the Moredo fault and the practically unfolded equivalent south of it can perhaps be explained from the presence of the Marimaña granite which prevented strong compression of the covering above this granite ridge. The longitudinal fault of the Moredo and the group of faults around Campans, of which only the former has a large throw (sections XII and XIII, fig. 18), are all marked by ½—20 m thick sterile quartz layers as far as they run in the limestone, while outside the limestone the amount of quartz on the fault plane decreases rapidly and the faults can no longer be recognized as such. The fault zones are not tectonized and slickensides were nowhere observed. They are probably tension faults, originating from the continued upward pressure of the granite batholith. They must have originated after the folds of the Campans. The movement must have been chiefly a vertical one.

The Marimaña granite is for the greater part surrounded by the metalliferous limestone. On the north-west border the fault of the Río Malo cuts out a part of the limestone. The limestone border is otherwise almost complete except for the south-eastern corner where the granite breaks through the limestone (see p. 105). The important fault of Estany Pudó cuts through the granite and is consequently a late phenomenon. A less important branch here borders the granite over a short distance. The main fault presents a net throw of ca 600 m (section XXIV, fig. 21), causing a repetition, which is the same in detail in both fault blocks. The northern block, which is surrounded on three sides by the granite, may be considered as a roof pendant, as can the smaller limestone block somewhat further west.

The Estany Pudó drains along the fault contact in the metalliferous limestone. A smaller sink-hole occurs some 750 m to the south-west. On the same side of the Puerto de la Bonaigua there are no springs in the limestone, so that the author is of the opinion that the drainage areas above the sink-holes are drained to the Garonne, where some rather big springs just above the river drain the limestone.

Below the metalliferous limestone of Estany Pudó (south of the Marimaña granite) lies a coarse, discontinuous conglomerate. On the south side of the upper Bonaigua valley a conglomerate identical to the former is outcropping, and is separated from it by a zone of parallel limestones and sandy slates. So the conglomerates are not directly connected with each other, but they must represent the same conglomerate. Although in the field no data concerning a structural cause for this repetition were found, the author attributes the doubling of the conglomerate to an anticlinal structure, of which the axis exactly cuts the Col. A syncline is not probable in connection with the proximity of the large Tírvia-Espot syncline, a remnant of Silurian is still present in the Refugio-Baños fault zone on the crest south of Puerto de la Bonaigua. Moreover, the most-southerly fold axis in the metalliferous limestone mass of Estany Pudó is a syncline. The hypothesis is illustrated in sections XXIV and XXV of fig. 21.

The Refugio-Baños fault borders the Basiero granite, part of the large Maladeta granite to the north. The contact-metamorphic zone is here very narrow or absent. Probably this very late tectonic fault for the greater part cuts out this contact zone. It is in any case remarkable that on the crest south of the Puerto de la Bonaigua the rock is practically unmetamorphosed up to the fault, but after it consists purely of hornfelses and spotted slates.

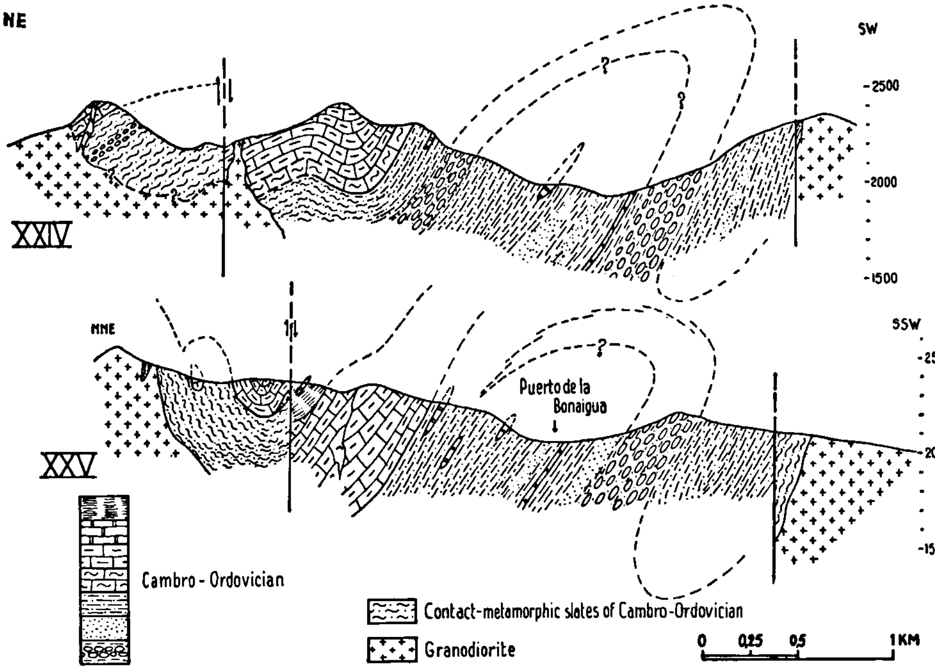


Fig. 21. Cross-sections near the Bonaigua pass between the Marimaña granite batholith in the north and the Basiero granite batholith in the south.

Towards the south-east the fault terminates in the northern flank of the Tírvia-Espot syncline, while towards the west the fault continues part of the way into the Maladeta granite as a broad mylonite zone (see fig. 1).

All the granites show to a greater or lesser extent a shouldering-away effect (see the geological map). Besides this shouldering-away the granites have occasionally caused some extra folding, a folding which is very irregular and in which no definite directions can be recognized. This secondary folding is usually found only up to some tens of metres from the contact (e. g. Teso crest; north of Monseny, section I, fig. 17; and on the western border of the Auzat-Bassiès granite). Also the two small synclines with north-south axes west of Pic de Argulls 2656 m (Marimaña granite) have been brought into their present position by granite movements.

b. The folding of the Devonian in the northern zone

The folding of the Devonian in the northern zone is quite well known. This sequence lends itself admirably to a mapping of a sequence of calcareous and non-calcareous sediments. The Devonian of the Salat-Alet area south of the fault zone of Couflens-Arigail was mapped in detail, in order to comprehend the folding characteristics of this formation.

Owing to the comparatively high degree of exposure, the limestone layers can be easily traced even in the woods. The vegetation often gives valuable indications for the mapping, especially in the more cultivated region in the north: hay-fields here lie almost exclusively on the slates, although many are now in a neglected state and covered with thick brushwood of bracken; the limestones, on the contrary, are usually covered with beech-woods.

Indispensable for this mapping were the first class aerial photographs of the Institut Géographique National of the Ministère des Travaux Publics et des Transports, which were closely analysed. The Devonian area in question is covered by photos 92 to 102, 160 to 173 and 292 to 308 of the run Aulus-Vicdessos-Ax.

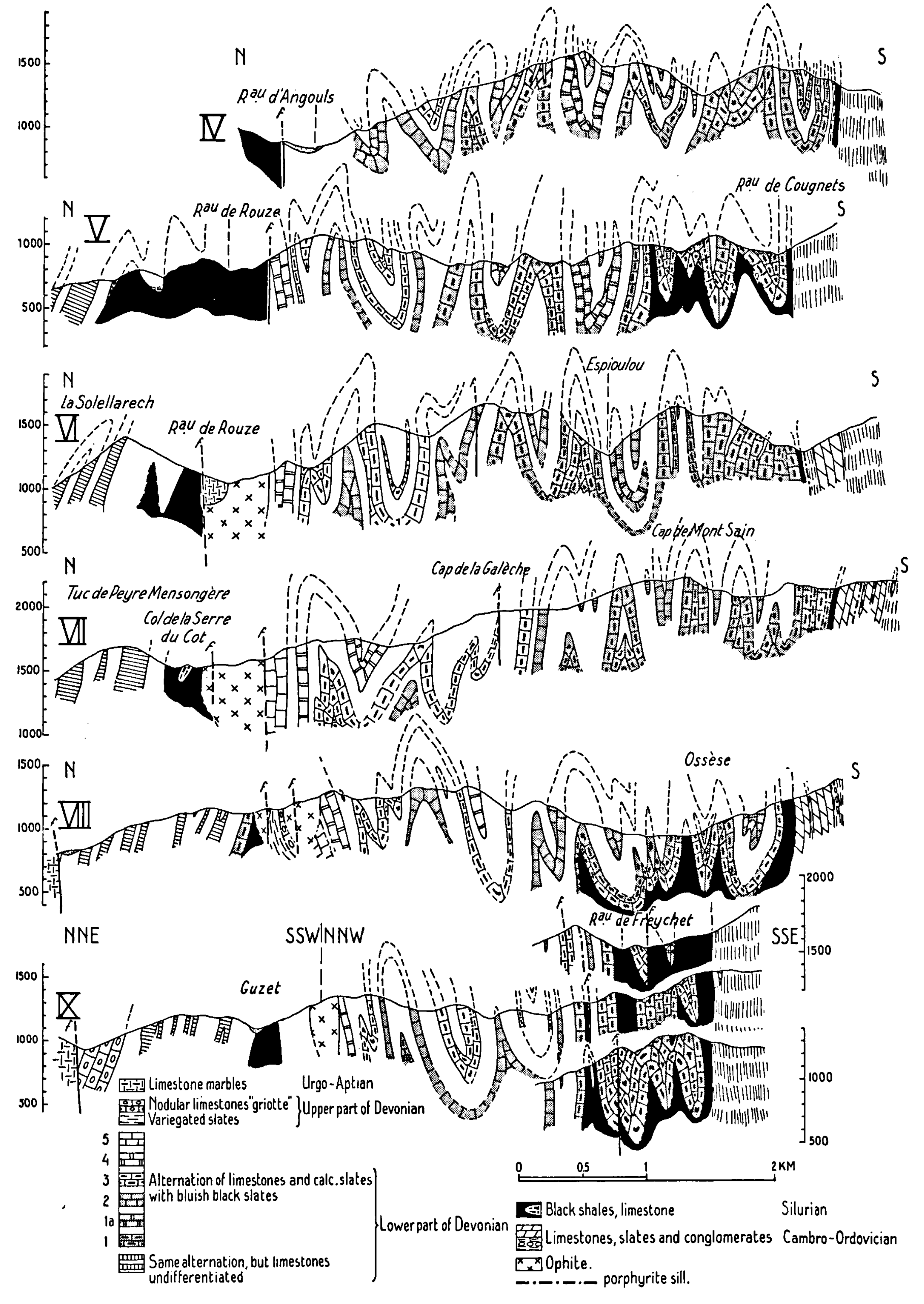
With the help of these data the 1:20.000 map was drawn and the sections constructed and from these two the stratigraphy was deduced. This method of deducing the stratigraphic sequence from the tectonic map was used in this case because the lithologic differences between the limestone bands are too small and at least as variable in a lateral sense as in a vertical sense.

One difficulty lies in the peculiar character of many closures. These can rarely be observed from close by. Owing to the coarse development of the cleavage in them, such a closure often looks like a stratified limestone band, which wedges out towards the top or towards the bottom. Seen from a long distance these "wedging-out" limestones are recognizable as closures probably because the cleavage is not longer conspicuous, and the stratification then becomes clear.

The Devonian south of the fault zone of Couflens-Arigail appears to consist of some five limestone layers alternating with slates (see p. 50 ff). This sequence of limestones and slates is very intensively folded, practically without faults. Between the above-mentioned fault zone and the steep border of the Lower Palaeozoic dome some 8 or 9 compressed and steep anticlines and synclines are present over a distance of only 4½ km (fig. 22). Moreover, the folding is such that steadily younger Devonian beds outcrop in the cores of the synclines in the direction of the fault zone. From this younging it can be deduced that the regional dip of the base of the Devonian (and with that the top of the Cambro-Ordovician) plunges some 15 to 20° towards the north.

The amplitude of the folds is at least equal to the fold length. Usually, however, the amplitudo is much larger, especially in the south against the steep border of the Lower Palaeozoic dome. There the folding is strongly isoclinal and the layers are practically vertical. The shortening must be very great, but cannot easily be calculated exactly owing to the strong stretching of the limestones and the slates. A layer in the hinges is, on the whole, 1½ to 4 times as thick as in the flanks . Supposing that thinning occurs to the same extent in the flanks as thickening in the hinges, then the flattening

Fig. 22. N—S cross-sections through the northern Devonian zone of the Salat-Alet region



N
S
Rau d'Angouls
IV
N
V
Rau de Rouze
Rau de Cougnets
S
N
La Solellarech
VI
Rau de Rouze
Espioulou
S
Cap de Mont Sain
N
Tuc de Peyre Mensongère
Col de la Serre du Cot
Cap de la Galèche
S
VII
N
VIII
Ossèse
S
NNE
SSW|NNW
Rau de Freychet
SSE
Guzet
IX
Limestone marbles
Urgo-Aptian
Nodular limestones "griotte"
Variegated slates
Upper part of Devonian
5
4
3
2
1a
1
Alternation of limestones and calc. slates with bluish black slates
Lower part of Devonian
Same alternation, but limestones undifferentiated
0
0.5
1
2 KM
Black shales, limestone
Silurian
Limestones, slates and conglomerates
Cambro-Ordovician
Ophite.
porphyrite sill.

as well as the stretching come to a factor $\sqrt{1\frac{1}{2}}$ to $\sqrt{4}$, that is to say a flattening of 18 to 50 % and a stretching of 22 to 100 %. Taking this into account there is a real shortening of some 40 or 50 % for the Devonian in the immediate neighbourhood of the flexure border in the south. Towards the north the deformation decreases, and the folds are less isoclinal and less steep. This tendency is continued also north of the fault zone of Couflens-Arigail: the Carboniferous of the Vallée d'Estours lies in rather flat and broad folds (Wissink 1956).

The folding is certainly not of the concentric type (there are clear differences in thickness in hinge and flank) but neither does not represent pure cleavage folding, although cleavage is very well developed in the slate component.

In the limestones themselves the cleavage is poorly-developed except in the hinges, where at least in the isoclinal folds a coarse cleavage occurs. The compact limestones in the flanks do not give the impression of being tectonically cleaved. In places where thin slate layers occur in the limestone the limestone appears to be always stratified parallel with the contact.

Cleavage is as said very finely and regularly developed in the slates, hence they show no more traces of stratification. The cleavage can clearly be recognized as such at the contact with limestone in the slight discordance between limestone surface and slaty cleavage. In a wide sense, however, the orientations of cleavage (in the slate) and of the stratification (in the limestone) hardly differ from each other. Fig. 23 gives two Schmidt contour diagrams, one with about 100 cleavage planes (slate) and one with about 100 stratification planes (limestone) from the surroundings of Pic de la Tèse, plotted on the upper hemisphere. At the same time some eight stratification planes found from the outcrop pattern were set in the diagram with the limestone measurements. About 50 % of the slate measurements represent dips towards the north, and of the limestone measurements about 40 %. In brief it appears:

1. that the strike of the cleavage in the slates is oriented on an average some 5 or 10° more east-south-east than the dominantly E—W bedding strikes of the limestones
2. that the dip of the bedding of the limestones shows a maximum about the vertical and that the dip of the cleavage in the slates has a maximum between 80 and 90° N.
3. that the distribution area of cleavage and of bedding orientations is about equally large.

The isoclinal character of the folding also appears from this.

c. Stretching and thinning of the Devonian rocks

On p. 68 we draw attention to the thickening of the Devonian limestones in the hinges of the isoclinal folds as compared to the flanks, a relation of 4:1 or a flattening of 50 %. Since cleavage is badly developed in the hinges of the limestone we assume that the stretching and flattening is at least partly due to intergranular slip and perhaps also to some solution and recrystallization. In fact the limestones are often finely crystalline and the calcite crystals show a clear twinning.

The stretching of the flanks is also proven by the stretching of thin slate layers which become broken up in small fragments parallel to the bedding. The surface area becomes greater and if the volume remained constant the thickness decreased accordingly.

In order to get some idea about the stretching exemplified by this broken-up slate a set of exposures along the Salat river near Salau was analysed. Thin limestones alternate here with very thin broken-up slates. The slate fragments are approximately square, regularly distributed on the limestone surface and have drifted away from one another in about equal distances vertically and horizontally as has been drawn schematically in fig. 24. The space between the slate fragments was filled up by recrystallized calcite, which weathered away again in the outcrop surface.

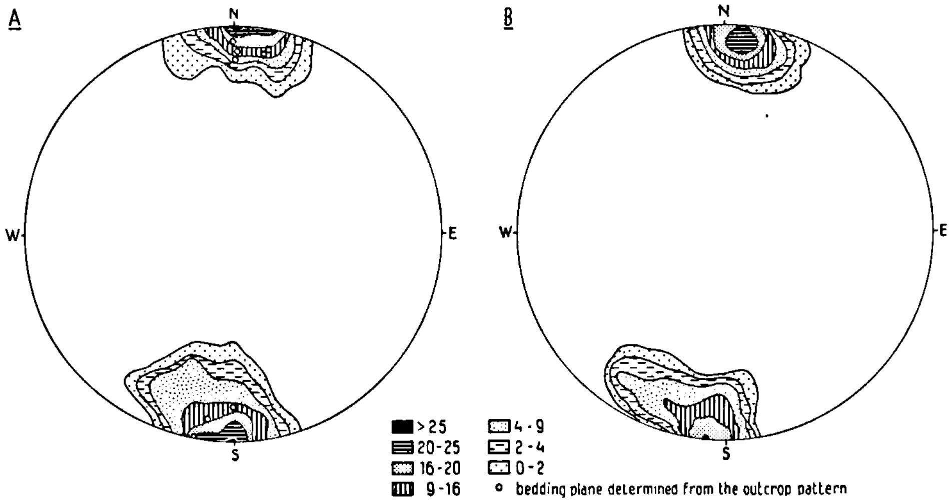


Fig. 23. Contour diagrams of bedding and cleavage plane poles from a small Devonian area round Pic de la Tèse, Salat-Alet region;
A. 98 limestone bedding-plane poles;
B. 104 slate cleavage-plane poles.

The measured data are as follows:

thickness of the limestone layers (t): 6—10 mm, average: 8 mm
average thickness of the slate layers (t′): 1 mm
average distance between the slate fragments in both directions (d): 1 mm
area of the practically square fragments (p^2) amounts to about 50 mm^2, this is at the same time the area parallel to the bedding of the limestone before the deformation
area of the limestone after the deformation $(q^2) = (p + d)^2 = (\sqrt{50} + 1)^2$.

When we consider the volume of the limestone as invariable then the thickness of the limestone layers before deformation was as an average:

$$\frac{\text{limestone volume}}{\text{surface of slate fragments}} = \frac{(t + t') \cdot q^2 - p^2 \cdot t'}{p^2} =$$
$$= \frac{(8 + 1) \cdot (\sqrt{50} + 1)^2 - 50 \cdot 1}{50} = 10.7 \text{ mm}$$

The thinning of the limestone layers by compression is then 10.7 — 8 mm = 2.7 mm, or $\frac{2.7}{10.7} . 100$ = ca 25 %. The reduction in thickness of the whole complex of slate and limestone is $\frac{2.7}{10.7 + 1.0} . 100$ = ca 23 %.

Comparison of these thinning values with those of 18 to 50 %, calculated from the ratios of the thicknesses of limestone in flank and hinge gives good agreement, especially since the 23 % of thinning found, is a minimum value, to which must be added the thinning, which had already taken place

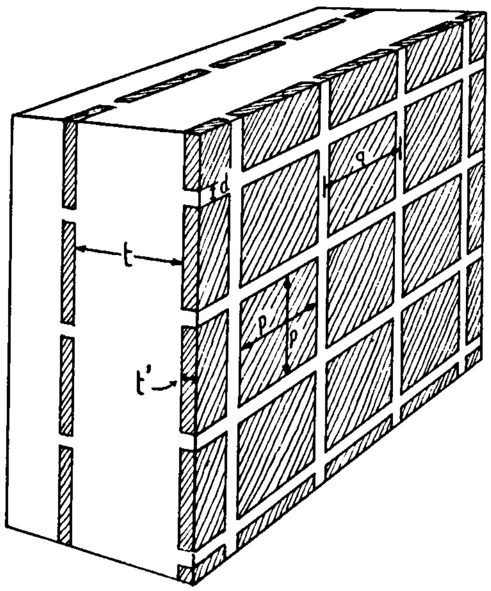


Fig. 24. Schematic representation of broken-up slate layers (shaded) enclosed in limestone (white).

before the slate layers broke. The slate fragments always show an intense slickensiding; in the same bedding plane it is parallel for all fragments, though sometimes a little undulating. In a subsequent slate layer the slickensides often have a somewhat different orientation. The slate layers are mostly broken up in such a way that one of the sides is parallel with the slickensides, the other side lying perpendicular to it. In our opinion this is an indication that the slickensiding and the fracturing are due to the same cause, viz stretching owing to thinning of the limestone layers. In the beginning this only led to slickensiding; at a further compression the slate layers could no longer follow the stretching and broke up. In keeping with this is the fact that the fracturing is especially common in the steep isoclinal structures against the northern border of the Lower Palaeozoic dome where the deforming stress perpendicular to the bedding was greatest. Slate layers of 1 cm thick or more are rarely broken. In these thicker layers the stretching can probably be taken up by shear along the cleavage planes, which are well-developed in these slate layers. The cleavage in the slightly thicker slate layers may result from the extension of the adjacent limestone beds. Thin slate layers are not or hardly cleaved.

Stretching not only took place in the direction of the a-axis (leading to an enlargement of amplitude), but also parallel to the fold axis (b-axis), (leading to steeper plunges).

Stretching in the a-axis may also be responsible for the origin of the characteristic isoclinal anticlines with pinched noses where the two limestone limbs directly lie on each other over may tens, or even hundreds, of metres, without slates in between. The stretching in the direction of the a-axis

Fig. 25. Folded siliceous cleavage fillings in marble (dark bands) from adjacent competent arenaceous layers.

amounts in the case described above to:

$$\frac{q-p}{p} \cdot 100 = \frac{1}{\sqrt{50}} \cdot 100 \text{ or about } 14\,\%.$$

Part of the great shortening in the south must consequently be ascribed to this phenomenon.

Stretching also took place in the recumbent folds of the Devonian of Monseny. This Devonian consists of rather pure limestones alternating with thin, very silicious and indurated layers. The latter are also broken and often drawn apart, but in a very irregular way. Moreover, these hard layers are often very irregularly folded, independently of the somewhat thicker hard layers. These parasitic folds give the impression of possessing parallel axial planes. Here, too, folded cleavage as a result of late drag was found (fig. 25). The cleavage is indicated by very thin layers of the material of which the hard layers consist. As the limestone layer is thicker, these cleavage fillings are thinner and especially these thinnest layers are often folded. The fold axes of these parasitic folds may even be parallel to the bedding. The angle between cleavage layer and bedding plane is always smaller than the same angle in the case of unfolded cleavage fillings. Thinning also probably played a part in it.

d. The Tór and Tírvia-Espot synclines and the recumbent folding of Monseny

In the mapped area the two synclines are uncomplicated, deep-reaching, isoclinal structures. The Tór syncline with Silurian and Devonian sediments outcrops for the greater part outside the present area in Andorra (see fig. 1). Sections III fig. 17, XXVII and XXVIII fig. 26 give a picture of this syncline.

More interesting is the Tírvia-Espot syncline (sections I, II, fig. 17, and XXVI fig. 27); the structure is characterized over the whole length of nearly 50 km by Carboniferous slates and sandstones outcropping in the core. The axial plane of the syncline dips 40 to 60° towards the north to north-east. In the west the syncline is cut off by the Maladeta granite and the Refugio-Baños fault; some 15 km further west the syncline reappears against the northern border of the granite in question. The Carboniferous of the Plá dels Estañys here occupies the core of the fold. Towards the east the syncline continues as far as the Lles-Aristot granite on the border of Andorra and Spain. Where the rivers Pallaresa and Cardós cross it the syncline looks a simple isoclinal structure, without any secondary folding and with only partly-developed cleavage. Further east as well as towards the west there is a gradual increase in the occurrence of secondary folding. Already in the neighbourhood of the Col de Mánega the Devonian limestones on the two flanks show numerous folds of the order of n. 10 m. On the map this appears in a substantial widening of the Devonian outcrop. The unvarying lithology of the Carboniferous sediments causes these secondary folds to be little conspicuous. Still further east the syncline becomes rather complicated owing to the increasing importance of the secondary folding (see fig. 1). This is also the case in the west around the crest of Teso. Here, however, it is an irregular folding owing to the shouldering away of the Basiero granite which has resulted in a widening of the northern Devonian flank. The axis of the syncline turns off towards the west, continues into the Mauricio valley and there forms a deep wedge of sediments in the Maladeta granite. The southern flank shows secondary isoclinal folds in the Caregue area, and in the more western Berasty valley the low-amplitude folds are clearly recumbent, especially on the western slope. Towards the granite these recumbent folds gradually become steeper and more isoclinal, the flanks of the folds are also longer here (see section I fig. 17).

The Devonian of Monseny 2881 m is exposed in low-amplitude recumbent folds with an axial plane dip of some 15 to 20° towards the north-east. The anticlines are generally sharp and the synclines more rounded (fig. 28). The folds are not isoclinal; the upper and normal flank is on the whole rather flat, while the lower recumbent flank is rather steep. In some cases it could clearly be determined that this steep flank is shorter than the more horizontal flank (see section I fig. 17, in which this feature is drawn schematically). To what extent these can be regarded as real gliding structures, cannot be decided. Also the transition of the 40 to 60° sloping Devonian

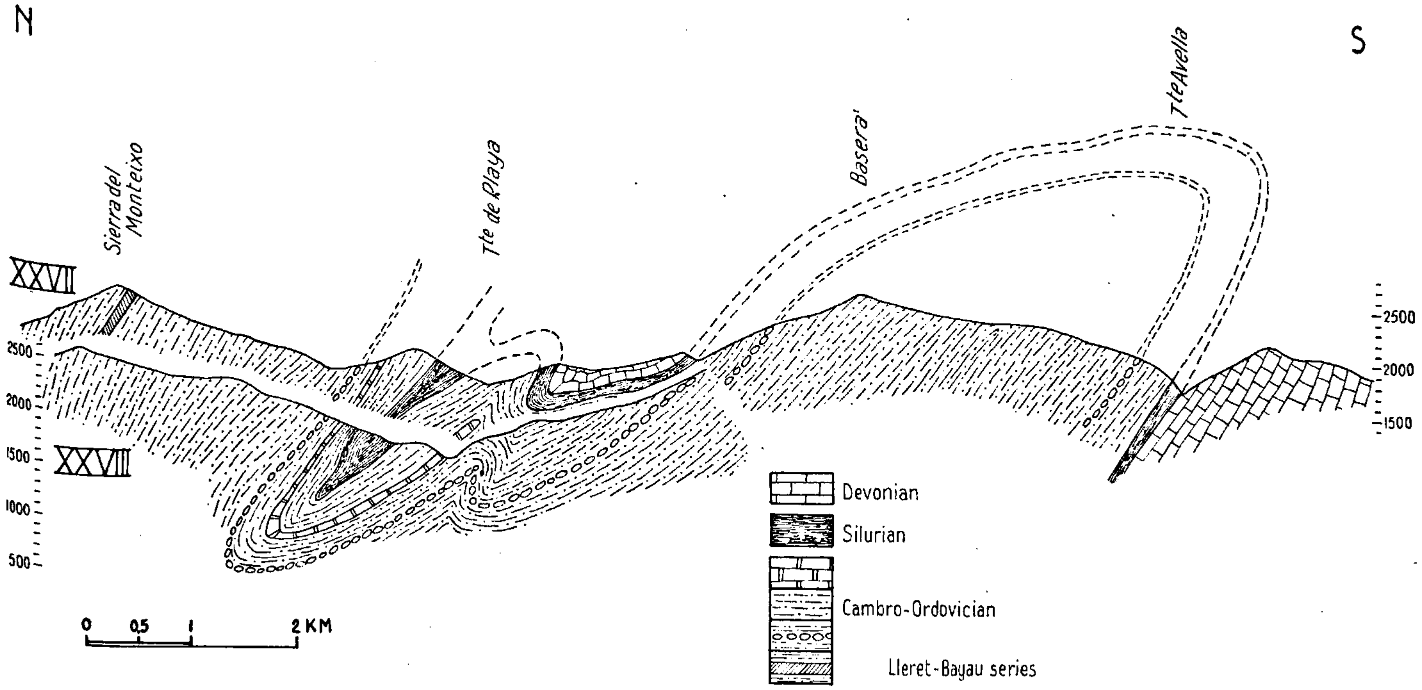


Fig. 26. N—S cross-sections through the syncline of Tór.

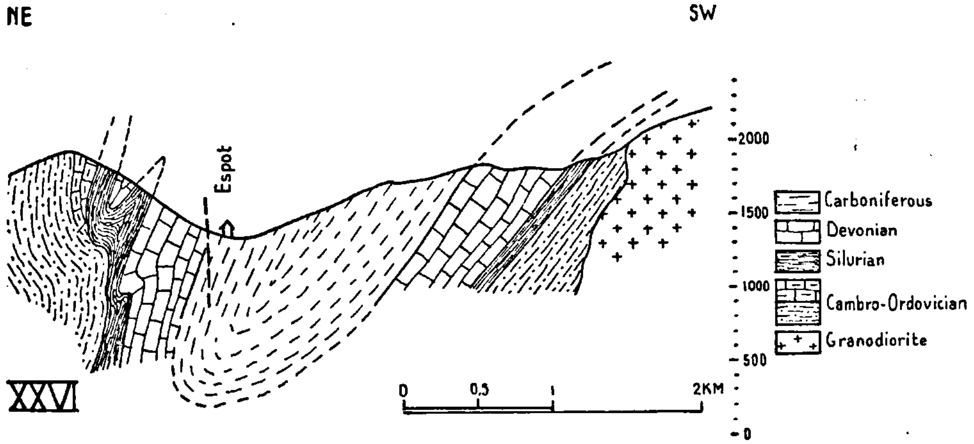


Fig. 27. Cross-section through the Tírvia-Espot syncline near Espot.

of the southern flank of the Tírvia-Espot syncline to the recumbent folds of the Monseny area requires further investigation. It must be added that along the whole southern border of the Maladeta granite the upper Palaeozoic lies in recumbent folds, and it seems probable that this phenomenon is genetically related to the uplift of the Maladeta granite.

Fig. 28. Recumbent fold. Intercalated limestone of the Devonian. Western slope of Monseny.

e. The tectonic significance of the Silurian black shales

From the description of the folding-characteristics of the Lower and Upper Palaeozoic given in this chapter it is clear that there is a structural disharmony between these two sediment successions. This large-scale disharmony results from the presence between these two sequences of a thin layer of black shales of Silurian age which have reacted plastically to the Hercynian orogeny. As they are situated between two comparatively competent sedimentary sequences, they form a detachment plane allowing these units to react independently to the tangential stresses, each according to its own nature and position. Plasticity during folding is undoubtedly the most typical property of the Silurian deposits, which consequently display a number of characteristic features which are little seen or absent in the other Palaeozoic sediments:

1. the Silurian pelites are often very much crumpled in directions not directly connected with the tangential folding-stress. There are also many undulating slip-planes
2. distinct tectonic thickenings and thinnings of the beds, especially north of the main dome and round Monseny

3. diapiric Silurian structures in the Devonian of the upper Pallaresa, (e. g. between Bonabé and Pic de Montaud) and along dislocations (e. g. along the fault of Couflens-Arigail in the upper Pallaresa area and along the Refugio-Baños fault at the top of the Bonaigua pass).

Although the exceedingly fine grain and the homogeneity of the sediment tend to obscure the stratification, the writer believes that the Silurian in its typical development is little, if at all, affected by tectonic cleavage. Indications for this conclusion are the locally numerous undeformed graptolites on planes which are inferred to be bedding planes. No trace of any other fissility can be discovered. A more or less well-developed cleavage would have either entirely destroyed the graptolites or have deformed them. However, the black shales often show a pronounced slickensiding, a feature which corresponds well with the above-described tectonic process.

The graptolites in the less fissile and less fine-grained and hence more competent Silurian deposits of Escaló occasionally give a somewhat deformed impression. The imprints of the graptolites sometimes show small, parallel cracks along which a very small offset has taken place. The orientation of these small cracks could not be fixed with any certainty. They possibly represent a coarse, poorly-developed cleavage.

Although the Silurian pelites are on the whole rather distorted, there occasionally occur outcrops in which the Silurian has retained parallel boundaries even very close to a hinge, for instance south of the mapped area near Pujol, (north-west of Sort), where the author found undeformed graptolites only a few dm away from a hinge.

The author sees the explanation for the absence of a well-developed cleavage in the Silurian pelites in their highly incompetent nature.

f. The North-Pyrenean fault zone

This deep-reaching, long, N 100°E-trending fault zone separates the axial zone of the Pyrenees from the North-Pyrenean zone. The latter consists of Mesozoic rocks surrounding cores of Palaeozoic, the satellite massifs. These massifs form blocks and have tilted to the north along faults of the North-Pyrenean fault zone.

The history of this fault-system reaches back into the Hercynian orogeny although at present the pre-Cenomanian activity accompanied by basic rock intrusion is by far its most prominent character (De Sitter 1954b).

Often the Cretaceous faults obliquely cut the Palaeozoic structure pattern. In our region the fault zone occurs in two localities:

1. The fault zone in the Mesozoic of Auzat-Seix
2. The fault zone of Couflens-Arigail

1. *The fault zone of Auzat-Seix*

In the north-eastern corner of the map the North-Pyrenean fault truncates the E—W trending Palaeozoic structures and thus forms the boundary with a Mesozoic synclinal structure.

This Mesozoic (Jurassic and Lower Cretaceous) syncline still forms a high WNW-trending ridge, which is followed by the rivers running south of it — i. e. the Saleix river and the Garbet river between Castel-Minier and Aulus, the Trape river, the Alet river between

Sérac and the Salat river, and even the Salat river itself over a short distance.

2. *The fault zone of Couflens-Arigail*

This fault zone, running from Aulus to Couflens and further westward can be considered as an E—W Cretaceous aged branch of the North-Pyrenean fault. This is suggested by the occurrence of numerous ophites along the fault plane and one remnant of Cretaceous limestone.

The blocks both north and south of the fault line are tilted northwards as shown by the rocks, which gradually become younger northwards. The movement of the blocks is therefore similar to that of the satellite massifs north of the North-Pyrenean fault, in relation to each other and to the axial zone.

In this Couflens-Arigail area the fault zone consists for 4½ km of its length of two parallel faults, the enclosed strip consisting of Upper Devonian rocks. The fault planes themselves are largely masked by ophite intrusions. Near Rouze occurs a marmorized block of Urgonian-Aptian limestone between an ophite stock and the Silurian. The limestone no longer shows any bedding and the contact with the Silurian is badly exposed thus we do not know whether it is a fault contact or an unconformity. On the map it looks as if this Cretaceous block does not belong to the strip between the faults, lying north of the faulted zone, but as the faults are nowhere very straight this impression may be erroneous. On the other hand the strip between the faults has certainly been downthrown in relation to its borders and the Cretaceous block would fit in better if it belonged to this subsided strip.

In fig. 29 some possible constructions are suggested for the origin of the downthrown strip and the occurrence of the Cretaceous block. In fig. 29a it is supposed that the Cretaceous limestone lies unconformably on the Silurian and is a remnant of the normal Mesozoic cover of the northern block. This explanation is similar to the one suggested by Casteras (1933) for the Lower Cretaceous north of the fault zone between the Arize massif and the Trois Seigneurs massif. In fig. 29b the Cretaceous block forms part of a compressed syncline between the faults, a solution suggested in many similar constructions by Casteras for the main North-Pyrenean fault zone. In fig. 29c and 29d the Cretaceous block has been downthrown by fault movements, in fig. 29c along convergent faults and in fig. 29d along parallel vertical faults. In all these cases the throw must have been considerable. The convergent set of faults suggests a tension field which seems rather improbable. The fault pattern might also be more complicated with an en echelon arrangement, enclosing the Cretaceous block between two branches. Taking the different possibilities into consideration a combination of the compressed syncline and the tilting along the faults appears most probable (fig. 29b).

At the eastern extremity of the fault near the Col de la Trape some ophite occurrences indicate that the Couflens fault can be followed this far and that it there finds its origin as a branch of the main fault plane.

West of Couflens the fault zone continues unbroken until the neighbourhood of the Liesca river. The fault is flanked occasionally by Silurian

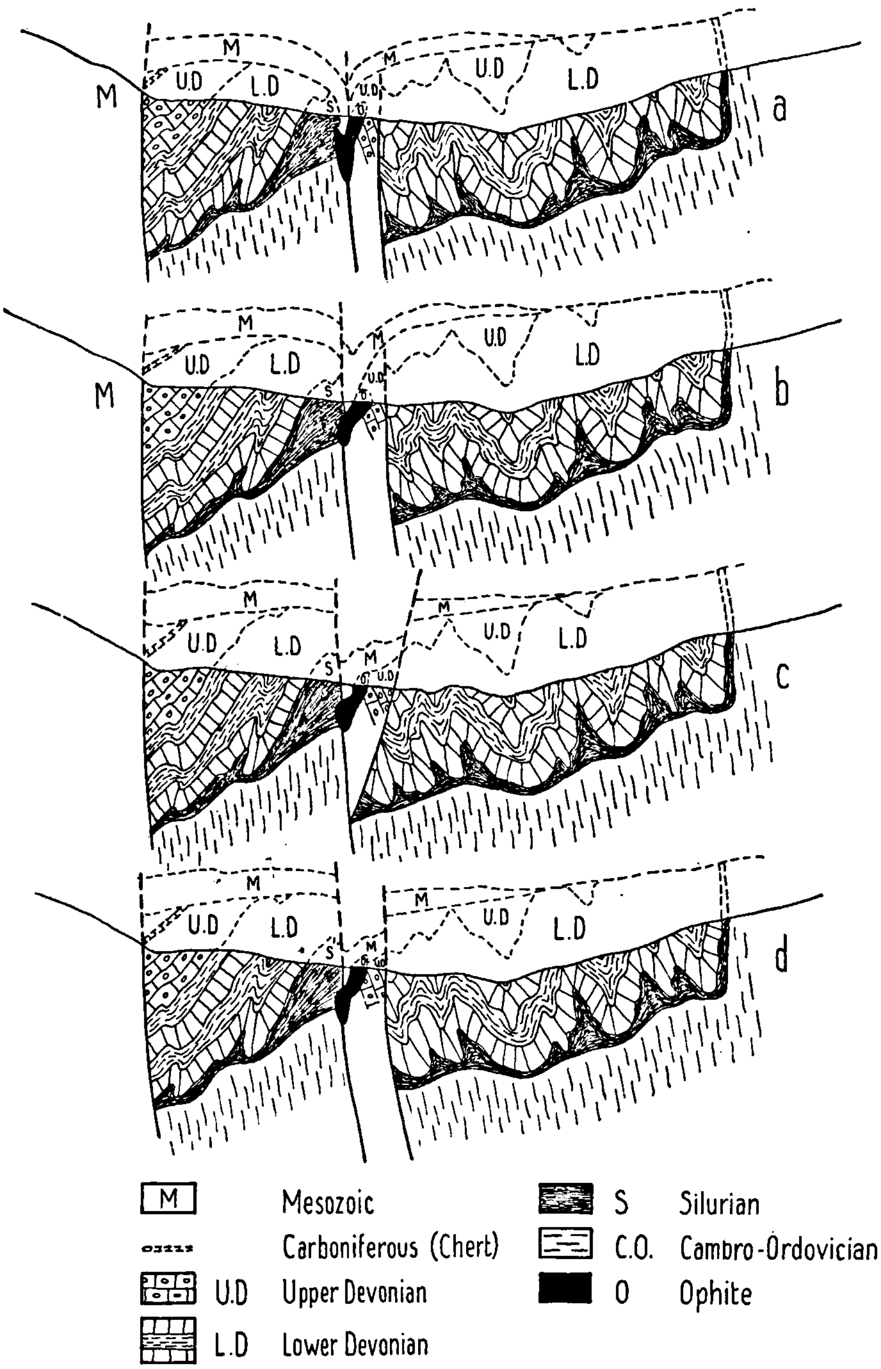


Fig. 29. Schematic N—S section (ca 1: 80.000) of the fault zone of Couflens-Arigail giving four different solutions to explain the occurrence of the Mesozoic limestone outcrop of Rouze. See text p. 78.

outcrops, the one in the valley of the Artigue lying clearly north of the fault again indicating the tilting of the northern block.

When the fault reaches the Devonian of the upper Pallaresa this tilt movement is no longer discernable. Although the fault is still accompanied by Silurian outcrops the latter are strongly squeezed and slickensided and have probably risen diapir-like along the fault plane. Further west the fault is lost in a number of splays and its function is taken over by a number of smaller parallel faults.

The occurrences of ophites along the fault plane poses a problem in itself because they always occur on the crests and not in the valleys except in the shallow Guzet valley. The Guzet ophite shows this peculiarity very clearly by the shape of its outcrop. i. e. narrow in the valley and widening towards the crests. The small ophite bodies between Alet and Bielle are completely restricted to the crests and the Rouze ophite shows the same tendency of widening upwards, although less clearly. The ophites apparently wedge out downwards and it seems possible that the outcrops together form the lower remnants of a large ophite body. It appears as though a post-intrusion pre-Cenomanian or Lutetian compression forced the ophite upwards.

B. STRUCTURAL DETAILS AND LATER DEFORMATIONS

a. The sm-cleavage

All the sedimentary Palaeozoic rocks of our region show a more or less pronounced, very regularly E—W to NW—SE-trending cleavage which is synchronous with the main Hercynian orogeny.

This synchronism is proved by the following facts:

1. The cleavage planes are always parallel to the axial planes of micro-folds
2. The intersection-lineation cleavage/bedding is parallel to the micro-fold axes
3. The trend of the cleavage is parallel to the general structural trend
4. The micro-folds themselves can only be understood as shapes at least partly due to slip along the cleavage planes.

In the Cardós basin only the very competent quartzites and some micro-conglomerates do not show any cleavage and in some instances the rather incompetent Devonian limestones split along the bedding plane. This latter feature is even more prominent in the sandy and rather incompetent Carboniferous sediments of the Tírvia-Espot syncline. These sandy slates sometimes do carry cleavage traces but never become really schistose and apparently the cleavage diminishes upwards in the stratigraphic sequence. The total thickness of the Upper Palaeozoic (at the end of the Palaeozoic) was never more than 2000 m in the centre of the axial zone (Devonian 1000 m max., less than 1000 m Carboniferous). A load of 5 to 6000 m, necessary for the development of cleavage as advocated by Fourmarier (1952) never existed here. As De Sitter (1956b, p. 475) pointed out, the factor of a rising temperature due to migmatization as suggested by Fourmarier cannot explain this deviation from Fourmariers hypothesis in the Pyrenees.

In many thin slides of the Cambro-Ordovician slates there occurs a younger less-pronounced and differently-oriented cleavage offsetting the primary one.

In the field this younger cleavage is much less pronounced and does not lead to misinterpretation of the syn-kinematic cleavage.

The cleavage map (fig. 30) and the sections (fig. 31) show that the cleavage orientation in the axial zone has the same fanning-out character as that often occurring in a simple anticline, the planes converging towards the core of the structure. Thus, the cleavage becomes flatter towards the

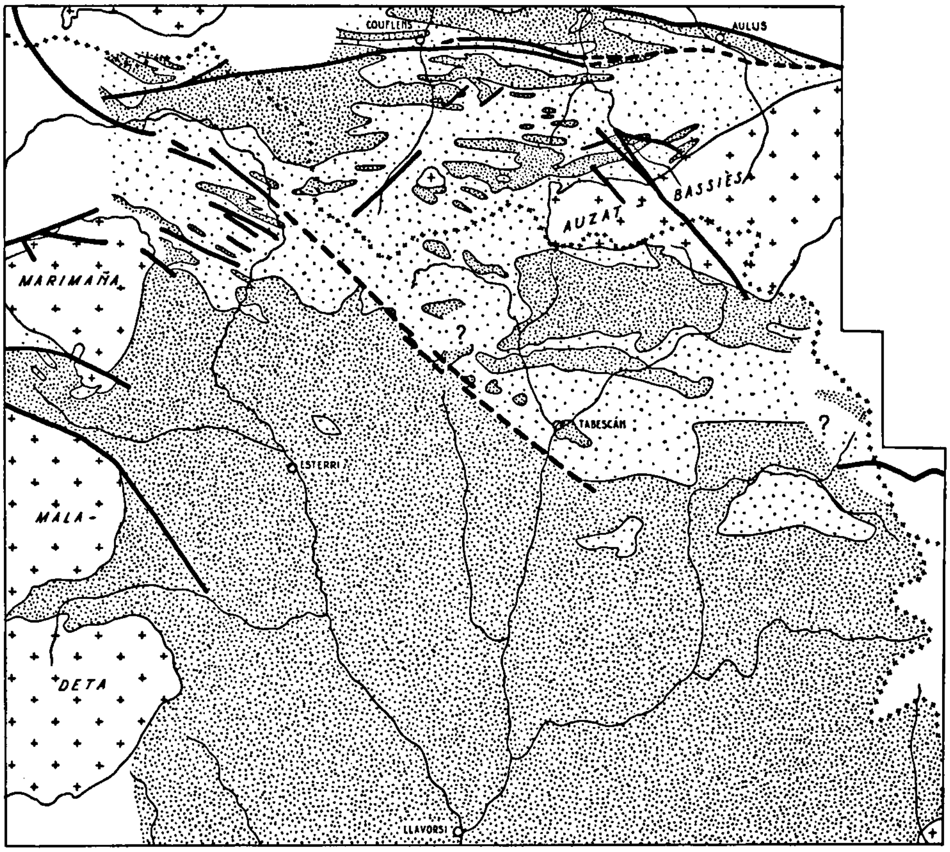


Fig. 30. Map of areas with north- and south-dipping cleavage. Dense stipple: north-dipping cleavage. Sparse stipple: south-dipping cleavage. Thick lines: faults. Broken lines cleavage "faults". Scale ca 1: 310.000.

boundaries of the axial zone, especially towards the south. Towards the northern boundary this flattening is much less pronounced and there the south-dipping cleavage zones alternate with north-dipping zones.

A closer scrutiny of the change from a north dip to a south dip reveals that it is not a transition but takes place in distinct stages so that whole intervals of dips are lacking. The average magnitude of the missing angle has been calculated for the Cardós region round Lake Naorte, always comparing dips as near as possible to the limits of the units of equal cleavage. The missing sector is 32° near the northern limit, 31° near the western

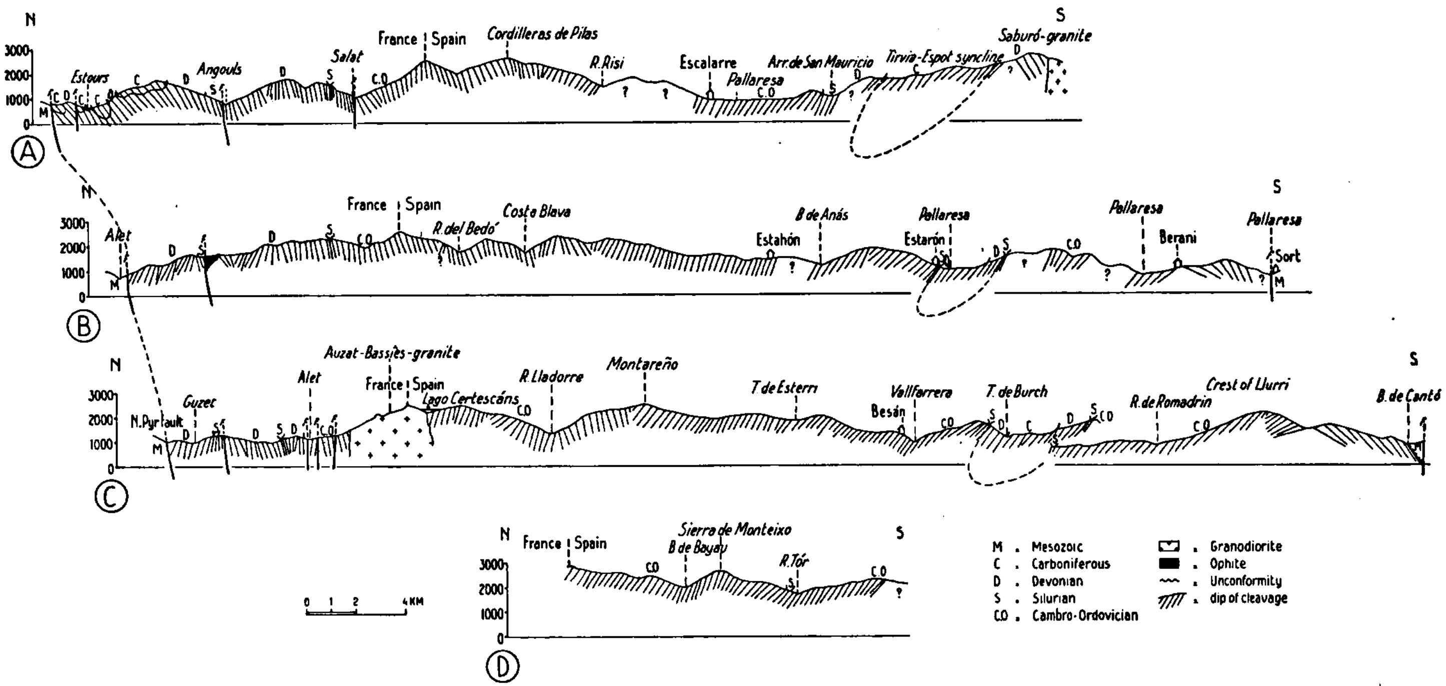


Fig. 31. N—S cross-sections showing cleavage dips (for location see fig. 1).

boundary and also 31° near the southern limit. A similar field, a little further south round Graus and the Turó Caubó gave a missing sector of 37° in the north and 31° in the south. Other cleavage units give the same result. The missing sector is considerably larger than the maximum difference in dip in one unit.

The southern limit of south-dipping cleavage is a line running along the northern boundary of the Marimaña granite over Alós de Isil and Ayneto and further towards the east. Along the line we find a missing sector of ± 20° west of Lago Lagola and of 41° east of this lake. East of Lago Lagola the boundary line between south- and north-dipping cleavage has a NW—SE strike cutting obliquely through the cleavage trend with an angle of about 45°. Further west this line runs into a fault along the Pallaresa maintaining the same strike towards the NW and which perhaps can be linked up with the fault of Liesca-Riberot (fig. 30). Apparently one must assume that the above-mentioned limit between cleavage units is a fault although no mylonitization or dislocation other than the cleavage offset can be found between the Pallaresa and Lago Lagola and hardly offset in the outcrop of the Upper Cambro-Ordovician limestone.

This structural line is curiously enough not reflected in the lineation units; neither do the cleavage dip units seem to be related in any way with the lineation plunge units. Even the N 18°E running lineation boundary line of Bohavi (p. 91) is independent and not related to variation of cleavage-dip. North of the oblique division of the north- and south-dipping cleavage fields we find first a zone of southern dip with irregular and rather extensive zones of northern dip and then, north of the steep flexure along the northern borders of the Auzat-Salau-Marimaña granites in the folded Devonian again an alternation of north- and south-dipping cleavage. In addition to the fields of northern dips there are isolated north-dipping cleavage measurements which make up about 4 % of all the measurements in the south-dipping zones (south of the oblique division line this percentage of individual contrasting dips is only 0.5 %). Also in this northern zone the missing sector between north and south dips is only 20°, or about equal to the maximum variation in dips within the south-dipping fields. The bedding plane dip variation in the isoclinal Devonian is also of the same order.

As has been mentioned before (p. 68), the Devonian becomes gradually younger across the folds in northerly direction, which fact suggests that the top of the Lower Palaeozoic dips with some 15 to 20° northwards. The gradual character of this phenomenon indicates that this basement inclination is not disturbed by further flexures or faults, except of course the Couflens fault zone (p. 78). The axial planes of the Devonian folds are rougly perpendicular to the inclined Lower Palaeozoic top, but they vary considerably in the same way as the cleavage dips.

It is noteworthy that as the flexure is approached, the folds become steadily tighter and the cleavage steadily less well-aligned parallel to the fold axial planes. This would seem to indicate a genetic relationship of these features with the flexure.

b. *Knick-zones*

South of the line Esterri—Lladrós—Tór a secondary coarse cleavage cuts through the finely-cleaved Palaeozoic sediments. This secondary cleavage always occurs in pairs of cleavage faces which enclose a narrow zone in which the sm-planes are knicked or flexured (fig. 32, 33). This long and narrow flexuring is essential to this kind of secondary deformation and therefore we will call it knicking cleavage and designate it by the symbol sk. In German literature we find it described as "Knitterung" (Kienow 1934, p. 69), "Knickbänder" (Simpson 1940, p. 34), "Knickzone" (Kienow 1951, p. 41, Hoeppener 1955, p. 34 ff) and "Flexuren" (Hoeppener 1956, p. 268 ff).

The strike of the sk is roughly parallel to that of the first cleavage, E—W to ESE—WNW, but its dip is uniformly to the south in contrast to the northern dip of the sm-planes.

The table on p. 88 gives some dimensions of the knicks, they were gathered mostly from some ten short (2 to 30 m long) sections in the Cambro-Ordovician slates of the Cardós and Valfarrera areas.

In general the sm-micro-lithons of non-calcareous slates have been broken by the knicking but without any slip along the planes, but the very fissile slates between Esterri and Guingueta are bent and not broken like the calcareous slates of the Devonian. In the massive Upper Devonian griottes the bending is more rounded giving the impression of micro-folds but the similar orientation of the axial planes suggests that this micro-folding belongs to the flexuring.

The sk-planes are somewhat wavy and never show any slickensiding or drag crinckles. When they can be followed over some distance they appear remarkably plane and parallel. Between Lladorre and La Fábrica they rise from the bottom of the valley up to the crest and some of the southern slopes of E—W crests along the Noguera Cardós are certainly dip slopes of sk-planes.

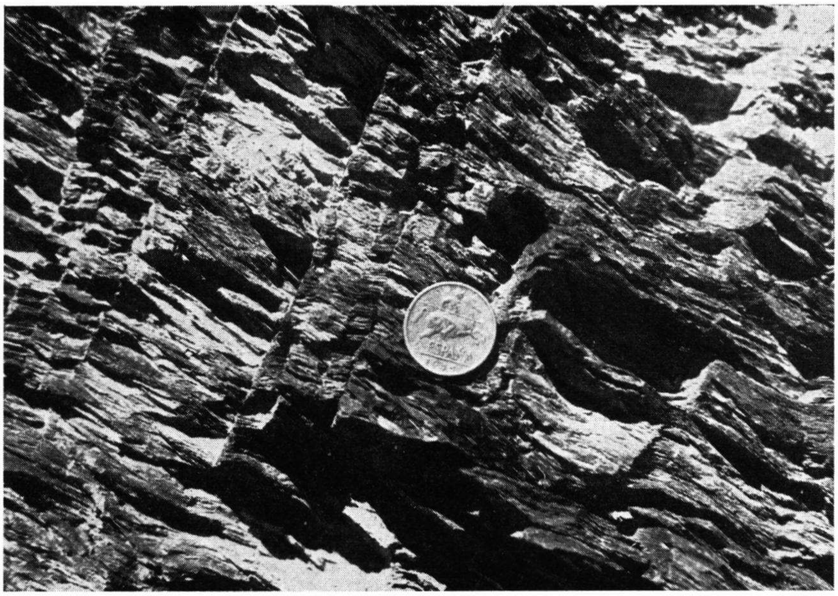
The knick-zones have in general a lens shape because the sk-planes approach one another and finally merge at the extremities of a lens and then they may show some slip in the same sense as the knick. Sometimes the knick-zone simply dies out. In that case the sk-planes gradually turn to a perpendicular position to the sm-planes and at the same time the knick movement diminishes (fig. 34). The width of the knick-zone is in general much smaller than that between two knicks, but when their frequency increases the two widths may become almost equal (see the table on p. 88).

The knick-zones become more frequent in the finer-grained slates, and in the paper shales of La Fábrica some 28 knicks were noted per horizontal meter. When the sand content of the slates increases, the sk-cleavage becomes coarser, less regular and less frequent. The quartzites and micro-conglomerates which do not show primary cleavage do not show any knicking either.

The conclusion that only well-laminated rocks, which allow differential movement of the laminae, can adopt knicking cleavage seems warranted, because knicking cleavage is only possible by slip within the knick-zones. This slip is also demonstrated by occasional recrystallization of calcite (in the Devonian) or quartz only within the knick-zones and not outside.

As mentioned before, the strike of knicking cleavage planes is parallel to that of the primary cleavage, roughly E—W and the rotation is always southwards, that is the southern limb has subsided or the northern limb has been lifted up. Which of these alternatives actually happened, uplift or subsidence cannot be deduced from the knicks themselves.

Almost everywhere the knicking cleavage plane bisects the angle between the rotated and non-rotated primary cleavage (fig. 35 A). This is always true when the rotation is less than 25°, but when the angle between sk and sm



S N

Fig. 32. Knick-zones: sm-cleavage (dipping 40° N) fractured and rotated between pairs of sk-fracture planes (dipping 60° S). Cambro-Ordovician slates, 750 m north of La Fábrica.

S N

Fig. 33. Exposed sk-planes dipping to the south. Cambro-Ordovician slates, 1200 m north of La Fábrica.

is small and therefore the angle of rotation must be large to reach this bisecting position of sk it often fails to do so. When the sk-plane is a bisector the width of the knick-zone is the same before and after the knicking, the movement is then something like the twinning in a stressed calcite crystal fig. 35 B). This geometrical connection between sk- and sm-planes also means that the dip of the sk-planes is dependent on that of the sm-cleavage, that the knicking happened only once and in one direction only and is therefore the result of only one stress field.

In analogy with twin-gliding in calcite, intensification of deformation can be affected only by increasing the number of knick-zones and by further rotation or slip.

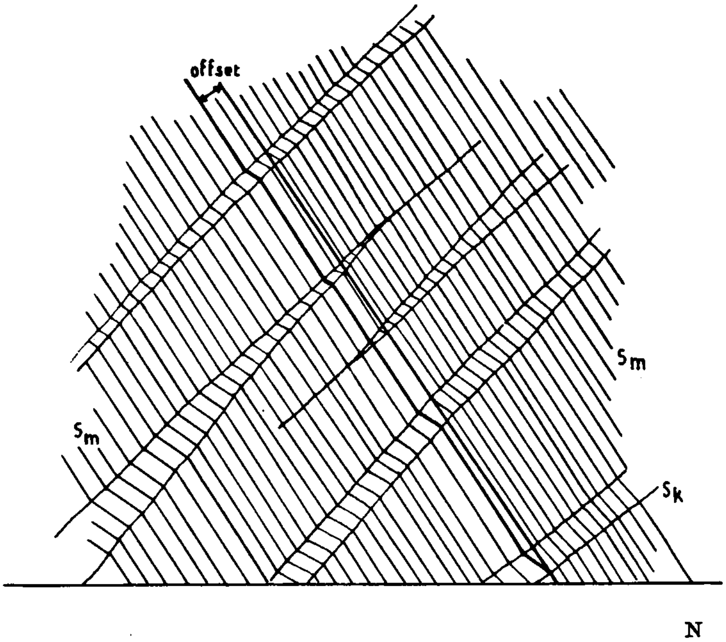


Fig. 34. Schematic representation of knick-zones.
sm = cleavage of the mainfolding.
sk = knick-planes.

When the angle between sk and sm approaches 90°, the rotation angle is small and the temporary dilatation is also small (for 20° rotation only 0.8 % of the knick-width), but when this sk/sm angle is smaller the necessary rotation angle becomes larger and the temporary dilatation increases (with a 40° rotation to 6.3 %, and with a 60° rotation to 16.4 %) a value which certainly is difficult to surmount. In those cases the rotation is apt to stay behind, and remains less than half the angle necessary to make the sk-plane a bisector. This feature is observed in particular on the northern border of the area of knicked cleavage where the knicks become more and more irregular not only in dip but also in strike and also become less frequent. North of Lladrós the sk-cleavage disappears altogether.

Along the Cardós river the knick-zones were measured accurately along a 9250 m long section between Lladrós and the Silurian of La Fábrica. The dip of the sk-planes is about 40 to 45° south in the north and about 65° south in the south, averaging 54° south. The sm-dip decreases southwards from 60 to 45° north, averaging 56° north. The integrated offset caused by all the knicks amounts to only about 300 m dilatation in a horizontal sense and a compression of about 415 m in the vertical sense in this 9 km section. The orientation of the stress field is therefore certainly not a horizontal compression but its exact orientation remains doubtful as the sense of movement was strongly influenced by the existing sm-cleavage.

The final analysis of the knicking cleavage in our area gives a dilatation in the horizontal sense and therefore strongly suggest a horizontal tension field and certainly not a further horizontal compression.

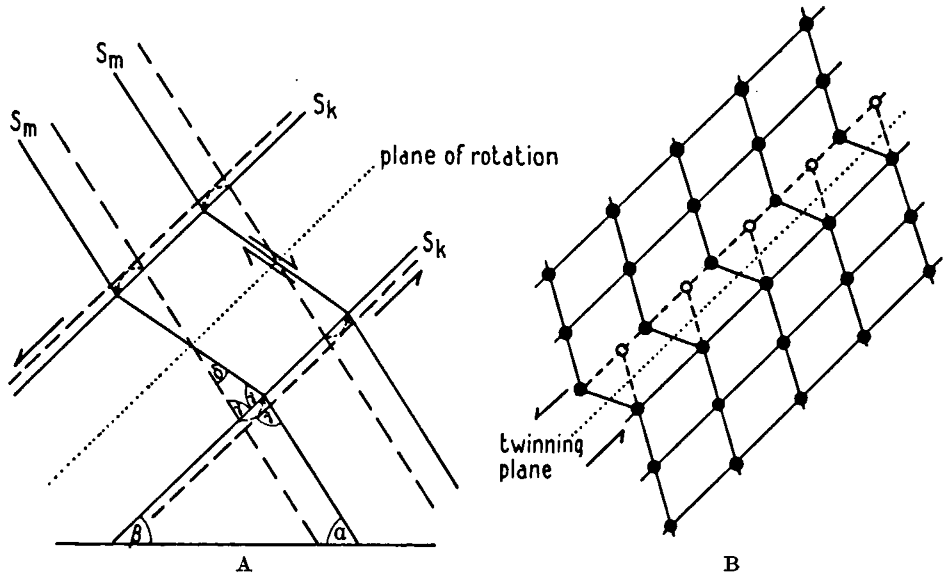


Fig. 35 A. Schematic representation of the movements taking place during the formation of a knick-zone, with the implied dilatation of this zone.

Fig. 35 B. After Fairbairn 1949. Twingliding in a simple rhombohedral lattice compared with the formation of knick-zones (A).

c. *Lineation and lineation sections*

Lineations in their geological sense have been comprehensively described by E. Cloos (1946). The only lineations taken into account in the present region are formed exclusively by the intersection of cleavage (sm) and bedding (ss). This lineation is well exposed in the whole area, mostly on the cleavage planes as a banding of different colour, grain size or composition (fig. 36).

When the rock splits along the bedding plane, which it rarely does, a fine lineation, less than 1 mm apart and always perfectly parallel is evident. The fact that this intersection lineation is present nearly everywhere shows that true parallelism between bedding and cleavage is never attained, although the angle between the two may be very small.

If the cleavage is parallel to the fold axial planes, this lineation ought to be parallel to the plunge and of use in determining the general plunge.

Some data concerning the knick-zones

section number	length of section	average strike and dip of sm-cleavage	average strike and dip of sk-fracture planes	width range of knick-zones	average width of knick-zones	range of distance between knick-zones	average distance between knick-zones	offset range of knick-zones	average offset of knick-zones	number of knick-zones per m (horizontally)	average aggregate offset per m (horizontally)	average offset per cm width of knick-zone	angle of rotation (knick-angle, δ in fig. 35 A)	necessary rotation angle for bisector position of sk
I^1	200	356/66	222/68	0.6— 2.4	1.5	2.5— 49.0	22.9	0.2— 1.0	0.5	3.5	1.8	0.3	19°	76°
I^2	200	8/58	215/68	0.8— 5.0	2.4	0.6— 44.0	13.7	0.2— 0.9	0.5	5.0	2.3	0.2	11°	66°
I^3	200	6/57	205/60	0.4— 8.2	2.6	3.8— 38.5	7.0	0.1— 1.8	0.9	9.0	8.0	0.3	20°	46°
I^4	200	7/57	200/62	0.7— 3.3	2.5	1.1— 19.8	3.8	0.2—19.8	1.6	11.5	18.7	0.6	$38^1/_2$°	49°
I^5	240	8/57	203/56	0.6—26.0	6.3	2.3— 54.0	25.3	0.2— 9.0	2.1	3.3	8.4	0.3	$19^2/_3$°	39°
$I^{1,2,3,5}$	840	5/59	209/64	0.4—26.0	3.1	0.6— 54.0	14.6	0.1— 9.0	1.0	5.1	4.9	0.3	18°	58°
II	178	4/36	215/63	0.3— 2.4	0.8	0.6— 6.7	2.6	0.1— 0.5	0.2	21.3	5.1	0.3	$17^1/_2$°	14°
III^1	1500	12/42	197/64	0.5—13.2	2.9	3.8—159.0	57.5	0.2— 3.1	1.0	1.5	1.5	0.4	$20^1/_2$°	21°
III^2	1500	18/42	196/66	0.2—12.0	2.7	3.1—126.0	32.8	0.1— 4.1	1.4	3.5	5.0	0.5	23°	22°
IV^1	1300	5/57	170/45	0.7—25.0	3.6	—	—	0.3—12.0	1.4	3.1	4.3	0.4	$22^1/_2$°	24°
IV^2	3000	5/57	170/45	1.0—25.0	5.0	—	—	0.5—16.0	2.1	2.4	5.1	0.4	$24^1/_2$°	24°

All dimensions in centimetres, strikes and dips in degrees.

I^{1-5}: 500—700 m north of Ribera de Cardós along the road towards Lladorre.
II : 750 m north of la Fábrica along the road towards Ribera.
III^{1-2}: 1250 m „ „ „ „ „ „ „ „
IV^{1-2}: 6000 m upstream from Alíns along the road towards Tór.

N.B. La Fábrica is situated there where the Vallfarrera and Cardós rivers unite.

Where micro-folds were exposed we observed that the cleavage forms the axial plane and that the above-mentioned parallelism actually exists. It remains difficult to ascertain in how far these micro-fold axes are parallel to the main fold axes, but in general the lineation plunge is in agreement with the general structure. West of the region the Cambro-Ordovician plunges below the Devonian of the Valle de Arán and the western part of this region is characterized by a westerly plunge of the lineation. The east-plunging Maubermé anticline north of Montgarri is flanked by easterly-plunging lineation. The westerly plunge of the Massif de l'Aston is also expressed by a similar plunge of the lineations in the east towards Estats. But in some cases the general structure is unrelated to or only poorly reflected by the

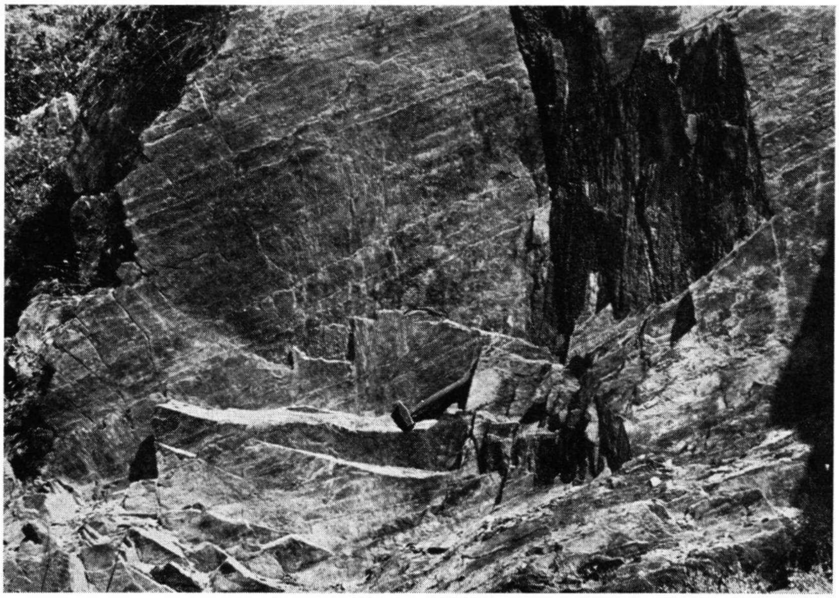


Fig. 36. Lineation on cleavage plane representing the intersection of the stratification. Cambro-Ordovician banded sandy slates, Alós de Isil.

lineation plunge. The Tór syncline for instance certainly broadens and plunges eastward. But where the Devonian outcrops in its eastern part we find a western plunge of the lineation and in the west, above Norís, we find unexpectedly Silurian with westerly-plunging lineations, but apart from these deviations the Tór syncline is accompanied by an easterly plunge.

The strike of the lineations is E—W or ESE—WSW, they plunge either to the west or the east, and the occurrence of the westerly and easterly plunges is not arbitrary but is the same in relative large areas which might be termed lineation units (fig. 37). Within these units a contradictory plunge does occur occasionally (varying from 8 % in unit II A to none in units III and IV). We will neglect these exceptions and state that a lineation unit plunges exclusively in one direction.

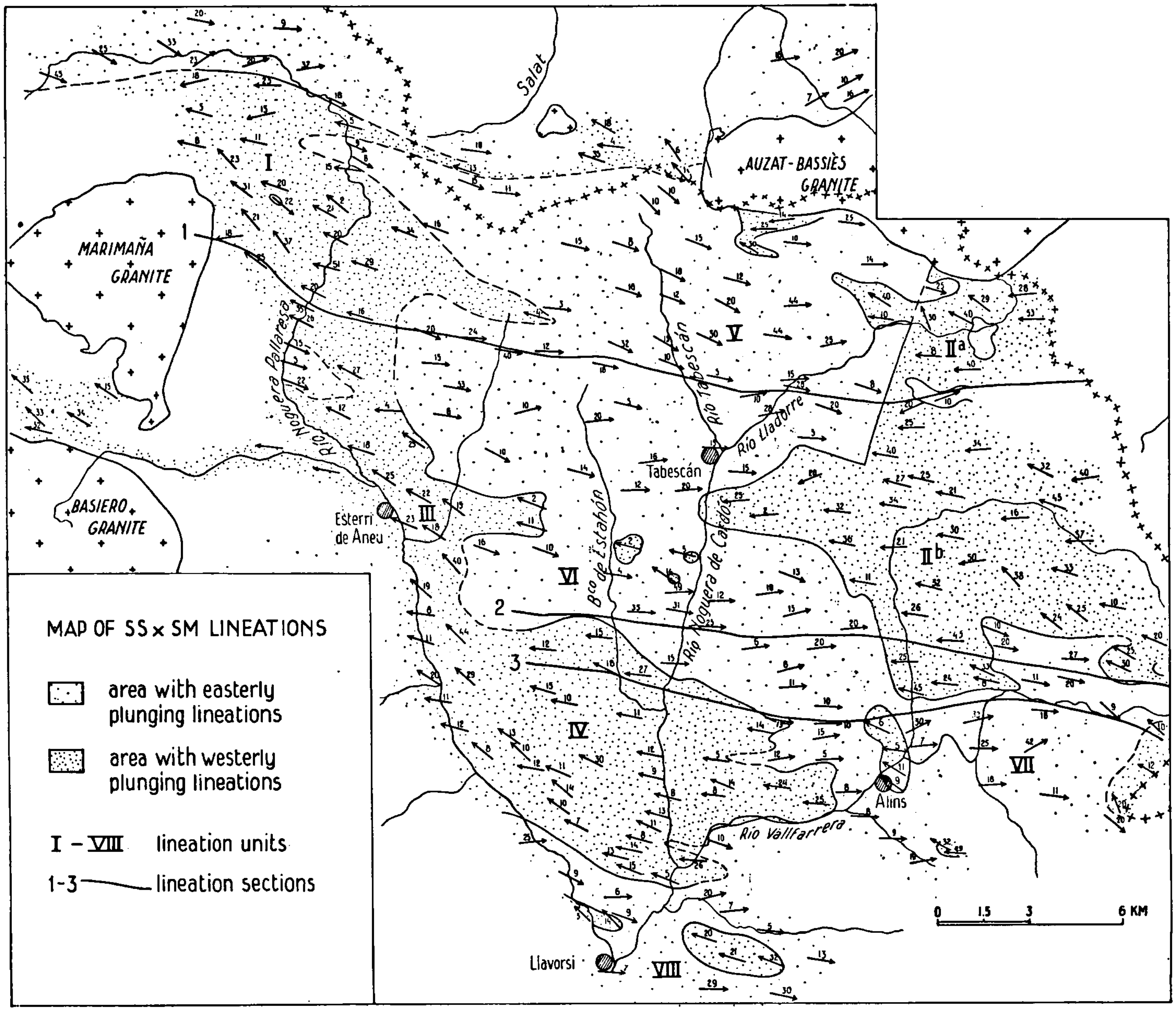


Fig. 37. Lineation map.

We distinguish eight units:

I	unit	of	Moredo-Alós-de-Isil	with	W-plunge
IIA	„	„	Brohate	„	W- „
IIB	„	„	upper Vallfarrera	„	W- „
III	„	„	Esterri-de-Aneo	„	W- „
IV	„	„	Berrós-Ribera-de-Cardós	„	W- „
V	„	„	ríos Risi-Tabescán-Lladorre	„	E- „
VI	„	„	Lladrós	„	E- „
VII	„	„	Arrahós-Tór	„	E- „
VIII	„	„	syncline of Tírvia-Espot	„	E- „

In order to facilitate a comparison between these units we have assembled them in histograms (fig. 38), where their pertinent facts are given in percentages and where the average plunge of a unit has been calculated as the arithmatic mean by adding the tangents of all plunge angles of one unit and dividing the sum by their number. This method is warranted because their strikes are parallel and the measurements are equally spread over the surface of one unit.

The following results may be of interest.

1. The average easterly plunge in the units is rather uniform

   unit V: 20° from 447 measurements
   „ VI: 20° „ 171 „
   „ VII: 19° „ 136 „
   „ VIII: 17.5° „ 41 „

   Only unit VIII, the Tírvia-Espot syncline, is slightly different, but this is the only unit not belonging to the large Lower Palaeozoic dome and the average is calculated from 41 measurements only.

2. The westerly plunges are more variable and on the whole steeper than the easterly plunges

   unit I : 28° from 198 measurements
   „ IIA: 35° „ 55 „
   „ IIB: 28° „ 269 „
   „ III : 23° „ 82 „
   „ IV : 16° „ 214 „

   IIA and IIB belong probably to one unit and III and IV perhaps also.

3. The average of all easterly plunges (795 measurements) is 20° and for all westerly plunges (818) 24.5°.

4. Very steep plunges occur in unit IIA, where 7 % have a plunge of more than 60° and 34 % are steeper than 40°. In the other units these percentages are much lower, for plunges steeper than 40° maximum 15 %, mostly only 5 %.

The lineations in IIA are less trustworthy than elsewhere, mostly because they are difficult to discern. Round Pic d'Estats the lineations are lacking altogether in these hard, irregular and metamorphic rocks with poorly-developed cleavage.

The easterly and westerly lineation units are separated by a remarkable straight 5 km long line trending N 18° E from Montareño (2550 m alt.) to Pleta de Bohavi. North of Bohavi the line is interrupted but beyond the lake of Rumedo it can be followed a little further in the same direction.

This line is in no way reflected in the topography. (It is interesting to note that the line can perhaps be followed into the Auzat-Bassiès granite as large joints visible on the aerial photographs).

In general the boundaries of the units are not straight lines and between

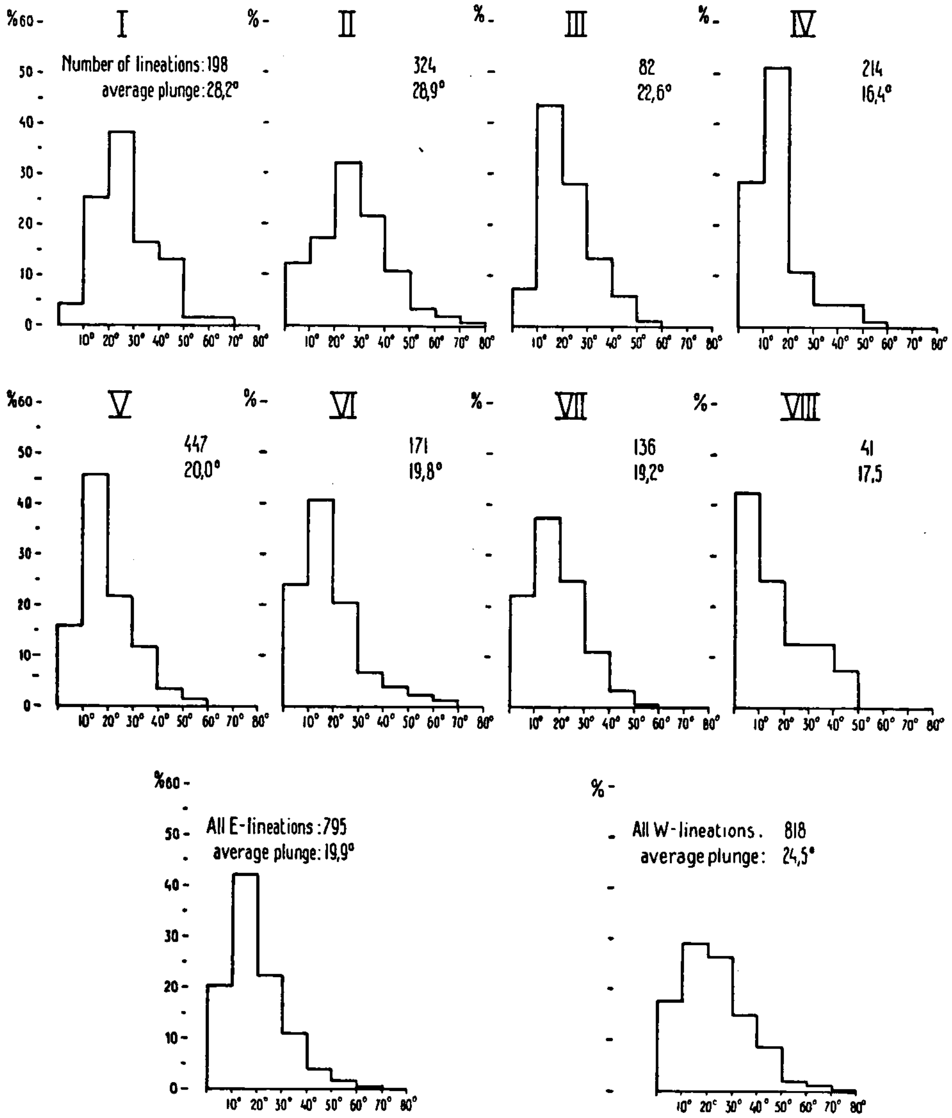


Fig. 38. Lineation plunge histograms. Lineation units I to IV with westerly plunges, V to VIII with easterly plunges.

the units we do not find a transition zone where a westerly plunge gradually changes into an easterly plunge passing through the horizontal. Only along the Vallfarrera north of Alins we find a gradual change, here we measured going from north to south over a 1½ km distance the following values: 39° E, 30° E, 15° E, 7° E, 2° W, 5° W, 7° W, 11° W and 19° W. The angle between average plunges of different adjoining lineation plunge units varies from 35—50°, average about 44° (from 20° E to 24.5° W). For measurements near the boundary lines less than 50 m apart this angle is still 35° (from 16° E to 19° W). Steep plunges are rare near the boundary lines and flat plunges somewhat more frequent than elsewhere.

From a mechanical point of view the function of the sudden changes in plunge along the boundary lines remains rather obscure, a change from 30—35° in plunge in less than 50 m horizontal distance is a rather catastrophic phenomenon, but no fault indications could be found. Apparently a kind of cross-folding with axis either perpendicular to the strike (Bohavi line) or making a sharp angle with it developed in a late stage of the synkinematic cleavage folding, probably accompanied by intensive micro-folding

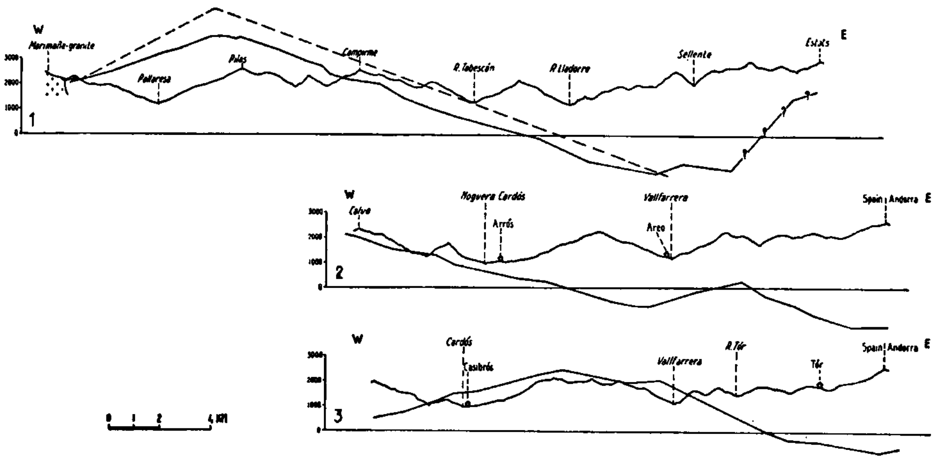


Fig. 39. Lineation sections. Broken line represents average plunge of lineation units I and V. For location see fig. 37.

with strongly variable axial plunges. The sudden variations in lineation plunge is completely independant of cleavage dip or strike variation, because the cross-folding movement itself took advantage of the cleavage planes.

Longitudinal lineation sections. — As the intersection lineation cleavage/bedding is parallel to the fold axis a longitudinal section can be constructed with conventional methods (De Sitter, 1941, p. 14 ff) making use of the lineation plunges. Particular stratigraphic horizons can thus be projected from one end of the section to the other and thicknesses be measured even in regions where in general mapping of stratigraphic horizons is no longer possible.

In the present area we constructed such a section from Moredo to Estats with the base of the Cambro-Ordovician metalliferous limestone as a marker (fig. 39). It is seen that the limestone ought to reappear in the ridge of the Campirme, but in actual fact it is totally absent in the general neighbourhood of this crest. Even if the heteropic character of the reef facies of this limestone causes its absence in the Campirme region the sandstones and slates of the Cardós region do not belong to the upper part of the Cambro-Ordovician.

When we construct the section with averages of lineation plunges (28° west in unit I and 20° E in unit V) we arrive at roughly the same result. Two other sections further south were constructed; they contain no marker horizon but give only the shape of the longitudinal section which varies considerably although the sections are only 1200 to 2700 m apart in the horizontal sense.

The discrepancy between the lineation sections and the stratigraphy may be due to a considerable thickening of the section from west to east or to other disturbing factors. Unrecognized faults for instance or large differential movements along cleavage planes in a late stage of the folding process causing non-parallelism between the lineation and the large-scale structural plunges. In view of the fact that the lineation plunges show remarkable irregularities indicating a late type of cross-folding it rather appears that this is the disturbing factor. Nevertheless from a general point of view the lineation plunges give a good picture of the structure even in the longitudinal section.

d. Jointing

N.B. The study of jointing in the sheet 5 area has been restricted to those joints which are visible on aerial photographs.

The aerial photographs of the Palaeozoic region of the Pyrenees everywhere show a more or less distinct pattern of straight lineaments. The more pronounced ones are visible also in the field as deep grooves, sometimes eroded into deep and narrow valleys. They cross crests and valleys without change in strike and must be vertical planes of fracturing. The drainage pattern makes use of them to a considerable degree (see p. 31). The smaller lineaments do not show in the relief, although on the photographs they show up as discontinuous thin lines.

These fractures may attain a length of 10 km (the lineament of the Ars valley for instance is 8.5 km long), their minimum mappable length is determined by the scale and sharpness of the photographs.

In the field the lineament fractures are not well exposed, usually being covered by scree, but in the high-altitude areas with little vegetation such as the granite and some Cambro-Ordovician massifs (Estats, Campirme) the fracture planes are vertical but none of them showed any slickensiding. These planes are in general accompanied by smaller parallel joints, together forming narrow clefts, often of variable width but always limited on both sides by fracture planes. Obviously it is in these bare regions that the frequency of visible joints is greatest. The maximum frequency was 80 joints per sq km with a total length of 15 km.

The length and strike of the fractures have been measured on the aerial photographs and transferred to the maps and have been assembled in groups of 5° strike interval on rose diagrams. The method of using a length unit in the diagrams instead of the frequency has the advantage of avoiding a predominating influence of the smaller joints. However, frequency and length unit diagrams for the same area show little difference (fig. 41a and b), and the comparison of long (> 1 km) fractures with short (< 1 km) ones show the same maxima for both, although the long-fracture diagrams show a more simple picture (fig. 41 c and e).

In fig. 40, more than 100 diagrams have been assembled, all together representing almost 1700 km of fracture. The general picture does not show

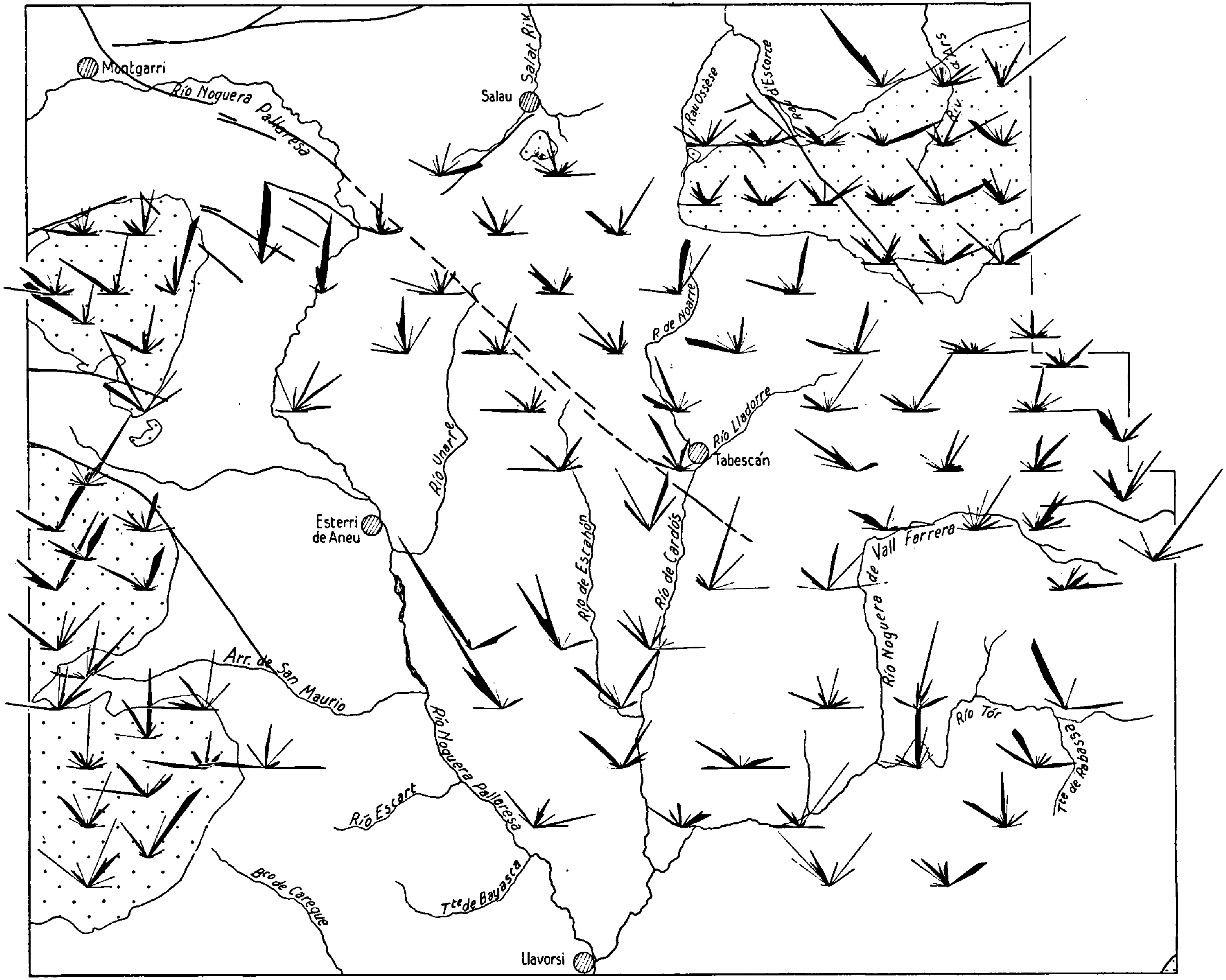


Fig. 40. Roses based on the strike and length of joints visible on the aerial photographs. Dots: granite batholiths. Broken lines cleavage "faults". Scale ca 1: 200.000.

a general consistency in joint orientations over extensive areas as has been observed for instance by Mollard (1957) and Blanchet (1957) in Canada. All of the diagrams show a rather large number of maxima and one can almost always match each set with another one roughly perpendicular to it. An E—W set occurs nearly always, while a N—S set is much less pronounced and often dubious. The N 10—20° E strike is common to many and a pronounced maximum occurs at N 25—35° E with a conjugate perpendicular set at N 50—60° W. Another conjugate set is presented by the maxima of

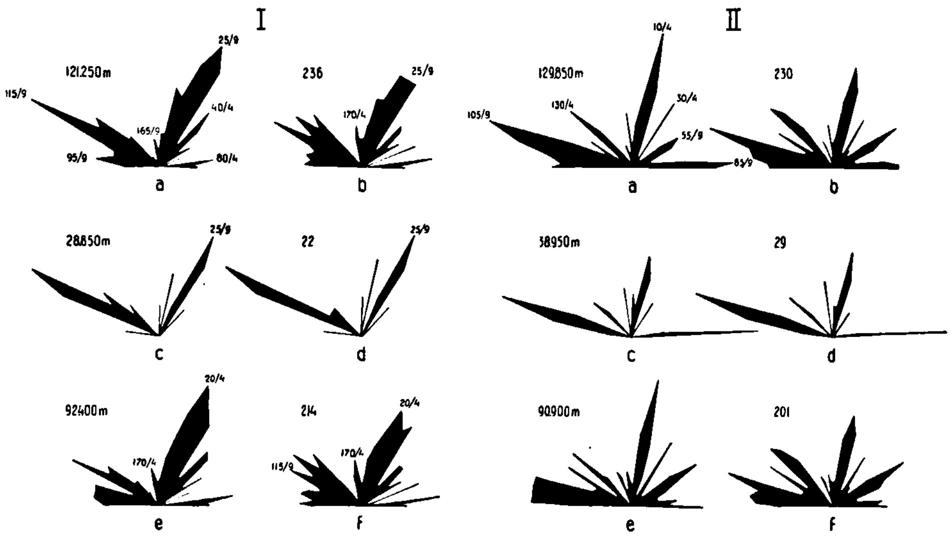


Fig. 41. Joint diagrams of a granite area (I, Basiero batholith) and of a sedimentary area (II, 60 sq km, south of Auzat-Bassiès batholith).

a strike-length diagram of all joints
b „ -frequency „ „ „ „
c „ -length „ „ joints > 1 km
d „ -frequency „ „ „ > 1 km
e „ -length „ „ joints < 1 km
f „ -frequency „ „ „ < 1 km

Total lengths and numbers are indicated for each diagram.

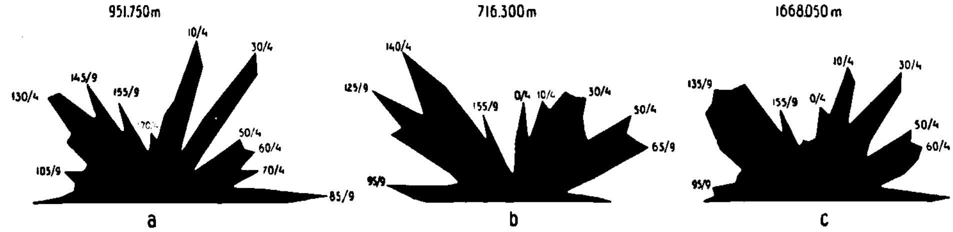


Fig. 42. Strike-length joint diagrams of
a all joints in sedimentary rocks
b „ „ in granite rocks
c „ „ both in sediments and in granite

Total lengths are indicated above each diagram and the 5 degree azimuth interval maxima are labelled.

N 50—65° E and N 20—25° W. None of the sets consequently offsets another one. but some of the larger fractures of the N 35—40° W system in the Auzat-Bassiès massif offset the granite contact boundary, but not the joints of another set. Whether or not later movement took place, cannot be decided. Often joints of three different sets converge in one point, for instance east of Lago Rumedo sup. (Auzat-Bassiès granodiorite), where they cross one another with an angle of 60°.

Obviously with so many maxima there is no difficulty to find coincidences of fault and joint strikes. The Moredo-Campans faults, north-east of the Marimaña granite for instance are parallel to a pronounced N 55° W fracture set in the Marimaña granite, and the strike of the Bohavi line (a boundary line between east and west lineation plunges, see p. 91) can be found back in the N 10—20° E joint set. The cleavage fault between Lago Lagola and Ayneto, striking N 45—50° W, is parallel to numerous joints of the same strike. It is difficult to decide whether such parallelisms are simply fortuitous or indicate a causal relation.

It is, in general, believed that joints in igneous rocks originate during the rise and emplacement of such rock bodies and that there exist mutual relations between the form of a plutonic mass, its primary flow structures and its fracture systems (e. g. Balk 1937). Although this also holds good for the small shear and tension joints one sees in an outcrop, it is questionable, whether all these small joints are genetically related to the vertical fracturing visible on aerial photographs, since the former jointing is often far from vertical. The same question arises with regard to the joints outside the igneous rocks.

On the south-western slope of the Beartooth Mountains (Montana & Wyoming, USA) Spencer (1958) has recorded and compared the strike patterns of joints and dykes in outcrops and on aerial photographs. There appears to be a great deal of agreement between the most prominent recurring dyke trends, the most prominent lineaments on aerial photographs and three of the four most distinct trends obtained by measurement of fractures in outcrops. The latter type of joints was exclusively measured in granitic gneiss (metamorphic sediments). It is striking that the joint pattern on the aerial photograph of a granitic gneiss area, published in Spencer's paper, is very similar to the fracture systems present in the intrusive granodiorite massifs of the Pyrenees and in the low-grade metamorphic Lower Palaeozoic sediments where this fracturing is well developed (cf. fig. 43 with fig. 2 of Spencer).

A noteworthy difference, however, is that both Balk and Spencer mention thick intrusions trending in the same direction as the joints. The granite and aplite dykes in the present area are restricted to the granodiorite massifs and a narrow zone outside them, they do not show a single coincidence with the trends of the big fractures. Differences cannot even be made out between the joint systems in the crystalline massifs and outside them; the two rock types show entirely comparable joint patterns (fig. 43, 42). Some longer joints, moreover, cut crystalline as well as sedimentary rock, and the boundary between the two rock types is sometimes dislocated (Auzat-Bassiès massif). All these data indicate that the fracture patterns on the aerial photographs are the result of one and the same stress field.

This regional joint pattern is clearly a later phenomenon than the emplacement of the granodiorites and its genetically related joint system in the sense of Balk. The joints are vertical over the whole region and there-

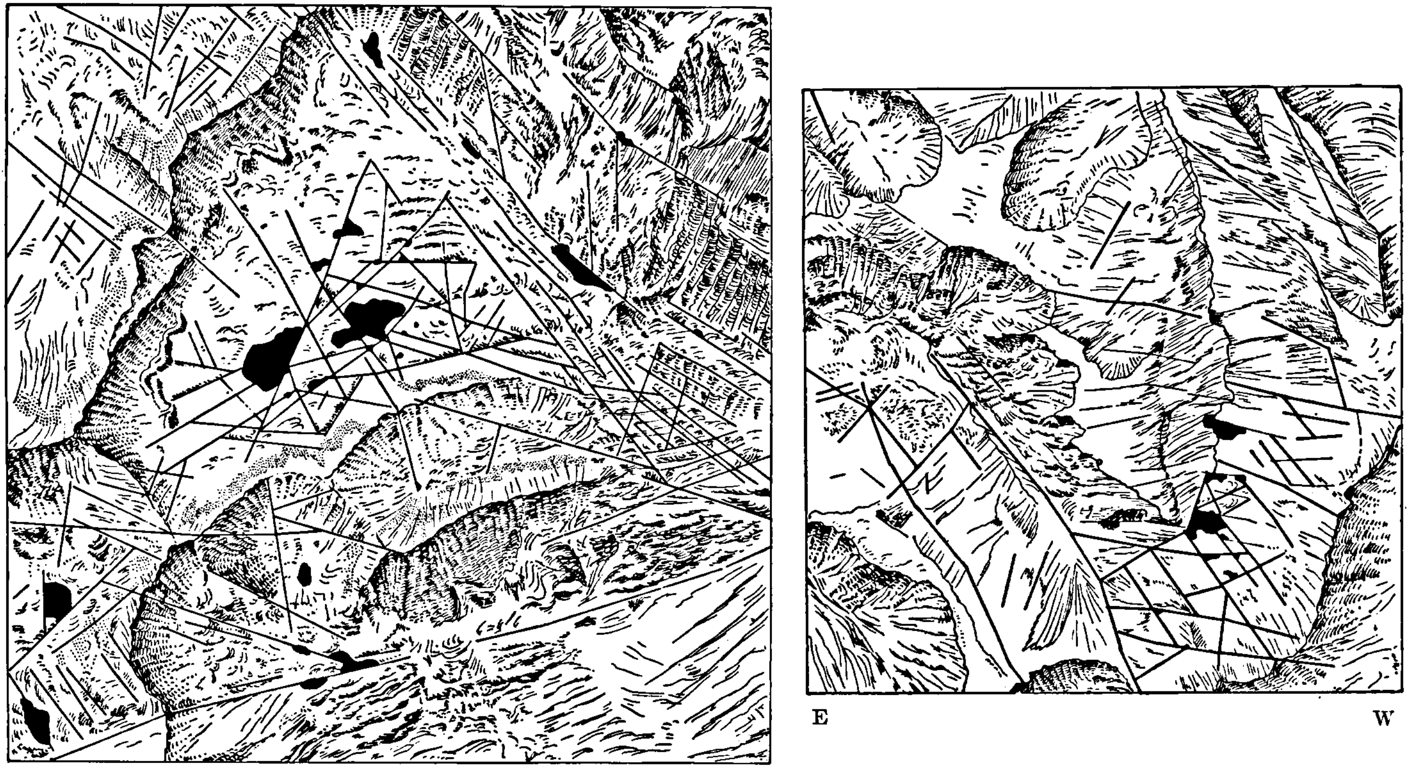


Fig. 43. Joint patterns. Left: granite area (Marimaña batholith); Right: sedimentary area (Cambro-Ordovician, upper Unarre valley). Scale ca 1: 40.000.

fore are certainly later than the main phase of the Hercynian orogeny and also later than the cleavage fan. As far as is known the Cretaceous cover does not show the same fracture pattern, — certainly not round Pobla de Segur — and therefore an Alpine age for the fracturing probably can be excluded. Consequently it seems to be an early post-Hercynian phenomenon, perhaps an elastic release feature connected with the pre-Triassic faulting. The irregularities of the pattern as a whole could be due to the inherent inhomogeneity of the Hercynian structure, i. e. granite massifs in an isoclinally folded sedimentary series of different rock types. On the other hand the relative regularity of small areas, of some 100 sq m, indicates that the stress field responsible for the fracturing was rather simple.

C. SUMMARY AND CONCLUSION

In the foregoing treatment of the tectonic features in the axial zone on sheet 5 little has been said about their possible mutual relation and original order of appearance. Since all these tectonic phenomena generally do not occur together in the same area, these relations are difficult to determine with any certainty. Moreover, the area under consideration is too small for the solution of some of these problems. A careful tectonic analysis of the northern and southern margins of the axial zone, especially also of the Stephanian beds, would be required for that purpose.

Let us recapitulate first the principal tectonic features which stand out in this chapter:

1. the Hercynian orogeny has led to strong cleavage-folding, the intensity of which increases towards the core of the axial zone and to a distinct disharmony between a supra- and an infra-structure, made possible by the plastic nature of the Silurian black shale facies during folding (p. 76)
2. the northern boundary of the Lower Palaeozoic dome is formed by a steep flexured zone (p. 63), immediately followed to the north by a more completely isoclinal Devonian with imperfect parallelism between axial planes and cleavage (p. 68 ff). This zone is moreover accompanied to the south by granite batholiths and a swarm of parallel porphyrite sills
3. in the sm-cleavage fan a sector of cleavage dips near the vertical is lacking (p. 81 ff)
4. the Upper Palaeozoic of the Monseny area (and further west in many localities between the Nogueras zone and the Maladeta granite) lies in recumbent folds, which originated to the north (p. 74 ff)
5. in the southern part of sheet 5 there is a late fracture-cleavage (sk-cleavage, knick-zones), which displaces the sm-cleavage in such a way as to indicate a dilation in a north-south direction rather than a compression (p. 83 ff)
6. there are folded areas, which, in their entirety plunge in a certain direction. They alternate with similar areas, with an opposite regional plunge. The various areas are not connected with each other via gradual plunge changes (p. 87 ff)
7. there is a late cross-folding, which is represented on a small scale by a fine ridging on the cleavage and bedding planes in directions not parallel with the Hercynian trends (p. 93)

8. both igneous and sedimentary rocks of Palaeozoic age are, especially in the higher parts of the chain, uniformly cut by numerous vertical joints, which are independent of the orientation of the cleavage planes (p. 94 ff).

The situation at the end of the main phase of the folding showed intensively folded Palaeozoic of which the axial planes — and therefore also the cleavage — stood almost vertical. It is at least, unlikely that in this phase the axial planes were oriented other than perpendicular to the compression or that the stress field itself should have had a different orientation. At a later stage the axial zone must have been slightly uplifted into an arch. In consequence of this arching the cleavage was disoriented, except in the centre where the lift was greatest (the cleavage stands perpendicularly on the arch). A stretching in the arch caused a further lowering from the vertical both northward (northern flank of the axial zone) and southward (southern flank).

It may be imagined that in the transitional zone (the central part where the cleavage was originally vertical) longitudinal blocks were slightly downfaulted more or less along the cleavage planes and at the same time tilted, one to the north, and the other to the south. In this central zone therefore areas of opposite cleavage dips occur side by side. Since this faulting apparently took place along the cleavage, the longitudinal boundaries of those blocks naturally show no change in the strike of the cleavage.

The fanning out of the cleavage in the axial zone is certainly Hercynian. This is demonstrated by the relation which exists between the orientations of the cleavage and the Mesozoic unconformity on the southern border of the axial zone. This unconformity dips about 50 to 60° southward on both sides of the Pallaresa. In the sm-cleavage section C (fig. 31) a striking phenomenon is that at about 4 km north of the unconformity the rather gently northward dipping cleavage suddenly changes into a similar one sloping southward. The same break can also be found in section B. If this unconformity plane is pushed back into its original horizontal position it appears that the cleavage corresponds exactly with the normal northward dipping cleavage of the southern flank of the axial zone. Thus the cleavage fan must have originated before the deposition of the Permo-Triassic.

At what moment in this final or post-main folding period the arching took place, is more difficult to decide. The area directly east of the Maladeta granite only shows a change in the strike of the cleavage (which need not have been caused by the Maladeta intrusion itself) and no change, in the direction of dip of the cleavage, not even in the wedge which the Tírvia-Espot syncline makes in the Maladeta granite. This may indicate that the cleavage fan is older than the emplacement of the batholith, unless the whole Maladeta granite also rotated southward. This latter opinion accounts for the many longitudinal faults of the Valle de Arán to the north of this batholith and the many marked longitudinal mylonite zones in the batholith itself. The recumbent folding of the zone south of the Maladeta granite — including the Monseny area — can then also be understood. This is also undoubtedly a Hercynian phenomenon, the straight border of the Permo-Triassic-Stephanian on the axial zone would seem to exclude an origin during or after the Alpine orogeny. There are indeed small post-Hercynian (Alpine) nappes known in the Nogueras zone. These, however, make an altogether different impression from the recumbent folding north of above-mentioned zone. In the northern part of the axial zone no gravity folding took place and the fanning was less marked and less regular here than in the south. Possibly this has some connection with the formation and tilting of the satellite massifs in this period.

It is highly probable that the origin of the steep northern flexure border has some connection with the granodiorite intrusion, as has been argued above. The presence of the swarm of porphyrite sills, which is restricted to this flexure zone and their chemical composition which is the same as that of the granodiorites suggest this. What the exact causal connection between the flexure border and the intrusion is, however, remains unsolved.

The knicking would appear to have resulted from the arching, for it has affected an already tilted sm-cleavage. It is unlikely that the sk-planes in an originally sub-horizontal position — a vertical cleavage — should have been rotated with the cleavage into their present position. The author seems a late-Hercynian origin as a direct result of the fanning-out more probable than an Alpine age. The influence of younger orogenies on the folded and consolidated axial zone consisted rather of a re-activation of pre-existing faults than such small-scale deformations.

Concerning the lineation units it has already been stated that a late, but still Hercynian cross-folding might explain these. Cross-folding is certainly present on a small scale. Nearly everywhere the cleavage planes show a fine wrinkling which strikes the attention chiefly in thin sections of the Infra-Ordovician slates. Sometimes small micro-folds of the sm-cleavage are also present. The axial planes of these folds have a different orientation from those which originated during the main phase of the Hercynian folding. It is, however, doubtful whether this cross-folding caused the coarser undulation represented by the lineation units. There are no further indications for this. The lineation units are, as stated above, not reflected in the distribution of the cleavage dips. This indicates that these phenomena are not related and have a different origin. The lineation units may be regarded as rotations on transverse axes while the cleavage fan and cleavage units represent rotations on longitudinal axes. A rotation on a longitudinal axis has little effect on the lineation plunges — and therefore on the axial plunges — because these are comparatively flat and lie on the steep cleavage planes, along which the movement took place. Conversely: rotation round transverse axes brings little change in the orientation of the cleavage, provided this is not gently-dipping. Which of these two rotations took place first, is unknown, but both are older than the regional jointing, for the latter is independent on both phenomena. The exact manner of origin of the lineation units is also for the present obscure. To solve this problem it would be necessary to survey an area several times larger than the present one.

As stated above, the regional jointing is a very late phenomenon, also nevertheless still Hercynian, for the Mesozoic nowhere shows this fracturing. Besides this late regional jointing an early local jointing may be distinguished. The latter was formed only in the granites and possibly in the contact-metamorphic aureoles during the emplacement of these batholiths. They resulted from partial cooling and further intrusion of the semi-congealed mass. The poorly consistent character of the regional pattern, however, makes it impossible to distinguish the local and regional patterns. Whether the regional jointing was caused by the same stress field as the cross-folding and whether the two phenomena are late manifestations of the main phase of the Hercynian folding or are the result of the second Hercynian (post-Stephanian) phase remains an unsolved problem. Fig. 44 gives a schematic representation of the relative ages of a number of tectonic processes as outlined in the above discussion.

Conclusion

It is highly probable that an arching of the axial zone took place in a late stage of the Hercynian orogeny and that a number of tectonic pecularities of the axial zone found their origin in this fact. The influence of the Alpine orogenies on the axial zone was slight and probably consisted only in a re-activation of pre-existing faults.

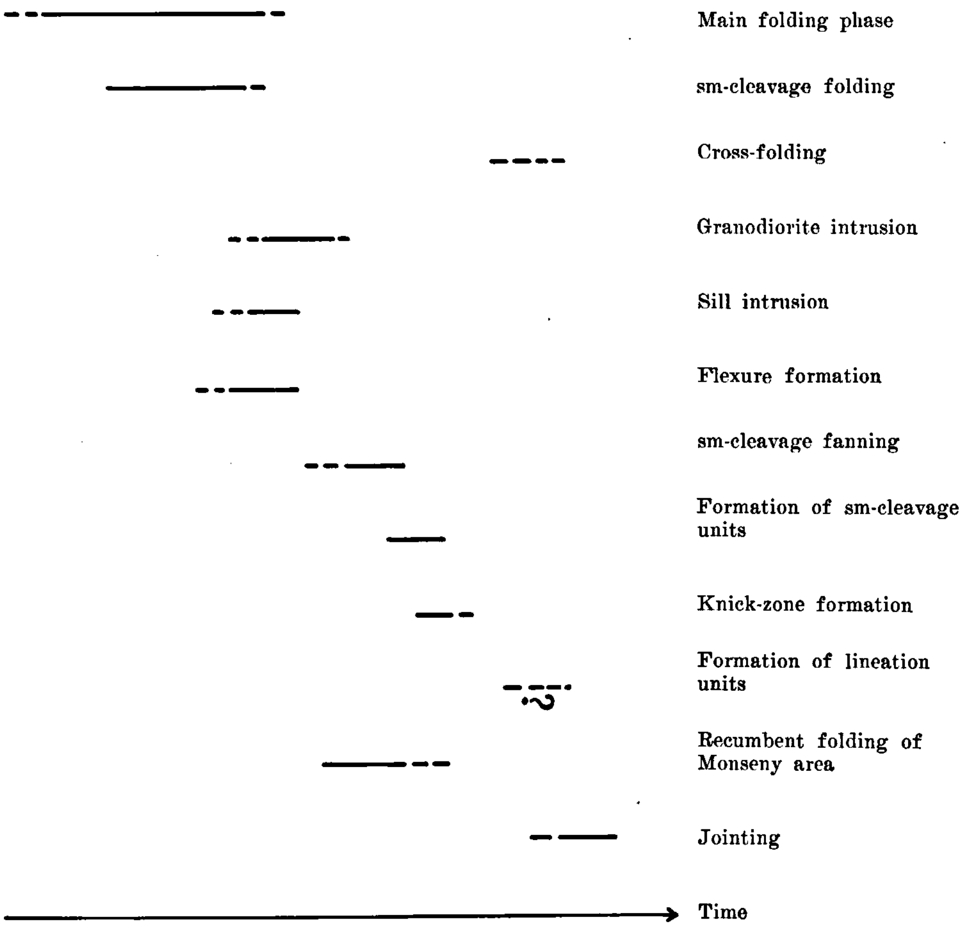


Fig. 44. Time relations of tectonic processes within the Hercynian orogeny.

CHAPTER IV

IGNEOUS ROCKS AND ORES

INTRODUCTION

The area of sheet 5 contains a considerable variety of igneous rocks as late-tectonic granodiorite batholith complexes, a number of dykes of various types, of which the group of concordant quartz-diorite-porphyrites is interesting from a tectonic point of view, and basic rock (ophite) intrusions of Lower Cretaceous age along an east-west branch of the North-Pyrenean fault system. The granodiorites alone among the igneous rocks have produced an appreciable thermal metamorphism in the host rocks. The author will give only a brief description of these rocks without going into pure petrological and mineralogical problems. More attention will be given to the relation between the emplacement of these igneous rock bodies and the folding.

The metamorphic rocks of the Aston massif, such as the mica-schists outcropping east of the Spanish boundary between Pic de Montcalm and the fault of Mérens in the core of the main dome are not discussed. (See, however, the map description of sheet 6, at present in preparation.) Only the western limit of the epi-metamorphic zone characterized by a somewhat coarser crystallinity and that of the post-kinematic biotite zone are indicated on the 1:50.000 map.

Finally, some occurrences of ore and talc will be briefly discussed.

A. IGNEOUS ROCKS AND THEIR THERMAL INFLUENCE

a. The biotite-granodiorite batholiths

Batholiths of biotite-granodiorite are common in the Hercynian belt of the Pyrenees. They are characterized by their homogeneity, by sharp and distinct contacts with the host rocks ("granite en massif circonscrit" according to Raguin, 1938) and a thermo-metamorphic aureole which occasionally reaches as high as the pre-Hercynian Carboniferous.

In the area in question some six batholiths are present either wholly or only partly, viz:

1. The Riberot batholith: ca 13 sq km outcropping in Devonian rocks.
2. The Marimaña batholith: ca 28 sq km in Cambro-Ordovician deposits and for the greater part in the metalliferous limestone, partly bordered by fault contacts.
3. The Salau stock: ca 1 sq km in the metalliferous limestone.
4. The Auzat-Bassiès batholith: ca 85 sq km in Cambro-Ordovician rocks and more or less flanked by the metalliferous limestone. Cut off by the North-Pyrenean fault in the north-east.

5. The Basiero and Saburo batholits: part of the large Maladeta batholith (nearly 400 sq km), in the area in question mainly situated in Devonian limestones.

6. The Santa Coloma batholith: ca 10 sq km in the extreme south-east of the map area.

These batholiths almost always show the above-mentioned characteristics. Only the apophyses in the north-west of the Auzat-Bassiès granodiorite in the Ossèse valley and the broad apophysis towards the south of the Marimaña granodiorite show a more diffuse transition into the surrounding sediments (the contact between the two types of rocks, although generally distinguishable in the field from a distance by a distinct difference in colour, is here difficult to map).

The intrusion of these granodiorite masses must have taken place in a late phase of the Hercynian orogeny. This is indicated by several different sources of evidence.

1. The batholith contacts are all discordant with the structural trends. The positions of the batholiths are not however altogether independant of the structural pattern (cf. the geological map). The southern border of the Auzat-Bassiès granodiorite for instance distinctly cuts off the Hercynian structure in many places, although the axis of the batholith diverges only slightly from the structural trends.

2. The batholith rocks are sheared only in a few apophyses, never in the batholith bodies proper. The sediments in the contact aureole and also fragments of host rock in the granodiorites generally show a normal cleavage.

3. The batholiths have caused extra folds around them by shouldering their way into the host rock (see p. 67).

The author has made a non-detailed study concerning the mineralogical composition of the Marimaña, Salau and Auzat-Bassiès granodiorites and their contact aureole.

The batholith rock — except for a comparatively narrow marginal zone — is very homogeneous and nearly always has an unoriented texture. Only in the two above-mentioned apophyses of the Marimaña and Auzat-Bassiès massifs parallel orientation of the biotite plates can be discerned. The rocks are generally medium-grained, leucocratic and fresh-looking. They consist of the following minerals, also recognizable in the field:

Plagioclase: sodic andesine (30—40 % An), hyp-idiomorphic, mostly oscillatory zoned, the more basic core of the crystals is generally a little altered to sericite (muscovite) and clinozoisite. In the hand specimen the plagioclase is white. It forms 25—50 % of the rock volume.
Potash-feldspar: mostly with microcline twinning, rarely idiomorphic, white to pink in the hand specimen. Occurs always in smaller quantities in the rock than the plagioclase, often in big crystals up to a few cm. Forms 15—20 % of the rock.
Quartz: irregular crystals with wavy extinction, sometimes cataclastic, forms 20—50 % of the rock.
Biotite: often altered to chlorites and ore, sometimes to muscovite, 5—15 %.

As accessories we find zircon and apatite. According to the classification of P. Niggli (1946) this rock can be named a granodiorite.

The table below shows two analyses of this rock type, one of the Salau granodiorite, and one of a more quartz-dioritic composition (with amphibole) of the Auzat-Bassiès batholith.

Towards the margins of the batholiths the plagioclase becomes on the average a little more basic (30—60 % An) and at the same time in greater quantities. The potash-feldspar remains present, as a rule, but decreases in quantity in the presence of limestone in the contact zone. The same is true of the biotite which in this case may even be absent. On the other hand it may form 25 % of the rock in a sandy-clay environment. Amphibole (cummingtonite and green hornblende) and pyroxene (diopside mostly) strongly increase in significance in limestone environments, as does titanite as an accessory mineral.

TABLE

Analysis no.	(92)	(93)	(94)	(95)
SiO_2	67.50	67.18	70.89	69.09
TiO_2	0.61	0.46	0.31	0.65
P_2O_5	0.22	0.21	0.27	0.21
Al_2O_3	14.44	14.74	14.70	14.82
Fe_2O_3	1.35	2.17	1.37	0.47
FeO	3.13	2.30	1.26	1.32
MnO	0.04	0.05	0.03	0.06
MgO	1.71	1.18	0.99	1.01
CaO	3.68	2.43	3.07	3.57
Na_2O	2.36	3.55	3.98	3.33
K_2O	4.00	2.88	2.36	4.46
H_2O-	0.79	1.82	0.67	0.99
H_2O+	0.12	0.29	0.11	0.19
CO_2	—	0.56	—	—
	99.95	99.82	100.01	100.17

92 Quartz-diorite, Auzat-Bassiès massif near Etang d'Alet.
93 Quartz-diorite-porphyrite, Salau.
94 „ „ „ , Val d'Ossèse.
95 Granite, Salau massif.

Analyst: Dr. C. M. de Sitter-Koomans.

Mineralogically the marginal zone is therefore of a more dioritic gabbroic composition. This zone is generally a little darker and is usually greener in colour on account of chloritized biotite (Auzat-Bassiès granodiorite) or of pyroxene and amphibole than the leucocratic central part of the pluton. Here and there in this marginal zone, (which greatly varies in breadth from nil to a few hundreds of metres) are found fragments of host rocks which are mostly small to very small and sharply bounded (generally near the contact) or have a somewhat diffuse contact and may or may not be oriented in situ. On the southern margin of the Marimaña granodiorite north of Lago Pudó occur a few large bodies of limestones and slates which may be considered as roof pendants in the granodiorite.

The broad granodiorite apophysis south-east of the lake mentioned is a fairly inhomogeneous rock with numerous strongly metamorphosed slate fragments in the same trend. These gradually pass over into the granodiorite, which in this whole apophysis is strongly sheared with conspicuous eyed structure. The eyes are mostly a few mm in diameter and consist of strongly undulous quartz, andesine or potash-feldspar, often occurring together in one eye and a fine-grained schistose matrix of quartz, sericite, clinozoisite-epidote, chlorites, ore and, occasionally, calcite in considerable amount. In the interior part of the batholith (e. g. even near the Etang d'Alet in the Auzat-Bassiès batholith) there sometimes appear rounded (up to a few dm in diameter) dark, fine-grained xenoliths, containing up to 50 % biotite and sometimes a little coarse amphibole (probably secondary). The feldspar is mostly andesine (40—45 % An).

The xenoliths are thus of a more quartz-dioritic composition than the leucocratic granodiorite.

b. The contact aureoles of the granodiorite batholiths

The batholiths have, on account of thermal influence, metamorphosed the enclosing rocks, in a zone varying greatly in breadth. The argillaceous rocks changed into hornfelses and spotted slates and the calcareous rocks into calcite marbles and lime-silicate rocks. As a general rule batholiths penetrating non-calcareous rocks show a considerably broader (up to 1½ km wide) recognizable contact aureole than those intruded into calcareous rocks.

In the area in question the dolomites have nowhere been affected by a distinct thermal contact-metamorphism.

In the Devonian area ENE of Montgarri, in an isolated area far outside the contact-aureole of the Riberot granodiorite, hornfelses and spotted slates occur. This is probably nevertheless an emergence in a southern direction of the thermal aureole of this batholith.

The marbles and lime-silicate rocks

Where pure limestones were present, as east of the Marimaña and round about the Salau batholiths no new products are present in the contact-aureole. The calcite is, however, completely recrystallized often into coarse crystals. The so-called "barrégiennes", a rapid alternation of limestone and less calcareous layers, are transformed near the contact into lime-silicate rocks. These rocks are especially well-exposed in the neighbourhood of the Estany Pudó south of the Marimaña massif, also in the valley of the B^{co} de Marimaña on the northern contact of this batholith. The following minerals are found there: albite, potash-feldspar (microcline), basic plagioclase (bytownite), clino-pyroxene ($+ 2\,V = 50$—$55°$), green hornblende, actinolite, wollastonite, red garnet (grossularite), idocrase (sometimes with small diopside inclusions), epidote-clinozoisite, zoisite, prehnite and titanite, of which especially diopside, idocrase, epidote-clinozoisite and titanite are frequently found in large crystals.

The rapid changes in the Ca-content of the barrégiennes layers together with metasomatism occasionally led to the formation of an alternation of thin monomineralic layers parallel with the stratification (epidote, idocrase, wollastonite or garnet). The metamorphic rocks containing feldspar, pyroxene, hornblende and titanite, are found nearest to the igneous rocks. The other minerals are found somewhat further out. Further than about 20 metres from the contact no new products in the barrégiennes can be distinguished by the naked eye.

The hornfelses and spotted slates

The transition of granodiorite into the argillaceous host rocks is usually sharp. In the well-cleaved slates we sometimes find lit-par-lit injection in a transition zone of not more than 3 dm width. These non-calcareous contact rocks with varying proportions of sand and clay are well-exposed around the Auzat-Bassiès granodiorite to which the following description is also applicable.

The argillaceous sediments have in the contact-aureole been altered to andalusite and cordierite hornfelses. The presence of these minerals is usually only recognizable by elliptic or elongate sericite masses with a little biotite and ore and occasionally some remnants of andalusite or cordierite. The andalusites occasionally have a rather fresh core, surrounded by a rim of sericite. The cordierite is always very much altered. The matrix is usually fine-grained with up to 50 % biotite (often with rutile needles), quartz and sometimes clinozoisite. Close to the contact the hornfelses contain some plagioclase (20—25 % An) and sometimes potash-feldspar in rather large crystals with numerous biotite inclusions or in veins and patches. Corundum was also found in the contact-aureole of the Auzat-Bassiès granodiorite, both on the north and south borders up to about 1 m from the contact in beautifully developed euhedral crystals up to 1½ mm in diameter. This mineral, also recognizable in the hand specimen, mostly occurs associated with andalusite. The cores of the corundum crystals are in some cases of a pronounced blue colour with a marked pleochroism from indigo-blue to purplish-violet or green, while in other crystals the coloured parts are irregularly distributed giving the mineral a patchy appearance. However, some crystals are colourless. The rims of the corundum crystals are always sericitized and the minerals are often set in a halo consisting of granoblastic quartz and feldspar but practically devoid of phyllo-silicates. Where Silurian black shales are present in the contact zone (east-north-east of Montgarri and north of the Basiero granodiorite) chiastolite occurs. Accessory tourmaline, zircon, apatite and ore minerals occur in these hornfelses, but titanite is seldom found. Large late muscovites have sometimes replaced these minerals.

When the sediment was extremely argillaceous the hornfelses in the hand specimen are also distinctly spotted and noduled. The nodules, oval to square in diameter and up to 1½ cm long, are satiny whitish-grey coloured sericite aggregates having a random orientation in the plane of the schistosity. Here and there the sericite aggregates are weathered out and long cavities are left. Andalusites are seldom recognizable in the sample.

In the contact zone the original cleavage has often been obliterated by the recrystallization, a proof for the post-kinematic age of the granodiorites.

The more arenaceous sediments contain little or no cordierite and andalusite, but much biotite between the recrystallized grains of quartz, which show a simple mosaic texture. Much feldspar also occurs in these rocks near the contact. Ore-minerals, sericite, muscovite, apatite and zircon are present as accessories.

The contact of the Auzat-Bassiès granodiorite west of the Certescâns lake shows some pecularities, viz: plumose mica (muscovite nests in a mosaic of quartz crystals) and a tourmaline-quartz rock with 40 % of brown tourmaline. Tourmaline also occurs here as radial aggregates on joint planes in the granodiorite and in the adjacent rock. These pneumatolytic modifications are, however, rare and restricted to the contact.

Real spotted slates (concentration of black particles in patches) occur more often on the outer border of the metamorphic aureole, although somewhat rare. The outer limit of the contact aureole has been drawn where hornfelses or spotted slates are no longer recognizable to the naked eye.

c. The hypabyssal rocks

The hypabyssal rocks of the area may be divided into several groups:

1. Dykes with the same composition as the granodiorite but more porphyritic in texture occur as far as 1000 m outside the batholith bodies, especially round those of Auzat-Bassiès and Salau. The granodiorite porphyrites which range in thickness up to 20 m, sometimes pass over into apophyses of the batholith, which have no porphyritic texture. The potash-feldspar is sometimes slightly rosy tinted. They are of the same age as the batholiths.

2. Aplitic dykes occur exclusively near or in the batholith bodies, they are much thinner than those mentioned above, being generally not more than a few dm wide and may sometimes be traced from the granodiorite into the country rock. These dykes are younger than the batholith intrusions.

3. A rather uniform group of sill-like intrusions is seen on either side of the steep northern border zone of the main dome (see geological map 1:50.000). Towards the west the sills continue in the same trend in the Valle de Arán (Kleinsmiede 1960).

Fig. 45. Quartz-diorite porphyrite sill in slates of upper part of Cambro-Ordovician. Ossèse valley.

The rock is very uniform in aspect, namely grey to greenish-grey, sometimes brownish on account of weathered pyrite, porphyritic with distinct phenocrysts of quartz and less distinct ones of biotite and feldspar. It is massive, jointed, occasionally schistose (north of the Marimaña massif and further to the west) and occasionally boudinaged (west of the Salau massif in the metalliferous limestone). They are conformable with the host rock. The difference in angle between cleavage and stratification is, however, too slight in the flanks of the folds — where these sills were exclusively found — to decide for certain whether they are cleavage or stratification sills. In places the sills transgress the bedding and cleavage planes over short distances.

Microscopically the rocks are porphyritic, occasionally non-porphyritic, unoriented, with a fine-grained matrix in which sericite and quartz predominate. As phenocrysts are found:

Plagioclase, andesine (mostly between 30 and 40 % An) sometimes oscillatory zoned, mostly altered to sericite-muscovite, clinozoisite-epidote and calcite. Sometimes the plagioclase is undefinable, the sericite and calcite then form together aggregates in the shape of a feldspar crystal. Up to 1½ mm in diameter.

Quartz, usually with wavy extinction, sometimes well-formed bipyramidal sections, often square with rounded corners, sometimes with highly irregular outlines, on account of magmatic corrosion. Up to 3 mm in diameter.

Biotite, rarely absent, frequently discoloured and altered into chlorite, sometimes with clinozoisite and ore, or muscovitized lamellae, often curved, sometimes with zircon. Up to 3 mm in diameter.

The matrix can generally not be determined on account of the intense seritization but where this has been less intense, plagioclase, calcite (as an alteration product of the plagioclase), quartz and in smaller degree biotite and ore, apatite and zircon may be recognized. The sills have therefore mineralogically a quartz-dioritic composition, they can be called quartz-diorite porphyrites. There is hardly a trace of a chilled margin.

The fact that these are sills, which generally have not been tectonically affected, indicates their origin at the time of a regional or a local tension. The chemical composition of batholith and sill rock is about the same (see analyses p. 105) which might indicate a similar magma and synchronous origin. As has been stated, the sills only occur in the vicinity of the flexure border of the main dome. This steep border is genetically connected with the batholith formation (p. 65, 101). The author's opinion therefore is that the porphyrite sills are synchronous with the origin of the flexure border and the intrusion of the granodiorite bodies and that these phenomena are genetically very closely related to each other.

4. Related to the above-described diorite-porphyrites there is a less frequent type of dyke which shares a few characteristics with the porphyrites, namely the conformable character, the high plagioclase percentage, little or no potash-feldspar and a distribution area which is situated within that of the porphyrites. A few occurrences are: in the Devonian of the upper Pallaresa 1 km east of the Port de Salau, in the western slope of the Cirque d'Anglade in Infra-Ordovician, near the mines of Carbauère in the Ossèse valley and 600 m east of the Port de Marterat in Infra-Ordovician.

Unlike the diorite-porphyrites, however, this type is never porphyritic, is less altered and darker in colour (lamprophyre) on account of a higher percentage of dark constituents, often among which are amphibole and rather more basic plagioclases. The ½ to 7 m thick dykes are generally massive, not sheared, dark greenish-grey (amphibole) or dark brownish-grey (biotite) in colour and rather fine-grained. Microscopically they consist of: quartz 5—30 %, plagioclase (andesine-labradorite) 45—60 % less altered to sericite and clinozoisite than in the porphyrites, biotite 5—35 % which may be discoloured and/or chloritized and light green amphibole 0—30 %. Dykes that contain amphibole have less biotite than dykes that do not contain amphibole, to such an extent that the percentage dark constituents is about the same in both cases, viz. 25—35 %. In the amphibole-containing dykes, however, the plagioclase is rather weathered. Only rarely is there a little potash-feldspar present. Mineralogically therefore these sills are quartz-diorites or quartz-gabbros with or without amphibole.

d. The ophite stocks

In the E—W faulted zone of Couflens-Arigail there are 8 stocks of a basic igneous rock, known in the Pyrenean literature as "ophites". It is a massive rock, medium to fine-grained and of a pale green to dark green spotted colour, which generally has a rather weathered appearance. Its resistance to erosion is usually no greater than that of the surrounding limestones and slates. The ophites therefore do not make a distinct relief-feature in the landscape. When fresh this rock consists of rather large pyroxene crystals which lie in a matrix of plagioclase and pyroxene. The plagioclase (basic andesine to labradorite) is lath-like and is not zoned. The pyroxene ($+ 2V =$ ca $50°$) sometimes shows diallage parting. Quartz and olivine are not found.

The rocks, however, is rarely as fresh as this. When it is weathered, the plagioclase has been altered into clinozoisite-epidote, or often, completely albitized. The pyroxene is either completely ouralitized, or only on the rims (tremolite or actinolite) or altered into clinozoisite. The pyroxene and/or its alteration products constitute about 60 to 70 % of the rock mass. Some hematite is often present, for instance in the ophite of the Guzet where it is locally even abundant, otherwise the rock of the various stocks is identical. The ophite therefore is a rock of a gabbro-dioritic to gabbroic composition.

The albitization of the ophite also extends to the host rock. In a contact-zone which may attain several metres in width the Devonian limestones contain perfectly fresh porphyroblasts and idioblasts of albite (mostly Carlsbad twins).

The opinion of most French geologists (cf. Viennot 1927, p. 23 and 45) that ophites are sheet-like intrusions in the Keuper is not supported in this area. The ophite occurrences are here clearly genetically associated with the North-Pyrenean fault system, the main movement of which took place during the pre-Cenomanian folding (p. 77 f). The ophite intrusion must therefore have originated during this Laramide fault movement.

B. ORE AND TALC DEPOSITS

a. The galena and sphalerite occurrence of Carbauère (Ossèse valley)

At many points in the Lower Palaeozoic along the northern boundary of the Auzat-Bassiès granodiorite, are found lead, zinc, iron and copper ores, mostly in a limestone or dolomite environment. Among these chiefly sulphidic ores galena (PbS) and sphalerite (ZnS) occur most frequently. A few of the more important occurrences have been indicated on the map, i. e. those of Carbauère (left bank of the Escorce river), Bocard (Escorce), le Pas d'Enfer (Ars), Castel Minier (Garbet), Lacore and Pic des Argentières (east of the Garbet). A few of these were already known in Roman times and were exploited. The vein-like occurrences, however, are irregular, discontinuous and contain many times more gangue than ore therefore being difficult to exploit with profit. The mines have therefore long since been abandoned. Recently, however, (in 1957) renewed prospecting has taken place in the Carbauère area.

As far back as 1864/65 and again later in the early years of this century exploitation on a small scale of silver-containing galena and sphalerite took place here. The ore produced was taken by means of a funicular to Estillon, where it underwent an initial process. The ore was obtained from six horizontal galleries — one above the other — between 1300—1750 m altitude. The exploitation was not carried on very far, however, the longest gallery extending only 30—35 metres.

In a breccia zone of limestone, slates and veins of quartz, galena and sphalerite occur chiefly on the borders, sometimes combined with pyrite, chalcopyrite, siderite and hemimorphite (rare). The centre of this area of 1—3½ m breadth is often sterile. The ore is found in irregular layers, at most 2 to 3 cm thick, surrounded by much gangue and is intimately intergrown.

15 to 20 m north of this zone there is another 15—40 cm thick productive layer, which higher up the slope joins the former. This vein has also been worked in some places. Mussy (1869) mentions yet another vein, 120 m further north. This author gives a partial analysis of each of the veins, which he considers as representative:

southern vein	Pb 39 %	Ag: 0.00065 %
middle „	Pb 57.80 %	Ag: 0.00080 %
northern „	Pb 68.40 %	Ag: 0.00040 %

The ore always contains a trace of gold.

Traces of this mineralization are also found on the western slope of the Pic de Carbauère. The veins, at least the southern ones, are more or less discordant. At an altitude of about 1400 m this vein is bordered to the north by Silurian pelites and fragments of black shales are found in the gangue. Higher up, a steadily greater thickness of calcareous slates lie between the Silurian and the vein. The mineralized breccia zone of the Carbauère is in fact a fault zone, for lower downward the fault cuts out the Silurian. As often occurs in the Pyrenean Palaeozoic (Upper Lez, Pic de Maubermé) the mineralization depends on a fault.

Near to the entrance of the mine at an altitude of 1650 m, rock fragments presumed to originate from the mine consist of 70—90 % of albite and the rest of ore. This albitite, however, could not be found in situ.

The mineralization has probably been caused by hydrothermal aftereffects of the granodiorite intrusion. The fault is probably connected with the flexured northern border of the main dome. The NW—SE cross faults, which also cross the granite and are certainly younger than the flexure, show no traces of mineralization.

Whether the other occurrences of ore, here mentioned, are connected with faults, the author does not know. It is, however, quite possible since all these instances occur in the flexured northern border zone of the main dome, where longitudinal faults may be expected.

The longitudinal faults of the Campans are accompanied by thick but sterile quartz veins. The cross faults immediately north of these have, in a few places, a distinct fault breccia-mylonite, and have been mineralized on a small scale (sphalerite especially being present).

The Carboniferous south and south-east of Tírvia shows many unconformable quartz veins in which there occurs much siderite, a bright green chlorite, and sometimes some talc.

b. The talc occurrence of the Fonta (Ossèse valley)

A steep cliff, partly consisting of talc (see fig. 46) occurs on the left bank of the Fonta (Ossèse valley). It is situated to the west of the Auzat-Bassiès granodiorite, at an altitude of about 1800 m, 40 m above the river. This is the only talc occurrence in the area. It is difficult to examine the outcrop, as only a small portion is directly accessible. The talc scree underneath indicates that a large part of the wall must consist of this mineral. The talc occurrence is on all sides surrounded by dolomite from the upper part of the Cambro-Ordovician, and within the talc mass there occur irregular

Fig. 46. The talc cliff cut by numerous quartz veins in the upper Ossèse valley.

dolomite patches. The outcrop can be recognized from a distance by the thick veins of quartz running criss-cross through the talc. Occasionally they enclose large calcite masses, in other places talc is only found in between the quartz. In the hand specimen the talc often has the same texture as the dolomite and the thin bluish-black irregular layers, so characteristic for dolomite, normally continue in the talc. The pyrite, which occurs in the dolomite in crystals up to 5 cm in diameter, similarly occurs in the talc.

The change of dolomite to talc is very gradual over a few dm. First an occasional talc lamina occurs among the dolomite crystals, then some talc laminae join into patches. These patches increase in number until the whole mass of rock finally consists of talc crystals. This condition is, however, seldom reached in this occurrence, generally the talc is polluted by a large number of carbonate crystals, chlorite and opaque matter in which the chlorite occurs exclusively together with the talc. The colour is consequently mostly grey to dark grey and even greenish-black.

No basic dykes occur either in the talc mass or nearby. Nor are pegmatites, which Zwart (1954 p. 172) mentions as ocurring in the talc of Trimouns, present here.

In the extension of the talc cliff there is a vertical fault, separating the dolomite from the slates, i. e. hornfelses. Along this fault, probably only a vertical movement took place with a downthrow to the west. On the face of these data it seems probable that this talc, like that of Trimouns, came into being by metasomatic addition of large quantities of hydrothermal quartz. This quartz in watery solution has its origin in the granodiorite intrusion, which is situated a few hundred metres to the east. The watery solution could easily reach the dolomite along the fault plane and convert it into talc according to the following reaction (Zwart).

$$\underset{\text{dolomite}}{3\,CaMg(CO_3)_2} + \underset{\text{quartz}}{4\,SiO_2} + H_2O \rightarrow \underset{\text{talc}}{Mg_3Si_4O_{10}(OH)_2} + \underset{\text{calcite}}{3\,CaCO_3} + 3\,CO_2 \uparrow$$

The generated calcite is now found in large quantities associated with the quartz. Apart from the talc occurrence only dolomite and no limestone is to be found in a wide surrounding area (see geological map). Whether the talc body is of large dimensions cannot be ascertained on account of the two-dimensional character of the outcrop. The fact that talc is found along a fault does not make it probable that the quartz and therefore the talc metasomatism has spread far beyond the fault. Why talc does not occur along the other faults is not clear, since there is apparently sufficient quartz, although not quite so much as in the talc wall.

In former years the talc was exploited and was transported down by mules.

SUMARIO

El relieve

El relieve actual de la región del Salat-Pallaresa tiene rasgos Hercinianos, glaciarios y post-glaciarios. Fuera de la estructura y naturaleza generales de las rocas, se evidencia la influencia Herciniana en la presencia de un sistema bien desarrollado de grandes diaclasas que ha determinado enteramente o en parte el curso de muchos ríos pequeños y algunos grandes (Hillette, Turguilla, Arce, Cardós, Montalto, Vallfarrera). Este sistema de diaclasas ha influído también considerablemente en la formación de muchos lagos glaciarios (dib. 43) y en la formación de algunos puertos importantes.

En la época inmediatamente anterior a la glaciación, partes de la cordillera fueron denudadas y luego levantadas en unos ciclos morfogenéticos. Se han preservado aún — sobre todo al lado español — restos de estos viejos niveles de denudación en la forma de mesetas subhorizontales, cimas y crestas truncadas y terrazas de erosión situadas a grandes alturas. Pueden distinguirse dos niveles importantes, a saber, uno situado ea 2400—2600 m y uno entre 1850—2350 m. La altura media de estos niveles decrece algo hacia el oeste. Existen superficies de aplanamiento a alturas inferiores de los 1800 m, pero están poco extendidas (ver mapa separado de 1:100.000). La existencia de un nivel de crestas como nivel de denudación de más edad de los aún existentes en la región estudiada, es, cuando menos, dudosa. Algunas de las cimas más altas son truncadas por declives usados en ángulo de 15—35°, unos hacia el norte, otros hacia el sur. Estos declives han de ser de la misma edad que la del nivel situado a una altura de 2400—2600 m, en el que a veces se continúan. Si acaso hubiese existido tal nivel de crestas, hubiera estado a una altura mayor que la cresta que constituye la divisoria principal de las aguas que se encuentra a 2700—3141 m, y debe haber desaparecido hace mucho. Para el establecimiento de la edad de los niveles de denudación, la región estudiada no ofreció nuevos datos.

En la región del Salat-Pallaresa se produjo por lo menos una glaciación (ver mapa de 1:100.000). Nuestros resultados están muy conformes con los de Nussbaum (1935). De los siete glaciares de esta región, la del Pallaresa fué la más larga (51 Km), pese al hecho de que este valle fué captado por el Garona ya antes de la última glaciación. Los glaciares del Bonaigua y del Arreu impidieron, sin embargo, el derrame del hielo en el curso superior, que, a causa del estancamiento del hielo y aunque éste se encontraba en altas acumulaciones, presenta poca erosión glaciaria. Por el Puerto de Beret y el Puerto de la Bonaigua, el glaciar de Pallaresa estaba en comunicación con el glaciar del Garona. Este fué también el caso con los glaciares del Alet y del Garbet por la vía de la Trape y del Plateau de Géou.

La posición del límite de las nieves perpetuas de la última glaciación pudo determinarse en muchas partes del nivel inferior de nichos de nivación. Resulta de ello, que el límite de las nieves perpetuas ascendió de 1500 a

1600 m en el norte de la región hasta de 2100 a 2200 m en el sur (dib. 5) y que el límite de las nieves perpetuas descendía generalmente algo más en las pendientes septentrionales y orientales que en las pendientes meridionales y occidentales, comparables con aquéllas. Aquéllas han experimentado, por consiguiente, una glaciación mucho más fuerte. A pesar de esta marcada tendencia de descenso del límite de las nieves perpetuas hacia el norte, los glaciares de la parte francesa han sido de muy poca extensión. La fuerte inclinación de fondo de los valles y la pequeña superficie que había encima del límite de las nieves perpetuas de la última glaciación, a causa de la escasa anchura de los Pirineos franceses en esta parte, impidieron una acumulación considerable de hielo. Como glaciarias tardías han de considerarse también las acumulaciones de deyecciones semicirculares hasta de forma de V de un radio de 50—300 m, que se encuentran en muchas partes a alturas superiores a los 2100 m (epiglaciares). Terrazas de acumulación se hallan en los valles glaciarios de la parte española en las pendientes hasta a 600—700 m encima del fondo de dichos valles. Las deyecciones eráticas que se encuentran contra el declive de las terrazas entre otros materiales, indican que la formación se produjo durante la glaciación. Son, en efecto, una clase de terrazas de "kame", pero éstas están compuestas de material fluvio-glaciario que procede del declive que está directamente encima.

La erosión postglaciaria dió principalmente lugar a una obliteración de características de relieve de más edad; así se produjeron en los fondos y los escalones de valles glaciarios abruptas incisiones y se colmaron lagos situados en sitios bajos.

Estratigrafía del Paleozoico

Las rocas Paleozoicas pueden distinguirse de rocas de menos edad, por un marcado clivaje tectónico. A causa de la casi total ausencia de fósiles, no es posible establecer una estratigrafía detallada de tiempo. Sólo el Gothlandiense se caracteriza bastante bien por graptolitas y, por tanto, sirve de capa guía para esta región en que se ha basado la entera correlación de la región en cuestión. La consecuencia es, sin embargo, que las unidades empleadas no son unidades puras de estratografía de tiempo, sino de estratografía de rocas, cuyos límites probablemente no coinciden perfectamente con las unidades generales de estratografía de tiempo. Sin embargo, aquí se emplea la terminología de estas últimas. No se encontraron indicios de discontinuidad o discordancias en la serie Paleozoica. Su espesor en el núcleo de la zona axial es de por lo menos 4500 m, sin que se pueda hablar de un marcado desarrollo geosinclinal.

Las rocas de más edad de la región estudiada pertenecen al Cambro-Ordoviciano. Este Cambro-Ordoviciano puede dividirse en tres. La serie de más edad — la de Lleret-Bayau — que en el centro del domo principal aflora en una estructura anticlinal, es una sucesión de calizas grises, cuarzitas, pizarras y un estrato característico, de color oscuro, duro y rico en hierro. Es probablemente el equivalente de la "série de Canaveilles" de Cavet (1951), que se encuentra en los Pirineos Orientales. La serie de Pilas-Estats que sigue a ésta es una sucesión que no contiene caliza y de un espesor de por lo menos 2000 m, de facies nerítica, y que está compuesta de pizarras, areniscas, cuarcitas y micro-conglomerados con alternaciones delgadas tanto en sentido horizontal como vertical. (Cotéjese el mapa de 1:50.000). En la parte superior del Cambro-Ordoviciano se presentan conglomerados, calizas, dolomías y pizarras calcáreas alternados con pizarras oscuras, pobres en arena (dib. 15). El conglomerado que contiene guijarros de cuarzo y cuarcita, se ve en todas partes de los Pirineos, al contrario del horizonte de caliza blanca, compacta y cristalima y dolomía de color pardo claro, de origen biostroma, que no se encuentra sino en la región limítrofe del norte de la región axial, entre el Garona y el macizo de St. Barthelemy (dib. 14). Este "calcaire métallifère" es, sin duda, de más edad que las pizarras calcáreas y las grauvaces que se encuentran en muchas partes de los Pirineos centrales y orientales y que datan del Caradociano. El equivalente de esta última serie de rocas se halla probablemente en las pizarras calcáreas grises que se presentan en casi todas las partes de la región de que se ha trazado el mapa, en niveles inmediatamente inferiores al Gothlandiense. A esta serie de calizas calcáreas siguen lutitas, de color negro y ricas en carbón, hierro y aluminio, de un espesor de 30 a 200 m y que presenta la facies de lutita negra. En el valle del Ossèse y más hacia el este, se encuentra una capa de cuarcita delgada, que presenta amenudo en su partè superior lentes delgados de caliza de distintos tipos. El Devónico de la zona axial es bastante uniforme, al contrario del de las zonas marginales. Según la litología, puede dividírselo en esta región en dos series. La serie inferior, al norte del domo principal, se compone de calizas de color gris azulado, pizarras calcáreas y pizarras oscuras no-calcáreas. Hacia el oeste, las pizarras se presentan un poco más arenosas. En la región del Salat-Alet esta serie inferior está compuesta de cinco a seis capas de caliza y pizarra, cuyos espesores varían de 30 a 150 m, sumando a un espesor total de unos 1000 m. Hacia el norte el espesor va decreciendo rápidamente. En el sinclinal de Tírvia-Espot, la serie inferior contiene más caliza y es mucho más delgada (300 m en la zona del Cardós), excepto hacia el oeste, donde se halla un desarrollo semejante al del norte. A éste sigue una serie superior de calizas nodulares abigarradas y pizarras. Esta facies de calizas nodulares sólo se presenta en su forma típica en el sinclinal de Tírvia-Espot, al este del Pallaresa, en la zona fallada de Couflens-Arigail; pero es amenudo muy difícil de reconocer. El espesor de esta serie superior es, en la zona del Cardós, de 30 a 40 m, en la región del Salat-Alet, en cambio, de unos 400 m. En dicho sinclinal, la serie superior de aspecto abigarrado y que no contiene liditas ni concreciones de fosfato, cambia gradualmente en una sucesión de pizarras monótonas, no calcáreas, de color gris oscuro hasta negro que contienen mica y datan del Carbonífero, y cuyo espesor es de por lo menos 500 m. Estas pizarras van presentándose, al oeste del Cardós más arenosas. Cerca de Espot, están intercalades en estas pizarras arenosas grauvaces y areniscas de color grisverde hasta negro y que contienen mica. Este Carbonífero es, sin duda, de más edad que la fase principal del plegamiento Herciniano y, por tanto, también de más edad que el Westfaliense Superior de Aguiró.

Estructura

La orogenia Herciniana ha dado lugar a un plegamiento de clivaje que produjo un acortamiento pronunciado, cuya intensidad aun aumenta hacia el núcleo de la zona axial, y asimismo a una clara desarmonía entre la supra-estructura y la infra-estructura, a causa de una capa delgada y plástica del Gothlandiense que se encuentra entre las dos estructuras. La infra-estructura es muy complicada y está formada de anticlinorios y sinclinorios muy complejos. Los buzamientos regionales de los pliegues mayores son relativamente planos; al contrario en los pliegues secundarios se presenta la estratificación siempre muy inclinada. El acortamiento es principalmente el resultado de fuertes micro-plegamientos isoclinales cuyas amplitudes generalmente ascienden a un múltiplo de la mitad de las longitudes de onda. Este micro-plegamiento, la mitad de cuyas longitudes de onda son del orden de $n.10^{-1}$ — $n.10^{-2}$ m, es un verdadero plegamiento de clivaje. Los pliegues isoclinales del orden de $n.10^2$ m de la supra-estructura no se hallan amenudo en el Infra-Paleozoico. Al revés, el micro-plegamiento es relativamente raro en la supra-estructura. Allí el acortamiento es efectuado por el plegamiento isoclinal de dimensiones mayores, que ce ha mencionado arriba; a éste se añade un acortamiento adicional causado por un proceso de adelgazamiento y distención de las calizas durante la compresión. Aunque hay, pues, un número muy grande de pliegues de distintas dimensiones en los afloramientos del Cambro-Ordoviciano, sus delineaciones según el eje b son relativamente constantes por grandes extensiones. Entre las distintas unidades de delineación constante se encuentran hundimientos opuestos. No se ha comprobado la existencia de transiciones en las inclinaciones de tales humidientos; la distribución regional de delineación no depende de la de las pendientes de clivaje (cotéjese dib. 37 con dib. 30). Puede ser que las unidades de delineación sean causadas por un plegamiento cruzado tardío o que constituyen el resultado de la adapción a la elongación producida por el plegamiento en la dirección del eje b de la zona axial.

El Paleozoico Superior de la región de Monseny presenta pliegues volcados, que indican que su origen se encuentra en el norte. Este plegamiento causado por la gravedad guarda probablemente relación con el surgimiento del granito de Maladeta y es, por tanto, aún de origen Herciniano. Una formación Alpina es improbable, ya que el límite del Permo-Triásico-Estefaniense en la zona axial es perfectamente recto.

El límite norte del domo central del Infra-Paleozoico lo constituye una zona de flexura abrupta, que da un rechazo vertical de al menos 2000 m. A lo largo de los dos lados de esta zona de flexura se encuentran numerosas capas intrusivas de pórfido, y del alto lado sur además un borde angosto de granito. Hay una relación genética entre estos tres fenómenos.

De menos edad que la fase principal del plegamiento Herciniano son el abanico de clivaje, las diaclasas regionales y el "knicked cleavage" (knickzones). El primer fenómeno puede explicarse mediante un arqueo de la zona axial. Es, sin duda, pre-Alpino; lo indica la relación entre la posición del plano de discordancia que se puso muy inclinado durante el plegamiento Alpino, y el cambio de la dirección del clivaje a que dió lugar (dib. 31). El abanico de clivaje es de menos edad que las diaclasas regionales, que pueden verse en las fotos aéreas, cortan igualmente el granito y los sedimentos y están verticales en todas partes. Estas diaclasas no se conocen en el Mesozoico. Las "knickzones" se limitan a la parte sur de la región y consisten de un clivaje de fractura pareada, con inclinación hacia el sur; entre los pares de clivaje de fractura el clivaje primario ha experimentado un movimiento de rotación, de tal forma que o se ha bajado la parte sur de la "knickzone" con relación a la parte norte, o al revés. Fué posibilitado este movimiento por la existencia de un clivaje volcado: las "knickzones" son, por tanto, probablemente de una edad alge menor que el abanico de clivaje, o de la misma edad. El mecanismo de aquéllas indica una dilatación on dirección N—S, y no una compresión.

La zona fallada de Couflens-Arigail, en que afloran unas ocho intrusiones de ofita, es un ramo E—W del llamado "Sistema de fallas de los Pirineos Septentrionales", que se produjeron durante el plegamiento pre-Cenomaniense. A consecuencia de ello, los dos bloques de fractura se volcaron hacia el norte, tal como sucedió en los macizos Paleozoicos satélites. Un bloque de caliza Urgo-Aptiense cerca de Couflens es probablemente la parte inferior de un sinclinal comprimido, que se ha conservado debido al vuelco a lo largo de las fallas (dib. 29b).

Rocas ígneas y minerales

Algunos batolitos de granodiorita tectónicos tardíos hasta post-tectónicos, afloran en el borde septentrional del domo central contra la zona de flexura muy inclinada. Son caracterizados por su homogeneidad, el contacto claro y nítido con las rocas circundantes y una aureola térmica, que se extiende localmente hasta muy entrado el Carbonífero pre-Herciniano. Las rocas son de grano mediano, leucocratas y compuestas de 25—50 % de plagioclasa (30—40 % An), 15—20 % de feldespato potásico, 20—50 % de cuarzo y 5—15 % de biotita. La zona marginal, de color algo más oscuro, es de una composición más diorítica hasta gabbroíde y contiene anfíbole y piroxeno en un ambiente calcáreo, y mucha biotita en un ambiente arcilloso. Estas intrusivas han ejercido una fuerte influencia termal en las rocas circundantes. Rocas arcillosas se han convertido en rocas córneas andalucitas y cordieritas, ricas en biotita, que, cerca del contacto contienen también corindón y feldespato. Andalucita carbonácea (chiastolita) se encuentra en pelitas Gothlandienses metamórficas de contacto. Las rocas ricas en calcita se han convertido en mármoles de calcita y rocas de silicato de cal. Cerca del granito se encuentran: albita, feldespato potásico, bitownita, diópsida, hornablenda verde, actinolita y titanita. Algo más apartado del granito: wollastonita, grossularita, vesubiana, epidota-clinozoisita, zoisita y prehnita, formando los primeros cuatro de estos minerales en "barrégiennes" amenudo capas monominerales. Modificaciones neumáticas, tales como cuarzo con turmalina y moscovita "palmé" se encontraron al sur del granito de Auzat-Bassiès.

Aplita discordante y diques de granodiorita se hallan en las zonas de contacto de los batolitos; estas aplitas son de menos edad que las intrusiones graníticas. Un grupo de capas intrusivas de cuarzo-diorita-pórfido con fenocristales de cuarzo, biotita y plagioclasa fuertemente alterada (30—40 % An) se encuentra a ambos lados de la zona marginal, muy inclinada, del norte del dome central. Están relacionados con esta flexura, como también a las intrusiones de granodiorita, con las que tienen en común su composición química (p. 105).

En la zona fallada de Couflens-Arigail se encuentran ocho intrusiones de ofita. Estas rocas de tipo gabbro-diorítico hasta diorítico se componen de grandes cristales de piroxeno ($+ 2\,V = \pm 50°$, amenudo con rotura dialaga) en una matriz de plagioclasa (40—60 % An) y piroxeno. El porcentaje de piroxeno es de 60—70. No se encuentran cuarzo ni olivino. La plagioclasa se ha convertido, amenudo, en albita, al igual que las calizas Devónicas en la inmediata proximidad de las intrusiones de ofita. Estas están genéticamente relacionadas con el "Sistema de fallas de los Pirineos Septentrionales", que obtuvo su configuración actual durante el plegamiento Laramide.

En el "calcaire métallifère" o sus inmediaciones al este del Ossèse se hallan en muchas partes algunos minerales, principalmente galenita y esfalerita en filones con mucha ganga y poco mineral. La mineralización de Carbauère está relacionada con una falla longitudinal y fué efectuada por la influencia hidrotermal de la intrusión granítica. Talco se halla en el valle del Ossèse (Fonta); éste fué formado de dolomía por adición metasomática de grandes cantidades de cuarzo hidrotermal. La calcita que así se produje, se encuentra en la masa de talco entre los filones de cuarzo.

In order to facilitate map-reading a list of all the topographic names — present in the sheet 5 area and used in this paper — is given together with their grid references or geographical coordinates. References in the French part and on the border crest are based on the Lambert grid used on the French 1 : 20.000 maps. Coordinate references to rivers and granites relate to the position of the names of these on the map.

(c) = col
(d) = depression
(g) = granite
(l) = lake
(mi) = mine
(m) = mountain
(p) = plateau
(v) = village

Agneserre (d)	521 G 8 — 52 G 9
Alet (l)	518 G 4 — 49 G 6
Alet (r)	512 G 8 — 53 G 8
Alins (v)	5°00′20″ — 42°32′50″
Alós de Isil (v)	4°47′10″ — 42°42′00″
Anglade (Cirque d') (d)	507 G 0 — 48 G 0
Anglade (r)	507 G 1 — 47 G 5
Angouls (r)	504 G 0 — 53 G 7
Angouls (v)	505 G 0 — 54 G 3
Anterrius (Pleta de) (p)	4°54′25″ — 42°38′20″
Arahós (v)	4°56′50″ — 42°32′05″
Aréau (l)	501 G 5 — 53 G 1
Areo (v)	5°00′50″ — 42°35′15″
Argentières (mi)	522 G 8 — 53 G 4
Argulls (m)	4°40′45″ — 42°41′00″
Arigail (v)	512 G 6 — 54 G 1
Arreu (r)	4°44′45″ — 42°40′10″
Ars (r)	520 G 5 — 51 G 4
Artigue (r) (tributary of the Ariège)	522 G 4 — 46 G 2
Artigue (r) (tributary of the Salat)	501 G 5 — 56 G 1
Aubières (m)	502 G 2 — 53 G 5
Aula (c)	499 G 8 — 52 G 2
Aulus (v)	518 G 5 — 54 G 9
Auzat-Bassiès (g)	519 G 0 — 49 G 0
Aydí (v)	4°53′05″ — 42°31′20″
Ayneto (v)	4°56′05″ — 42°38′15″
Baborte (l)	5°03′05″ — 42°38′35″
Barbaña (d)	4°46′55″ — 42°38′35″
Basibé (l)	4°41′25″ — 42°42′10″
Basiero (g)	4°40′30″ — 42°37′10″
Bayau (l)	5°07′10″ — 42°35′45″
Bedó (r)	5°54′10″ — 42°42′00″
Berasty (r)	4°44′20″ — 42°30′30″
Berrós Josa (v)	4°50′25″ — 42°34′25″
Bielle (r)	511 G 4 — 55 G 0
Bielle (v)	512 G 0 — 56 G 1
Bleu (l)	521 G 9 — 49 G 4
Bocard (mi)	514 G 2 — 51 G 0
Boet (c)	5°06′45″ — 42°37′05″
Bohavi (Pleta de) (d)	5°00′50″ — 42°40′50″
Boldís (r)	4°59′30″ — 42°38′35″
Boldís (v)	4°57′30″ — 42°36′45″
Bonabé	4°43′45″ — 42°45′25″
Bonaigua (c)	4°40′20″ — 42°39′40″
Bonaigua (r)	4°45′00″ — 42°38′30″
Bor(r)en (v)	4°46′10″ — 42°39′30″
Brohate (m)	520 G 8 — 44 G 1
Brohate (r)	5°02′50″ — 42°40′45″
Burch (v)	4°57′30″ — 42°30′15″
Cabaña (r)	4°43′50″ — 42°37′45″
Cabrís (c)	5°26′25″ — 42°32′45″
Cagateille (Cirque de) (d)	515 G 1 — 49 G 8
Campans (m)	4°46′50″ — 42°43′00″
Campirme (m)	4°52′50″ — 42°39′30″
Carbauère (mi)	513 G 0 — 50 G 5
Cardós (r)	4°55′00″ — 42°33′30″
Caregue (r)	4°47′30″ — 42°29′40″
Casibrós (v)	4°55′10″ — 42°34′25″
Castel-Minier (mi)	521 G 4 — 53 G 5
Cerdá (m)	517 G 1 — 50 G 7
Certescáns (l)	4°59′20″ — 42°42′30″
Certescáns (r)	5°00′10″ — 42°41′40″
Cireres (r)	4°45′20″ — 42°44′00″
Couflens (v)	506 G 0 — 54 G 5
Cougnets (r)	506 G 6 — 50 G 5
Courret des Etangs (d)	500 G 9 — 53 G 0
Cruzous (d)	514 G 7 — 55 G 3
Cuenca (m)	4°44′30″ — 42°41′55″
Encantats (m)	4°42′10″ — 42°34′05″
Escalarre (v)	4°49′40″ — 42°36′55″
Escaló (v)	4°50′35″ — 42°32′50″
Escorce (r)	513 G 8 — 51 G 4
Escots (c)	516 G 1 — 52 G 2
Espós (Coma de) (m)	4°41′35″ — 42°30′25″
Espot (v)	4°46′30″ — 42°34′30″
Estahís (r)	4°48′05″ — 42°34′25″
Estahón (r)	4°53′50″ — 42°36′00″

Estangento (l)	4°41′25″ — 42°30′35″
Estangento (r)	4°41′05″ — 42°30′15″
Estarón (v)	4°52′40″ — 42°31′45″
Estats (l)	5°04′30″ — 42°39′25″
Estats (m)	523 G 4 — 41 G 2
Esterri de Aneo (v)	4°48′40″ — 42°37′35″
Estillon (v)	513 G 1 — 53 G 2
Fàbrica (La) (v)	4°55′20″ — 42°31′15″
Farrera (v)	4°57′35″ — 42°29′45″
Faup (v)	505 G 2 — 54 G 9
Flamissel (r)	4°40′45″ — 42°30′15″
Fondo (l)	5°04′50″ — 42°38′45″
Fonguera (m)	4°43′15″ — 42°33′30″
Fonta (d)	511 G 1 — 48 G 1
Fonta (m)	503 G 7 — 56 G 3
Fontaret (r)	508 G 2 — 48 G 5
Forsa (La) (v)	5°00′40″ — 42°35′30″
Fouillet (r)	517 G 5 — 53 G 7
Founiérous (d)	502 G 9 — 52 G 1
Fraychet (r)	514 G 5 — 51 G 6
Galèche (Cap de la) (m)	508 G 6 — 52 G 3
Garbet (l)	522 G 1 — 50 G 4
Garbet (r)	519 G 2 — 54 G 3
Garbettou (Cirque de) (d)	522 G 3 — 51 G 4
Géou (p)	515 G 5 — 56 G 3
Gérac (r)	515 G 5 — 50 G 6
Gerbel (l)	4°41′00″ — 42°37′40″
Gerbel (r)	4°41′40″ — 42°38′20″
Glorieta de Montesclado (v)	4°56′40″ — 42°30′15″
Graus (v)	4°55′20″ — 42°40′30″
Guingueta (v)	4°49′10″ — 42°35′35″
Guzet (r)	514 G 0 — 54 G 7
Hillette (l)	515 G 8 — 48 G 4
Hillette (r)	515 G 7 — 48 G 6
Isabarre (v)	4°47′10″ — 42°39′00″
Isil (v)	4°46′25″ — 42°40′40″
Lacore (mi)	522 G 5 — 53 G 4
Lagola (l)	4°51′30″ — 42°41′05″
Lausanas (r)	4°44′10″ — 42°45′20″
Lauze (r)	521 G 6 — 53 G 9
Liesca (r)	4°41′50″ — 42°46′15″
Lladorre (r)	4°57′05″ — 42°38′55″
Lladorre (v)	4°56′10″ — 42°37′10″
Lladrós (v)	4°55′40″ — 42°36′20″
Llaguña (r)	5°02′05″ — 42°38′00″
Llavorsí (v)	4°53′50″ — 42°29′40″
Lleret (v)	4°55′20″ — 42°37′05″
Mail (l)	505 G 0 — 47 G 2
Maillat (d)	511 G 0 — 49 G 0
Maladeta (see Saburo & Basiero) (g)	
Mánega (c)	5°01′05″ — 42°30′10″
Marimaña (g)	4°40′40″ — 42°41′50″
Marimaña (r)	4°43′00″ — 42°44′00″
Marterat (c)	510 G 7 — 47 G 2
Mauricio (= Escrita) (r)	4°43′40″ — 42°34′55″

Mérens fault	5°07′30″ — 42°37′30″
Mire Picou de la) (m)	515 G 9 — 52 G 7
Monseny (m)	4°42′45″ — 42°29′25″
Montalto (r)	5°00′40″ — 42°39′35″
Montareño (p)	4°59′40″ — 42°38′15″
Montaud (m)	501 G 4 — 50 G 2
Montcalm (m)	524 G 2 — 41 G 7
Montgarri (v)	4°41′30″ — 42°45′35″
Moredo (m)	4°44′15″ — 42°43′05″
Naorte (l)	4°59′10″ — 42°41′25″
Negro (l)	4°43′50″ — 42°32′20″
Niña (m)	4°43′50″ — 42°33′55″
Noarre (r)	4°56′15″ — 42°41′15″
Noarre (v)	4°55′45″ — 42°41′00″
Norís (v)	5°01′55″ — 42°33′50″
Os de Civís (v)	5°07′35″ — 42°30′40″
Ossèse (r)	512 G 5 — 52 G 7
Ovella (c)	5°06′35″ — 42°33′10″
Pala Pedregosa (m)	4°42′40″ — 42°30′55″
Pallaresa (r)	4°51′00″ — 42°32′40″
Pas d'Enfer (Le) (mi)	519 G 9 — 53 G 2
Pause (c)	502 G 4 — 54 G 8
Peguera (r)	4°41′30″ — 42°33′45″
Perefita (p)	4°52′05″ — 42°37′25″
Pilás(m)	4°49′05″ — 42°40′55″
Plana (p)	4°43′20″ — 42°39′55″
Pudó (l)	4°41′45″ — 42°40′20″
Portet (v)	512 G 6 — 54 G 6
Pouil (r)	503 G 0 — 49 G 3
Quer Ner (r)	507 G 8 — 48 G 1
Rabé (r)	507 G 5 — 50 G 2
Refugio-Baños fault	4°45′20″ — 42°35′40″
Renacha (p)	494 G 0 — 53 G 5
Ribera de Cardós (v)	4°54′45″ — 42°33′45″
Riberot (g)	494 G 5 — 55 G 9
Riberot (r)	490 G 2 — 55 G 3
Rieu (v)	505 G 7 — 55 G 0
Riús-Saburedo (see Basiero & Saburo) (g)	
Romadera (Coma) (p)	4°37′30″ — 42°30′55″
Rosario (l)	4°42′10″ — 42°40′55″
Rouze (v)	507 G 7 — 54 G 8
Rumedo Media (l)	5°00′35″ — 42°42′15″
Rumero (r)	5°01′35″ — 42°41′40″
Saburo (= Saburedo) (g)	4°42′10″ — 42°32′55″
Salat (r)	505 G 7 — 50 G 0
Salau (c)	501 G 4 — 49 G 8
Salau (g)	506 G 4 — 49 G 3
Salau (Pleta del) (p)	4°45′05″ — 42°37′35″
Salau (v)	506 G 2 — 50 G 7
Sallente (r)	4°41′10″ — 42°30′05″
Santa Coloma (g)	5°08′10″ — 42°29′30″
Sellente (c)	5°02′25″ — 42°38′50″
Sens (Mayor de) (l)	4°52′05″ — 42°41′40″
Sérac (= Trape) (r)	515 G 3 — 55 G 9
Sérac (v)	512 G 8 — 56 G 2

Serre du Cot (c)	509 G 3 — 55 G 2
Servi (v)	4°50'10" — 42°38'45"
Son (r)	4°47'20" — 42°37'15"
Son del Pino (v)	4°47'05" — 42°37'10"
Sottlo (r)	5°04'00" — 42°38'20"
Spioulou (r)	507 G 5 — 51 G 8
St. Lizier (v)	512 G 4 — 55 G 5
Surri (v)	4°54'30" — 42°33'55"
Tabescán (c)	510 G 7 — 47 G 2
Tabescán (r)	4°55'20" — 42°40'25"
Tèse (m)	509 G 3 — 51 G 5
Teso (m)	4°43'40" — 42°36'30"
Tírvia (v)	4°55'50" — 42°30'50"
Tór (r)	5°03'55" — 42°34'15"
Tór (v)	5°05'10" — 42°34'10"
Trape (c)	516 G 4 — 55 G 2
Trape (= Sérac) (r)	515 G 3 — 55 G 9
Turguilla (r)	516 G 3 — 49 G 4
Turó Caubó (m)	4°57'05" — 42°40'20"
Unarre (r)	4°50'05" — 42°37'35"
Ustou (Trein d') (v)	511 G 7 — 57 G 1
Vallfarrera (r)	5°03'05" — 42°37'40"

REFERENCES

ALLAART, J. H., 1954. "La couverture sédimentaire septentrionale du Massif Ax-Montcalm". Leidse Geol. Med., 18 (1953), p. 254—269.

ALMELA, A. & RIOS, J. M., 1947. "Explicación al mapa geológico de la provincia de Lérida. 1: 200.000". Madrid, Rev. Inst. geol. minero España, 193 p.

ALPHEN, G. J. VAN, 1956. "Structural features round las Bordas, Valle de Arán, central Pyrenees". Leidse Geol. Med., 21/2, p. 485—489.

AUTRAN, A. & GUITARD, G., 1955. "Stratigraphie du Paléozoïque inférieur de la zone axiale pyrénéenne au Sud de St. Girons (Ariège)". C. R. Acad. Sci. Paris, 241, p. 969—971.

BALK, R., 1937. "Structural Behavior of Igneous Rocks". Mem. Geol. Soc. Am., 5, X + 177 p.

BARROIS, C., 1887. "Sur les faunes siluriennes et dévoniennes de la Hte-Garonne d'après les découvertes de M.M. Gourdon". C. R. Assoc. franç. avanc. Sci. Toulouse, 16, p. 507—514.

BERTRAND, L., 1907. "Contribution à l'histoire stratigraphique et tectonique des Pyrénées Orientales et Centrales". Bull. Serv. Carte géol. France, 17 no. 118, p. 365—547.

——, 1911. "Sur la structure géologique des Pyrénées occidentales et leurs relations avec les Pyrénées orientales et centrales; essai d'une carte structurale des Pyrénées". Bull. Soc. géol. France, (4) 11, p. 122—153.

BLANCHET, P. H., 1957. "Development of fracture analysis as exploration method". Bull. Am. Assoc. Petrol. Geologists, 41, no. 8, p. 1748—1759.

BOISSEVAIN, H., 1934. "Etude géologique et géomorphologique d'une partie de la vallée de la Haute Sègre (Pyrénées Catalanes)". Bull. Soc. Hist. nat. Toulouse, 66, p. 32—170.

BRESSON, A., 1903. "Etudes sur les formations anciennes des Hautes- et Basses-Pyrénées (Haute-Chaîne)". Bull. Serv. Carte géol. France, 14, no. 93, p. 45—322.

CAPDECOMME, L., 1943. "Sur les procédés d'étude des roches charbonneuses, à propos des schistes graphitiques de la région de Marignac". Bull. Soc. Hist. nat. Toulouse, 78, p. 181—190.

CARALP, L., 1888. "Etudes géologiques sur les hauts massifs des Pyrénées Centrales (Haute Ariège, Haute Garonne, Vallée d'Aran)". Thesis Toulouse, Durand, Fillons & Lagarde, 512 p.

CAREZ, L., 1903—09. "La géologie des Pyrénées françaises". Mém. Serv. Carte géol. France, 6 vol.

CASTERAS, M., 1933. "Recherches sur la structure du versant nord des Pyrénées centrales et orientales". Bull. Serv. Carte géol. France, 37, no. 189, 525 p.

CAVET, P., 1951a. "Sur la stratigraphie du Paléozoïque de la zone axiale des Pyrénées-Orientales". C. R. Acad. Sci. Paris, 232, p. 743—745.

——, 1951b. "Découverte du Cambrien à Archaeocyothidés dans la zone axiale des Pyrénées-Orientales; interprétation de cette zone". C. R. Acad. Sci. Paris, 232, p. 858—859.

——, 1958. "Stratigraphie du Paléozoïque de la zone axiale pyrénéenne à l'E de l'Ariège". Bull. Soc. géol. France (6) 8, p. 853—867.

CAVET, P. & GUITARD, G., 1958. "Aperçu sur la géologie de la zone axiale des Pyrénées-orientales entre le Roussillon et la Cerdagne". Livret guide, Réunion extraord. Soc. géol. France. Pyr.-or., 13 p.

CHARPENTIER, J. DE, 1812. "Mémoire sur la nature et le gisement du pyroxène en roche, connu sous le nom de lherzolite". Jour. Mines, 32, p. 321—340.

——, 1813. "Mémoire sur le terrain granitique des Pyrénées". Jour. Mines, 33, p. 101—156.

——, 1823. "Essai sur la constitution géognostique des Pyrénées". Strasbourg, Levrault, XVI + 633 p.

CHAUBET, M. CH., 1937. "Contribution à l'étude géologique du Gothlandien du versant méridional de la Montagne Noire". Trav. Lab. géol. Montpellier, Mém. 1 (hors sér.), 223 p.

CHEVALIER, M., 1954. "Le relief glaciaire des Pyrénées du Couserans". Rev. géogr. Pyrénées & Sud-Ouest, Toulouse, 25/3, p. 97—156.

CLARKE, F. W., 1924. "Data of Geochemistry". U. S. Geol. Surv., Bull. 770, 841 p.

CLOOS, E., 1946. " "Lineation" a critical Review and annotated Bibliography". Geol. Soc. Am., Mem. 18, 122 p.

DALLONI, M., 1910. "Etude géologique des Pyrénées de l'Aragon". Ann. Fac. Sci. Marseille, 19, 444 p.

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

——, 1930. "Etude géologique des Pyrénées Catalanes". Ann. Fac. Sci. Marseille, 26/3, 373 p.

——, 1952. "Découvertes du terrain silurien dans la région de Laruns B.-P." C. R. Acad. Sci. Paris, 234, p. 740—741.

DALY, R. A., 1912. "Geology of the North American Cordillera at the Forty-ninth Parallel". Canada Geol. Surv., Mem. 38/2, 857 p.

DEBYSER, J., 1955. "Etude sédimentologique du système lagunaire d'Abidjan (Côte d'Ivoire)". Rev. Inst. franç. Pétrole, 10/5, p. 319—334.

DÉLÉPINE, G., 1931. "L'âge des schistes de Mondette". C. R. Soc. géol. France, 1931/12, p. 157—158.

——, 1937. "Le Carbonifère du Sud de la France (Pyr. et Mont. Noire)". C. R. 2e Congr. stratigr. Carb. Heerlen (1935), 1, p. 139—158.

DÉLÉPINE, G., DUBAR, G. & LAVERDIÈRE, J. W., 1929. "Observations sur quelques gisements du Carbonifère des Pyrénées". C. R. Soc. géol. France, 1929/16, p. 236—238.

DESTOMBES, J. P., 1949. "Stratigraphie de l'Ordovicien des Hautes Pyrénées". C. R. Soc. géol. France, 1949/7–8, p. 138—140.

——, 1950. "L'Ordovicien de la Haute Ariège". C. R. Soc. géol. France, 1950/5, p. 75—77.

——, 1953. "Stratigraphie des terrains primaires de la haute Garonne". C. R. 19e Congr. géol. internat., Alger (1952), sect. 2, fasc. 2, p. 107—129.

DESTOMBES, J. P. & RAGUIN, E., 1955. "L'étude de la partie occidentale du massif de l'Aston". Bull. Soc. géol. France, (6) 5, p. 101—113.

DESTOMBES, J. P. & VAYSSE, A., 1947. "Sur le Gothlandien de la Vallée de la Pique (Hte-Garonne)". Bull. Soc. géol. France, (5) 17, p. 403—409.

DREYFUSS, M., 1948. "Contribution à l'étude géologique et paléontologique de l'Ordovicien supérieur de la Montagne Noire". Mém. Soc. géol. France, n. sér., 27/2–3, no. 58, 63 p.

DUFRÉNOY, 1831. "Note sur la position géologique des principales mines de fer de la partie orientale des Pyrénées". Bull. Soc. géol. France, (1) 2, p. 69—74.

DUFRÉNOY & ELIE DE BEAUMONT, 1830/38. "Mémoires pour servir à une description géologique de la France". 4 t. Paris, t. 1 (1830), VIII + 469 p., t. 2 (1834), 474 p.

DURAND, J., 1935. "Remarques sur quelques formations de calcaires anciens dans les Pyrénées". C. R. Soc. géol. France, 1935/14, p. 222—223.

DURAND, J. & RAGUIN, E., 1943. "Sur la structure du Massif du Maubermé, dans les Pyrénées Ariégeoises". Bull. Soc. géol. France, (5) 13, p. 9—19.

DURAND DELGA, M. & LARDEUX, H., 1958. "Les lydiennes à Tentaculites de Cascastel (massif de Mouthoumet, Aude)". C. R. Soc. géol. France, 1958/2, p. 18—20.

DUROCHER, J., 1844. "Essai pour servir à la classification des terrains de transition des Pyrénées, et observations diverses sur cette chaîne de montagnes". Ann. Mines, (4) 6, p. 15—112.

ELLES, G. L., 1925. "The characteristic assemblages of the graptolite zones of the British Isles". Geol. Mag., 62, p. 337—347.

ELLES, G. L. & WOOD, E. M. R., 1901/18. "Monograph of British Graptolites." pts. I—XI. Palaeont. Soc. London, p. 1—539.

FAIRBAIRN, H. W., 1949. "Structural petrology of deformed rocks". Cambridge, Mass., Addison Wesley, 2nd ed., IX + 344 p.

FAUCHER, D., 1935. "Surfaces d'aplanissement dans les Pyrénées ariégeoises". C. R. Congr. internat. Géogr., Varsovie 1934, 2, p. 472—476.

——, 1938. "Surfaces d'aplanissement dans les Pyrénées ariégeoises". Congr. internat. Géogr. Amsterdam. Rapp. Comm. Cartogr. Surfaces d'aplan. tertiaires. Paris, p. 34—38.

FAURA Y SANS, M., 1928. "Résumé de nos connaissances sur l'Anthracolitique de la Catalogne et ses relations chronologiques avec les formations similaires de la péninsule Ibérique". Congr. Avanc. Etude Strat. Carbonifère, Heerlen (1927), p. 821—852.

FEARNSIDES, W. G., 1905. "On the geology of Arenig Fawr and Moel Llyfnant". Quart. Jour. Geol. Soc. London, 61, p. 608—637.

FLINT, R. F., 1947. "Glacial geology and the Pleistocene Epoch". New York, J. Wiley & Sons. XVIII + 589 p.

FONTBOTÉ, J. M., 1948. "Sobre la presencia de terrenos cámbricos en el Valle de Ribes (Pirineos Catalanes)". Est. Geol. Madrid, no. 8, p. 209—220.

——, 1949. "Observations au sujet de métamorphisme dans le haut bassin du Ter. (Pyr. Catalanes)". C. R. Soc. géol. France, 1949/4, p. 76—78.

FOURMARIER, P., 1952. "Aperçu sur les déformations intimes des roches en terrains plissés". Ann. Soc. géol. Belg., 75, p. 181—194.

FRANÇOIS, J., 1841. "Aperçu sur l'étude géologique de l'Ariège". Ann. agric., litt. & industr. Foix, 9, p. 125—150.

——, 1843. "Recherches sur le gisement et le traitement direct des mines de fer dans les Pyrénées et particulièrement dans l'Ariège". Paris, X + 404 p.

GARCÍA SAÍNZ, L., 1933. Vestigios de época glaciar en el valle del Flamisell". Bol. Soc. Geogr. Nac., Madrid, (B) 73, no. 21, p. 211—237.

——, 1935. "Morfología glaciar y preglaciar de la región de La Noguera (Cuenca-Cinca-Segre)". Bol. Soc. Geogr. Nac., Madrid, 75, p. 64—110.

——, 1940a. "Las superficies de erosión que preceden a los glaciares cuaternarios del Pirineo Central y sus recíprocas influencias". Est. Geogr. Madrid, 1, p. 45—73.

——, 1940b. "Morphologische Skizze des Zentralabschnittes des spanischen Pyrenäen in der Tertiär- und Quartärzeit". Peterm. Geogr. Mitt., Gotta, 86, p. 366—371.

——, 1941. "Las fases epiglaciares del Pirineo español". Est. Geogr. Madrid, 2/3, p. 209—268.

GÈZE, B., 1949. "Etude géologique de la Montagne Noire et des Cevennes Méridionales". Mém. Soc. géol. France, 29, no. 62, p. 1—192.

GORON, L., 1937. "Les unités topographiques du pays ariégeois. Le rôle des cycles d'érosion tertiaires et des glaciations quaternaires dans leur morphologie". Rev. géogr. Pyrénées & Sud-Ouest, Toulouse, 8, p. 300—333.

GUITARD, G., 1954. "Classification, position et âge des gîtes métallifères dans les Pyrénées-Orientales". C. R. Soc. géol. France, 1954/9, p. 172—177.

HOEPPENER, R., 1955. "Tektonik im Schiefergebirge, eine Einführung". Geol. Rundschau, 44, p. 26—58.

——, 1956. "Zum Problem der Bruchbildung, Schieferung und Faltung". Geol. Rundschau, 45, p. 247—283.

JACOB, CH., 1927. "Considérations tectoniques sur les Pyrénées". Bull. Soc. Hist. nat. Toulouse, 55/2, p. 289—298.

——, 1930. "Zone axiale, versant sud et versant nord des Pyrénées". Bull. Soc. géol. France. Livre jubil. (1830—1930) 2, p. 389—410.

JACOB, CH., FALLOT, P., ASTRE, G. & CIRY, R., 1927. "Observations tectoniques sur le versant méridional des Pyrénées centrales et orientales". C. R. 14e Congr. géol. internat., Madrid 1926, 2, p. 335—411.

JACQUOT, E., 1887. "Sur la constitution géologique des Pyrénées: le système Cambrien". C. R. Acad. Sci. Paris, 104, p. 1318—1321.

——, 1890. "Note sur la constitution géologique des Pyrénées. Le système Cambrien". Bull. Soc. géol. France, (3) 18, p. 640—672.

JELGERSMA, S., 1957. "Investigaciones palinologicas de lignitos terciarios procedentes de Cerdaña y del Valle de Arán (Pirineos españoles)". Cursillos y Conferencias Inst. "Lucas Mallada", Madrid, 4, p. 159—162.

KEIZER, J., 1954. "La géologie de la couverture sédimentaire du Massif de l'Arize". Leidse Geol. Med., 18 (1953), p. 229—253.

KIENOW, S., 1934. "Die innere Tektonik des Unterdevons zwischen Rhein, Mosel und Nahe". Jahrb. Preuss. geol. Landesanst., Berlin, 54, p. 58—95.

——, 1951. "Experimentelle Untersuchungen über das Fertigkeitsverhalten und den Deformations Mechanismus von trockenen Ton unter verschiedenen allseitigen Drucken". Beiträge zur Geomechanik II. N. Jahrb. Geol. & Pal., 1951/2, p. 39—52.

KLEINSMIEDE, W. J. F., 1960. "The geology of the Valle de Arán. Central Pyrenees". Leidse Geol. Med. 25.

LACVIVIER, C. DE, 1882. "Note sur le Dévonien et le Trias du département de l'Ariège". Bull. Soc. géol. France, (3) 10, p. 434—456.

——, 1884. "Etudes géologiques sur le département de l'Ariège et en particulier sur le terrain Crétacé". Ann. Sci. géol., 15, 305 p.

——, 1886. "Note sur le terrain primaire du département de l'Ariège". Bull. Soc. géol. France, (3) 14, p. 613—628.

LAMARE, P., 1939. "La série Paléozoïque du Massif du Baygoura et de la Vallée de la Nive. Ses relations avec les terrains secondaires environnants". Bull. Soc. géol. France, (5) 9, p. 163—184.

LARTET, L., 1884a. "Sur le terrain carbonifère des Pyrénées centrales". C. R. Acad. Sci., Paris, 99, p. 250—253.

——, 1884b. "Découverte de divers fossiles dans les Pyrénées". Bull. Soc. Hist. nat. Toulouse, 18, p. XVII—XVIII.

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

LEITH, C. K. & MEAD, W. J., 1915. "Metamorphic geology". New York, Henri Holt & Co.

Livret guide de l'excursion N 1, Pyrénées. 5me Congr. internat. Inqua. Madrid-Barcelona 1957, 107 p.

LLOPIS LLADÓ, N., 1946. "Los movimientos corticales intracuaternarios del NE de España". Est. Geol. Madrid, 3, p. 181—236.

——, 1947. "El relieve del Alto Valle del Aragón". Pirineos (Rev. Estación Est. Pirenaicos, Zaragoza), 3, no. 5, p. 81—166.

MAGNAN, H., 1871. "Coupe générale des Pyrénées de l'Ariège et des environs d'Esterri, sous le méridien du porte de Salau". Bull. Soc. Hist. nat. Toulouse, 5, p. 41—44.

——, 1874. "Matériaux pour une étude stratigraphique des Pyrénées et des Corbières". Mém. Soc. géol. France, (2) 10/1, p. 1—112.

MOLLARD, J. D., 1957. "Aerial mosaics reveal fracture patterns on surface materials in Southern Saskatchewan and Manitoba". Oil in Canada, Aug. 1957, 11 p.

MUSSY, M., 1868. "Description de la constitution géologique et des ressources minérales du Canton de Vicdessos et spécialement de la mine de Rancié". Ann. Mines, (6) 14, p. 57—112.

——, 1869/70. "Les ressources minerales de l'Ariège". Ann. Mines, (6) 16, p. 547—595, (6) 17, p. 237—287 and p. 459—518.

——, 1870. "Carte géologique et minéralurgique de l'Ariège. Texte explicatif". Foix, Pomiès, 275 p.

NIGGLI, P., 1946. "Tabellen zur Petrographie und zum Gesteinsbestimmen". Zürich, Min. Petr. Inst. E. T. H., VI + 149 p.

NUSSBAUM, F., 1935. "Die Seen der Pyrenäen. Eine geographische Studie". Mitt. Naturf. Ges. Bern (1934), p. 1—184.

——, 1938. "Die eiszeitliche Schneegrenze in den Pyrenäen". Verh. 3. internat. Quartär-Konf., Wien 1936, p. 198—200.

——, 1956. "Observations morphologiques dans la région de la Noguera Pallaresa". Pirineos (Rev. Estación Est. Pyrenaicos, Zaragoza), 12, p. 57—98.

PALASSOU, 1784. "Essai sur la minéralogie des Monts Pyrénées". Paris, XVI + 330 p.

PANNEKOEK, A. J., 1935. "Evolution du bassin de la Tet dans les Pyrénées Orientales pendant le Néogène". Thesis Utrecht, Oosthoek's Uitgevers Mij., 72 p.

——, 1937. "Die jungtertiäre morphologisch-tektonische Entwicklungsgeschichte der östlichen Pyrenäen". Géol. Pays Catalans (Assoc. Etud. géol. Méditerr. occ. Barcelona), 3/4, 25 p.

PENCK, A., 1883. "Die Eiszeit in den Pyrenäen". Mitt. Ver. Erdk. Leipzig, p. 163—231.

PETTIJOHN, F. J., 1949. "Sedimentary Rocks". New York, Harper & Brothers, XV + 526 p.

RAGUIN, E., 1938. "Contribution à l'étude des gneiss des Pyrénées". (observ. de P. Lamare). Bull. Soc. géol. France, (5) 8, p. 11—36.

——, 1946a. "Découverte des roches éruptives ordoviciennes dans les Pyrénées centrales". C. R. Acad. Sci. Paris, 223, p. 816—817.

——, 1946b. "Géologie du granite". Paris, Masson & Cie, 198 p. (2me éd. 1957, 275 p.).

ROUSSEL, J., 1885. "Le Dévonien et le Carbonifère de Larbont et de St. Antoine". Foix, Gadrat, 20 p.

——, 1893. "Etude stratigraphique des Pyrénées". Bull. Serv. Carte géol. France, 5, no. 35, 306 p.

——, 1904. "Tableau stratigraphique des Pyrénées". Bull. Serv. Carte géol. France, 15, no. 97, p. 23—141.

SCHMIDT, H., 1931. "Das Paläozoikum des spanischen Pyrenäen". Abh. Ges. Wiss. Göttingen math.-phys. Kl. 3. Folge, H 5, p. 981—1065.

——, 1944. "El Paleozoïco del Pirineo español". Publ. alem. Geol. España, 2, p. 95—195. (Translation of Schmidt, 1931).

SERMET, J., 1949. "Les influences structurales hercyniennes dans les Pyrénées Ariégeoises Occidentales". Mélanges géogr. Faucher, Toulouse, 2, p. 667—676.

——, 1950. "Réflexions sur la morphologie de la zone axiale des Pyrénées". Pirineos (Rev. Estación Est. Pirenaicos, Zaragoza) 6, no. 17—18, p. 323—404.

SIMPSON, S., 1940. "Das Devon der Südost-Eifel zwischen Nette und Alf". Abh. Senckenberg. Naturforsch. Ges. Frankfurt/M., 447, 81 p.

SITTER, L. U. DE, 1941. "De tectoniek van eenvoudige structuren". Leiden, E. J. Brill, 63 p.

——, 1954a. "Note préliminaire sur la géologie du Val d'Aran". Leidse Geol. Med., 18 (1953), p. 272—280.

——, 1954b. "La faille Nord-Pyrénéenne dans l'Ariège et la Haute-Garonne". Leidse Geol. Med., 18 (1953), p. 287—291.

——, 1954c. "Note provisoire sur la géologie primaire des Pyrénées ariégeoises et garonnaises". Leidse Geol. Med., 18 (1953), p. 292—307.

——, 1956a. "A cross Section through the Central-Pyrenees". Geol. Rundschau, 45/1, p. 214—233.

——, 1956b. "Structural Geology". New York, McGraw Hill, 552 p.

SITTER, L. U. DE & ZWART, H. J., 1950. "Le calcaire de Bentaillou dans les Pyrénées ariégeoises". C. R. Soc. géol. France, 1950/13, p. 229—231.

——, 1959. "Voorlopige resultaten van de karteringen in Noord-Spanje en Zuid-Frankrijk verkregen in 1958 door de afdeling structurele geologie". Leidse Geol. Med., 24, p. 1—16.

SNOEP, J. P., 1956. "Stratigraphy and structural geology of the district west of the Marimaña granite, Valle de Arán". Leidse Geol. Med., 21/2, p. 504—515.

SOLÉ SABARÍS, L., 1951. "Los Pirineos, el medio y el hombre". El mundo y los hombros 8, Barcelona, Alberto Martín, 624 p.

SPENCER, E. W., 1958. "Structural trends in the Beartooth mountains, Montana and Wyoming". Billings Geol. Soc. Guidebook, 9th Ann. Field Conf., 8 p.

VERNEUIL DE & KEYSERLING DE, 1861. "Coupes du versant méridional des Pyrénées". Bull. Soc. géol. France, (2) 18, p. 341—357.

WISSINK, A. J., 1956. "La géologie des environs d'Estours". Leidse Geol. Med., 21/2, p. 516—520.

ZIRKEL, J., 1867. "Beiträge zur geologischen Kenntniss der Pyrenäen". Zeitschr. deutsch. geol. Ges., 19, p. 68—215.

ZEILLER, R., 1886. "Note sur des empreintes houillères recueillies par M. Gourdon dans les Pyrénées centrales". Bull. Soc. géol. France, (3) 14, p. 328—329.

ZWART, H. J., 1954. "La géologie du Massif du Saint-Barthélemy, Pyrénées, France". Thesis Leiden, Leidse Geol. Med., 18 (1953), p. 1—228.

Unpublished reports present in the Geological Institute of the State University of Leiden, Holland:

ALPHEN, G. J. VAN, 1955. "Verslag over de doctoraal kaartering 1954 in de omstreken van de Puerto de la Bonaigua".

BOERSMA, J. R., 1959. "De geologie langs de bovenloop der Río Flamisell".

DESSAUVAGIE, TH. F. J., 1960. "De geologie van zuid-west Andorra en aangrenzend Spanje".

GULDENAAR, P. M., 1955. "De geologie van de Vallée d'Angouls".

HEYDE, C. VAN DER, 1956. "Scriptie over veldwerk in de Pyreneeën, gebied tussen May du Bulard en Pic des Trois Comtes (Frans-Spaanse grens)".

LAPRÉ, J. F., 1959. "De geologie van het westelijk gedeelte van het Ax-Montcalm massief".

OP DE KAMP, J., 1957. "Verslag van de geologische kaartering in de omstreken van de Aneu (Lérida, Spanje)".

PRINS, J., 1952. "De geologie van het Palaezoïcum van de zone axiale van de Pyreneeën, onder de gemeenten Aulus en Ustou, tussen Port de Saleix en de rivier de Alet".