

## GEOLOGY OF THE UPPER RIBAGORZANA AND TOR VALLEYS, CENTRAL PYRENEES, SPAIN

SHEET 8, 1 : 50,000

BY

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## ABSTRACT

The geology of the region of map sheet 8, Ribagorza-Tor, of the 1 : 50,000 geological maps of the Central Pyrenees, is described. The map covers part of the southern axial zone, the Nogueras zone, and the northern part of the Southern Cretaceous zone. The stratigraphic sequence consists of marine Palaeozoic rocks from the Cambro-Ordovician to the Carboniferous, unconformably overlain by mainly fluvial deposits ranging in age from the Upper Westphalian to the Lower Triassic. The Middle and Upper Triassic is represented by a lagunal evaporite sequence. The fully marine Mesozoic rocks, mainly limestones and marls, range in age from Liassic to Upper Cretaceous. The entire sequence is again unconformably overlain by fluvial piedmont deposits of the Upper Eocene-Oligocene.

Hercynian and Alpine orogenies acted on this intercontinental mountain chain. In the axial zone at least five individual deformation phases with different trending axes can be attributed to the Hercynian stress field. The second, or major, folding phase produced a generally northward-dipping axial-plane slaty cleavage. Granodiorite batholiths and numerous dykes intruded almost at the end of this compressional phase.

The major Alpine deformation shortened the axial zone by means of north to south up- and overthrusts along the Hercynian cleavage or fault planes, which at the same time caused asymmetric folding of the post-Hercynian strata above the unconformity. The Nogueras zone is interpreted as a steep flexure zone that collapsed due to the vulnerability of the easily deformed Keuper series filling the space between this flexure and the rigid mass of the overlying Mesozoic calcareous rocks; the latter is thought to have moved towards the south mainly under the influence of the gravitational pull.

The post-Miocene erosion following the Alpine uplift led to various aplanation levels, some remnants of which are preserved in our area. Glacial forms developed during the Pleistocene, and subsequent river erosion modified the glacial morphology.

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## INTRODUCTION

Mapping of the Palaeozoic of the Central Pyrenees by staff members and students of the Department of Geology of Leiden University under directorship of Prof. Dr. L. U. de Sitter started in 1948 with the Arize massif. Since then the survey has been extended first eastwards, 1949 St. Barthélemy massif; then westwards, 1950 Salat valley, 1951 Riberot valley, 1952 Garonne valley, and then southwards, 1953 and 1954 Valle de Arán; and then southwards again to the Spanish part of the Pyrenees between the Esera and Llobregat rivers. Various publications including descriptions of map sheets 1—7 (see reference list), have appeared and other reports are deposited in our files. In recent years special petrographical and micro-structural studies have been undertaken by Prof. Dr. H. J. Zwart and his students.

Most of the above-mentioned studies have been restricted to the Palaeozoic, the younger deposits only being mapped when they form the boundaries or cover of the Hercynian-folded basement; equally, the structural analysis is mainly concerned with the Hercynian structure, although the Alpine deformation had to be analysed sufficiently to know how much

must be subtracted from the present configuration to arrive at the Hercynian structure.

For the area covered by sheet 8 (Fig. 1), the first provisional mapping by undergraduates of our university started in 1955 (Tor valley), and since then seven studies have been completed within that area: Stevens (1956), Steffelaar (1959), Mey (1960), Burkens (1960), Smit (1961), Hofker (1961), and Nagtegaal (1962). Frijlinck, who was mainly concerned with the remapping of the Tor area, had a fatal accident just before completing his promising Master's thesis. In November 1967, the first map and detailed description of this southern part was published in the form of a Ph.D. thesis (Mey, 1967), mainly treating the stratigraphy and structure of the axial zone west of the Tor River. The present map is thus a compilation of the above-mentioned works (see index map on sheet 8), recently supplemented by the results of about two months' additional fieldwork by the author.

Use was made of the 1:50,000 scale topographic maps of the Instituto Topografico y Catastral at Madrid, including the sheets for Benasque (180), Bisaurri (213), Sort (214), and Esterri de Aneu (181), enlarged to a scale of 1:25,000, and for some local

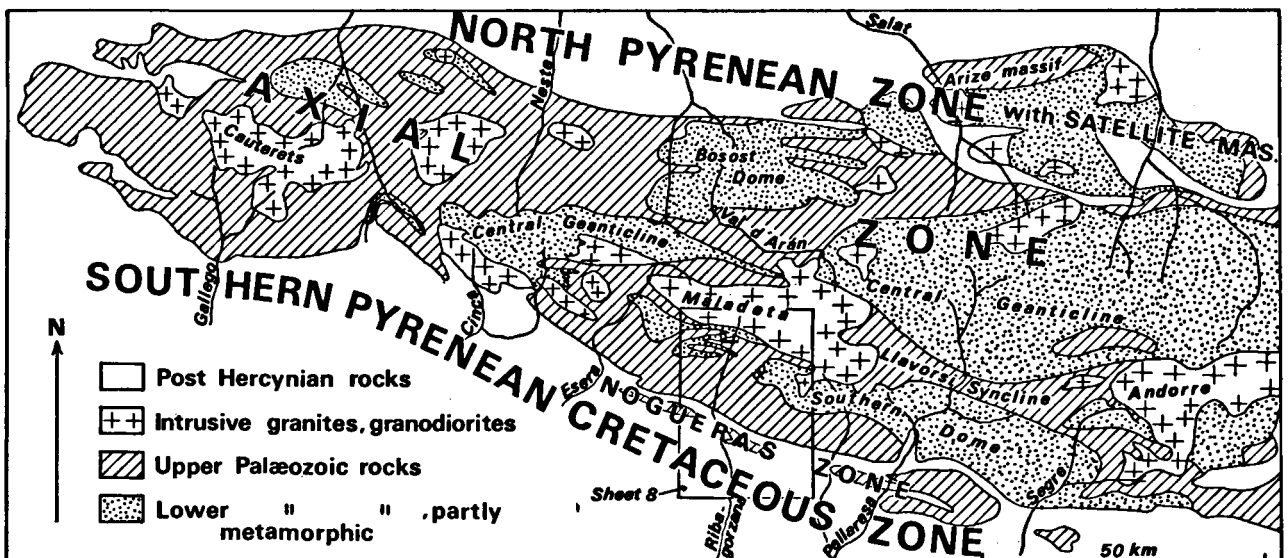


Fig. 1. Geological sketch-map of the Central Pyrenees (modified after de Sitter, 1964) showing location of sheet 8.

cases even to 1 : 7,500. In the northern Tor area and southeastern part of sheet 8, mapping was carried out on the much better 1 : 25,000 scale maps of Sarroca de Bellera (military map) and Montardo (tourist map of the Alpina series, Grenollers).

Good aerial photographs were available for the area east of the Tor, and were used not only to improve the topographical maps and fill in the Quaternary deposits but also to solve some incomprehensible structures. They were most useful for the granodiorite area and its aureole of hornfels and marble, and in the Nogueras zone in the south.

The area is thinly populated and its villages are small. Many of the higher mountain villages, isolated by bad roads, are doomed to die out and others are already completely depopulated. The population supports itself mainly by breeding horses and cattle (cows, sheep, and goats) as well as a little farming (grain and potatoes), in the lower part of the main valleys or on the flat fluvio-glacial terraces (e.g. east of Tahull and Durro).

The town of Pont de Suert (almost 3000 inhabitants), situated where the Ribagorzana flows into the Escalles reservoir, is the centre of the district with prosperous shopkeepers, hotels and restaurants, and a weekly market. It is also the local headquarters of the ENHER, a company producing hydro-electrical energy in the Ribagorzana valley and its major tributaries. Lead-zinc and coal mining are also of some importance in the Bono and Malpas areas, respectively, and there is an important cement factory in Cheralló.

Within the map-area there is an appreciable amount of indigenous timber including beech, black pine, ash, and oak, with poplars along the river banks. Although

locally important, forestry was formerly not profitable as a long-term enterprise, but the activities of the ENHER have inaugurated a program of checking and preventing soil erosion, and reafforestation is now in progress on a large scale.

Since the map-area is one of rugged and isolated beauty, it is not surprising that it attracts many foreign tourists (especially from France). As main tourist attractions we may mention:

1. Modern church and weekly market in Pont de Suert.
2. Mountain climbing centred on Bohi and recently also Barruera.
3. Trout fishing in the upper waters of Ribagorzana, Tor, and their tributaries.
4. Aiguas Tortes National Park in the San Nicolás valley.
5. The hot springs of Caldas de Bohi (Balneario de Bohi).
6. The high dams and artificial lakes such as Cavallers and Escalles, and the underground power station at Caldas.

However, the lack of camping sites in the southern area and the uncertain weather in the northern areas (1200—2000 mm of annual precipitation and 40 % of the year total cloud cover) are considerable limiting factors.

For details concerning the general geography, physiographic aspects, and the climate of the map-area, we refer the reader to a very recent Master's thesis by Roger W. Penn, an internal report of the Geographic Department of the Nottingham University, England.

## CHAPTER I

### STRATIGRAPHY

#### CAMBRO-ORDOVICIAN

In the map-area the oldest outcropping rock sequence is of pre-Silurian age. These rocks, which consist of a rather monotonous, non-fossiliferous alternation of mainly slates, locally phyllitic, with numerous thin quartzite bands, are exposed over large areas in the northeastern corner of the map-area immediately south of the Maladeta granodiorite (Muro and Payasos Domes). The outcrops in the upper part of the Baliera valley are smaller and less well exposed. Folding is very intense (main folding and refolding), which makes the stratigraphical succession hard to establish. For the upper part of the sequence, Mey (1967-b, pp. 158—159) has given a detailed section in the upper course of the Baliera valley that can be correlated with a similar succession of Cambro-Ordovician age in the Segre valley (Fig. 2). A somewhat comparable succession has been observed locally in the upper part of the Muro and Payasos Domes. Conglomerates or pebbly mudstones have not

been encountered anywhere, however. They either were never deposited in this area or erosion does not reach deep enough in the sequence to expose them. The typical "*calcaire métallifère*" does not occur in this region either.

Since most of the Cambro-Ordovician outcrops consist of a typical warped-like series ("*schistes rubanés*"), which in other Pyrenean areas is characteristic for the rock sequence below the conglomerate, we must assume that in our area the conglomerate is not developed. Absence of both the conglomerate and intense folding makes the estimation of the thickness very unreliable, but a minimum of several hundred metres seems plausible.

The slates are more or less phyllitic, and have a dark-grey fracture if pure; they may be silty to locally fine sandy, especially where they alternate with lighter grey, generally thin, quartzitic sandstones. In the lower, warped-like series, about two-thirds to three-quarters of the rock consists of pelites and the rest

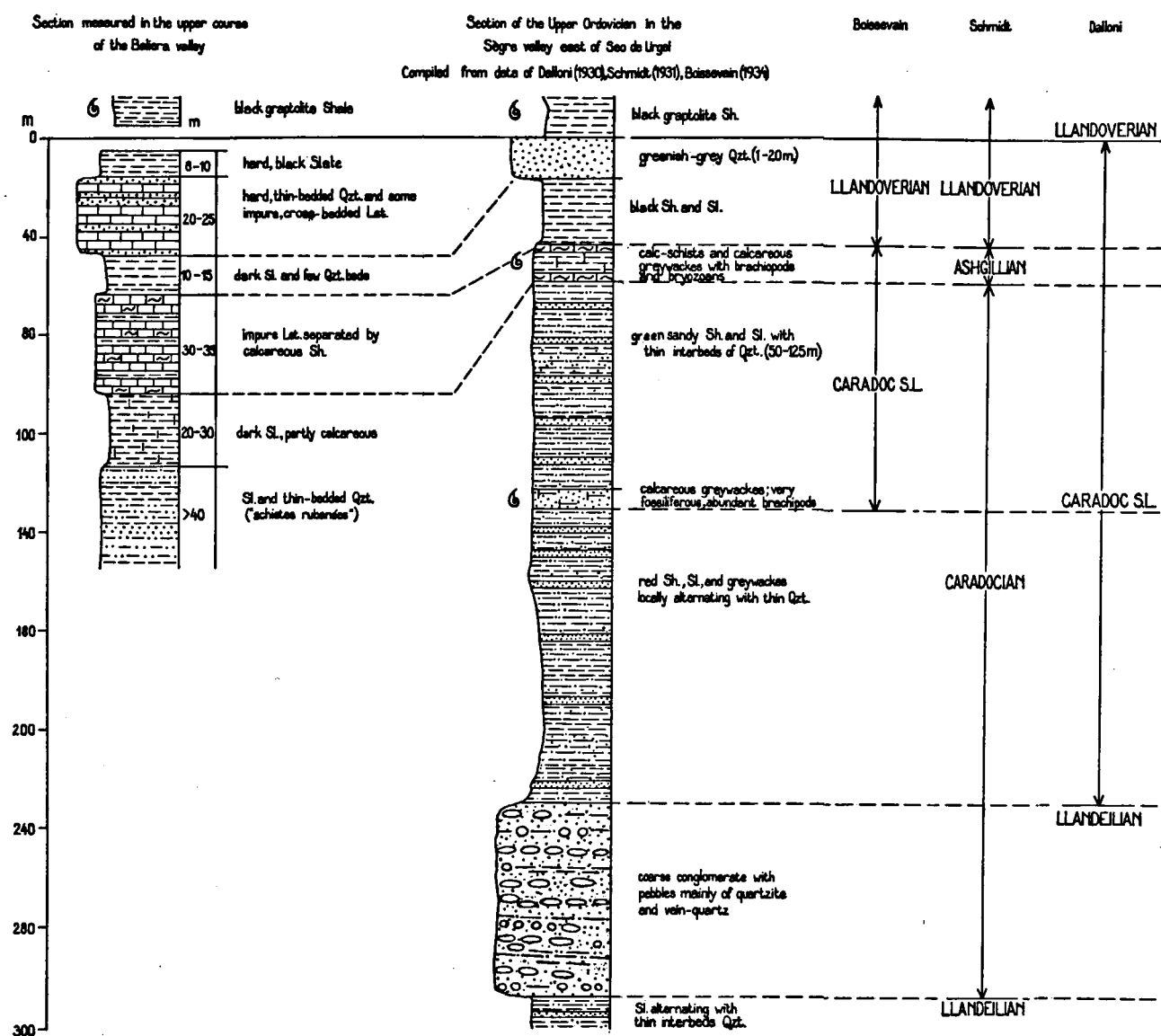


Fig. 2. Comparison of a columnar section of the Upper Ordovician, taken in the Baliera valley, with a schematic section in the Segre valley east of Seo de Urgel.

of very fine graded sandstones. The alternating bands vary from about 1 mm to 3 cm.

The weathering colour of the sandy pelitic sequence as well as the thicker quartzite in the upper part, is generally rusty brown; the few occurrences of impure limestones, which are mainly thin bedded, show a more yellowish-brown weathering colour. All Cambro-Ordovician outcrops are covered with vast fields of debris.

This neritic formation is everywhere intensely folded, mainly isoclinally with a small wavelength. The strongly developed slaty cleavage, which in the finer sediments runs more or less parallel to the bedding, is very often strongly folded in the eastern outcrops (NW-SE and E-W refolding, Figs. 30, 31).

#### SILURIAN

In the Pyrenees and even far beyond this mountain range, the most uniformly developed and best-dated horizon is the Silurian, which consists of friable, black, carbonaceous shales/slates, the upper part containing thin-bedded intercalations of dark-grey to black limestones. The shales and locally the limestones frequently delivered graptolites (mainly *Monograptus*) and the limestones contain many orthocerids, *Cardiola interrupta*, and crinoid ossicles. In the map-area graptolites are rare and strong tectonization has made them hard to recognize; orthocerids are usually present. From the Sierra Negra area, in the northwestern corner of our map, Dalloni (1910) collected the following graptolites:



*Monograptus priodon*, Bronn.  
*Monograptus sedgwickii*, Barr.  
*Monograptus spiralis*, Hisinger  
*Monograptus becki*, Barr.  
*Monograptus convolutus*, His.  
*Monograptus helli*, Barr.  
*Monograptus nilsonni*, Barr.

The fossil discoveries from all over the Pyrenees indicate that the black slates and Orthoceras limestones range from Middle Llandovery to Lower Ludlow (e.g. Dalloni, 1910, 1930; Schmidt, 1931; Keizer, 1954, and Destombes, 1953).

The main field characteristics of the so-called "*schistes carburés*" are well described in the literature (e.g. Kleinsmiede, 1960, p. 144). The thickness of these strongly flattened and sometimes diapyrically intruded slates is hard to evaluate, but in the Sierra Negra area a thickness of at least 50 to 100 m seems very likely.

## DEVONIAN

### Introduction

The thick limestone-shale alternation following conformably upon the Silurian black slates can now be satisfactorily subdivided and mapped. For the southern Pyrenees this was first done successfully in the Baliera area (Mey, 1967-b) and this subdivision has since been proved to be fully applicable in a much wider area south and southeast of the Maladeta batholith. From bottom to top it embraces the following five rock units (Fig. 3):

Mañanet Griotte	— nodular limestone, limestone and calc-schists
Fonchanina Formation	— slate with rare limestone intercalations
Basibé Formation	— limestone and dolomite with an intercalated quartzite member in the Baliera-Ribagorzana area
Gelada Formation	— sandy slates, quartzwackes, and few impure limestones
Aneto Formation	— mainly shale and slate with few marly limestone intercalations.

A detailed analysis showed the existence of a southern Baliera facies area, in which the above-mentioned formations have a thick development and the Basibé quartzite member is locally up to about 50 m thick, and a northern Sierra Negra facies area in which the formations are much thinner and some parts are even missing: both the Basibé quartzite member and the main part of the dolomites and nodular limestones are missing and the Aneto slate formation can no longer be mapped as a separate rock unit. For the inseparable lower unit (Gelada + Aneto Formation), Mey (1967-b) introduced the name Rueda Formation and called the limestone unit above it the Castanesa Formation (local equivalent of the Basibé Formation). Both these facies areas are

separated by an important longitudinal Alpine overthrust (Senet thrust) that probably cuts out the transition zone between the two facies.

More recent investigation of the area east of the Tor River has shown that the Basibé Formation gradually loses its characteristic quartzite member and accompanying dolomite beds. The remaining formation then consists of pure limestones, the upper part of which is regularly bedded and slightly bituminous and the lower part represents a more nodular carbonate showing some resemblance to the more massive parts of the Mañanet griotte. It is evident from the foregoing that it would be inconvenient to use a new formation name when the quartzites and dolomites are no longer present in the Basibé Formation. We therefore reject the name Castanesa Formation and continue to speak of the Basibé Formation even when no quartzite and dolomite are present. The same has been done in the description of the Devonian of sheet 7 (Wennekers, 1968).

In the area east of the Tor River and south of the Cambro-Ordovician Muro and Payasos Domes it was no longer possible to map the Aneto and Gelada Formations separately, first of all because the Aneto Formation is there much thinner than in the west, and secondly, because it shows more limestone intercalations, thus more or less resembling the Gelada Formation. Therefore, it is here again much more convenient to use the term Rueda Formation for the entire rock unit of shales, sandy shales, impure limestones, and quartzwackes.

In the small triangle located immediately south and southeast of the Muro Dome (throughout this article we will refer to this area as the Durro Triangle), it often proved impossible to map all the tiny outcrops of Silurian slates separately. Consequently, Silurian slates together with the lowermost Devonian limestones and slates, both intensely tectonized, have been mapped as one unit (see map).

The above-mentioned Devonian formations will be discussed briefly below; for more detail we refer to Mey (1967-b, pp. 160–173).

### Aneto Formation

In the Baliera facies area, especially west of the Tor River, the Devonian starts with a thick series of slates bearing a slight resemblance to the Silurian black slates. Their Gedinnian age was first demonstrated by Dalloni (1910) in an outcrop near to Pico Cerler (sheet 7, coord. 42°34' . 4°15') and later by Schmidt (1931) in the same locality and in the Sierra de Monros, some 10 km east of our map-area.

The formation has been defined by Mey (1967-b, p. 167); as type section he selected the well-exposed and typical sequence in the vicinity of the village Aneto in the Ribagorzana valley (coord. 42°33' . 4°26'). These slates are mainly exposed near the northern boundary of the Baliera facies area. The sequence is thick (50–200 m) and contains only a few intercalations of marly limestones in the northwestern part of the map-area, but it is probably thinner (50–80 m)



show a brownish weathering and have a relatively higher competence. The "basal limestone" in the Durro Triangle consists of a frequent alternation of thin platy and sometimes nodular limestones and marly pelites. When freshly fractured this rock is blackish to rather dark blueish grey; the weathering is dull and lighter coloured.

Locally in this formation there are black, spherical or elliptical nodules with a diameter of 3 to 20 cm. These nodules sometimes contain fossil remains of orthocerids, brachiopods, and tentaculites. For an extensive list of fossils (mainly brachiopods), we refer to Dalloni (1910, pp. 62–63) and Schmidt (1931, pp. 41 and 44).

#### *Gelada Formation*

The Gelada Formation, a series of sandy slate, impure limestones, and quartzwackes lying on top of the Aneto Formation, has been defined by Mey (1967-b, p. 168); as type section he selected the well-exposed and non-disturbed sequence immediately north of the Pico Gelada (coord.  $42^{\circ}31\frac{1}{2}'$  .  $4^{\circ}27'$ ).

The Gelada Formation proved to have a very constant development in the entire Baliera facies area, where it varies little in thickness, i.e. from about 100 to 120 m. In the west and center of the map-area, i.e. in the area where the Aneto slates are thickest, the Gelada Formation starts with some 25 m of marly limestone, followed by a monotonous alternation of sandy slates, impure limestones, and quartzwackes, the latter being most frequent in the upper part of the rock sequence. In the area east of the Tor River a gradual change in attitude of the upper part of the formation becomes noticeable: the amount of quartzwackes decreases along with an increase in the number of marly limestone beds. The tendency to an increasing number of limestones and decreasing sand content from west to east is general for the Lower Devonian of the entire Southern Pyrenees (Dalloni, 1910, 1930; Mey, 1967-a).

Typical for the Gelada Formation are black, irregular streaks which are sometimes bifurcated or branched and may strongly resemble carbonaceous plant

remains, but they may also be feeding trails of *Chondrites*. The shape of these streaks depends on the angle between the bedding and the cleavage plane (Fig. 4). The results of chemical analysis make an organic origin quite probable.

The bedding of the different rock types can be easily distinguished from nearby, but from a distance it is hardly visible. The thickness of the individual beds of the fine-grained rocks varies from 1 to about 50 cm; the coarser quartzwackes are thicker bedded (up to 2 m). The  $s_1$ -cleavage, which is not very strong in this formation except for the pure slates, cuts the bedding at widely varying angles. The individual cleavage planes show an irregular, rough surface and are seldomly lustrous. The weathering of the rocks is a dull dirty brownish colour (due to the high iron content).

The sandy pelites, like the marly limestones, contain many fossils of mainly orthocerids, brachiopods, corals, trilobites, and tentaculites. Their poor preservation and distortion due to cleavage flattening, however, prevent closer determination of this material. For a fossil list of this formation, we refer to Schmidt (1931, p. 43) who assigned a Lower Coblenzian age on the basis of the assemblage.

#### *Rueda Formation*

The Rueda Formation comprises the Aneto and the Gelada Formations. This term is applied to situations in which the Aneto Formation is very thin or entirely absent, as in the northern Sierra Negra facies area, or, as in the area east of the Tor River, where no proper and mappable limit can be drawn between both formations.

#### *Basibé Formation*

The Gelada or Rueda Formations are conformably overlain by a competent unit consisting of spatic and nodular limestones, which in the western part of the Baliera facies area are accompanied by dolomites and quartzites, together forming a single conspicuous ridge in the field. This valuable marker group has been defined by Mey (1967-b, p. 170); it is named after the



Fig. 4. Carbonaceous black streaks typical of the Rueda and Gelada Formations  
a. on the bedding plane; b. on the cleavage plane (the bedding makes a low angle with the cleavage).

Basibé massif (coord.  $42^{\circ}32\frac{1}{2}'$  .  $4^{\circ}17'$ ), the dominant mountain ridge of the Upper Isabena valley.

In the Baliera facies area, south of the Senet thrust zone, the Basibé Formation is thickest and most typical. The greatest thickness occurs near the Liri faultzone in the west (sheet 7), where this formation and the entire Baliera facies ends abruptly. There (Barranco de Urmella) a total thickness of 175 m, and for the quartzite member 60 m, was measured. In an easterly direction the amount of quartzites and dolomites diminishes gradually along with the total thickness of the entire formation (see Mey, 1967-b, Fig. 10). Fluctuations in a N-S direction are of minor importance. East of a line running NNW over the village Durro in the north and Sas in the south, the Basibé Formation has lost its quartzite member entirely and consists of one limestone unit whose lower part is nodular and light coloured, the upper part being regularly thin bedded and dark coloured.

In the Sierra Negra facies area, north of the Senet thrust and north of the Payasos Dome, the Basibé Formation consists of a rather thin limestone unit (20–60 m) which is thick bedded to massive. The lowermost part (4–8 m) is locally dolomitic.

In the isoclinal synclines south of the Montañeta, the lower part of the formation contains a few sandy dolomites as well as nodular limestone beds, thus forming a more gradual transition into the southern facies with intercalations of quartzites.

In its most typical development the Basibé Formation consists of an evenly-bedded alternation of mainly nodular limestone, grey spatic limestone, and dolomites with very thin intercalations of calc-schists. The thickness of the individual beds varies between 15 and 30 cm. The general weathering colour is yellowish-brown. On fresh cuts the nodular limestones are white, cream, pink, greenish, grey, or light-brown. The dolomites are also light-brown, have a sugar-like texture, and contain many quartz veins. The cleavage is only well developed in the calc-schists; it is irregular and widely spaced in the limestones and most often absent in the dolomites. The thickness of this lower member varies between 20 and 55 m.

The following quartzite member starts at its base with an alternation of evenly-bedded quartzites and dolomites, both showing a brownish weathering. When a cleavage is present it is of the fracture cleavage type. Higher up, the number and thickness of the quartzites increase and the amount of dolomites diminishes until they disappear altogether in the upper part of this member, in which quartzites are thick bedded and massive (up to 10 m). The quartzites are very homogeneous and show no sedimentary features such as cross-bedding, grading, burrows, or wave ripples. Their internal colour varies between blueish-white, grey, cream, and light-brown; their weathering colour is always light-grey. This member is about 60 m thick in the west near the Liri faultzone, about 40 m in the Baliera valley, about 15 m in the Ribagorzana valley, 4–8 m in the Tor valley, and disappears completely east of the Durro-Sas line. Although thin beds of

quartzite are still present in the area between the Tor and the Durro-Sas line, the dolomites have already disappeared several hundred metres west of the Tor River.

Mey (1967-b, p. 171) explains the sand wedge (Basibé quartzite) with accompanying dolomites as a shifting beach barrier, and suggested as place of origin a river delta somewhere near the southern end of the Liri faultzone.

The origin of the sand and the depositional environment of both the quartzites and dolomites of the Basibé Formation are now under study by M. A. Habermehl of the Leiden Department of Sedimentology and his report will be published in due course.

The upper member of the Basibé Formation consists of a series of blackish to dark-blueish, thinly-stratified limestones with a total thickness of 20 to 50 m. In the uppermost part thin intercalations of dark-coloured calc-schists are frequently found. The cleavage is rather well developed in both calc-schists and limestones; its planes show a lustrous surface.

In both the upper and lower members of this formation we observed the following fossils: brachiopods, orthocerids, stromatopores, solitary corals, and many crinoid ossicles. None of them are sufficiently well preserved to permit closer determination. For a more detailed fossil list, we refer to Dalloni (1910, pp. 78, 80) and Schmidt (1931, p. 48). The fossil assemblage points to a Middle Devonian age.

#### *Fonchanina Formation*

The Fonchanina Formation, as defined by Mey (1967-b, p. 165), comprises a series of fissile dark slates with only a few or no thin-bedded limestone intercalations. In both facies areas these slates overlie the Basibé Formation and are in turn overlain by nodular, multi-coloured limestones; lower as well as upper boundaries are conformable and rather gradual.

This slate formation is thickest in the western part of the Baliera facies area (50–70 m) and contains only one horizon (5–10 m) of black, thinly-bedded limestones in the middle part. East of the Tor River it is probably thinner (30–40 m) and the number of thin limestone intercalations increases in an eastward direction. In the Sierra Negra facies area north of the Senet thrust the formation is even thinner (15–35 m) but lacks limestone intercalations. Only rough estimations of the thickness are possible since the cleavage flattening, which may well attain a value of 0.45 to 70 %, may vary from one place to another. Another disturbing factor is the great mobility of these slates.

The slates are fine, fissile, and dark coloured with lustrous cleavage planes. Even when they are black they never stain the finger as do the Silurian black slates. Going from west to east, it is in these slates that the first indications of refolding are noticeable in the form of faint crenulations of the cleavage plane. In the area east of the Peranera River the refolding may be so strong in these slates that macroscopically

the  $s_1$ -cleavage has vanished almost completely, being replaced by a secondary cleavage.

The intercalated limestones are thin bedded, dark coloured, and slightly bituminous. They show great resemblance to those of the upper member of the Basibé Formation.

Determinable fossils are scarce in both slates and limestones. We observed only a few strongly deformed crinoid ossicles and tentaculites.

#### *Mañanet Griotte*

The Fonchanina slates are conformably overlain in both facies areas by mainly nodular, multicoloured limestones (griotte). This rock sequence has been defined by Roberti (pers. comm., 1967) from the Mañanet valley and was first described by Mey (1967-b, pp. 166, 173).

The total thickness of this formation varies considerably in a N-S as well as in an E-W direction. In the southwestern part of the Baliera facies area it is about 80 m thick (Isabena), in the Baliera and Ribagorzana valleys 200 to about 300 m thick, diminishing again to about 100 to 150 m in the Tor area and east of this river. In the area north of the Senet thrust thicknesses of only 35 to 70 m were measured.

One of the primary general characteristics of this formation is the variety in the colours of the fresh rocks apart from their usually yellowish-grey to greenish-brown surface weathering. They range from white, reddish, greenish, to violet-grey with either an abrupt or a gradual transition.

The second characteristic, from which the rock derives its name, is the pseudo-nodular internal structure of the limestone. Tectonical flattening with accompanying elongation in another direction has led to the often rod-like structures lying parallel to the b-axes of the slaty cleavage folds. The cross-sections of these rods are still roundish to flat ellipsoid, a shape which is also prominent in the non-cleaved griotte occurrences, e.g. in the Pallaresa valley near the hamlet of Compte (Nagtegaal, 1968).

Besides these typical nodular limestones, the formation shows many intercalations of vividly coloured calc-schists, pure shales, and dark limestones which become especially prominent in the area of sheet 7, where they are thought to represent the Visean limestone of the Western Pyrenees (Mey, 1967-a; Wennekers, 1968).

The bedding of the various rock types can be easily distinguished. In the griotte the bedding planes are undulant and the transition from one bed to another may be either gradual or sharp. The thickness of the individual beds ranges from 1 to 10 cm for the calc-schists, to about 10 to 50 cm for the limestone and griotte beds. The cleavage is well developed in both griotte and calc-schists, and less so in the dark limestones. The individual planes are smooth and lustrous. We found only a few fossils, including large orthocerids (up to 50 cm), a few solitary corals, a small trilobite, and the usual crinoid ossicles. The goniatites, so frequently seen in the Northern Pyrenees and in the Noguerras zone east of sheet 8, were not observed in

this region. Neither Dalloni nor Schmidt mention typical fossils of this formation in our area. However, a correlation with the well-dated griotte formations in the adjoining areas (Pallaresa valley, Segre valley, Northern and Western Pyrenees, see review in Mey, 1967-a) makes it very probable that the Mañanet Griotte represents both the Frasnian and Famennian periods.

#### PRE-HERCYNIAN CARBONIFEROUS

In both facies areas the Mañanet Griotte is conformably overlain by a mainly pelitic sequence of probably Carboniferous age. The transition is often gradual. Bedded cherts, calcium phosphate nodules, or ash layers, which are known to represent the base of the Carboniferous in the Pyrenees (Ziegler, 1959; Krylatov, 1963-a & b), have unfortunately not been found here. The locally encountered cherty, pebble-like concretions at the very base of this pelitic sequence may, however, correspond to the radiolaria-bearing bedded cherts.

The original thickness of this shale-sandstone sequence is unknown, because it is unconformably overlain by Westphalian, Stephanian, Permian, or Triassic, but the exposed thickness may well attain several hundred metres.

The monotonous sequence of shales/slates, siltstones, and calcareous sandstones lacking any continuous markers allows no subdivision. The dull-toned colours range from greyish-brown, light-grey, to dark-grey. As elsewhere, this Carboniferous can be readily recognized by the constant presence of much detrital mica, often in flakes of up to 1 mm, mostly in the silty and more sand-bearing rocks.

The bedding of the pelitic rocks is often hard to recognize; the cleavage is the most prominent feature, being a true slaty cleavage in the few scattered outcrops in the north and a weak, almost fracture cleavage in the south. Pencil-shales occur quite frequently in the south.

The calcareous sandstones, seldomly mappable for more than a few hundred metres, are between 5 to 10 m thick. Their bedding is irregular, and when cleaved it is by a coarse fracture cleavage.

Fossils are rare in these detrital rocks and when present are heavily deformed. Besides some solitary corals and small brachiopods, a very small trilobite was found, but reliable determination was not possible because of their strong distortion due to cleavage flattening. Since we know, however, that the first post-tectonic conglomerate is of Upper Westphalian age (Aguiró, p. 239) the detrital sequence described above must be older.

#### POST-HERCYNIAN CARBONIFEROUS

##### *Introduction*

The oldest post-Hercynian rocks, all of them terrestrial, are restricted to the northern border of the Noguerras zone. In the area west of the Tor River these are

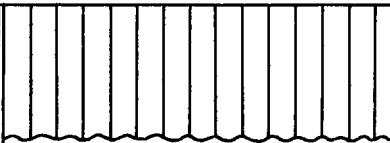
TIME - STRATIGRAPHIC UNIT		LITHO-STRATIGRAPHIC UNIT	GENERAL LITHOLOGY (thickness not on scale)	THICKNESS VARIATION on sheet 8
PERIOD	EPOCH			
TERTIARY	Oligocene Upper Eocene	Collegats Conglomerates	coarse conglomerate with silt/sandstone intercalation	>500m
	Pyrenean folding phase concentric folding			
CRETACEOUS	Lower Eocene Paleocene			
	Danian Maestrichtian Campanian Santonian Coniacian	Vallcarga Formation	marls, nodular argillaceous limestone and calcareous silt/sandstones	>500m
	Turonian	Santa Fé Limestone	massive micritic limestone and nodular argillaceous limestone with slump horizons	150-250m
	U. Cenomanian	Sopeira Marls	sandy marl and nodular argillaceous lst.	50-500m
	Austurian phase epeirogenetic movements			
	L. Cenomanian	Aulet Orbitolina Limestone	orbitolina-rich, marly and sandy limestone	10-900m
	Albian	San Martín fm	sandstones and conglomerates	0-80m
	Aptian	Llusa Marls	mainly marls and nodular marly limestone	20-1000m
	Barremian Hauterivian Valanginian	Prada Limestone	massive, micritic, fossiliferous limestone	0-800m
	Malm	Bonansa Formation upper member	massive dolomite	50-500m
JURASSIC	Dogger	Bonansa Formation middle member	very fossiliferous dark shale and marl	about 60m
	Liassic	Bonansa Formation lower member	finely-stratified marly dolomite and lst./dol.	about 60m
TRIASSIC	Keuper	Pont de Suert Formation upper member	mudstone, dolomitic marl and gypsum with ophite bodies	180-220m
	Muschelkalk	Pont de Suert Formation lower member	mainly dolomitic limestone and dolomite	20-50m
	Buntsandstein	Bunter	red mud- and siltstones, sandstones and conglomerates	180-220m
PERMIAN	Saalic phase epeirogenetic movements			
		Peranera Formation	red mud- and siltstones, sandstones dolomitic beds and breccias	0-700m
CARBONIFEROUS	Stephanian	Malpas Formation	siltstones, sandstones with intercalations of coal seams and arkosic beds	0-300m
		Erill Castell Volcanics	mainly pyroclastics with locally andesite streams	0-350m
	Upper Westphalian	Aguiró Formation	mainly coarse conglomerate, with locally in the lower part coal stringers and breccias	0-130m
	Hercynian folding phase cleavage folding, metamorphism, acidic intrusions			

Fig. 5. Table of post-Hercynian formations for the area covered by sheet 8.

Permo-Triassic rocks, a few outcrops of which also occur in the axial zone. North of the hamlet Iguerri (coord.  $42^{\circ}26' . 4^{\circ}27'$ ), the Permo-Triassic rocks are underlain by a post-tectonic, very coarse conglomerate, and still further to the east the first post-Hercynian sediments are tuffs and lava overlain by a sequence containing coal beds of Stephanian age. East of the village of Avallanos (coord.  $42^{\circ}24\frac{1}{2}' . 4^{\circ}35'$ ), a conglomerate comparable to the one north of Iguerri is found below the Stephanian tuffs; coal beds intercalated between these conglomerates could be dated further to the east (near Aguiró, just outside the map-area) as Upper Westphalian (Roussel, 1904; Dalloni, 1930; Nagtegaal, 1967 pers. comm.).

Determinable plant imprints are restricted to a few coal-bearing horizons in the lower part of the conglomerates and the sedimentary series above the tuffs and lava. Since our mapping was based on different rock types and not on dated horizons, we prefer to use here the formation names introduced in a separate article by Mey *et al.* (1968).

The post-Hercynian Carboniferous is subdivided into 3 formations (Figs. 5 and 8); these are from top to bottom:

Malpas Formation	} Stephanian
Erill Castell Volcanic Formation	
Aguiró Formation	} Upper Westphalian

These three formations will be briefly described below; for a more detailed description, their mode of origin, and conclusions concerning climate and palaeogeography, we refer to the thesis of Nagtegaal to be published shortly in the *Leidse Geologische Mededelingen*.

#### *Aguiró Formation*

The Aguiró Formation, as defined by Mey *et al.* (1968), is the rock unit composed of mainly coarse conglomerates with locally at the base a horizon of breccia and some coal stringers. It rests unconformably on folded and cleaved Devonian and Lower Carboniferous rocks. The upper boundary is drawn at the base of the first pyroclastics (Erill Castell Volcanics) as near Aguiró, or at the base of the red sequence of sandstones and mudstones (Bunter) as north of Iguerri.

The most highly differentiated outcrop of this formation occurs a few hundred metres west of the village of Aguiró, just outside our map-area. The total thickness is there about 130 m. A simplified section is shown in Fig. 6. The lower breccia member is badly sorted (Fig. 7A) and contains angular blocks measuring up to 2 m. Stratification is absent and the colour is grey to brownish. The fact that the breccia does not contain fragments other than of the Devonian Rueda Formation lying immediately below the unconformity, the bad sorting, and the angularity of the fragments, strongly suggest that these rocks represent a fossil slope breccia (as interpreted by Nagtegaal, *ibid.*), indicating a (locally) steep relief.

This breccia member was not found at the base of

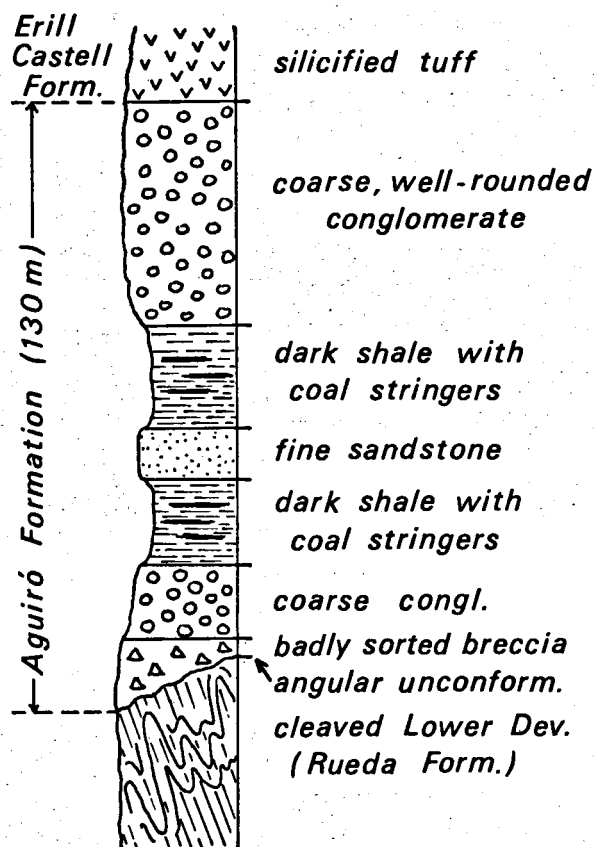


Fig. 6. Columnar section of the Aguiró Formation, taken west of the village of Aguiró.

the Iguerri outcrop, but the same kind of rock is sometimes present at the base of the Erill Castell Volcanics, e.g. immediately north of Erill Castell.

The basal breccia is overlain by a coarse, well-rounded conglomerate (Fig. 7B), the pebbles having an average diameter of about 15 cm (max. diameter about 50 cm). These pebbles originated from many different Palaeozoic strata, ranging from Ordovician quartzite and hornfels to Upper Devonian cleaved grotte. No pebbles of igneous rocks were found. These conglomerates are true river deposits. The very irregular lower contact and the general character of these conglomerates indicate that both their source area and the area of deposition still had a pronounced relief.

The conglomerates are followed by black shales with a few coal stringers and a horizon of fine-grained sandstone. This member, composed of fine detrital sediments and coal, is badly exposed. The coal was formerly mined by the inhabitants of Aguiró. The best-preserved plant imprints come from the lower coal stringers, which deliver an Upper Westphalian flora (Roussel, 1904, p. 20; Dalloni, 1930, pp. 95–96). For an extensive fossil list, we refer to these authors and to Nagtegaal (*ibid.*).

#### *Erill Castell Volcanic Formation (Stephanian)*

The Erill Castell Volcanics, as defined by Mey *et al.* (1968), represent the rock unit composed of mainly





A. Lower breccia member.



B. Upper conglomerate member.

Fig. 7. Typical appearance of the Aguiró Formation.

light-coloured tuff containing bombs of up to 1 m, which in the Erill Castell-Peranera area is overlain by a dark-green, massive andesite sheet. The lower boundary is drawn at the Hercynian unconformity or, as in the Castelnou-Aguiró area in the east, at the contact with the fluvial Aguiró Formation. The upper boundary is drawn at the contact with the fluvial sediments of either the Malpas Formation or the Peranera Formation.

A thick and well-exposed section of this formation can be studied in the surroundings of the village of Erill Castell (coord.  $42^{\circ}25\frac{1}{2}'$  .  $4^{\circ}29'$ ), after which this formation is named. Between the unconformity on top of the Devonian and the first pyroclastics there is a lense-shaped, badly sorted breccia about 10 to 30 m thick and continuing over a distance of a few hundred metres. The angular components,

mainly slates and limestones, are surrounded by a matrix of quartz sand and small chips of shale. The good exposures show that this breccia fills cracks and channels in the erosion surface on top of the Devonian. The breccia does not show signs of stratification and Nagtegaal considers it to represent a fossil slope breccia comparable to the basal member of the Aguiró Formation.

The pyroclastics, then, start at their base with a few metres of probably reworked light-coloured tuffs, locally showing cross-bedding. They are followed by non-reworked tuffs which are porous and poorly cemented.

Fragments occurring embedded in an indistinct, possibly devitrified matrix, consist of twisted Palaeozoic slates, and bombs of basic igneous rock measuring up to 1 m. Small lenticular intercalations of tuff breccia and reworked tuff are present. The colour of the tuffs varies from white, yellow, and brownish to pink, greenish, and violet. The entire sequence is stratified, locally graded, and about 150 m thick.

A comparable sequence has been observed elsewhere in the map-area. Locally, the tuffs are silicified (e.g. south of the beautiful Aguiró section) and then the rocks become very hard and brittle.

In the Erill Castell-Peranera area the tuffs are overlain by a dark-green andesite sheet, nearly 200 m thick, in the lower part of the Peranera valley. South of the village Peranera this lava flow is conformably overlain by the fluvial Malpas Formation, but south of Erill Castell another tuff occurrence is intercalated between the two rock units. According to Nagtegaal (pers. comm. 1968) the dense, dark-green rock of the lava sheet is a basaltic andesite (average  $\text{SiO}_2$ -content: 53.5 %). In thin section the rock shows a porphyritic, pilotaxitic texture. In the matrix, consisting mainly of labradorite microlites, one finds small phenocrysts of hypersthene and large crystals of labradorite and a fully altered mafic mineral, probably originally olivine.

For a more detailed mineralogical description as well as the results of a number of chemical analyses of this rock, we refer to the thesis of Nagtegaal (in preparation).

Elsewhere in the Nogueras zone, smaller andesite lenses were observed locally between the tuffs, but these are too small for individual mapping.

#### *Malpas Formation (Stephanian)*

The Malpas Formation has been named after the village of Malpas (coord.  $42^{\circ}24\frac{1}{2}'$  .  $4^{\circ}29'$ ), the centre of the still active coal mining in this district. This formation, as defined by Mey *et al.* (1968), comprises a series of dark-grey and brownish fluvial sediments with intercalations of coal seams up to 6 m thick. The lower boundary coincides everywhere in the map-area with the contact with the Erill Castell Volcanics (tuff breccia or andesite) and the upper boundary may be either the first red fluvial sediments of the Peranera Formation or the angular unconformity with the also red Bunter. The thickest sequence measured



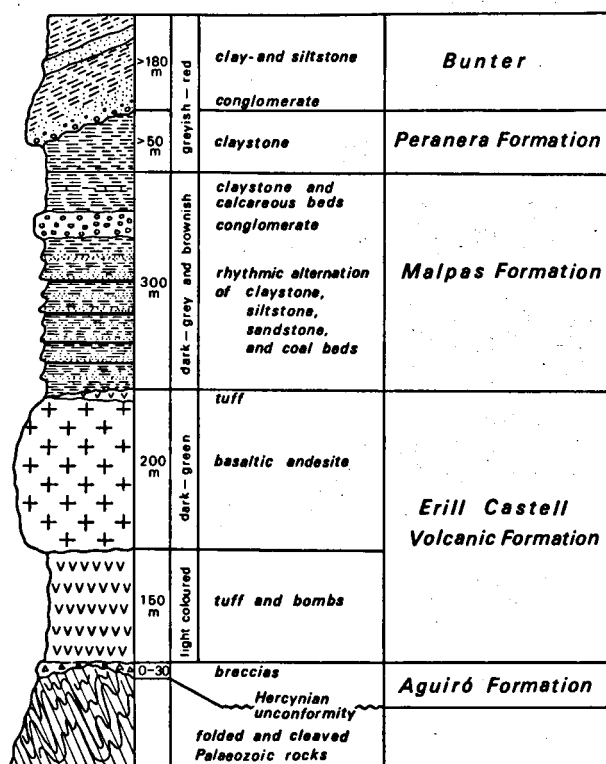


Fig. 8. Columnar section of the post-Hercynian Palaeozoic formations taken in the lower part of the Peranera valley.

occurs in the lower part of the Peranera valley (nearly 300 m thick), where this formation is conformably overlain by the greyish-red Peranera Formation. The best exposed and almost complete sections of this formation can be found south of Erill Castell (Fig. 8) and in the surroundings of the village Peranera. There, the formation consists of medium-grey to black micaceous shales and mudstones, light-grey sandstones and greywackes, a few reworked tuff beds, one horizon of coarse, well-rounded conglomerates, thin siderite beds and concretions, bedded cherts, some impure limestones, and a few actively mined coal seams. The thickness of the individual beds varies from 1 cm to several metres. The various lithofacies are distributed in an orderly way in a vertical sense: a considerable number of upward-fining cycles have been recorded by Nagtegaal (pers. comm., 1967). Vegetational debris abounds in most beds; marine fossils are entirely lacking. An abundance of plant material has been determined, and there can be no doubt that at least the lower part (first 200 m) of this formation is of Stephanian age. For an extensive list of fossils, we refer to Dalloni (1930, p. 94) and Nagtegaal (thesis, in preparation).

## PERMO-TRIASSIC

### Introduction

Some authors have treated the entire non-fossiliferous sequence of red fluvial sediments on top of the dated

Stephanian and below dolomitic limestones and gypsum-bearing marls as the Permo-Triassic (Misch, 1934; Ashauer, 1934) or as the Permian (Dalloni, 1910). Later investigations of Dalloni (1930) paid more attention to the often present angular unconformity in this sequence, as a result of which he considered the sequence below the unconformity to be Permian and the one above Lower Triassic. He even collected an extensive flora in the Pallaresa valley just south of the village Baro along the main road (Dalloni, 1930, p. 114). Schmidt (1931, p. 173) reported still other plants from the same outcrop. Mainly on the basis of the occurrence of *Walchias*, *Callipteridea*, and *Ullmannia frumentaria*, both authors attributed these brownish to blackish detrital rocks to the Permian. A more thorough analysis of their recorded plant material does indicate that the assemblage is somewhat younger than the one of certainly Stephanian age, collected in the Peranera valley (Dalloni, 1930, p. 94), but it cannot be considered to exclude the possibility that this sequence south of Baro still belongs to the Stephanian C (van Amerom, pers. comm., 1968). In spite of this uncertainty about its age, it is obvious that according to our definition this (not red) sequence south of Baro belongs to the Malpas Formation.

We therefore consider that in the map-area and adjoining Nogueras zone it has not yet been either proven or disproven by fossil evidence that the red beds below and above the unconformity represent the Permian and the Lower Triassic respectively. We consequently prefer, like many other authors (e.g. Mirouse, 1959; v. d. Lingen, 1960; Roger, 1965), to employ formation names instead of these time-stratigraphical names. Nagtegaal (1962, internal report) proposed the following denomination:

Bunter (used as a formation name)

~~~~~ unconformity

Peranera Formation

The lithological differences between the two formations are well exposed, so that even in areas where the unconformity is less visible, both formations can easily be mapped individually. The main macroscopical differences between these two types of red beds are listed below (Nagtegaal, 1962; Virgili, 1961; Roger, 1965):

### Peranera Formation

breccias with limestone fragments

commonly lime-bearing, often in the form of nodular beds

very few, small-sized micas

much reworked tuff material

### Bunter

well-rounded conglomerates without limestone pebbles

not lime-bearing

large-sized detrital micas common

no recognizable tuff material

|                                                             |                                                   |
|-------------------------------------------------------------|---------------------------------------------------|
| bad sorting and<br>angularity of rock-<br>forming materials | better sorting and round-<br>ness of the material |
|-------------------------------------------------------------|---------------------------------------------------|

|                      |                      |
|----------------------|----------------------|
| cross-bedding scarce | cross-bedding common |
|----------------------|----------------------|

|                                         |                     |
|-----------------------------------------|---------------------|
| many green reduction<br>spots or layers | few reduction spots |
|-----------------------------------------|---------------------|

The microscopic differences between the two formations are even more conspicuous, but for these we refer to the thesis of Nagtegaal (in preparation).

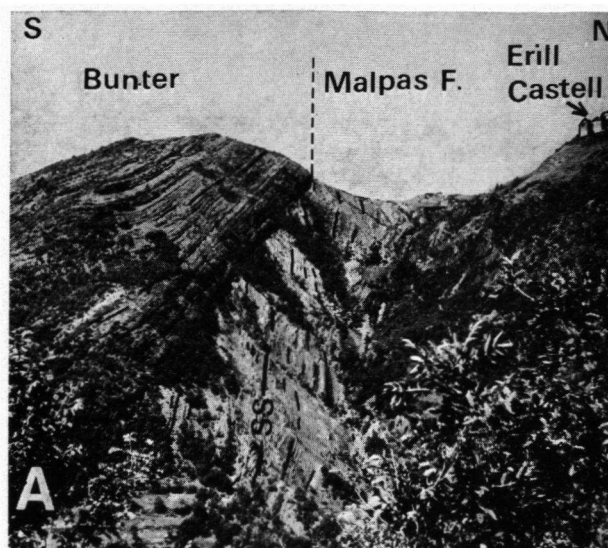
#### *Peranera Formation (Permian?)*

The Peranera Formation, as defined by Nagtegaal (1962, internal report; Mey *et al.*, 1968) and named by him after the Peranera River, consists of a monotonous, greyish-red sequence with alternating mudstones, siltstones, sandstones, tuffs and transported tuff material, breccia beds, and nodular limestone beds. In the area east of Erill Castell and in the Gotarta block, where this formation overlies the Malpas and Erill Castell formations in a conformable way, the lower boundary is drawn at the top of dark-coloured shales with intercalated brownish sandstones (Malpas Formation) or at the contact with the Erill Castell Volcanics. In the area west of the Tor River the lower boundary is the plane of unconformity on top of the Hercynian folded rocks. The upper boundary is drawn at the sharp contact with the first coarse quartz-sandstones or pebble beds belonging to the Bunter. East of Erill Castell this contact is a conspicuous angular unconformity, its angle varying from 10 to about 30° (Fig. 9).

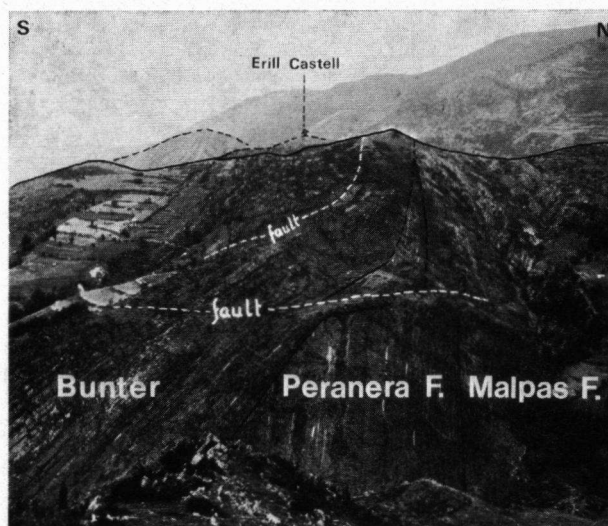
In the map-area the thickest measured sequence of this formation (about 700 m) is found in the Mañanet and Valiri rivers, north of Las Iglesias. In these sections the lithology of the lower part (about 200—300 m) is somewhat comparable with the Malpas Formation, differing in that the sequence is continuously greyish-red and lacks coal clays, coal seams, and the coarse conglomerate.

Breccia beds with angular fragments of mainly Palaeozoic limestones alternate with mud- and siltstones, arkosic sandstones and greywackes, thin tuff sheets or beds exclusively built up of transported tuff material, and nodular limestone beds. The thickness of the individual beds varies from several centimetres to up to several metres.

The upper part of the formation is less differentiated and one finds mainly mud- and siltstones and impure badly sorted sandstones, all of which are red. The fact that tuff relicts decrease in this upper part is an indication that the volcanism ceased and also that the Stephanian tuff deposits in the north have been eroded away. The absence of coarse material points to a more denudated relief of the source area. River transport and sheet wash under arid climatologic conditions are thought to have contributed to the formation of this typical rock sequence (Nagtegaal, pers. comm., 1967).



A. South of Erill Castell.



B. Southwest of Sas.

Fig. 9. Angular unconformity of the Bunter on Stephanian-Permian rocks (photos by Nagtegaal).

The Peranera Formation can be correlated with the "Série de Somport" (Mirouse, 1959) and the series "P<sub>1</sub>" and "P<sub>2</sub>" of van der Lingén (1960) in the Western Pyrenees, and with the Permian series (Lamare, 1936) in the Basque Pyrenees.

#### *Bunter (Lower Triassic)*

The Bunter (used as formation name) has been defined by Mey *et al.* (1968) as the rock unit consisting of a predominantly greyish-red sequence of conglomerates, coarse-grained quartzose sandstones with or without pebbles, finer psammitic sandstones, and silt- and mudstones, the medium grained rocks commonly being micaceous.

The lower boundary is almost everywhere an angular

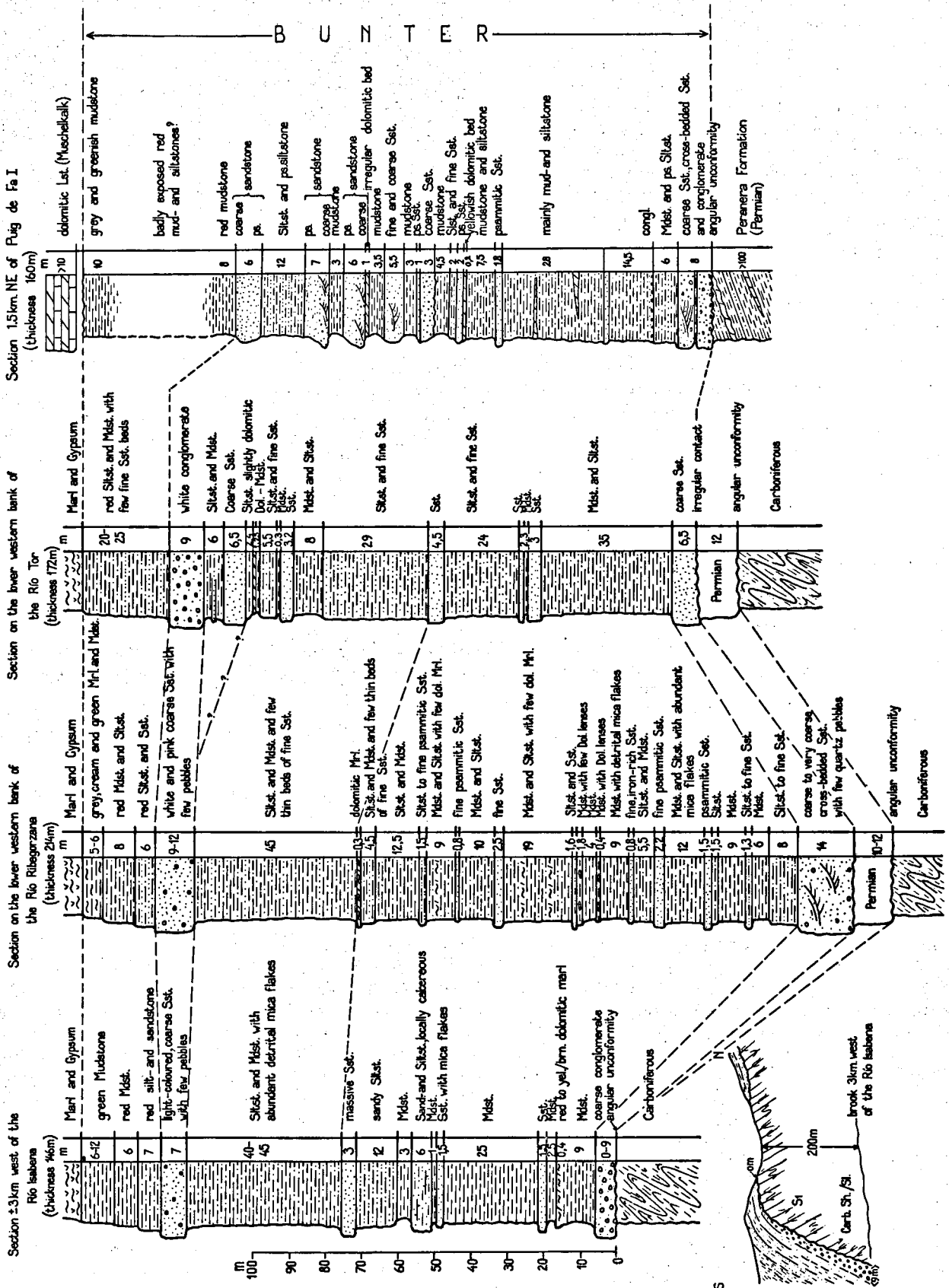


Fig. 10. Columnar section of the Bunter, taken in the northern border of the Nogueras zone.

unconformity on rocks ranging in age from Lower Devonian to Permian (?). West of the Tor River and in the Gotarta block, the lower boundary of this formation has been drawn at the sharp but conformable contact with the lime-bearing and breccious Peranera Formation. The upper boundary is drawn at the first occurrence of dolomitic marl, often yellowish and frequently showing light-coloured gypsum veins. Very often, however, the upper boundary is a fault, and then one can expect everything from Palaeozoic slates up to Lower Cretaceous limestones. For the main differences between the Bunter and the Peranera Formations, see pp. 241–242.

The Bunter, which in the studied area has an average thickness of 200 m and ranges between 150 and 300 m, is very widespread and rather uniformly developed in the entire Pyrenees and far outside this mountain range. The lowermost part of this formation is formed almost everywhere in the Noguera zone by a coarse-grained, cross-bedded quartz-sandstone (Fig. 10), with or without well-rounded pebbles of white vein quartz, quartzite, or black chert, or by a coarse, well-rounded conglomerate with mainly quartzite pebbles sometimes having a diameter of up to 15 cm. This competent basal unit is up to 15 m thick and shows grey to pink colours. These coarse detrital rocks are generally overlain by a red sequence of alternating mud- and siltstones with few psammitic sandstones, which only in the upper part become coarser to locally conglomeratic. Many sedimentary structures such as cross-bedding, wave and stream ripples, slumps, load-casts, graded bedding, and other bottom structures provide good evidence for the determination of the top and bottom of the beds; this information is essential for solving the structures in the more complicated parts of the Noguera zone.

The few outcrops of Bunter in the axial zone have in common that their basal layers are strongly dependent on the nature of the underlying Palaeozoic formations. The Bunter starts:

- a. with a very coarse quartzite-breccia, when it lies on Devonian quartzite (Fig. 11);
- b. with a fine limestone breccia, when it lies on carbonate rock;
- c. with silt- or sandstones, when lying on shale or slate.

Where there has been no tectonic disturbance, the uppermost part of the formation consists of red, green, grey, and blackish mudstones/shales with a total thickness of up to 25 m. These are in turn conformably overlain by yellowish and light-grey dolomitic marls with gypsum veins, which we assign to the Muschelkalk/Keuper sequence. This part of the formation is considered by some authors (e.g. Misch, 1934; Virgili, 1958) to represent the Röt, the transition layers between the Buntsandstein and the Muschelkalk in Germany.

So far, Dalloni (1930, p. 115) is the only one ever to have found a recognizable fossil in this formation in the Pyrenees, his now famous finding near Guils of imprints of *Equisetum arenaceum*, proving a Triassic age



Fig. 11. Basal breccia of the Bunter as it appears when lying on competent rocks in the axial zone, here on the Basibé quartzites.

for these rocks. Our own attempt to isolate pollen from the grey shales in the upper part of the Bunter was not successful.

The general lithology, the arbitrary mixture of coarse and fine sediments, and the abundance of sedimentary structures, all point to a fluvial origin of this formation under probably semi-arid climatological conditions. The Bunter corresponds to the last chapter of the Hercynian history of the Pyrenees, and to the first one of the Alpine history. It represents the peneplanation stage. Low relief and slow erosion permitted prolonged reworking and weathering before sedimentation, and produced the well-sorted, highly rounded sandstones and conglomerates of this formation.

## MIDDLE AND UPPER TRIASSIC

### *Introduction*

Owing to the lithological similarity to the Germanic Triassic, it seems reasonable to suppose that the limestone/dolomite unit on top of the Bunter represents the Muschelkalk and the overlying mudstone-marl-gypsum sequence the Keuper, but so far these sediments have yielded few and rather insignificant fossils in the Central Pyrenees.

Since east of the Segre valley several thick limestones alternate with gypsum and marl beds, the distinction between the two units becomes somewhat arbitrary. In the rest of the Noguera zone one often finds a chaotic mixture of an incompetent marl/gypsum matrix in which float slabs of the more competent limestone/dolomites. It was therefore decided to group these two units under one formation name, i.e. Pont de Suert Formation, and to treat the individual mappable units as members when possible.

### *Pont de Suert Formation (Middle and Upper Triassic)*

This formation is named after the village of Pont de Suert (coord. 42°24' . 4°26'), in the vicinity of which occur several well-exposed, only weakly deformed and almost complete sections of this formation (Fig. 12). The Pont de Suert Formation, as defined by Mey *et al.*

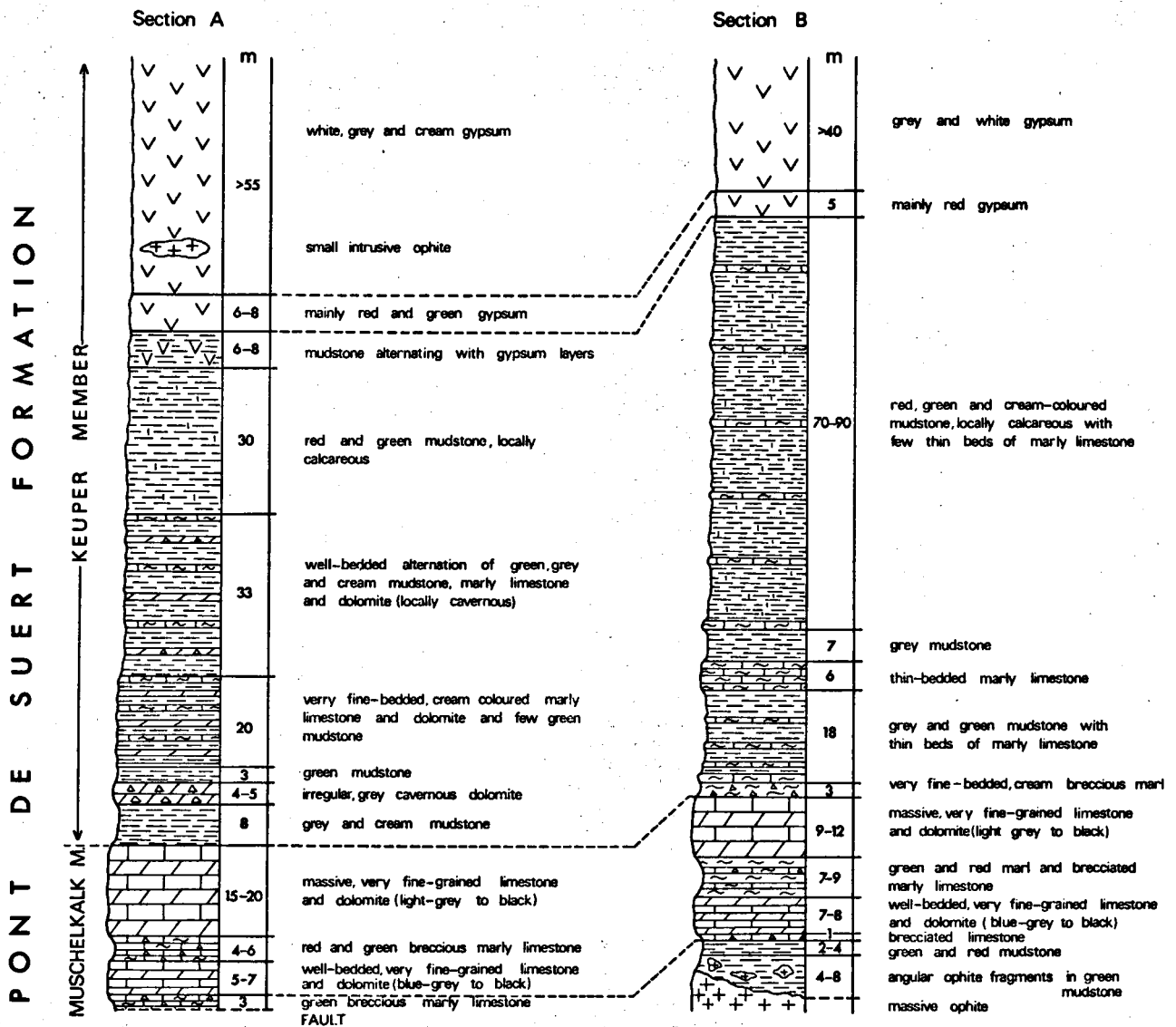


Fig. 12. Columnar sections of the Pont de Suert Formation;

Section A: taken near the road tunnel 2 km south of Pont de Suert.

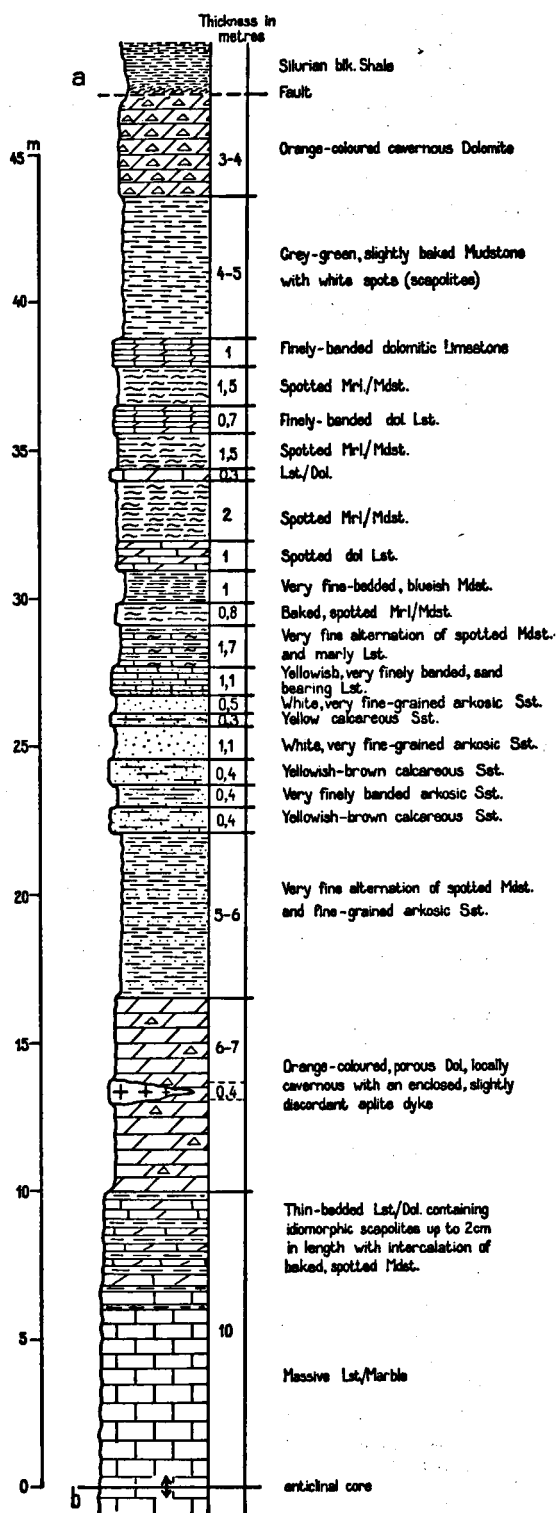
Section B: taken on the lower eastern bank of the Tor River, 800 m north of Pont de Suert.

(1968, p. 223), comprises in the lower part mainly dark-greyish to black micritic limestone and fine-grained dolomite (Muschelkalk member) and in the upper part a usually strongly tectonized alternation of grey, red, and green dolomitic marls, cavernous dolomite, vividly-coloured gypsum, and rock salt (Keuper member). Restricted to the incompetent upper member are small and large bodies of basic rock (ophites).

Almost invariably, the lower as well as the upper boundaries of this formation are tectonically disturbed. In the scarce outcrops, where an undisturbed transition from the Bunter is present, the lower boundary has been drawn at the base of yellowish dolomitic marl, gypsum, or limestone/dolomite lying above dark-grey, greenish, or reddish shales of the uppermost Bunter.

When not faulted, the limit between the Pont de Suert Formation and the overlying Bonansa Formation is drawn at the base of a competent carbonate unit in which thinly stratified, non-fossiliferous marly and dolomitic limestones and dolomites alternate.

In only a few outcrops is it possible to see an undisturbed transition from the Muschelkalk into the Keuper member. The thickness of the limestone/dolomite unit with an often intercalated marl horizon then varies between 25 and 50 m. More generally, however, thin and thick limestone/dolomite slabs float in a heavily distorted mass of vividly-coloured marls and gypsum. Viewed from a distance the competent limestone/dolomite units have a massive appearance, but on closer scrutiny they show a very characteristic sharp, close-spaced fracturing along the bedding and/or



Map and cross-section through the thrust-zone with enclosed Triassic rocks

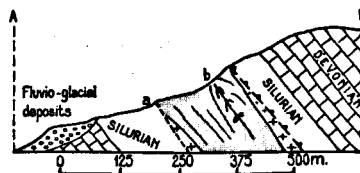
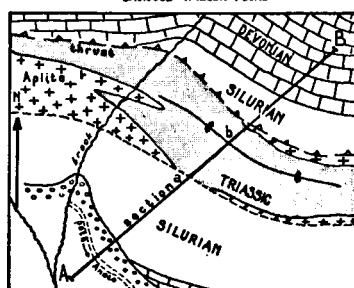


Fig. 13. Columnar section of the northern Triassic, taken 1800 m northwest of the village of Aneto.

joints. Near Pont de Suert cross-bedding has locally been observed in dolomitic limestones.

We found no fossils, but Dalloni (1930, p. 137) mentions from the Muschelkalk occurrence north of the Puig de Fa (coord.  $42^{\circ}23' . 4^{\circ}35\frac{1}{2}'$ ) the finding of *Nucula* sp. and *Lingula tenuissima* Bronn. Near Aulet he found many small fossils (e.g. *Natica gregarea*) in marls associated with these limestones.

The yellow series of dolomitic marls and thin, platy limestones lying in alternation on top of the massive Muschelkalk, grade upwards into mainly red and green mudstones having a total thickness of 30 to 90 m. These are followed by rather pure gypsum, the first few metres still vividly-coloured but becoming a monotonous white, grey, and black higher in the stratigraphy. At the surface, this Keuper member seldom contains salt and then only in fresh outcrops where a light salty taste betrays the presence of this mineral in the unweathered rock. In view of the salt springs of Gerri de la Sal, Cambrils, Salinas de Jaca, Salinas de Sin, etc., salt must still occur in larger quantities at greater depths.

No fossils were found, nor are there any mentioned in the literature.

The transition into the overlying thinly-bedded dolomitic limestones of the Bonansa Formation is nowhere a normal contact in the map-area. The contact is either a normal fault or the contact between a competent carbonate unit and diapyrically intruded Keuper series. According to the literature (Misch, 1934, and Virgili, 1958), the uppermost part of the gypsum member should contain yellowish, cavernous dolomites (*Zellenkalk*, *cagneules*, *carniolas*). In our area only small drifting bodies of this so-called *Zellenkalk* were encountered in the plastic gypsum-mudstone series, but these bodies were too small to be included on the map.

The matrix of the Keuper carries, besides the already-mentioned slabs of limestone/dolomite and *Zellenkalk*, small and large bodies of basic rock consisting of plagioclase and clinopyroxene, with epidote, prehnite, aerinite, and chlorite as alteration products. Owing to its ophitic texture, this basic rock is referred to in the French literature as ophite. Its colour is dark-green to greenish-brown. The surrounding sediments show no signs of contact metamorphism.

The time of emplacement of these ophites may have been either simultaneous with the deposition of the Keuper, in which case they could be extrusives, or later, which would make them intrusives. The only certainty is that they are pre-Alpine, since ophite fragments are frequently found in the middle and upper parts of the Campo Breccia in the Esera valley, which is of Santonian age (Souquet, 1963, 1964).

The narrow Alpine thrust zones occurring in the axial zone enclose locally in the northern area (e.g. Senet thrust zone) orange-coloured carbonate rocks bearing considerable resemblance to the lower and central part of the Pont de Suert Formation. A difference which is most probably due to the strong tectonization and the thermal influence of the accom-

panying aplite dykes is the scapolitization of these rocks. Pure marbles also occur. The scapolites occur either as tiny white spots (0.5–5 mm), or as black, idiomorphic crystals up to 2–3 cm long. The largest and best preserved section of these rocks can be seen in an outcrop some 1800 m northwest of the village of Aneto. This section, shown in Fig. 13, was taken in the overturned limb of an anticline. For more details on these rocks we refer to Mey (1967-b, pp. 178–179 and 214).

The suggested Middle and Upper Triassic age for the Pont de Suert Formation is not based on the scarce fossil discoveries but exclusively on its lithologic resemblance to the better dated Triassic outside the Pyrenees, such as the Triassic of the Catalan Mts. (Virgili, 1958) and the Triassic of the Celtiberic Chain southwest of Zaragoza (Würm, 1911). The entire Triassic thus shows a close affinity to the Triassic of most of Western Europe.

## JURASSIC AND CRETACEOUS

### Introduction

The younger Mesozoic rocks are predominantly marine calcareous deposits, commonly with an abundance of micro- and macrofossils permitting satisfactory age determinations. Detailed mapping of the various rock units has shown that considerable facies changes and diachronism occur in and near the map-area (Souquet, 1967).

For these reasons and the fact that the thick Jurassic dolomites yielded hardly any fossils, it was considered necessary to subdivide the younger Mesozoic rock sequence too into lithological units or formations in the sense of the Code of Stratigraphic Nomenclature (1961). This subdivision, which has proved its usefulness during several years of internal use, has been published separately by Mey *et al.* (1968).

Only the formations occurring within the area covered by sheet 8 (Fig. 5) will be described in more detail here. For the palaeogeography, the palaeoclimate, and the depositional environment of these rocks, we refer to Dalloni (1910, 1930) and to the recent publication of Souquet (1967).

### Bonansa Formation (*Liassic-Dogger*)

The strongly tectonized and often diapyrically intruded Pont de Suert Formation is overlain in the map-area by a marine sequence of mainly dolomites with an intercalation of highly fossiliferous dark marls and marly limestones. This rock unit is named after the village of Bonansa (coord.  $42^{\circ}25\frac{1}{2}' . 4^{\circ}21'$ ),  $2\frac{1}{2}$  km east of which a well-exposed continuous outcrop of this formation occurs north and south of the road tunnel (Fig. 14).

The Bonansa Formation, as defined by Mey *et al.* (1968, p. 223), is the rock unit whose lower part consists of a competent, fine-grained carbonate unit in which alternate thinly-stratified marly and dolomitic limestones and dolomites (which may locally be cavernous), a central part of dark shales, marls, and



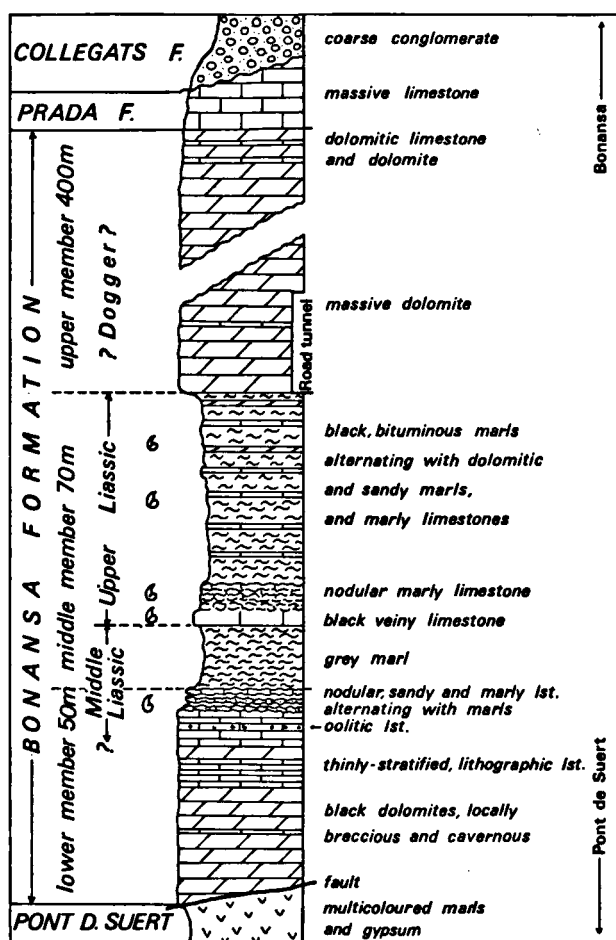


Fig. 14. Columnar section of the Bonansa Formation, taken along the road from Pont de Suert to Bonansa.

marly limestones (the latter being very fossiliferous), followed by a competent coarse-grained sequence of mainly massive dolomites with a characteristic dark-grey to black weathering colour.

The lower boundary of this formation is most often faulted, but when undisturbed it is drawn at the base of a competent carbonate unit below which dolomitic marls occur, sometimes locally with a cavernous texture. The transition into the overlying Prada Limestone Formation is rather gradual, but the boundary is drawn at the top of a sequence in which dolomites dominate over limestones.

The lower member of the formation, which is about 30 to 60 m thick, is the equivalent of what Misch (1934) calls Carniolas, a descriptive term based upon the sometimes cavernous texture of the weathered rock. More generally, however, the fine-grained rock is very thinly stratified, giving it the appearance of a lithographic limestone. Very often the limestones are marly and/or dolomitic, and are dark-grey on a fresh cut but show a characteristic beige, pink or slightly violet weathering colour. In the highest parts the limestones are oolitic locally. Especially in the lower part

there are veiny dolomites which generally are more massive, darker in colour, and greyish when weathered; locally, they are breccious and/or cavernous. Fossils are absent except for the uppermost beds, which are nodular, coarser, marly and slightly sandy, and weather yellowish-brown. Belemnites, lamellibrachiates, and brachiopods predominate.

Owing to its lithology and its stratigraphic position on top of the Keuper and below the fossiliferous horizon dated as Middle Liassic, Dalloni (1910, 1930) and the authors of the geological maps of Lerida and Huesca consider this rock unit to represent the Infra- and Lower Liassic.

The succeeding incompetent member, which is 40 to about 70 m thick, comprises a series of bituminous shales, marls, and marly limestones, the latter generally being very fossiliferous (mainly cephalopods, brachiopods and lamellibrachiates). The shales and marls have a dark-grey to greyish-brown weathering colour, contrasting with the beige to yellowish tones of the marly limestones. On fresh cuts all the rocks are dark-grey to black. For extensive fossil lists we refer to Vidal (1875, 1898), Mallada (1878), Dalloni (1910, 1930), Almela *et al.* (1947), and Alastrue *et al.* (1958). The fossils indicate a Middle and Upper Liassic age.

The upper member of the Bonansa Formation is a massive, coarse-grained, porous, and locally breccious dolomite, resting with a sharp but conformable contact upon the bituminous marls. They have a characteristic dark-grey to black weathering colour. On fresh cuts the saccharoidal rock ranges from light-grey to black. Except for a few indeterminable shell fragments in the upper part of the sequence, no fossils have been found. The bedding is hardly noticeable, except in the lowermost and highest parts where a few intercalations of dark dolomitic limestone beds occur. The total thickness varies widely from place to place; it is thickest in the Bonansa area, where we measured about 400 m. North of the Fallada de Malpas (coord.  $42^{\circ}22\frac{1}{2}'$  .  $4^{\circ}27'$ ) the dolomite member measures only 20 to 40 m.

In the literature this dolomite member is often referred to as the Dogger dolomite. This denomination is not based on fossil discoveries within the series but on datings of a 100 m thick sequence of lithographic limestones, which in the Sierra de Montsech is intercalated between the dolomite member and the overlying Prada Formation (Urgonian). Vidal (1915) was the first to describe the limnic fauna of mainly reptiles, fishes, and insects, and plant imprints, an assemblage pointing to a Kimmeridgian age. The dolomite below must therefore be Kimmeridgian or older, but younger than Upper Liassic.

The fact that in most of the map-area the Prada Limestone Formation (Urgonian) is in immediate contact with the Keuper member of the Pont de Suert Formation, does not imply that the Bonansa Formation is missing; in our opinion, this phenomenon is due to the diapyric intrusion of the incompetent gypsum-marl series (see cross-sections).

For a detailed palaeogeographic description of the Jurassic of the Pyrenees and neighbouring areas, we



refer to Dalloni (1930), Richter and Teichmüller (1933), and Misch (1934).

#### *Prada Limestone Formation (Urgo-Aptian)*

In an apparently conformable way the dolomite member of the Bonansa Formation is overlain by a thick series of massive limestones. This rock unit is named after the Sierra de Prada (sheet 9, in preparation, coord.  $42^{\circ}15' . 4^{\circ}55' - 5^{\circ}02'$ ) a mountain range west of the Segre River that is formed by these limestones. A very thick and characteristic section is exposed along the road from Orga a to Seo de Urgel.

The Prada Limestone Formation, as defined by Mey *et al.* (1968, p. 224), is the massive, dark-grey to black, fossiliferous and predominantly micritic limestone overlying the Bonansa Formation. In its lower part dolomitic and breccious intercalations abound, and in its upper part there are some intercalations of marly limestone and marl.

The transition from the underlying dolomites of the Bonansa Formation is gradual, but the boundary is drawn at the base of a sequence containing far more limestones than dolomites. The upper boundary is drawn at the top of a sequence in which limestones and marly limestones dominate over marls. The contact may be sharp or gradual.

Although the thickness of this formation changes considerably from place to place in the map-area, the extremes being zero to a few metres NW of Bonansa and 700 to 800 m in the Fallada de Malpas, we did not find direct field evidence for an unconformity below and/or above these limestones. An angular unconformity amounting locally to values up to  $30^{\circ}$  between these limestones and the underlying dolomites has, however, been described by Hup  (1954) in the area immediately west of the Isabena River, where this author also found evidence of an erosion period after the deposition of the Prada Limestone. The gradual thinning of the limestones in a northward direction, so well-demonstrated in the Bonansa area (from 375 m in the Cruz de Bonansa to zero some 1500 m north of Bonansa), where Albian sandstones rest directly on the Dogger dolomite, could, however, be taken to point equally well to a different rate of subsidence of the Urgonian basin.

Conforming to the definition, the lower part of the formation is a very massive, dark-grey to black micritic limestone, often breccious and veined with calcite. In some sections thin dolomite beds and irregular dolomite masses are intercalated. The weathering colour of both limestones and dolomites is a light-grey. Except for miliolids and on the weathered surface fragments of rudists and oysters, fossils are rare. Higher in the series the carbonates reflect an influx of fine terrigenous material: the limestones become thinner bedded and lighter in colour, are frequently marly, and show an abundance of micro- and macro-fossils (mainly orbitolines, *Toucasias*, *Reguienias*, and oysters). For some local cases, as for example north and south of the Barranco de Viu, the thickest marl intercalation (up to 50 m) has been mapped separately.

Some 200 m north of the hamlet of Pallerol (coord.  $42^{\circ}20\frac{1}{2}' . 4^{\circ}23\frac{1}{2}'$ ) such a marl horizon is accompanied by a thin intercalation of lignite, formerly exploited by the inhabitants.

The top layers of the formation are generally marly, have a nodular aspect with a gregish-brown weathering, and contain many oysters and orbitolines. The succeeding marls and marly limestones (Llusa Marl Formation) are generally darker in colour and have a more shaly appearance. Within 5 m a mappable boundary can always be drawn. In the Bonansa area the boundary between the two formations is a rather sharp contact between massive, micritic limestones overlain by shaly marls with many solitary corals and molluscs. For extensive fossil lists of this formation within the map-area and the immediate surroundings, we refer to Dalloni (1910, pp. 205—212; 1930, pp. 175—176, 179). For a general outline of the pre-Cenomanian palaeogeography we refer to the same author, to Misch (1934), and to Souquet (1967).

On the basis of macro-fossils it could only be demonstrated that the Prada Limestone represents the Lower and the main part of the Upper Aptian, and that the transition into the overlying marl formation lies in some sections high in the Upper Aptian and in others coincides with the limit between Aptian and Albian (Dalloni, 1910, 1930). With micro-fossils (mainly orbitolines and ostracodes), however, it could be shown for the Segre valley, where this formation is more than 1500 m thick, that the limestones also represent the Barremian and the Hauterivian, and probably even the Upper Valanginian (Pruvost, 1961). The lowest of these significant fossil discoveries still lies 400 to 500 m above the contact with the Dogger dolomite.

Just as this report was going to press we received a short article in which Peybern s (1968) demonstrates the presence of still older stages in the Prada Formation of the Segre section on the basis of micro-fossils of Neocomian, Portlandian, Kimmeridgian, and even Oxfordian age.

Hofker (1963), who made a detailed study of the evolution of the genus *Orbitolina*, demonstrated for two localities in and near the map-area (north of Pallerol and south of Senterada; Fig. 15), that the uppermost part (50—100 m) of the Prada Limestone is already of Albian age. This contradicts the datings based on macro-fossils, which point, at least in the case of the Senterada section, to an Upper Aptian age for the upper part of the Prada Limestone and the lowermost part of the overlying Llusa Marls (Dalloni, 1930, p. 179; Souquet, 1967, p. 57).

#### *Llusa Marl Formation (Albian)*

In a generally conformable way the massive Prada Limestone is overlain by a series of dark marls. This rock unit is named after the hamlet of Llusa in the Flamisell valley (Fig. 15, coord.  $42^{\circ}18\frac{1}{2}' . 4^{\circ}38'$ ), which is surrounded by a very well-exposed and typical sequence of these marls.

The Llusa Marl Formation, as defined by Mey *et al.*

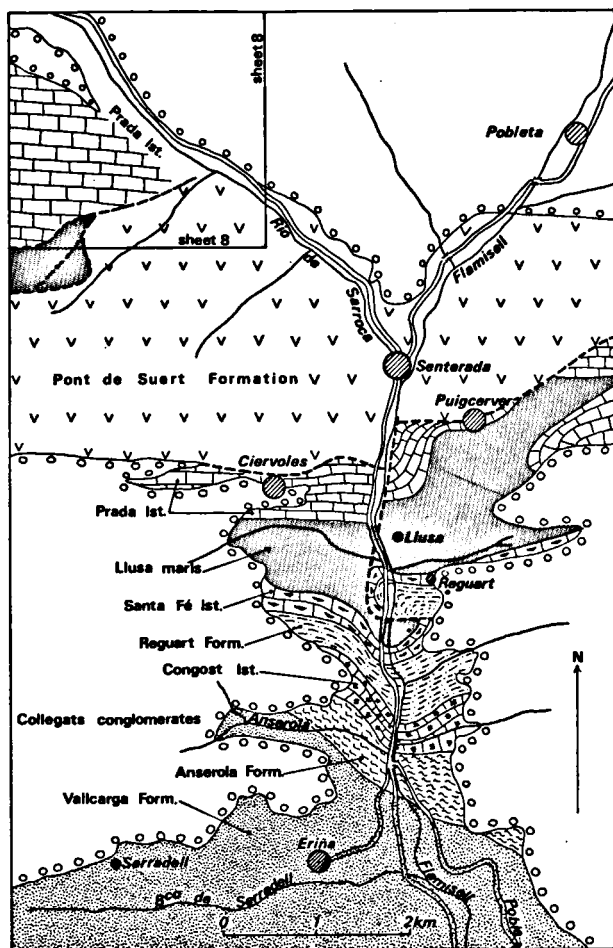


Fig. 15. Geological sketch-map of the Flamisell valley north and south of Senterada (after van Leeuwen, 1966).

(1968, p. 244), comprises the rock unit consisting of a greyish, monotonous alternation of fossiliferous shaly marls and silty marls (often nodular) with a few intercalations of thin-bedded marly limestones. In the Santorens-Aulet area the Llussa Marls include a few thin intercalations of sandier material.

In most of the map-area the transition into the underlying Prada Limestone is gradual, the lower boundary being drawn at the base of a sequence in which marls dominate over limestones. In the Bonansa area this boundary is a sharp but probably still conformable contact between dark micritic limestones below and shaly, very fossiliferous marls above.

Generally, the Llussa Marls are conformably overlain by *Orbitolina*-bearing limestones of the Aulet Formation, the boundary being drawn where dark, incompetent marls give way to more competent limestones. In the area northwest of Santorens and in the Bonansa area, however, the Llussa Marls are unconformably overlain by conglomerates, sandstones, and sandy limestones of the San Martin Formation. The Llussa Marls, which Misch (1934) called the "*Unterste Mergel*" (lower marls), displays, within the map-area, detrital sediments with a strong variation

in carbonate content, fossil assemblage, and thickness. The mainly dark-coloured rocks can vary from shale and shaly marl to argillaceous limestone. The weathering colour varies from dark-grey to a dull yellowish-brown. The finest and darkest marls and shales, which are often glauconite-bearing and slightly bituminous, are restricted to the upper part of the formation. Sponge needles and quartz grains are frequently occurring elements of all sediments.

The fauna may be benthonic and/or pelagic. The macro-fossils (mainly cephalopods and molluscs) are often pyritized. For an extensive list of fossils collected in and near the map-area, we refer to Dalloni (1910, pp. 209–212; 1930, p. 179).

The total thickness of the formation varies considerably within the map-area. It is thickest in the south, where it probably exceeds 1000 m. A minimum thickness of about 20 to 40 m is found south of the Urganian of the Fallada de Malpas; this low value may, however, be due to a tectonic thinning. In the Montiberri area and south of the Barranco de Cirés about 200 m of marls are exposed.

The older datings, based mainly on macro-fossils, indicate that the Llussa Marls, including the upper sandstone in the western areas, range in age from Upper Aptian to Upper Albian (Dalloni, 1910, 1930).

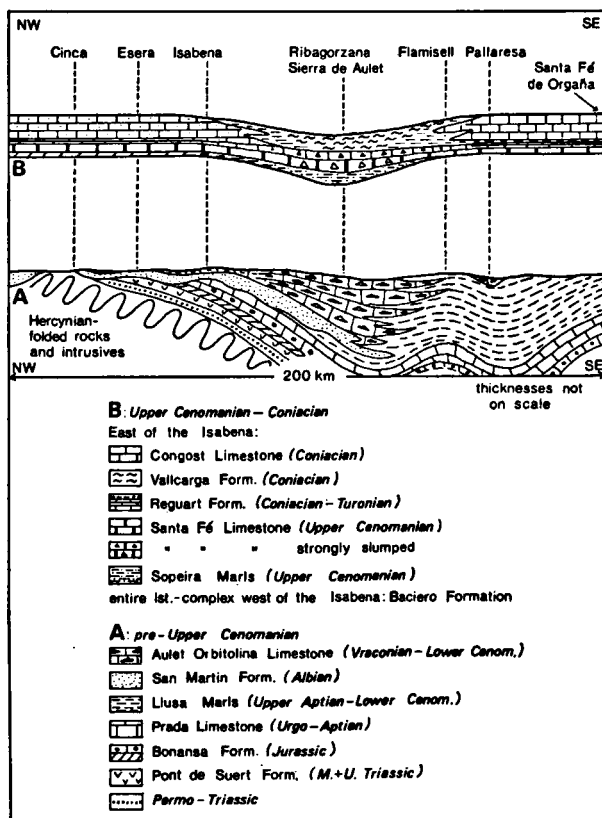


Fig. 16. Schematic NW-SE cross-section through the Mesozoic basin between the Cinca in the west and the Segre in the east (modified after Souquet, 1967; reproduced by permission of the author).

The datings using the evolutionary stages of the genus *Orbitolina* (Hofker, 1962, 1963) are not everywhere in agreement with the older data, however. In the Senterada area (Fig. 15) and near the village of Pallerol, for example, the uppermost part of the Prada Limestone already belongs to the Lower Albian (Dalloni, 1930: Upper Aptian). In the Ribagorzana valley, near the Barranco de Inglada (coord.  $42^{\circ}22\frac{1}{2}'$ .  $4^{\circ}25'$ ), Hofker found that the overlying *Orbitolina*-limestones (Aulet Formation) already start in the middle Albian. Souquet (1967, pp. 68–69) agrees with Hofker (1963) for the Sierra de Aulet, that the first intercalations of *Orbitolina*-bearing limestone in the upper part of the black Llusà Marls (Fig. 18) are of uppermost Albian (Vraconian) age. These limestones obtain their maximum thickness of about 900 m in the Lower Cenomanian and part of the Upper Cenomanian.

For a general outline of the pre-Cenomanian palaeogeography, we refer to Dalloni (1910, 1930), Misch (1934), and Souquet (1967). Fig. 16, taken from Souquet's thesis, shows the lateral facies change in a simplified NW–SE cross-section.

#### San Martin Formation (Albian)

In the most western part of the map-area (Bonansa and Santorens areas) the Upper Albian already shows elements of the more terrestrial development of the Albian further to the west ("Facies d'Utrillas", Souquet, 1967; Wennekers, 1968).

The sandstones and conglomerates intercalated between the Llusà Marls and the *Orbitolina*-limestones (Aulet Formation) form part of the San Martin Formation, which has its most typical development near the village of San Martin in the Nogueras zone (sheet 7, coord.  $42^{\circ}28'$ .  $4^{\circ}12'$ ). For the definition and boundaries of this formation we refer to Mey *et al.* (1968).

In the Bonansa area, the very fossiliferous dark-coloured Llusà Marls are unconformably overlain by some 10 m of light-coloured quartz conglomerates (Fig. 17). The succeeding, also light-coloured sandstones, varying in thickness between 30 and 70 m, are calcareous in the upper part, which contains fossils of *Orbitolina subconca* and *Orbitolina discoidea* Gras., indicating an upper Albian age (Dalloni, 1910; Souquet, 1967).

In the area northwest of Santorens, the coral and mollusc-bearing, dark Llusà Marls are overlain by violet-brownish-toned sandy limestones and sandstones, containing an abundance of the above-mentioned orbitolines of Albian age. There, the San Martin Formation is about 30–50 m thick.

#### Aulet Orbitolina Limestone Formation (Upper Albian-Cenomanian)

This formation is named after the Sierra de Aulet (Fig. 18), an E–W striking mountain ridge formed by a very thick (almost 900 m) sequence of these limestones, in which the Ribagorzana River has

| m     | Lithology                                                              | Fauna<br>mainly after<br>Souquet 1967                   | Formation<br>Mey <i>et al.</i> 1968 | Age<br>Dalloni 1910<br>Souquet 1967 |
|-------|------------------------------------------------------------------------|---------------------------------------------------------|-------------------------------------|-------------------------------------|
| >100  |                                                                        |                                                         |                                     |                                     |
| 30    | massive bioclastic limestone with black chert lenses and quartz grains | rudists<br>Vidalina<br>Valvulamina                      | SANTA FÉ Limestone FORM.            | CONIACIAN                           |
| 30/50 | nodular fine-grained limestone and marl                                | rudists<br>Stromiosphera<br>Pithonella<br>Globotruncana |                                     | TURONIAN                            |
| 30/50 | biomicritic and bioclastic limestone                                   | rudists<br>praevulgoles<br>Caprines<br>few orbitolines  |                                     | UPPER CENOMANIAN                    |
| 10    | iron-rich sandy limestone                                              | orbitolines (flat)<br>ammonites                         | AULET FORM.                         | LOWER CENOM.<br>=VRACONIAN          |
| 30/70 | calcareous sandstone                                                   | orbitolines (flat)                                      | SAN MARTIN FORM.                    | ALBIAN                              |
|       | sandstone with clay cement                                             | no fossils                                              |                                     |                                     |
| 10    | quartz conglomerate                                                    |                                                         |                                     |                                     |
| 25/40 | marl and marly limestone                                               | orbitolines (conical)<br>solitary corals<br>molluscs    | LLUSÀ MARL FORM.                    | ALBIAN-<br>UPPER APTIAN             |
| 30/50 | massive micritic limestone                                             | corals<br>rudists<br>orbitolines (conical)              | PRADA Limestone FORM.               | LOWER APTIAN<br>and older?          |
| >200  | dolomite                                                               | no fossils                                              | BONANSA FORM.                       | JURASSIC ?                          |

Fig. 17. Columnar section of the Lower Mesozoic near Bonansa.

carved a narrow gorge. In the upstream part the Escalles gravity dam has been constructed.

According to the definition (Mey *et al.*, 1968), this formation consists of a yellowish to reddish-brown, coarse-grained bioclastic limestone with an abundance of orbitolines. The limestones, in addition to being breccious and dolomitic, are generally sand-bearing and alternate with dark-coloured marls and marly limestones.

In the type section and in the surroundings of the Fallada de Malpas the contact with the underlying Llusà Marls is gradual, the boundary being taken at the base of a competent sequence in which *Orbitolina*-bearing limestones exceed marls and marly limestones. In the Bonansa and Santorens areas, where strongly orange-coloured *Orbitolina*-limestones rest on sandstones and conglomerates of the San Martin Formation, the contact is sharp and slightly unconformable.

The boundary with the overlying nodular Sopeira

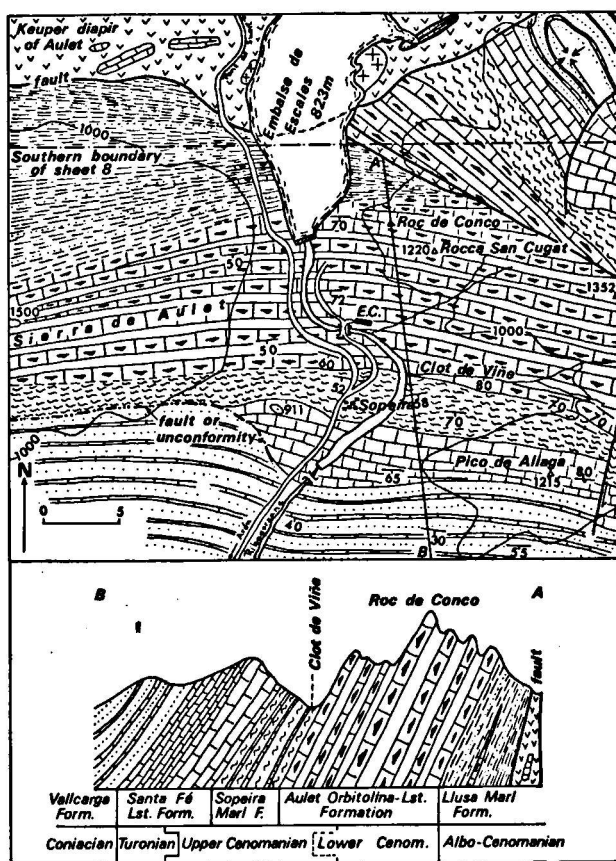


Fig. 18. Geological sketch-map and cross-section of the Sierra de Aulet.

Marls is everywhere taken on top of the highest, brownish or reddish massive *Orbitolina*-bearing limestone. Only in the area north and northwest of Bonansa, where the Sopeira Marls are missing, are the very thin and iron-rich *Orbitolina*-bearing limestones directly overlain by grey, *Prealveolina*-bearing limestones of the Santa Fé Formation (Fig. 17). The total thickness of the formation, which is about 900 m in the Sierra de Aulet, decreases quickly to about 70 m north of the Keuper diapir of Aulet. A minimum thickness of about 10 m is found NW of Bonansa.

In its most typical development, 35 to 40 % of the coarse-grained rock is composed of fossils and fossil fragments among which the orbitolines form the major part. Their shape varies from conical to very flat, the latter being as large as 4 cm. The other organic components are mainly algae, bryozoa, spines of echinoids, and fragments of molluscs and corals. Quartz grains, varying from 0.1 to 10 mm, may locally take up as much as 10 % of the rock volume. Small chert lenses are observed sporadically.

The predominantly reddish colour of the rock is due to the presence of hematite and limonite in small flakes, as pigment, or in the form of tiny iron oölites; these iron oxides form about 3 % of the rock volume.

The macro-fossils are mainly molluscs, echinoids, and locally cephalopods.

In the Escales area the limestones are, except for the uppermost part, very thick bedded, the individual massive banks varying in thickness from some metres to about 40 metres (Fig. 19). In the map-area itself, the formation is more regular and thinner bedded, the alternating beds of black marls and brownish *Orbitolina*-limestones varying from several centimetres to about 1 metre. Marl beds are entirely absent in the Bonansa area, where the limestones are very sandy and a vivid orange.

The abundance of macro- and micro-fossils permit a satisfactory age determination. The molluscs and echinoids collected by Dalloni (1930, p. 178) in the Escales area and the cephalopods from the area SW of Bonansa (Dalloni, 1910, pp. 210–211), both from the lower part of the formation, clearly indicate a Vraconian (Uppermost Albian) age. This is in agreement with the datings based on micro-fossils (mainly orbitolines) in the same and other sections within the map-area (Hupé, 1954; Hofker, 1963; Souquet, 1967). For a detailed description of the palaeogeography during this period and the depositional environment of the *Orbitolina*-limestone, we refer to Souquet (1967). Fig. 16A, taken from Souquet's thesis, shows the lateral facies change in a highly simplified NW–SE cross-section.

#### *Sopeira Marl Formation (Upper Cenomanian)*

This formation is named after the village of Sopeira in the Ribagorzana valley (Fig. 18), which is situated in a depression measuring several hundred metres in width between the *Orbitolina* limestones in the north

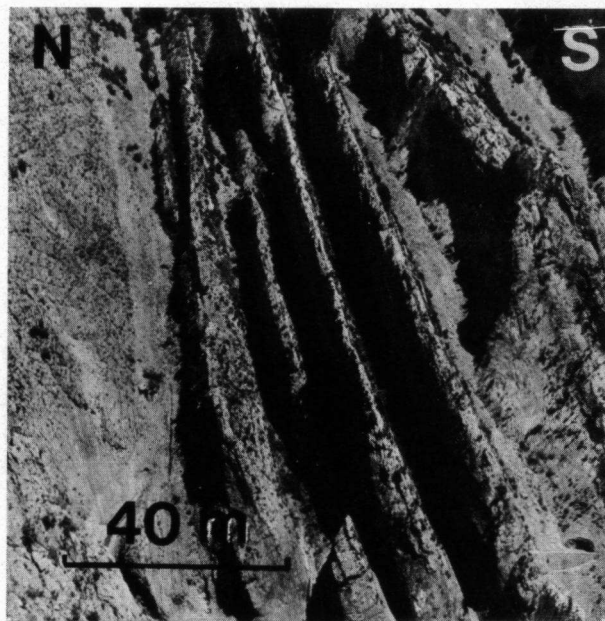


Fig. 19. Very thick-bedded and steep attitude of the *Orbitolina*-bearing limestones of the Aulet Formation south of the Escales reservoir (photo by Ditzel).

and the Santa Fé Limestone Formation in the south. The occurrence of this formation is mainly restricted to the Ribagorzana and Isabena valleys (Fig. 16B). The Sopeira Marl Formation, as defined by Mey *et al.* (1968), consists of a light-coloured sequence in which sandy marls and nodular argillaceous limestones, spotted with glauconite and pyrite, alternate in regular, thin beds.

The lower boundary of this formation is taken at the base of a sequence in which marls and nodular limestones dominate and which overlies massive bioclastic limestones characterized by an abundance of orbitolines (Aulet Formation). The upper boundary is taken where the nodular marly sequence gives way to massive, micritic limestones, characterized by an abundance of prealveolines (Santa Fé Formation). This contact is generally gradual.

The total thickness of this formation, which amounts to 250 to 300 m in the type area (Fig. 18), is some 500 m southwest of the Fallada de Malpas and diminishes to about 50 to 100 m southwest of Bonansa. Northwest of this village this formation is missing; there, the grey Santa Fé Limestones with prealveolines rest directly on the orange *Orbitolina*-limestones (Fig. 17).

In its most typical development, about 40 % of the formation consists of nodular, argillaceous, biomicritic limestones with a characteristic beige to yellowish weathering colour. The bedding is very regular (Fig. 20), the thickness of the individual beds varying from a few centimetres to several decimetres. The intercalated marls (60 % of the formation) are generally darker and often sandy. Faint cross-bedding has been observed locally. The marl beds vary in thickness from a few centimetres up to several metres. Both rock types, in addition to being spotted with

pyrite and glauconite, show many iron concretions with diameters of up to several centimetres.

The macro-fossils are represented by brachiopods, echinoids, and cephalopods, which occur throughout the formation. The macro-fossils collected by Dalloni from the Sopeira and Santorens areas and from the Barranco de Inglada (1910, pp. 225–226; 1930, p. 178) all point to an Upper Cenomanian age for this rock unit. This dating is in complete agreement with the findings of Hofker (1963, p. 194), who found within this formation orbitolines of the form-group V, which are representative for the Upper Cenomanian. Souquet (1967) shares this opinion. We also refer to the latter author for a general outline of the Cenomanian palaeogeography (see also Fig. 16).

#### *Santa Fé Limestone Formation (Upper Cenomanian-Turonian)*

This formation is named after the Peña de Santa Fé, an impressive mountain which dominates the Segre valley and is situated some 40 km southeast of our map-area (Mey *et al.*, 1968).

By definition, the Santa Fé Limestone Formation is a light-grey to beige micritic carbonate characterized by an abundance of prealveolines and miliolids. The upper part of the formation, in which fissurines and globigerines abound, is generally marly.

This strongly transgressive formation rests with a rather sharp but conformable contact on the nodular Sopeira Marls, except in the area NW of Bonansa where biomicritic limestones directly overly sandy *Orbitolina*-limestones of the Aulet Formation. Fig. 16 shows that the same formation overlies the black Llusà Marls of Albian to Lower Cenomanian age in the area between the Flamisell and the Segre. In the area west of the Esera River the lower boundary of this formation becomes a conspicuous unconformity on Permo-Triassic or Palaeozoic rocks.

In the map-area the micritic Santa Fé Limestones, which are slightly marly and frequently slumped in the upper part, are conformably overlain by also slumped marls, nodular marly limestones, and graded calcarenites of the Vallcarga Formation. Beyond the southern boundary of the map this contact is assumed to be a local unconformity (Souquet, 1967, p. 241).

South of the Sierra de Aulet (Fig. 18), where this formation is well exposed and 250 to 300 m thick, the lower third is, except for the first 20 m which contain few marl intercalations, a very massive, light-grey, biomicritic limestone spotted with pyrite and glauconite. The micro-fauna of mainly fissurines and globigerines accounts for 20–35 % of the rock volume.

The first slump structures occur some 110 m above the base of the formation and show a regular increase in frequency towards the top of the formation.

In the middle part of the formation, some 20 m of nodular limestones with echinoids are intercalated. Above this horizon the limestones become breccious, the angular fragments varying in size from less than 1 mm to several centimetres. These limestone breccias

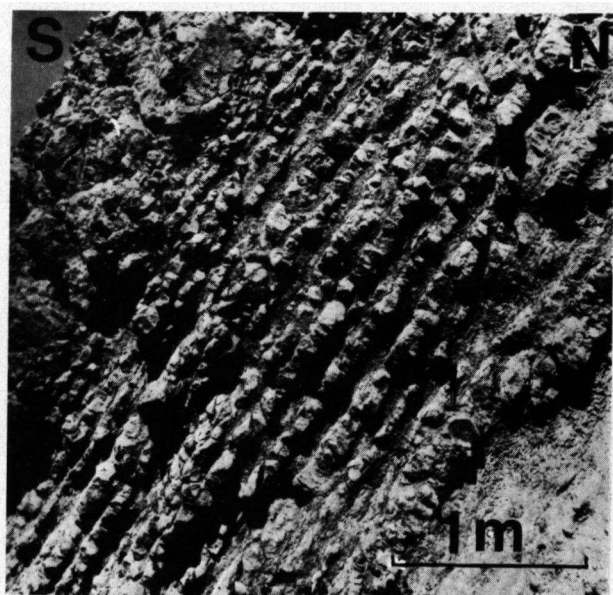


Fig. 20. Typical nodular habit of the Sopeira Marls (photo by Ditzel).



are succeeded by finely laminated limestones with a rather high content of organic material. Chert lenses are associated with this horizon. The rocks are biomicritic and strongly slumped.

The slumps vary in size from a few centimetres to as much as 3 m. Besides rounded fold shapes, angular blocks measuring several metres have been observed. The many measured slump axes display a strong maximum in an E—W direction, which is an indication that the sub-aquatic slope was E—W striking, dipping either north or south (Ditzel, 1966). The strong asymmetry of the folds and their often thrust habit point to a southward dip of the sub-aquatic slope.

Echinoids and rudist fragments were the only macrofossils observed. Under the microscope the biomicritic rocks contain many prealveolines (especially in the lower part), fissurines, globigerines, miliolids, algae, and fragments of echinoids and lamellibrachiates.

The Santa Fé Limestone bordering the Santorens syncline has more or less the same thickness, lithological composition, and fossil content. Slump horizons occur only sporadically. In the upper part a marly horizon, still with plenty of prealveolines, is intercalated, and the uppermost limestones are pseudooolitic with hardly any clastic material.

In the Bonansa area the lower part consisting of light-grey, biomicritic, and bioclastic limestones overlying the orange *Orbitolina*-limestone, is about 30 to 50 m thick (Fig. 17). Except for the presence of a few orbitolines, the micro-fauna is comparable to the sections described above. These lower limestones are overlain by 30 to 50 m of marly and finer-grained nodular limestones, their lithology and micro-fauna being comparable to those occurring in the Santorens section. In their turn they are succeeded by pseudooolitic, sandy limestones with chert lenses and many rudist fragments. Slump structures and limestone breccias are absent everywhere in the section.

The Santa Fé Limestone in the Ribagorzana area is the equivalent of what Misch (1934) calls the "*Ober Kreide Kalk*" (Upper Cretaceous limestone). Although it was erroneously dated as Coniacian by Dalloni (1910, 1930), Misch was the first to realize that in the Sierra de San Gervas and the Ribagorzana valley at least the lower part of this limestone is of Cenomanian age, therefore clearly older than the massive limestone immediately below the Senonian marls in the Flamisell and more eastern areas (Coniacian — Santonian). Thanks to the brilliant work of Souquet (1967), it is now known that these two limestone occurrences, although underlying the same formation, are not one and the same rock unit. The "*Ober Kreide Kalk*" of the lower Ribagorzana valley, formerly correlated with the Coniacian — Santonian limestone in the Flamisell valley (Congost Limestone), must now be correlated with the Cenomanian limestone in the same valley (Mey *et al.*, 1968). This mutual relationship is visualized on the schematic cross-section in Fig. 16B, which is a modified version of Souquet's Fig. 27. The same Figure shows the connection with the areas west of the Ribagorzana.

To summarize the data of Souquet (1967), the age of the Santa Fé Limestone Formation ranges from Upper Cenomanian to Turonian (probably including the lowermost Coniacian) in the area south of the Sierra de Aulet and in the San Gervas area; it certainly includes the entire Coniacian in the Bonansa area and probably in the Santorens area as well.

#### *Vallcarga Formation (Coniacian-Santonian)*

This formation is named after the small tributary of the Pallaresa River just north of the small town of Pobla de Segur (sheet 9, in preparation; coord.  $42^{\circ}15\frac{1}{2}'$ — $42^{\circ}16\frac{1}{2}'$  .  $4^{\circ}39\frac{1}{2}'$ ). In this streamlet an almost complete section of the lower part of the formation is exposed.

The Vallcarga Formation, as defined by Mey *et al.* (1968), consists in its lower part of characteristically yellowish-brown weathered turbidites (containing quartz and calcareous fragments of strongly varying coarseness). The middle part of the formation contains mudflows and olistostrome levels; in the upper part homogeneous blueish-grey marls predominate.

In the map-area, where only the lower part of the formation is exposed, the boundary with the underlying Santa Fé Formation is taken at the base of a sequence in which marls, nodular limestones, and calcarenites dominate over pure micritic limestone. Apart from being an abrupt change in competence, this boundary is also marked in the field by the colour change from light-grey (Santa Fé Limestone) to yellowish or beige. The abnormal contact between these marls and older formations seen in Fig. 18, is interpreted as an angular unconformity by Souquet (1967, p. 241) and as a fault by Hofker (1962). Within the map-area, this contact seems to be concordant.

Of the Vallcarga Formation, which is about 2000 m thick in the Ribagorzana valley south of Sopeira (Fig. 18), only about 600 m are exposed in the Santorens syncline south of the Fallada de Malpas and about 300 m in the area south of the Cruz de Bonansa. The lower part (first 30—70 m) consists of a regularly-bedded alternation of nodular, faintly graded, biomicritic limestones and darker marls and siltstones. The individual beds range in thickness from a few centimetres to about 1 m. Burrows and worm tracks can locally be observed on the bedding plane. Glauconite and wood fragments are characteristic components of both rock types. Microscopically, it is possible to distinguish: micro-forams up to 5 % (fissurines and *Globotruncana*), sponge needles, chert, and corroded quartz grains. Among the important macroscopic features of this rock sequence are the many slump horizons (Figs. 21, 22).

Higher in the stratigraphy the number of marl intercalations increases and the nodular biomicritic limestones become strongly graded, quartz-bearing calcarenites with many sedimentary features such as flute casts, parallel and wavy lamination, convolute bedding, etc. Slumps occur less frequently. The thickness of the individual beds varies from a few centimetres to a few decimetres, but locally calcareous

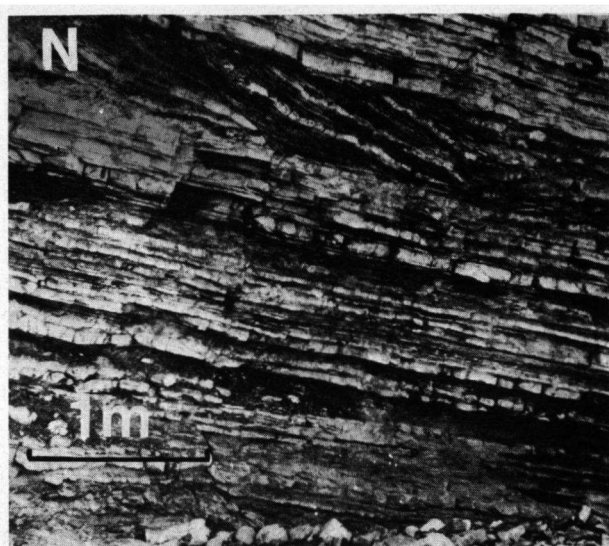


Fig. 21. Typical thin-bedded appearance of the Vallcarga turbidites with a slump horizon in the upper part.

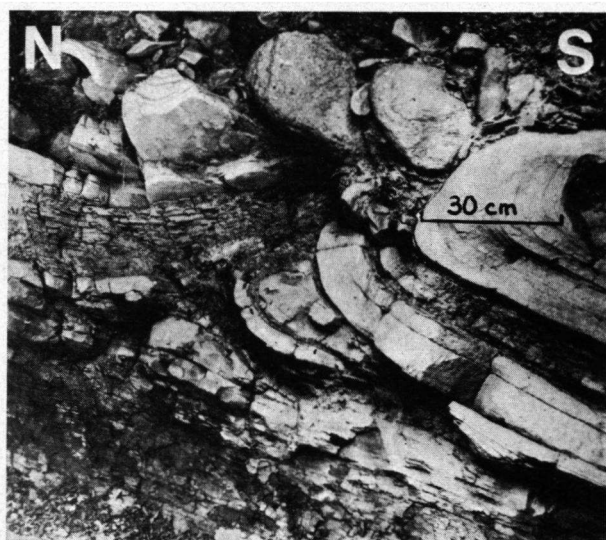


Fig. 22. Detail of a slump fold in the Vallcarga Formation (photo by Ditzel).

beds of up to 1.60 m have been observed. This rock sequence has been interpreted by Ditzel (1966) and Nagtegaal (1963) as a turbidite sequence.

Stereograms of the measured slump axes display, after deduction of the tectonical tilt, an E—W maximum, which is an indication of an E—W striking sub-aquatic slope (Ditzel, 1966). Measurements of the flute-cast directions reveal that the general sediment transport within the basin was from east to west (Ditzel, 1966).

Macro-fossils are scarce. The collected micrasters and ammonites of this lower part of the Vallcarga Formation point to a Coniacian and Santonian age (Dalloni, 1930, pp. 196—199). The mutual relationship between these detrital deposits and the lime-

stones of the same age occurring west and east of the Ribagorzana valley is shown in Fig. 16. For further details of the Senonian palaeogeography and for the depositional environment of this time, we refer to Souquet (1967).

### TERTIARY

After the Pyrenean folding phase (Late Eocene), the high areas underwent active erosion and the valleys and lower areas filled with extensive sheets of conglomeratic piedmont deposits. In the map-area these sub-horizontal conglomerates unconformably overlie folded Mesozoic and Palaeozoic rocks, but more to the south, where the folding was less severe, they overlie Paleocene and Lower Eocene deposits. This rock unit, which only yielded fossils of Upper Eocene age in one particular outcrop (Sosis), is called the Collegats Conglomerate Formation.

#### *Collegats Conglomerate Formation (Upper Eocene-Oligocene)*

This formation is named after the gorge in the Pala-resa valley situated some 10 km northeast of the small town of Pobla de Segur. According to the definition (Mey *et al.*, 1968), this rock unit consists of a thick sequence of largely conglomeratic piedmont deposits. Deep red marls and sandy siltstones are locally intercalated.

The unconformity at the base of the formation carries a sharp relief. The upper boundary, in the area under consideration, is the present topography. Generally, the base of these conglomerates consists of pebbles and boulders derived directly from the surrounding older rocks; higher up in the sequence the content of older rocks derived from the Palaeozoic of the axial zone increases.

Three individual patches of these unconformable conglomerates occur in the map-area. The westernmost of these outcrops, which is the largest, is at least 700 m thick and is composed mainly of well-cemented limestone conglomerates. The pebbles, which have a maximum size of several cubic decimetres, are mainly derived from Jurassic and Cretaceous rocks, including a few blocks of Bunter conglomerates. Devonian elements are rare; granite pebbles are absent.

The much smaller central outcrop is about 300 m thick and shows a fair amount of marl/siltstone intercalations, giving rise to the picture of so-called badland erosion. Of the eastern patch only a minor part is exposed within the map-area. Cemented limestone conglomerates alternate with reddish-brown marl/siltstone intervals. The exposed thickness exceeds 300 m. The pebbles in the central and eastern outcrops derived mainly from Bunter and Palaeozoic rocks. Orientation diagrams of the scour-and-fill structures, which were measured in the eastern outcrop, clearly indicate a N—S transport of the material (Nagtegaal, 1966). For further details of the depositional environment, we refer to this author.

The age of the Pyrenean piedmont deposits is not

known with any precision because of the general scarcity of fossils. Dalloni (1930, p. 246), de Sitter (1961), and Rosell & Riba (1966), found a Ludian age for a well-known level of fresh-water deposits and lignite beds near the base of the sequence at Sosis (coord.  $42^{\circ}15'.4''40\frac{1}{2}'$ ). It is generally assumed, however, that the major part of the sediments was deposited during the Oligocene (Misch, 1934; Alastrue *et al.*, 1957).

Since these conglomerates unconformably overlie clay- and sand stones of Lower Ludian age in the Tremp basin, which is a Tertiary structure situated a few kilometres south of Sosis, it is clearly established that the Pyrenean folding occurred during the Ludian period. That some local movement continued after the deposition of the Collegats Conglomerates is demonstrated by the steeply warped bedding in a few narrow zones situated above old fault lines (see structural sections). These fault lines are always associated with incompetent rocks of the Keuper.

#### QUATERNARY

Since our survey was not mainly concerned with either the physiographic features of the area covered by sheet 8 or with the sub-recent to recent deposits, we shall only indicate some of the major features and refer the reader to the existing literature, in the first place to Solé Sabaris (1962), García Sainz (1940a+b), and Nussbaum (1935, 1938).

The Quaternary deposits on sheet 8 are of glacial and post-glacial origin. Their distribution is essentially restricted to the valleys and cirques within the axial zone, once occupied by glaciers (see Fig. 49). Most of the glacial deposits consist of loose granite boulders embedded in argillaceous to fine sandy and/or arkosic

material. In the northern granite areas these boulders locally have diameters of up to 10 m, but in the southern areas less than 1 m. The upper parts of these erratics are often somewhat finer and show good stratification.

A special type of glacial and late glacial deposits are the rock glaciers, strongly curved ridges consisting of coarse angular blocks and occurring in the glacial cirques or in the stepped valleys behind rock bars. There are many such rock glaciers, especially in the Maladeta area and in the metamorphic surroundings, and they show up beautifully on aerial photographs.

Also conspicuous are the broad, gently downward-sloping fluvio-glacial terraces in the open valleys of Durro and Tahull. Smaller and less obvious are those occurring in the Rio Llausét and the streamlets of Artiga, Estét, Barruera, and Erill Avall. The two largest terraces, both of which are shown on the geological map, have slopes gradually decreasing downward and becoming almost horizontal above the villages Tahull and Durro, where they provide good land for agriculture. Near both villages the terraces are terminated by an erosional scarp. In many areas these sloping terraces have been incised by brooks, the incision showing that the terraces consist of loose, very inhomogeneous material, thickening towards the main valley. It seems probable that these accumulation terraces are kame terraces, but the greater bulk of their material is fluvio-glacial.

Alluvial stream deposits, scree slopes, and fans are of post-glacial age, since they overlie glacial deposits and partly fill up the U-shaped glacial valleys. The distinction between them and material of glacial origin is often difficult to make and must therefore sometimes be arbitrary.

## CHAPTER II

### STRUCTURE

#### INTRODUCTION

The Pyrenees Mountain chain is divided into a series of longitudinal zones of which, from a structural point of view, the outer and youngest ones on both sides are the marginal troughs, and the central axial zone is the oldest. Whereas the axial zone consists almost exclusively of Palaeozoic rocks, the marginal zones are filled with thick series of Upper Cretaceous and Tertiary strata.

On both the north and south flanks of the axial zone a particular internal zone intervenes between the axial zone and the marginal troughs (Fig. 1), which in the case of the northern internal zone is characterized by a thick development of Lower Cretaceous surrounding cores of more or less isolated Palaeozoic blocks, the so-called satellite massifs. The southern internal zone, also characterized by thick early Mesozoic strata, has a quite different structure without any satellite massifs but with conspicuous

gliding nappes, especially well developed in the area west of the Esera River. East of this river and west of the Segre, a narrow complex zone intervenes between the axial zone and the southern Cretaceous internal zone. This is the so-called Nogueras zone, which is characterized by relatively small, isolated Palaeozoic blocks, surrounded by Stephanian and Permo-Triassic rocks. It is assumed that most of these Palaeozoic blocks are autochthonous or sub-autochthonous, and that only few entered this zone by slipping down from the uplifted axial zone.

On sheet 8 of the maps of the Central Pyrenees the following three units are represented (Fig. 1):

- The axial zone, including part of the intrusive Maladeta granodiorite,
- The Nogueras zone,
- The southern Cretaceous zone.



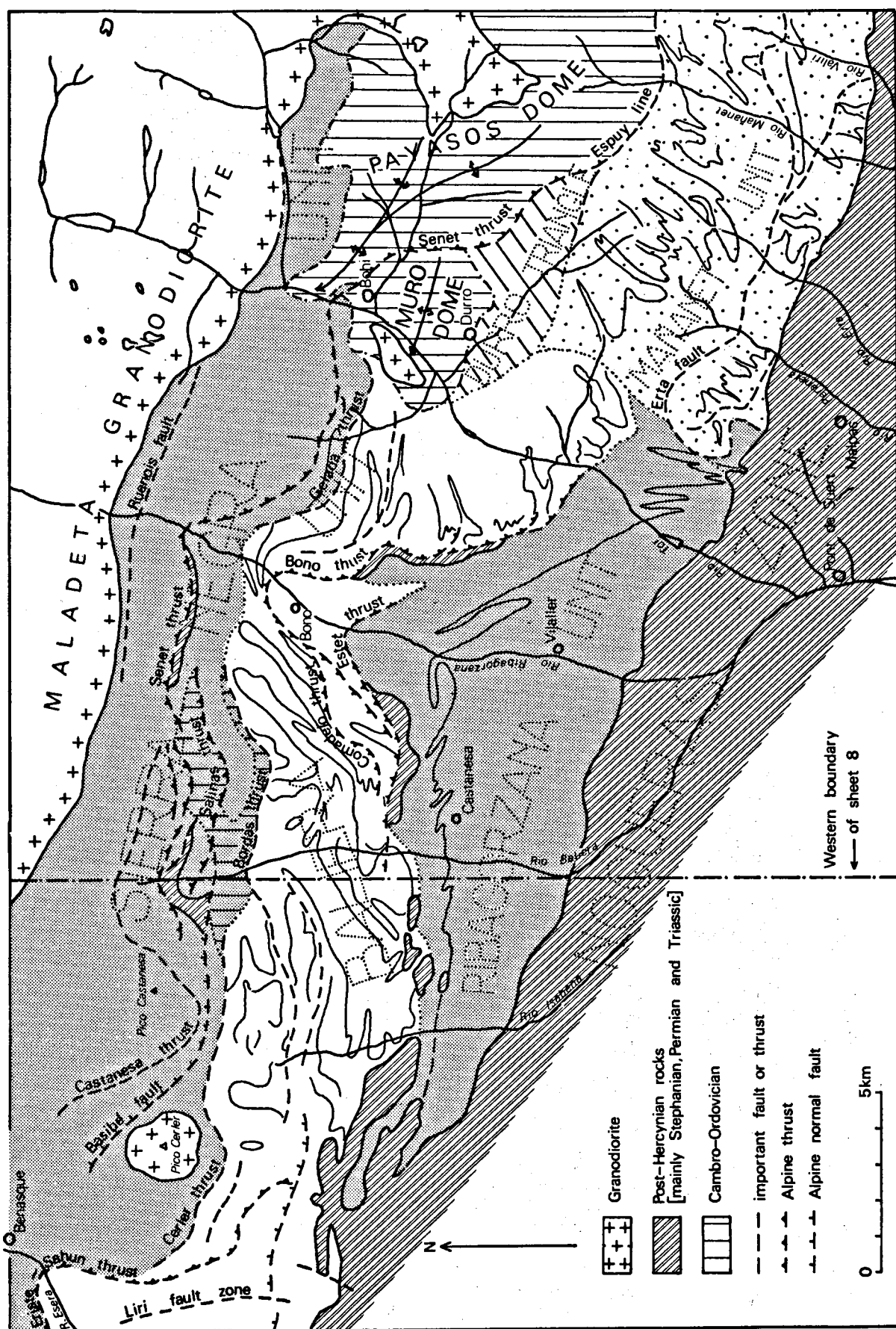


Fig. 23. Structural sketch-map of the axial zone of the Pyrenees south of the Maladeta granodiorite.

## AXIAL ZONE

Our map shows only about one third of the width of the axial zone. Its large over-all structure is here characterized by a rather consistent WNW—ESE to NW—SE strike, a trend which is also largely followed by the southern border of the intrusive Maladeta batholith. The general northward dip of the structural elements (axial planes of minor folds, cleavage, faults, and intrusive dykes) varies from about 60 to 80° immediately south of the Maladeta granodiorite to about 35 to 45° in the centre and south of the axial zone.

This parallelism between the major structures and the regional cleavage and faults does not hold for the bedding and larger folds, which, except for a narrow rim bordering the granodiorite batholith, diverge strongly from the regional cleavage trend. This is clearly evident from the geological map, which shows persistent NE—SW structures measuring several kilometres and at large angles to the regional trend.

Although the general cleavage trend follows that of the major structures, a more careful analysis reveals that the strike and dip of the cleavage are also somewhat irregular: they not only vary at passing through different rock types, but also show weak regional changes within one and the same rock unit. Moreover, the cleavage changes in intensity from one place to another.

East and southeast of a N—S line passing over the Barruera stock, the slaty cleavage is folded and cut by a secondary axial-plane cleavage ( $s_2$ ), somewhat weaker than the first slaty cleavage ( $s_1$ ). Its dip and strike, however, are comparable with those of the  $s_1$ -cleavage west of this line.

The individual structural styles of certain areas, such as the divergence of the bedding and large folds of the regional trend, the change in dip and intensity of the  $s_1$ -cleavage, and the occurrence of a secondary cleavage in the eastern part of the map-area has led us to subdivide the area into a number of structural units. These units are shown on the structural map (Fig. 23), which includes in the west a small part of the axial zone covered by sheet 7.

The structural units are, from north to south:

Sierra Negra Unit  
Payasos and Muro Domes  
Durro Triangle  
Baliera Unit  
Ribagorzana Unit  
Mañanet Unit

They will be described separately.

*Sierra Negra Unit*

This structural unit, which lies just south of the Maladeta granodiorite (Fig. 23), is characterized by isoclinal folding with a well-developed axial-plane slaty cleavage ( $s_1$ ). It includes rocks of Cambro-Ordovician to Lower Carboniferous age, but within the map-area pelitic and calcareous rocks of the Devonian predominate. The over-all dip of the

structures, which is always towards the north, varies from steep near the granodiorite border to moderate in the centre and south (Fig. 24). The intrusion of the Maladeta batholith formed a metamorphic aureole between 1.5 and 2.5 km wide, in which hornfels, marble, and spotted slate occur.

The southern boundary of this unit is formed by thrusts in the west and centre (Cerler and Bordas thrusts) and is therefore very sharply defined. But near and especially east of the Ribagorzana, this southern boundary is less distinct, because the flattening produced by the cleavage becomes gradually less important in that direction. The most logical boundary is the Gelada thrust with its enclosed slices of Cambro-Ordovician rock. East of the Tor, this structural unit is bordered by the 40 to 60° northward dipping flank of the Payasos Dome.

About 7 km east of the Tor this Devonian structure is truncated by a southward extending lobe of the granodiorite. Where the Devonian reappears a few kilometres further to the east, it is classified as the southern flank of the Llavorsi syncline (Fig. 1). To the west of the area covered by sheet 8, the Sierra Negra Unit widens and grades into the Perramó Dome, for details of which we refer to Wennekers (1968).

The swing in strike of the mean cleavage follows that of the entire structure, and moreover runs almost parallel to the southern margin of the granodiorite. Although a causal relationship seems likely, the post-main-phase intrusion of the granodiorite could hardly be considered responsible for this deviation because a similar, but far more pronounced, swing is present in the southern Baliera and Ribagorzana Units (Ribagorzana hinge-zone), which certainly lie beyond the limits of the "shouldering-aside" effect of the Maladeta batholith. Mey (1967-b, p. 204) advanced arguments for a possible relationship between this bend and the diagonal cross-folding (NE—SW and NW—SE), so well-known for many Cambro-Ordovician domes in the Pyrenees (Zwart, 1963, 1965).

With respect to the considerable change in the northward dip of the cleavage, which diminishes from about 80° locally in the north to some 45° in the south and locally is even smaller, it is assumed that this is an original tectonic feature. Warping of the cleavage from an original vertical position, as proposed by Boschma (1963), does not seem very likely here (Mey, 1967-b, pp. 199—201).

The tight to isoclinal cleavage folds observed in this area vary in size from a few decimetres to several kilometres. The smaller folds are visible in most outcrops. Probably the most striking example of large-scale isoclinal folds are visible on the steep southeastern slope of the Montañeta (coord. 42°33' . 4°30') when viewed from the road north of Bohi (de Sitter & Zwart, 1961, Fig. 16). The lighter-coloured limestones represent isoclinal synclines of the Basibé Formation and the surrounding greyish-brown rocks anticlines of the Rueda Formation. The top itself consists of Upper Devonian grotte.

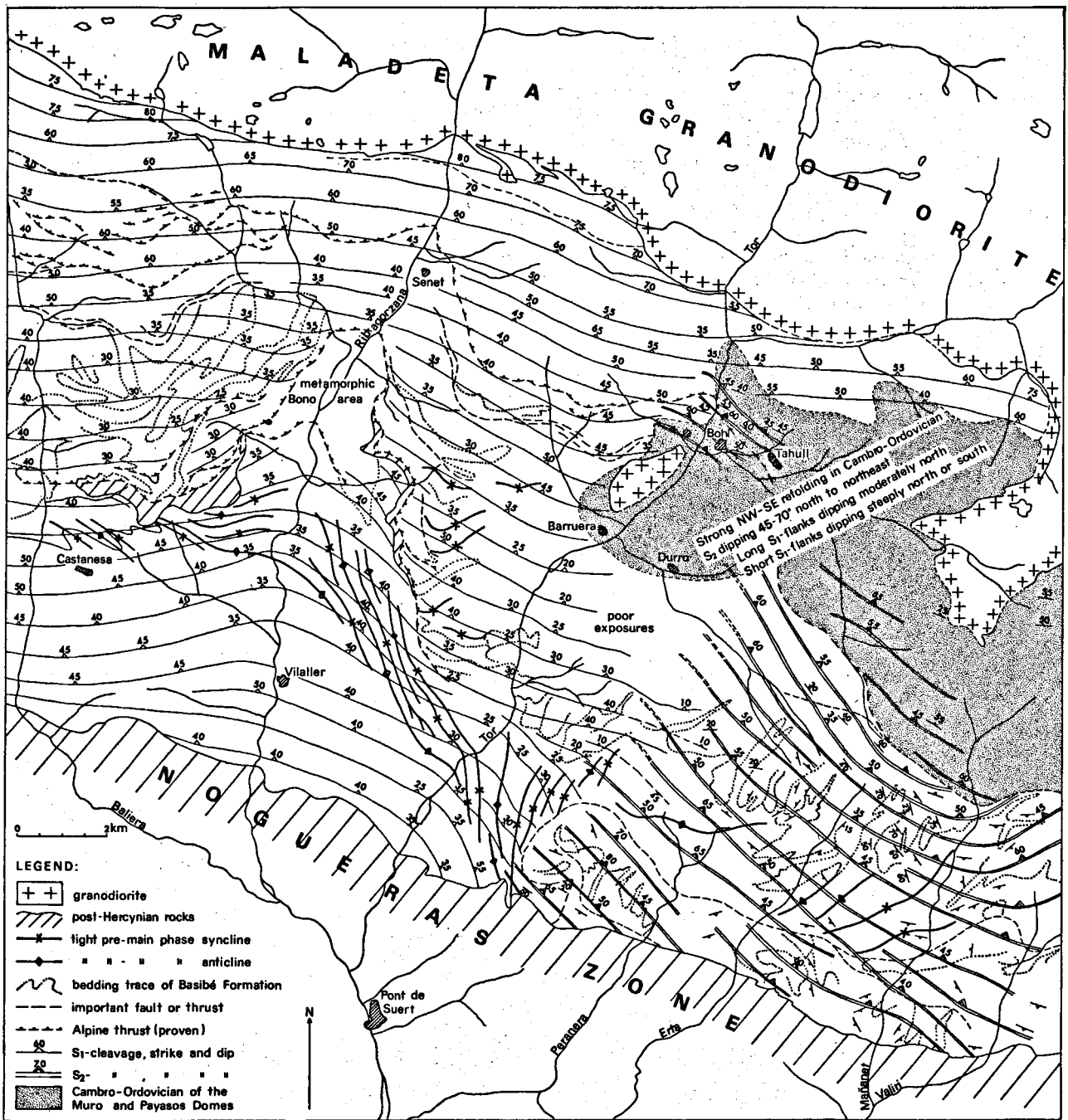


Fig. 24. Cleavage trend map.

Measurements of the isoclinal folds yield typical orientation diagrams, with the modal cleavage and bedding planes coincident. The minor folds clearly have the cleavage parallel to their axial planes and consequently the  $\delta$ -lineations parallel to the axes. This relationship is also clear in the diagrams (Mey, 1967-b, Fig. 19, A<sub>0</sub>-A<sub>6</sub>), which show the measured lineations and axes lying in complete girdles parallel to the modal cleavage plane. The northward-directed fold axes and  $\delta$ -lineations suggest, however, that the

folding preceding the cleavage development had a slightly different strike.

Within the map-area, folding of the slaty cleavage ( $s_1$ ) with development of a secondary crenulation cleavage ( $s_2$ ) has only been observed locally in the surroundings of Erill Avall. Most folds are of the chevron type, with an axial-plane cleavage ( $s_2$ ) generally dipping 45 to 70° NNE and with steeply plunging axes. For further details concerning these folds, which are thought to have originated during a

late Hercynian folding phase, we refer to Boschma (1963, pp. 147—155).

In the Sierra Negra Unit beyond the western boundary of our map these chevron folds having more or less the same attitude as near Erill Avall but with sub-horizontal axes, are frequently found in the gently northward-dipping Castanesa syncline and in the area of Benasque. In the Castanesa massif it can be demonstrated that this folding of the  $s_1$ -cleavage occurred later than the intrusion of the granodiorite and also later than the intrusion of the diorite dykes, since the latter are also folded (Mey, 1967-b, p. 187). This post-intrusional deformation is also demonstrated by the locally observed rotated angular hornfels chips in a carbonate matrix, the latter showing adaptation by flow. A comparable phenomenon is seen in thin sections where contact metamorphic cordierite crystals sometimes show a slight rotation as compared to its surrounding schistose matrix. This refolding may therefore be a late Hercynian phenomenon (E—W refolding) or an Alpine deformation. Wenckers (1968) favours the latter solution, although his arguments are, in our opinion, not very convincing.

The last Hercynian compression phase was followed by dilatation in a horizontal sense (horizontal tensional stress). This is shown by knick-zones, which occur frequently in larger, non-metamorphic slate belts such as those east and west of the village of Aneto. For the principle of movement within knick-zones in relation to the general stress field, we refer to Zandvliet (1960) and Boschma (1963).

Of the large, longitudinal thrusts, which follow the general cleavage trend, most were active during the Alpine orogeny. This is demonstrated by the Mesozoic (probably Upper Triassic) rocks, which are incorporated into their fault zone west of the Ribagorzana River. Undoubtedly, the most important of these is the Senet thrust, which splits up into a number of splays near the western border of sheet 8. In the east this Alpine thrust separates the Muro and Payasos Domes, and further to the southeast it links up with the Espuy line, the steep southern border fault of the Payasos Dome.

The Sierra Negra Unit has been treated extensively by Mey (1967-b, pp. 181—191), and we refer to this article for further detail.

#### *Payasos and Muro Domes*

These two anticlinoria, exposing a monotonous series of strongly cleaved Cambro-Ordovician rocks, form part of the much larger Southern Dome (Fig. 1). The Muro as well as the larger Payasos Dome both have a general northward dip of their structural elements and strongly westward-plunging dome axes. Both structures are separated by a narrow zone of Silurian black slates, a depression which is also followed by the Alpine Senet thrust (Fig. 23). Further to the southeast this thrust links up with the Espuy line, a steep fault zone bordering the Payasos Dome in the south for at least 12 km. The southern Muro Dome encloses in its western part the small Barruera granodio-

rite stock, which only produced a relatively narrow zone of contact metamorphism. The thermal influence of the southward-extending lobe of the Maladeta granodiorite is also restricted to only a few hundred metres.

A conspicuous feature of the northern flanks of both domes is the occurrence of several vein-like bodies of white quartz, which locally are as thick as 40 m. These bodies are often found at the very contact of Cambro-Ordovician and Silurian rocks; locally, they include well-orientated slabs of either formation or lie fully within the dark Silurian slates. Field indications point to formation by replacement rather than by hydro-thermal ascension. Secondary pyrite crystals, which are fairly abundant in the tectonized Silurian and Cambro-Ordovician of the steep southern border of the Muro Dome, are absent in the thick quartz veins.

Since slickensides are frequently observed in these quartz bodies and at their contact with their host rock, a relationship with a fault zone seems warranted. This is certainly true for the quartz veins in the northern flank of the Muro Dome, which can be linked up with the Gelada thrust, since this fault zone includes slices of Cambro-Ordovician rocks and similar lense-shaped quartz bodies. The string-of-pearls arrangement of slices of Devonian, Cambro-Ordovician and quartz lenses in a matrix of strongly sheared Silurian slates occurring in the northern flank of the Payasos Dome near the granodiorite border, is also very suggestive of a fault zone. Since this zone is truncated by the granodiorite, it is there unmistakably a Hercynian phenomenon, prior to the granodiorite intrusion.

The complex internal structure of both domes is undoubtedly due in the first place to the very finely laminated character of the sediments. The slightly higher grade of regional metamorphism as compared to the overlying Devonian rocks may have played a minor role. The slaty cleavage, which is very strongly developed in these fine sediments, can be traced down from the Devonian folds through the Silurian into



Fig. 25. Tight  $s_1$ -fold with axial-plane slaty cleavage in thinly-bedded shale-sandstone sequence of the Cambro-Ordovician.



the Cambro-Ordovician. This cross-cutting axial-plane orientation of the slaty cleavage in the Cambro-Ordovician differs widely from that in the higher metamorphic Cambro-Ordovician Garonne Dome, where the schistosity is mainly parallel to the bedding and consequently also to the roof of the dome (Boschma, 1963; Zwart, 1963).

Lineations due to the intersection of cleavage and bedding are well developed. In places where minor  $s_1$ -cleavage folds could be studied, the cleavage was always found to be parallel to the axial planes (Fig. 25) and the cleavage-bedding lineations parallel to the fold axes. The most important feature of the axial plane cleavage in the majority of the Cambro-Ordovician rocks is its high degree of parallelism, making it a good surface for the detection of subsequent folding. Its high fissility also promoted the formation of small secondary folds.

The slaty cleavage plane, which is parallel to the axial planes of the associated  $s_1$ -folds, generally dips approximately  $50^\circ$  N, although detailed observation shows many variations in dip and strike. Folds in the finely laminated rocks are generally tightly compressed and most often isoclinal. In many cases a parallel foliation of cleavage and bedding, as shown in Figs. 25 and 26, is developed. The  $s_1$ -folds of the slightly thicker bedded quartzites and marly limestones of the uppermost Ordovician are less severely compressed, most of the folds being open to close (Fig. 27);

consequently, cleavage and bedding in the marly limestones are far from parallel. In the thicker quartzites, where no slaty cleavage could develop, one finds a coarse fracturing perpendicular to the bedding.

Folding of the  $s_1$ -cleavage, which is strongest in the Muro Dome and the westward plunging part of the Payasos Dome, can be observed in many exposures. The most conspicuous folds are small asymmetrical chevron folds, comparable to those in the Sierra Negra Unit where they occur only very locally. The axial planes of these secondary folds generally have a moderate NNE to NE dip. When no clear folding of the  $s_1$ -cleavage is present, the lustrous surface of the slates may still show faint crenulations, striking parallel to the  $s_2$ -fold axes elsewhere.

Such minor secondary folds show a high degree of parallelism of the axial planes and crenulation cleavage, which is developed in both hinges and limbs. The folds are usually not very tight, but in some outcrops they are extremely compressed, becoming almost isoclinal. In the latter case a new fissility is produced of near parallel limbs and crenulation cleavage.

Although the folds appear to be inhomogeneous in several exposures, with fold axes varying from horizontal to very steep, the axial planes are parallel in uniform rock. When plotted on stereograms they clearly display one or two maxima.

Boschma (1963, pp. 147—162), who made a very



Fig. 26. Micro-photograph of isoclinal  $s_1$ -fold with accompanying axial-plane slaty cleavage; enlarged directly from thin section.



Fig. 27. Concentrically folded thick-bedded quartzite of the Upper Cambro-Ordovician with coarse fracture cleavage.

detailed study of these complex refolding phenomena in the Muro and western Payasos Domes, established for this area the following structural history:

1. Main-phase folding (probably E—W) with vertical axial planes and development of a strong slaty cleavage.
2. Diagonal refolding with steep axial planes, the NW—SE set occurring the most frequently.
3. Warping of the  $s_1$ -cleavage, so that it is now progressively flatter towards the south.
4. E—W refolding with steep axial planes.
5. Continued inclination, with very little new E—W refolding and finally with shear and knick-zones.

Warping of the  $s_1$ -cleavage from an original vertical position, as suggested by Boschma (*ibid.*), is doubted by the present author, for whose arguments, which are not entirely conclusive, we refer to Mey 1967-b (pp. 199—201). In any case, it is certain that the whole structure already had its present inclination at the time of the intrusion of the Maladeta granodiorite, since in the rigid hornfels rim surrounding the southward extending lobe of this batholith, the baked cleavage shows the same dip. This also holds for the cleavage in the contact metamorphic rim around the smaller Barruera stock.

The NW—SE and E—W refoldings are true Hercynian phenomena, although this has recently been doubted by our French colleagues (Mattauer, 1964, 1966). The following three well-established facts do not, however, leave any doubt as to the Hercynian origin of the above-mentioned refolding:

1. Baked folds of the NW—SE and E—W sets of refoldings occur in the hornfels rim of the late Hercynian granodiorites.
2. Microscopically, it is evident that the crenulation cleavage accompanying these folds was consumed by the contact metamorphic cordierites (Fig. 28).
3. In the unconformable Upper Westphalian conglomerates (Aguiró Formation) north of Iguerri, Devonian and Ordovician pebbles and boulders occasionally show slaty cleavage and two sets of crenulations.

The occurrence, however, of a compression phase (roughly N—S) subsequent to the intrusion of the granodiorite and accompanying dykes, can be inferred from the locally overthrust southern border of the Maladeta batholith with accompanying shear zones parallel to it (as e.g. N and NE of the Llebreja, coord.  $42^{\circ}32' . 4^{\circ}35'$ ), and the occurrence of a folded aplitic dyke in the Barranco de Ginebrel (coord.  $42^{\circ}29' . 4^{\circ}33'$ ). In the latter outcrop it can be seen that the dyke intruded parallel to the  $s_1$ -cleavage and was concentrically folded around moderately to steeply northeastward-dipping axial planes. The accompanying axial-plane crenulation cleavage is consistently followed by thin quartz veins. It is not yet known whether these movements formed part of the Hercynian orogeny or originated during the Alpine compression phase.

#### *Durro Triangle*

The Durro Triangle is undoubtedly the most chaotically disturbed structural unit of the map-area. Since outcrops are poor and surface creep a very common feature of most exposures due to the high fissility of the incompetent rocks, the over-all-structure is insufficiently understood. It may well be that this complex zone represents the incompetent detachment horizon between the Muro and Payasos Domes in the north and the Devonian-Carboniferous rock sequence in the south, but the arguments put forward in favour of this hypothesis seem weak.

This structural unit, which exposes exclusively incompetent Silurian shales, the thin "basal limestone" of the Devonian, and Lower Devonian shales, is in the northwest bordered by the steeply southward-dipping flank of the Muro Dome and in the northeast by the overturned flank of the Payasos Dome, which coincides here with the Alpine Senet thrust zone. The southern boundary is only well defined in the narrow eastern part, where it coincides with the Espuy line, which is thought to have a steep southward dip here. The central and western parts of the southern boundary, which are very badly exposed, probably represent a gradual transition into the more regular and better



Fig. 28A. Thin section showing bedding,  $s_1$ - and  $s_2$ -cleavages, and contact metamorphic cordierite crystals enclosing both these cleavages (enlarged directly from thin section).

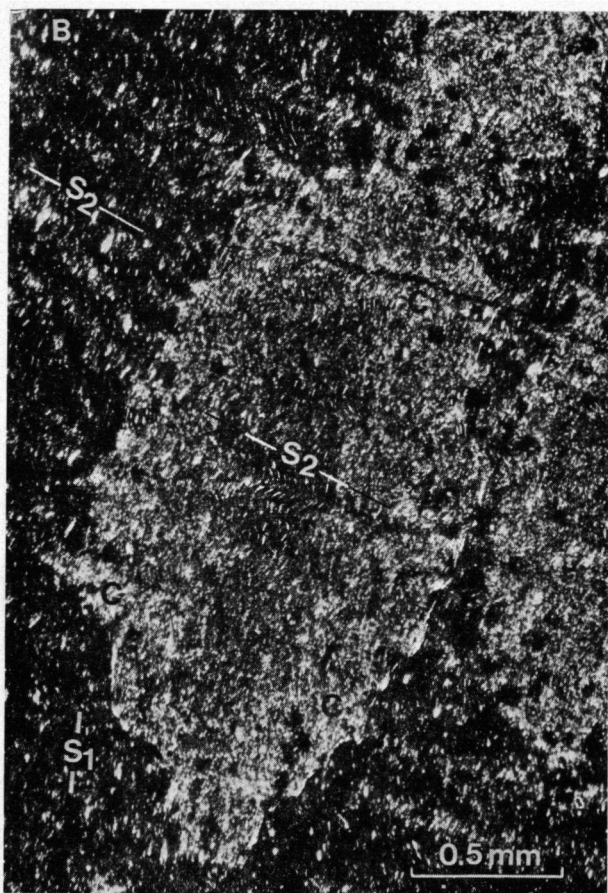


Fig. 28B. Micro-photograph showing detail of Fig. 28 A.

understood Baliera and Mañanet Units. It is questionable whether this transition zone still forms part of the Espuy line.

The western part of the area is characterized by a black matrix of Silurian shales having an irregular, generally sub-horizontal to moderately southward dipping  $s_1$ -cleavage, in which float isolated hinge-

zones of the "basal limestone". These folds are open to close and have the slaty cleavage ( $s_1$ ) parallel to their axial planes. The latter show dips in all directions and their inclination varies from horizontal to vertical. Plain refolding may have been partially responsible for this irregular pattern of the slaty cleavage and the attitudes of axial planes, but undoubtedly still later and less irregular deformation was also active, as suggested by the rather wide spread on diagrams of the secondary cleavage planes (Boschma, 1963, Figs. 88, 89, and 90).

The central and eastern part of the Durro Triangle is somewhat better exposed, contains a much larger percentage of Lower Devonian rocks, and shows a more regular pattern as far as the refolding is concerned. The relationship between main folding and refolding can best be studied in the two E—W-running streamlets which are crossed by a few intrusive dykes above the fluvio-glacial terrace. The two diagrams of this area (Fig. 29) clearly show that the  $s_2$ -folding, which has here a very constant NW—SE strike and sub-horizontal axes, is responsible for the spread in  $s_1$ -cleavage poles. Figs. 30 and 31 show how this refolding presents itself in the field.

The strongly sheared attitude of the aplitic dykes in the southeastern corner of the Durro Triangle (Espuy line) points to a shear movement subsequent to their intrusion. This movement may have occurred at the same time as the folding of the aplitic dyke in the Barranco de Ginebrel and the local thrust movement of the Maladeta granodiorite (see page 262). So far, we have no evidence indicating whether these movements were caused by a late Hercynian or an Alpine compression.

#### Baliera Unit

The Baliera Unit south of the Sierra Negra Unit discussed above (Fig. 23) is characterized by large folds with an oblique strike as compared to the general cleavage trend (Fig. 24). These predominantly NE-striking structures, involving exclusively Lower and Middle Devonian strata (including the Fonchanina slates), are overturned concentric folds without cleavage development. They were flattened and refolded by the main-phase cleavage folding. Refolding of the  $s_1$ -cleavage is only found in the area east of the Tor River. When this secondary folding becomes the most outstanding element in the field, one has entered the Mañanet Unit.

Like the northern boundary, most of the southern boundary was formed by a thrust (Alpine Bono thrust). However, near the western and eastern ends of the Bono thrust, the southern limit of the Baliera Unit coincides with an imaginary line roughly tangential to the southern termination of the large NE—SW-striking folds of the Lower and Middle Devonian. In the Ribagorzana Unit, south of this line, the Upper Devonian griotte has an E—W to NW—SE strike. In the western part of our structural map (Fig. 23), some 5 km west of the sheet boundary, the Baliera Unit gradually loses its characteristic, large NE-



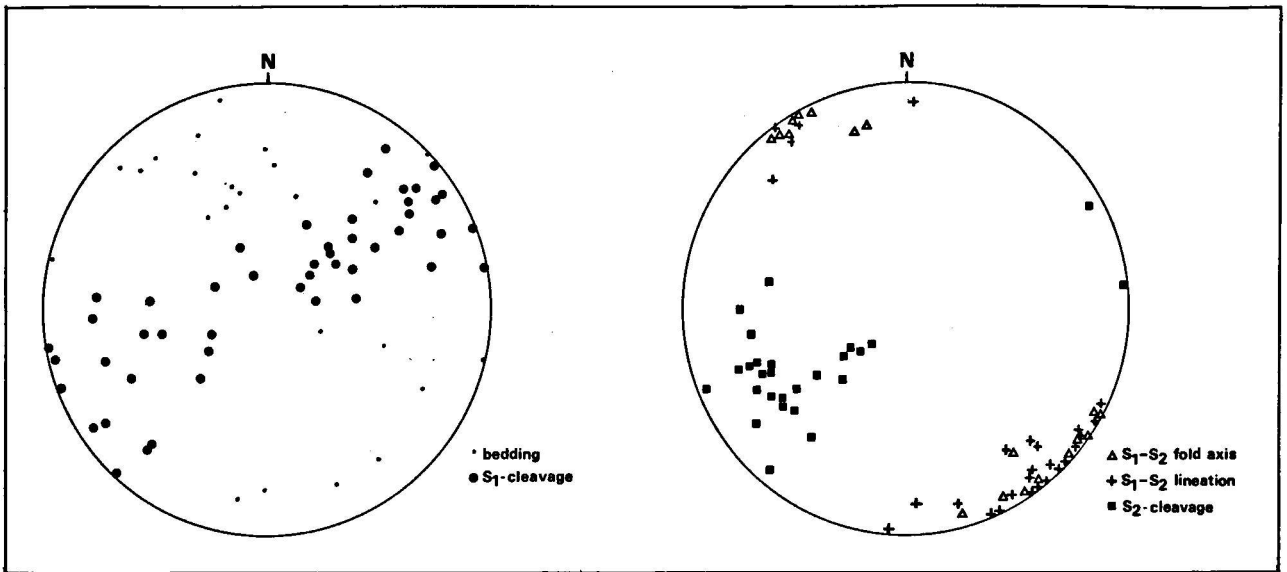


Fig. 29. Orientation diagrams of the eastern part of the Durro Triangle, showing the distribution of  $s_1$ ,  $s_2$ ,  $ss$ ,  $s_2$ -fold axes, and  $s_1$ - $s_2$ -lineations.

striking pre-cleavage folds. The area up to the Alpine Eriste-Sahun thrust is dominated mainly by E—W-striking main-phase cleavage structures, although N—S-directed minor folds and  $\delta$ -lineations indicate that here, too, the cleavage development was preceded by folding with an aberrant strike.

Before mapping was started, the presence of oblique pre-cleavage folds in this area was first detected from the distribution of  $\delta$ -lineations and fold axes of cleavage folds that did not show a concentration in an E—W direction as expected but had a girdle distribution coinciding with the mean cleavage plane (Boschma, 1963, Figs. 32—61; Mey, 1967-b, Figs. 30 and 31).



Fig. 30. Typical appearance of  $s_2$ -folding in the Lower Devonian of the Durro Triangle; axial planes dipping 60 to 70° towards the northeast.

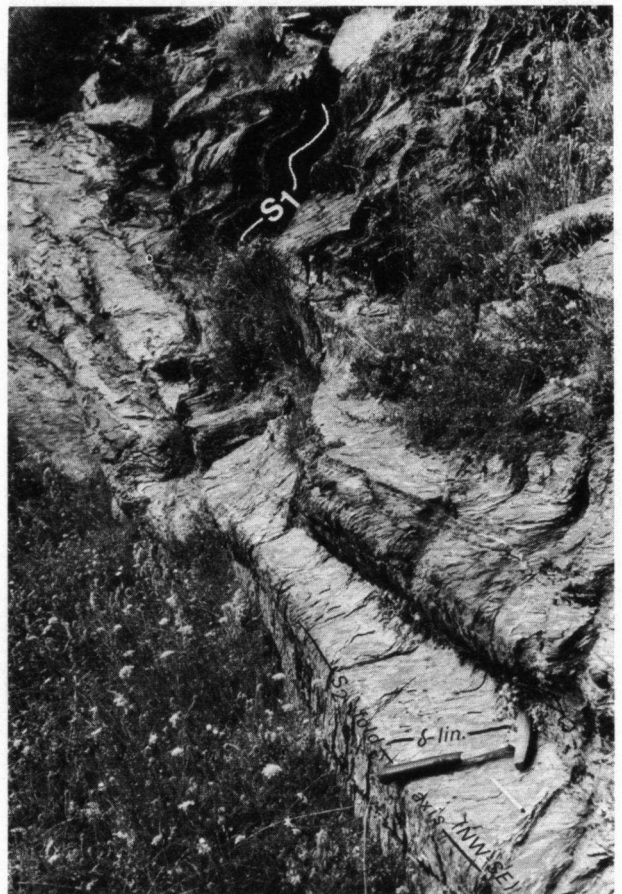


Fig. 31. Typical appearance of  $s_2$ -folding in the Lower Devonian of the Durro Triangle; axial planes dipping about 20° towards the northeast.



On the geological map these folds can be recognized by the peculiar course of the formation boundaries; these boundaries do not trend constantly E—W, as in the central part of the axial zone and most of the Sierra Negra Unit, but show a kind of “*Schlingenbau*”. The slaty cleavage, on the other hand, has a rather constant E—W to ESE—WNW trend (Fig. 24).

The direction and shape of these early folds is difficult to ascertain when the later cleavage deformation is strong and/or only slightly oblique, as for instance in the area between the Tor and Ribagorzana rivers. West of the Ribagorzana, however, the effect of the cleavage folding on the original folds was much weaker, which can be ascribed to the larger angle between cleavage and pre-cleavage structures, the considerably higher competence and thickness of the quartzite-bearing Basibé Formation, and thirdly, to a less strongly developed slaty cleavage. In that area, therefore, the cleavage folds can usually be subtracted from the pre-cleavage structures.

The size of these pre-cleavage folds is considerable (some kilometres) as compared to the superimposed cleavage folds (decimetre to hectometre dimensions), so that in the field one is only aware of the latter. The mean NNW to NW dip of the bedding planes, together with the numerous N—S-directed axes of cleavage folds and  $\delta$ -lineations on the generally E—W-striking cleavage plane, however, constitute clear evidence of early, obliquely-striking structures. Although the pre-cleavage folds show up beautifully on maps, they are, because of their very large size, seldom visible in the field as a whole. The relative dips of their flanks can, however, be deduced from the shape, axial plunge, and symmetry of the super-

imposed cleavage folds. We noticed that on one flank cleavage folding had produced normal anticlines and synclines but on the other steeply plunging antiforms and synforms in which the stratigraphy was the reverse of the normal sequence. Consequently, the latter cleavage folds must have been superimposed on an overturned flank of a pre-cleavage structure. This example shows that careful analysis of measurable structural phenomena such as cleavage,  $\delta$ -lineations, fold axes, fold symmetry, and the map contours, provides a rather good basis for a reconstruction of the original direction, size, and shape of the large pre-cleavage folds.

The change in dip and strike of the main cleavage ( $s_1$ ) is shown in Fig. 24, which was compiled from about 1300  $s_1$ -cleavage measurements. First, for a number of well-selected adjoining areas the modal cleavage plane was elaborated by means of contoured stereograms. Next, for each diagram area the modal cleavage plane was plotted on the map and then connected by smooth lines. The original stereograms for the smaller diagram areas are shown in our preceding publication (Mey, 1967-b).

To avoid any misunderstanding of Fig. 24, however,

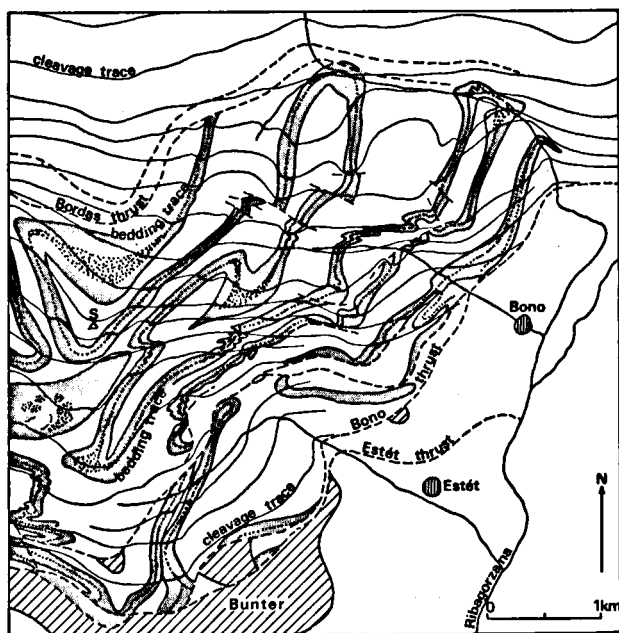


Fig. 32. Sketch-map showing bedding traces of pre-cleavage folds in the Baliera Unit, their strike clearly oblique to the cleavage traces.

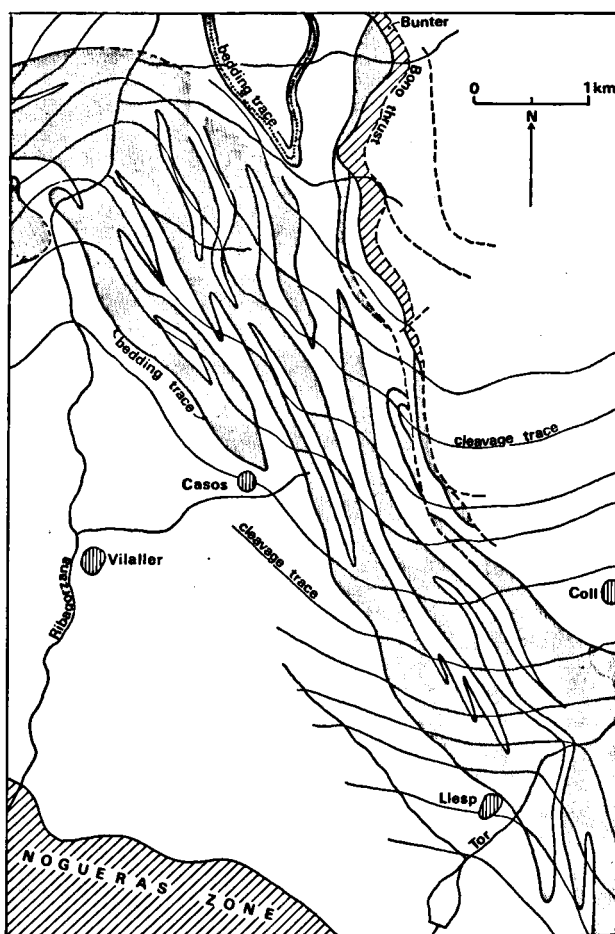


Fig. 33. Sketch-map showing bedding traces of pre-cleavage folds in the Ribagorzana Unit, their strike clearly oblique to the cleavage traces.

it must be stressed that in an accidented area like ours the angle between the bedding traces (intersection of the bedding with the topography) and the constructed cleavage trend does not represent the true angle between the cleavage and pre-cleavage folds. For this, either the bedding must be projected to a horizontal plane, which is almost impossible in view of the complex structure, or the regional cleavage plane intersected with the topography. This has been done for two relatively simple areas, as shown in Figs. 32 and 33.

For a more detailed description of this structural unit and for an analysis and elaboration of the following subjects, we refer to our preceding publication (*ibid.*, 1. bend in the cleavage traces, the so-called Ribagorzana hinge-zone, p. 204; 2. relation between pre-cleavage folds and main-phase deformation, pp. 193—197; 3. reconstruction of the pre-cleavage folds, pp. 198—199; and 4. dip of the cleavage, pp. 199—201).

The metamorphic Bono area and the hundreds of parallel dykes swarming out of it will be discussed in Chapter III.

#### *Ribagorzana Unit*

The Ribagorzana Unit, located south of the Baliera Unit, is bordered in the west and south by the Lower Triassic, which unconformably overlies the Carboniferous shale/sandstone sequence. South of this unconformity is the Nogueras zone. In the east, about halfway between the Tor and Peranera rivers, this structural unit grades into the Mañanet Unit, which is characterized by very strong folding of the  $s_1$ -cleavage.

The Ribagorzana Unit itself is characterized by the preponderance of schistose Upper Devonian and Carboniferous rocks with a rather constant dip in a northward direction. In the field one encounters many close to tight folds of metre to decametre dimensions, having the  $s_1$ -cleavage in their axial plane. However, the many northward-directed  $\delta$ -lineations and fold axes measured in the field also suggest the presence here of a pre-cleavage folding with an aberrant strike. This is confirmed by the map contours of the large folds, which generally are oblique to the over-all cleavage trend (Figs. 24 and 33). In the field one also observes locally minor folds with oblique axial planes with respect to the cleavage. The "Bono Dome", which Boschma described in his thesis (1963, pp. 140—143), is an excellent example of this.

Undoubtedly the most striking large-size pre-cleavage folds are those of the Upper Devonian griotte occurring between the Ribagorzana and Tor rivers. These folds have a constant NW—SE trend, but the mean cleavage strikes WNW—ESE (Fig. 24). From their map contours it follows that these folds must be tight and sub-vertical, since their map trace is hardly affected by the topography. The cleavage of this area, on the other hand, dips moderately (30—45°) and consequently makes a large angle with the axial plane of the folds, which therefore must be pre-cleavage folds. In the field this is confirmed by the orientation of the

$\delta$ -lineations (in the griotte developed as rods) and the axes of minor cleavage folds, both of which show a large spread in the cleavage plane. For the relevant diagrams we refer to Mey (1967-b, Figs. 43 and 44).

No plausible explanation has so far been given, however, for the large angle (about 90°) between the pre-cleavage trend in the Ribagorzana Unit (NW—SE) and that in the Baliera Unit (NE—SW) immediately north of it. Our earlier suggestion (Mey, *ibid.*), that the locally divergent trend in the Ribagorzana Unit is in some way related to a basin rim of the Baliera basin, was hypothetical, but this divergence can readily be explained if it is assumed that the transition zone between these two structural units coincides with the axial plane of a huge, roughly E—W-striking, overturned cleavage fold, the northern normal flank representing the Baliera Unit and the southern overturned limb the Ribagorzana Unit. The general NE—SW-trending pre-cleavage folds on the normal flank become E—W striking (with a strong westward plunge) on the crest of the fold, to attain a NW—SE trend on the overturned limb (Fig. 34). It is important to note that with the 90° change in strike the normal anticlines and synclines of the Baliera Unit change into synforms and antiforms, respectively (see coloured cross-sections).

In the area west of the Baliera River (sheet 7 and Fig. 23), where very few pre-cleavage folds occur on this overturned limb of Upper Devonian griotte, the attitude of this huge main-phase fold is clearly expressed. Within the map-area most of the hinge and the axial plane zone of this huge inclined fold was destroyed by the Alpine Bono thrust. However, the E—W to WNW—ESE-trending folds north and northwest of the village of Coll (coord. 42°28½', 4°27½') may well be interpreted as parasitic cleavage folds on the crest of the superfold. Equally, the pronounced

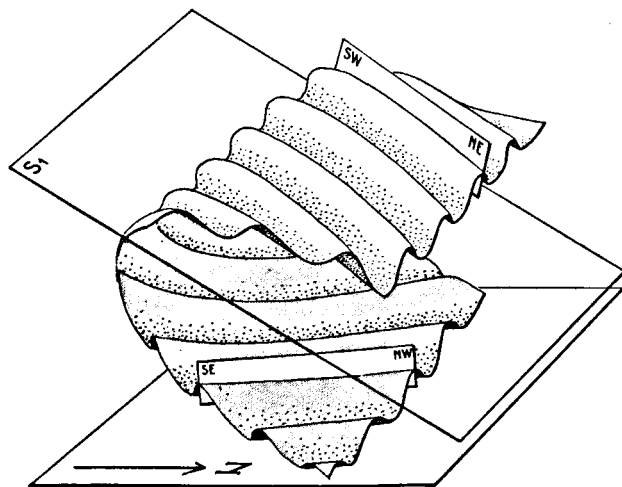


Fig. 34. Relation of pre-cleavage folds in the Ribagorzana Unit and those in the Baliera Unit, the former striking NW-SE on the overturned flank and the latter striking NE-SW on the normal flank of the huge main-phase fold.

N—S trend of the tight folds southeast of Irán (coord.  $42^{\circ}27\frac{1}{2}'$ ,  $4^{\circ}28'$ ) is thought to arise from the steep pre-cleavage folds situated on the vertical part of the overturned huge anticline. The thrusting or gliding of the Lower and Middle Devonian rocks along the Erta fault must have happened later, since it truncates these N—S structures.

Since the axis of this overturned major structure, the axial plane, and the cleavage parallel to it, show a strong bend in the Ribagorzana valley (Ribagorzana hinge-zone), it is thought to be a later phenomenon, but for a more detailed explanation of this deformation we refer to Mey (*ibid.*, p. 204). A somewhat comparable, but less pronounced bend occurs in the southern Tor area (Fig. 24).

In the Carboniferous shale belt, with its constant moderately northward-dipping cleavage, both axes of minor cleavage folds and  $\delta$ -lineations show a large spread in the cleavage plane, suggesting the presence here too of an oblique folding prior to the cleavage formation. This is confirmed by the constructed axes of the adjacent bedding planes, which point to a general NW trend of the early structures (Mey, *ibid.*, Figs. 43 and 44). This is another indication that most of the Carboniferous forms part of the overturned limb of the above-mentioned huge cleavage structure.

Besides, in the transition zone into the strongly refolded Mañanet Unit, minor folds of the  $s_1$ -cleavage with sub-horizontal WNW—ESE-trending axes also occur in one locality about 1 km northeast of the village of Castanesa. The hinge zone of these folds is angular to sub-angular in the slates and rounded in the griotte, but a distinct crenulation cleavage is absent. Although these shapes differ somewhat from the knick-zones encountered frequently in other localities, we do not exclude the possibility that they represent the same phenomenon.

For further details concerning cleavage folds,  $\delta$ -lineations, and knick-zones, as well as an analysis of these features, we refer to Boschma (1963, pp. 166—169) and Mey (*ibid.*, pp. 201—205).

#### Mañanet Unit

Together with the Durro Triangle, this Mañanet Unit is certainly the most complicated and least understood part of our map-area. In this article we will refer only to the most evident structural features, since we can refer the reader to a forthcoming thesis by Roberti (with a 1 : 25,000 scale map), the main subject of which is the structural history of this area and its eastern prolongation.

The Mañanet Unit is characterized by a very strong refolding of the  $s_1$ -cleavage now seen to have a sub-horizontal to roughly southward-dipping attitude. The  $s_2$ -cleavage, on the other hand, has a general NW—SE strike and a 40—70° northeastward-directed dip. This structural unit is bounded by the sub-vertical Espuy line in the north, and by the steeply southward-dipping wall of post-Hercynian formations in the south. In the west the Mañanet Unit grades into the Ribagorzana and Baliera Units,

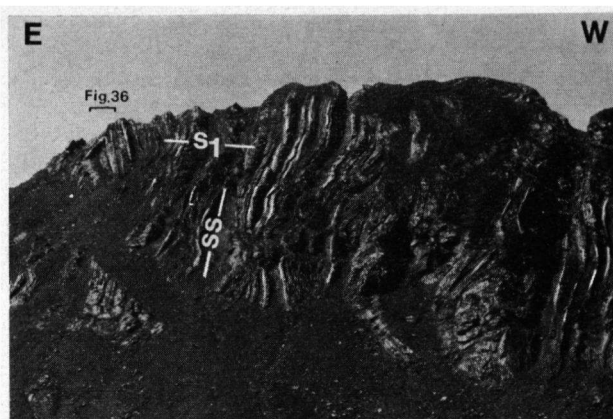


Fig. 35. Steep pre-cleavage structures in the Rueda Formation (western Corrunco area).

both of which have an  $s_1$ -cleavage with a northward attitude and do not show this refolding.

The geological map shows at first sight here again the peculiar NE—SW-trending structures, already demonstrated for the adjoining western areas to be pre-cleavage folds. In the field, most of these structures are seen to be steep and in the west almost isoclinal structures traversed by two sets of cleavages, both with an aberrant strike and dip as compared to the axial planes of these folds (Figs. 35, 36, 37).

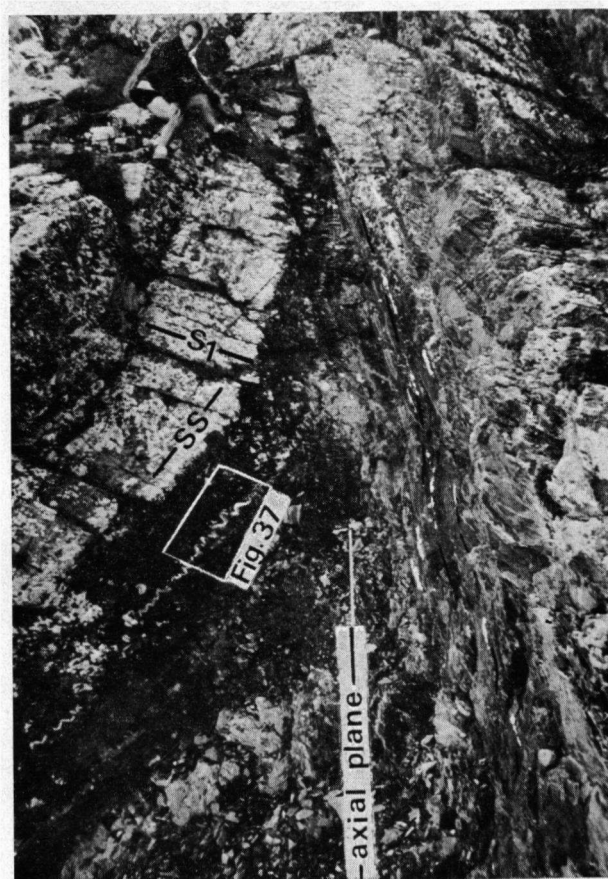


Fig. 36. Detail of Fig. 35.



Fig. 37. Detail of Fig. 36.

A more thorough analysis of these pre-cleavage structures showed us that a great number of these folds — especially those in the Corrunco area — are completely reversed, so that the exposed antiforms are in fact true synclines, and vice versa (Fig. 38). Since large areas with reversed structures alternate with smaller areas where the folds have a normal attitude (Roberti, pers. comm. 1968), two possibilities must be envisaged:

1. The pre-cleavage folds are superimposed on large flanks of still earlier recumbent folds.
2. The pre-cleavage folds are the earliest structures and they now occur in large, normal and reversed, flanks of flat-lying main-phase cleavage folds, striking perpendicular to the pre-cleavage trend.

The latter hypothesis seems the most likely, since in the surrounding, less complex, areas we found no trace of recumbent pre-pre-cleavage folds. Moreover, such very large-scale cleavage folds measured in kilometres are not at all uncommon in the rest of the axial zone. The only striking point is that these cleavage folds here have sub-horizontal to southward-dipping axial planes, an attitude only found so far in the higher-grade metamorphic areas of the infrastructure. It might, of course, be argued that the present

attitude of the main-phase cleavage is a phenomenon related solely to the very strong refolding, but this is highly unlikely because completely horizontal, isoclinal main-phase folds occur in the Flamisell valley further to the east, where refolding is almost absent. Moreover, an over-all Alpine tilt is out of the question (see Roberti, thesis in preparation).

In the field one can generally distinguish easily enough between the pre-cleavage structures and the cleavage folds when their size does not exceed a hectometre. It is not always possible to distinguish between them in the still larger structures, especially where the  $s_2$ -folding crumpled the  $s_1$ -cleavage to such an extent as to obscure the mean over-all attitude of the latter completely.

It must be noted that the  $s_1$ -deformation is much stronger in the rock units above the Basibé Formation (Fonchanina and Mañanet Formations, and Carboniferous shales), the incompetent Fonchanina slates often acting as detachment horizon between folds of the Rueda-Basibé complex and the overlying Mañanet griotte. The consequence of this situation is that the NE—SW-directed pre-cleavage structures have almost vanished in the griotte, where only northward-directed  $\delta$ -lineations reveal their former presence. This disharmony in folding is strongest in the area west of the Erta (see map).

The  $s_2$ -folding, the characteristic feature of the entire Mañanet Unit, requires far fewer assumptions than the preceding deformations. First of all, the reference plane, which is folded ( $s_1$ -cleavage), has been a relatively more constant plane (especially in the strongly schistose rocks) than the bedding plane in the case of the  $s_1$ -folding, which was already strongly folded when the  $s_1$ -cleavage originated. Secondly, the size of these  $s_2$ -folds is such that an entire fold can be studied in one outcrop, so that the measured axis, lineations, axial plane, and axial-plane cleavage form together a much more comprehensible picture. It is, moreover, very fortunate that the axial-plane  $s_2$ -cleavage has a very constant dip and strike over large areas (Fig. 24). For a great number of stereograms incorporating several thousand plane and line meas-

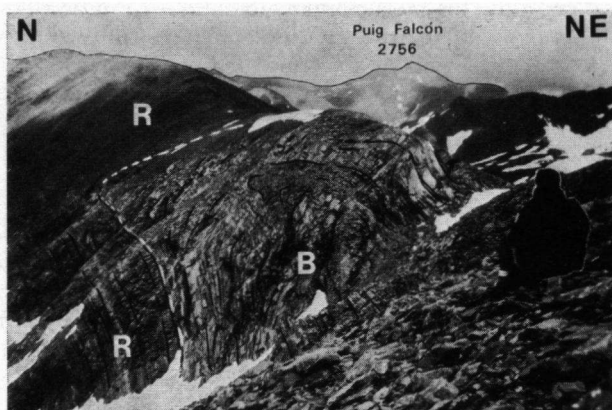


Fig. 38. Antiformal syncline of the Basibé Formation in the Corrunco area.



urements within selected areas, we refer to the thesis of Roberti. How this  $s_2$ -folding appears in outcrops, in the hand-specimen, and in thin section is shown in Figs. 39, 40, 41, 42.

Except for this ubiquitous  $s_2$ -cleavage with accompanying folds, the  $s_1$ -cleavage plane shows locally a faint crenulation with accompanying cracks orientated perpendicular to the already-mentioned major set. This NE—SW-trending feature is, however, hardly ever accompanied by macroscopically visible folds. It is not yet known whether these two directions are the result of one and the same stress-field and were therefore formed simultaneously, or whether they are the product of two separate deformation periods.

A major structural phenomenon requiring mention here, is the Erta fault bordering the Erta block in the north and west. This major fault, with its over-all southward dip, splits up into a northern and southern branch halfway between the Erta and the Mañanet rivers. The attitude of this fault seems strongly related to the local dip of the  $s_1$ -cleavage, varying from 10 to about 90° in a southward direction, the mean dip being about 60°. It is flattest in the west, where the Erta block may be interpreted as a nappe composed mainly of Lower Devonian lying on a substratum of Upper Devonian and Carboniferous. It is thought to

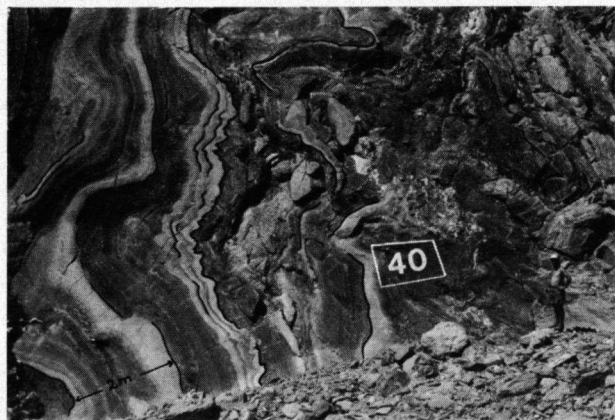


Fig. 39. Typical  $s_2$ -folding in the Mañanet Griotte (in the Mañanet valley).

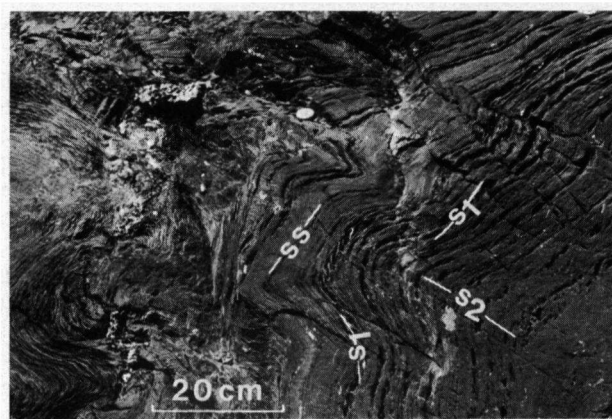


Fig. 40. Detail of Fig. 39.

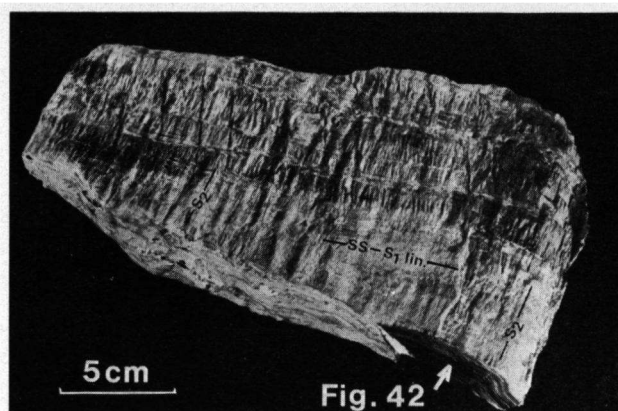


Fig. 41.  $s_2$ -folding in hand-specimen.

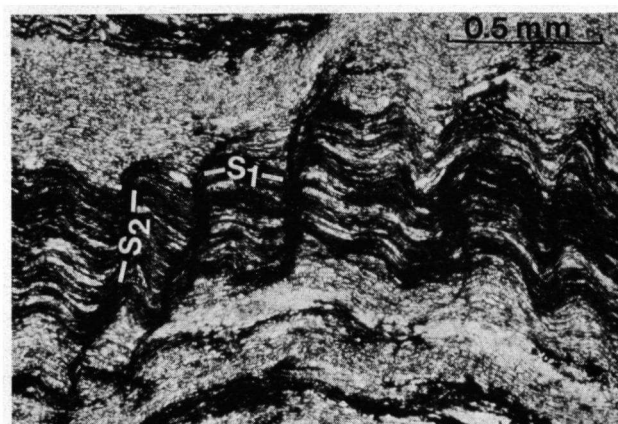


Fig. 42. Detail of thin section taken from hand-specimen shown above.

have slipped down along the  $s_1$ -cleavage plane from an elevated northern area prior to the  $s_2$ -folding (Roberti, pers. comm. 1968). Where this Erta fault is truncated by the post-Hercynian formations of the northern border of the Noguera zone it can be seen that minor fault movements along that plane took place posterior to the deposition of the Triassic, very probably a rejuvenation due to the Alpine deformation.

#### SUCCESSION OF THE HERCYNIAN DEFORMATION PHASES

At the conclusion of the separate descriptions and analyses of the six major structural units in the map-area, it may be useful to summarize the various phases and directions of deformation in chronological order as follows:

1. Pre-cleavage folding phase.
2. Main folding phase.
3. N—S deformation.
4. Major part of the  $s_2$ -folding.
5. Intrusion of the granodiorites, the dykes, and the accompanying metamorphism.
6. Local folding and thrusting.
7. Knick-zones.

1. *The pre-cleavage phase* is characterized by concentric folds without development of a cleavage. Traces of this folding are seen all over the map-area in Devonian and Carboniferous rocks; in the Cambro-Ordovician of this area the presence of this pre-cleavage folding is still doubtful. In the Sierra Negra Unit the folds are probably large, with steep E—W- to WNW—ESE-striking axial planes. In the area south of it, where these folds are preserved best, they are very large, NE striking, and inclined slightly to the southeast. These folds are close to tight in the Lower and Middle Devonian, but virtually isoclinal in the Upper Devonian grotto. The NW—SE trend of the pre-cleavage folds in the Ribagorzana Unit is due to their position on an overturned flank of a huge main-phase fold.

2. *The main phase* is characterized everywhere by tight to isoclinal folds with an axial-plane slaty cleavage. The axial planes have a generally WNW-ESE strike running parallel to the present mountain trend. In the Sierra Negra Unit the cleavage is strong and steep, becoming only moderately dipping near its boundary with the Baliera Unit and the Cambro-Ordovician Payasos Dome, where moderate northward-directed dips dominate. In the Ribagorzana Unit these moderate dips may locally become as low as 30°. In the Mañanet Unit the main-phase cleavage is strongly folded but has in general a southward attitude. The degree of flattening that produced the cleavage decreases from north to south. In the areas with dips of about 45° and less, the cleavage may be a shear cleavage. The Erta block probably slid down from a northern elevated area shortly after the main-phase folding.

3. *A post-main-phase bending* of the entire structure around an axis coinciding with the Ribagorzana River is very likely. A correlation with the second or third deformation phase of the folding scheme given by Zwart seems plausible.

4. *S<sub>2</sub>-folds*, which deform the s<sub>1</sub>-cleavage, are accompanied by an axial plane fracture cleavage. This secondary cleavage has a generally NW—SE strike, but all transitions to an E—W strike are also found locally. Its dip is moderately northeast. Fold axes vary from sub-horizontal (then parallel to the s<sub>2</sub>-trend) to locally steep in a northerly direction; the latter are most frequently found in the westward-plunging parts of the Cambro-Ordovician Muro and Payasos Domes. Less frequently seen is a secondary cleavage with a NE-trend, but folds related to this set have hardly ever been observed. Since the NW- to E—W-striking set of refoldings is found to occur in the hornfels rim of the granodiorites and also enclosed in the contact metamorphic cordierite crystals, the major part of this refolding is dated as pre-intrusion.

5. *The intrusion of the granodiorites* and their accompanying dykes in this compound structure was greatly

facilitated by the pronounced cleavage. The intrusions produced metamorphic aureoles of moderate widths, in which relicts of the s<sub>1</sub>- and s<sub>2</sub>-cleavages are found together with their accompanying folds.

6. *Local folding and thrusting* subsequent to the granodiorite intrusions can be inferred from folded dykes, rotated and cracked cordierite crystals, and the local occurrence of a thrust southern boundary of the Maladeta granodiorite (p. 262). The refolding has an attitude similar to the NW-set of s<sub>2</sub>-foldings mentioned above. This deformation phase may well be a late Hercynian stage, but an Alpine age cannot be altogether excluded.

7. *Knick-zones*, which occur sporadically throughout the area, indicate tensional stress, whereas the above-mentioned structures indicate compression. Consequently, they must be late- or post-tectonic. They may be related to minor tilt movements of the cleavage, although we do not believe that the predominantly moderate dips and the locally sub-horizontal to southward dips of the cleavage in this part of the Pyrenees can be explained by tilting from an originally vertical position. We favour the idea of the original development of an inclined or sub-horizontal cleavage.

#### ALPINE TECTONICS WITHIN THE AXIAL ZONE

The Alpine tectonics can be analysed from the deformation of post-Hercynian rocks; within the axial zone they are all of Triassic age, but in the northern border of the Noguera zone the sediments range in age from Upper Westphalian to Lower Triassic. According to the type of deformation involved, the outcrops of post-Hercynian rocks in the sheet area can be subdivided as follows:

1. The northern outcrops in the Sierra Negra Unit, which are always bound to thrusts.
2. The overthrust synclines located on the boundary between the Baliera and the Ribagorzana units.
3. The continuous steep wall of the northern border of the Noguera zone.

#### Northern area

The northern Triassic, which consists mainly of incompetent and slightly metamorphic marly limestone, cavernous limestone, and dolomite (Pont de Suert Formation), is only extant in thrust zones. The sediments, with a vividly orange-coloured weathering, have been strongly tectonized locally (internal folding, faulting, and brecciation, but no cleavage) and reorientated in the fault zone, so that the history of deformation is difficult to reconstruct.

The most important of these thrusts is the Senet thrust, which splits up into a number of splays about halfway between the Ribagorzana and Baliera rivers. The main thrust and its splays run roughly parallel to the general cleavage trend but with slightly lower dips locally. The amount of throw is difficult to evaluate, but must be at least several hundred metres.



Although east of the Ribagorzana this fault zone is no longer associated with Triassic rocks, the Senet thrust can easily be traced as far as the vicinity of the watershed between the Tor and Mañanet rivers. There it links up with the Espuy line, which may or may not have played a role during the Alpine orogeny. This also holds for the Bordas and Gelada thrusts (see Fig. 23). The Ruenois fault, on the other hand, is a true Hercynian phenomenon, as demonstrated by the fact that it is truncated by the granodiorite and traversed by intrusive dioritic dykes.

#### *Overthrust synclines*

The two large outcrops of Bunter on either side of the Ribagorzana valley are both situated on the thrust boundary between the Baliera and Ribagorzana units. The westernmost of these outcrops mainly represents a NNW-dipping flank of a huge syncline that was overthrust from the north. This thrust (Estét thrust) is rather steep near the Baliera River and has an oblique trend as compared to the Triassic syncline, whose northern limb is still extant about 1 km north-east of the village of Fonchanina. Towards the east, this thrust becomes flatter and also roughly parallel to the strike of the Triassic structure. The dip of this southern flank varies between 0 and 55° in a northward direction; some irregular undulations and one small anticline occur in the eastern part of this outcrop. Both the northern and southern flanks are cut by a well-pronounced coarse fracture cleavage lying almost parallel to the main-phase slaty cleavage in the Devonian below the unconformity.

The elongated outcrop of Bunter on the eastern slope of the Ribagorzana valley consists of NE- to ENE-dipping (10–45°) strata with a rather constant NW to NNW strike. A fracture cleavage with more or less the same strike but with a steep (65–90°) eastward dip is always present. Locally, we encountered tight minor folds of metre to decametre dimensions with the fracture cleavage lying parallel to their axial plane. The large overthrust (Bono thrust) limiting this outcrop in the east runs roughly parallel to the strike of the Triassic strata, but has a slightly steeper dip.

It is interesting to note that, regarded as a whole, the Bono thrust has a rather flat (30–45°) northwestward dip on the western slope of the Ribagorzana valley north of the village Estét, i.e. roughly parallel to the pre-cleavage structures but oblique to the general cleavage trend (see Fig. 24). Near the thrust plane, however, we observed a slight adaptation of the cleavage to this plane. Where the Bono thrust crosses the Ribagorzana valley it has an E–W strike, but it becomes NW–SE-striking (eastward dips between 30 and 65°) on the eastern slope of the valley, i.e. parallel to the pre-cleavage structures in the Ribagorzana Unit but oblique to the general cleavage trend (Fig. 24). Here too, we noticed near the fault plane a slight adaptation of the cleavage in the non-metamorphic area to the direction of the thrust.

Smaller but nonetheless comparable to the Bono and

Estét thrusts is the Comadelo thrust which forms the northern border of a small isolated patch of Bunter on the Pico Comadelo. The outcrop of Bunter in the wedge between the Bono and Estét thrusts, southwest of the village of Estét, is strongly folded and also accompanied by a fracture cleavage dipping north. The relatively poor outcrops do not permit a reliable interpretation of the complicated structure.

With incorporation of the data from the less severely compressed Triassic structures further to the west (Mey, 1967-b, pp. 207–209) the deformation history can be summarized roughly as follows: an initial Alpine compression (N–S) caused an upward shear movement along the pre-existing, moderately northward-dipping planes of the slaty cleavage ( $s_1$ ), producing at the same time in the Triassic strata above the unconformity undulations, shear folds and overthrust folds accompanied by an axial plane fracture cleavage. These shear movements were strongest in the best cleaved rocks, such as slates, and it is therefore comprehensible that the shear zones either lie parallel to the over-all cleavage or follow incompetent horizons of persistent Hercynian structures (cleavage or pre-cleavage structures). Further compression beyond the maximum point of simple folding initiated axial plane thrusts (Bono, Estét, and Comadelo thrusts), along which further movement could take place. Once the fault had traversed the rather competent Bunter, further gliding of the Triassic and Palaeozoic rocks over the sub-horizontal southern flank was greatly facilitated by the incompetent series of the higher Triassic (e.g. marl and gypsum). These and similar overthrusts may be considered as the roots of gliding masses encountered locally in the Nogueras zone.

#### *Northern border of the Nogueras zone*

The continuous steep wall of post-Hercynian formations at the northern border of the Nogueras zone consists predominantly of Lower Triassic rocks west of the Tor River, and of Upper Westphalian, Stephanian, and Lower Triassic rocks east of this river. The strike of this wall is constantly WNW–ESE and its dip varies from a 45° normal dip to a 50° overturned attitude locally, but the mean value is about 70° in a southward direction. The faint undulations of the strata have a wavelength of several hectometres; close folds of metre dimensions are only found sporadically. The lower part of the Bunter sediments is cut by a coarse fracture cleavage having roughly the same strike as the bedding, but a reverse (i.e. northward) dip.

In the western part of the area, where Permo-Triassic rocks unconformably overlie the Carboniferous shale belt of the Ribagorzana Unit, the Alpine fracture cleavage in rocks above the unconformity is strikingly parallel to the Hercynian main-phase cleavage below it (Mey, 1967-b, Fig. 51). Very often, the plane of unconformity is seen to be displaced over a few metres along small faults parallel to this fracture cleavage. In the Carboniferous shales the same movement must have taken place along the pre-existing  $s_1$ -cleavage.

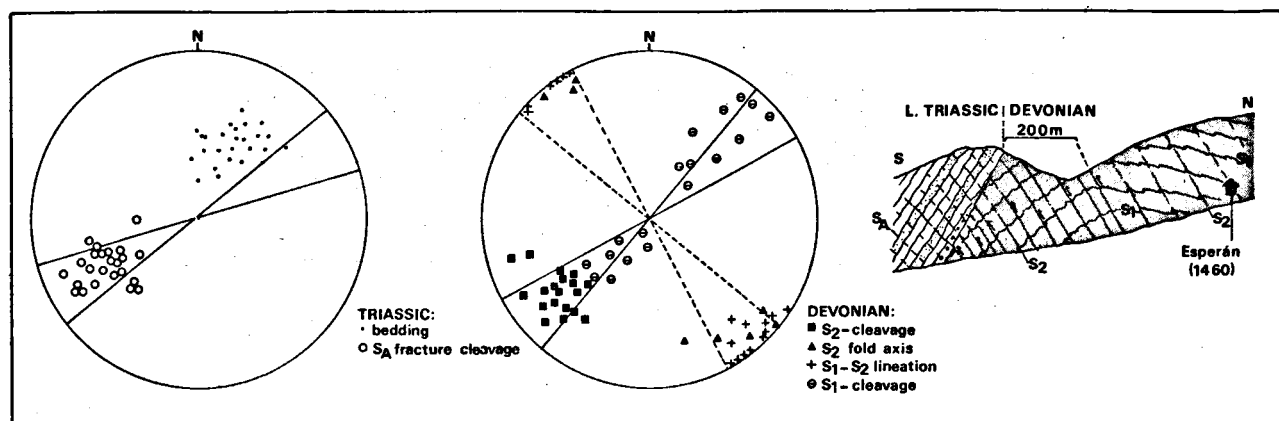


Fig. 43. Orientation diagrams and cross-section showing the relation between the Alpine fracture cleavage in the Bunter ( $S_A$ ) and the secondary Hercynian cleavage ( $s_2$ ) in the Devonian below the unconformity.

From the unconformity upward, this fracture cleavage becomes weaker, and it is completely absent in the top layers of the Bunter. It is therefore evident that the structures of the Triassic are strongly related to the well-cleaved Palaeozoic structure below. An upward shear movement along a large number of cleavage planes in the Carboniferous was most probably responsible for this fracture cleavage and the present position of Triassic strata. Unequal movements along certain zones would explain the undulations in the southward-dipping wall of Bunter rocks.

An important deviation from the generally south-dipping wall of Bunter is found on the lower slopes of the Ribagorzana valley, where there is a locally overturned northward attitude of the Triassic strata. This divergent structure is moreover accompanied by a number of N—S-striking faults with steep as well as flat dips and displacements of up to 50 m. It is interesting to note that in this overturned structure the fracture cleavage is horizontal to locally southward-dipping, although the cleavage of the Carboniferous has a normal dip to the north. This might be explained by a further rotation of the Triassic strata subsequent to the general steepening by shear.

In the eastern part of the area, where various post-Hercynian formations unconformably overlie Devonian rocks of the Mañanet Unit, the steep wall makes a low angle with the persistent  $s_2$ -trend in this structural unit (Fig. 24). A pronounced Alpine fracture cleavage is only present in the detrital Permo-Triassic rocks; this cleavage lies strikingly parallel to the  $s_2$ -cleavage in the Devonian below the unconformity. For the Esperán area this is shown in Fig. 43.

Geologists advocating the view of a strong supple deformation of the axial zone during the Alpine orogeny (e.g. Mattauer) might be inclined to interpret this secondary cleavage as an Alpine phenomenon, but since the wave of secondary folding can be traced continuously up to the Maladeta granodiorite, where it is found to occur in the hornfels rim, its Hercynian origin is now well established.

We assume therefore, that the Alpine compression here

too reactivated the potential planes in the Devonian, which, in the Mañanet Unit, are the secondary cleavage planes.

To summarize, it may be said that the Alpine orogeny caused a considerable shortening of this part of the axial zone, brought about by large overthrusts and upward shear movements along a large number of cleavage planes. Secondary flattening along pre-existing slaty cleavage planes of the Palaeozoic underground may have also played an active role locally, but this cannot be proven. At the same time, the unconformable post-Hercynian sediments were steepened, folded, cleaved, and overthrust from north to south.

## NOGUERAS ZONE

### Introduction

This major structural unit, which is separated from the axial zone by a steep flexure zone, is characterized by large slabs of normal and overturned strata ranging in age from Stephanian to Upper Triassic. The largest of these slabs incorporate in their northern part pre-Hercynian rocks (mainly Devonian). True folds of observable size are restricted to the marls and gypsum of the Upper Triassic. The diapirism of the incompetent Keuper series is undoubtedly the major cause of the irregular and sometimes chaotic attitude of this fault zone.

This Nogueras zone, which in the south is bounded by regularly southward-dipping strata of Jurassic and Cretaceous age, begins in the west in the Esera valley (sheet 7) and retains the above-mentioned characteristics up to the Segre valley in the east. The Palaeozoic blocks occurring in this zone have intrigued geologists since the first general mapping (Dalloni, 1910), and their origin is still the subject of active discussion.

Before we describe this zone in more detail and present our own interpretation, a short historical review of the former interpretations will serve as a useful introduction.

### Historical review

Dalloni (1910, 1913), who made the first geological map of the Southern Pyrenees, including part of our area, introduced the name "*Zone de Nogueras*" for the narrow stretch between the axial zone in the north and the Cretaceous zone in the south, in which he encountered Mesozoic rocks intimately folded together with Palaeozoic. He described several recumbent folds with a southward-directed attitude, and to explain the position and orientation of the Palaeozoic blocks within that zone he assumed a relationship between these blocks and a large thrust-sheet coming from the north, which he called "*Nappe de Nogueras*".

This interpretation raised many objections. We give here essentially the conceptions of Jacob *et al.* (1926), on the one hand, and Misch (1934) on the other. The former authors, who made a complete study of the transition zone between the axial zone and its Mesozoic southern cover, came to the conclusion that most of the Palaeozoic blocks were more or less in place, and that others, especially the smaller ones between the Flamisell and Esera rivers, formed part of recumbent folds *in situ*, with sub-horizontal to southward-dipping axial planes. Certain blocks, especially those of Santa Coloma, Malpas-Gotarta, Escané, and Las Paules, were considered to represent the same recumbent fold, the roots of which they assumed to lie in the Sas-Benés area ("*Synclinal de Sas*"). According to this interpretation, the north-to-south displacement of the different units was of little importance. These authors admitted that the latter, slightly allochthonous blocks represent the western extremity of the "*Nappe de Nogueras*" (Dalloni, 1913), but they definitely excluded from it the larger Palaeozoic masses between the Segre and the Flamisell (e.g. the Freixa block), which they interpreted as upthrust from the south.

Dalloni (1930) rejected this idea and continued to defend his nappe-theory on the basis of important north-to-south displacements of the different structural units.

Misch (1934), who made the first relatively detailed and very trustworthy map of this area (1 : 175,000), explained the structure as an originally virgating bundle of parallel folds, sometimes dislocated from their roots, creating in this way normal and overturned structures with north- as well as southward dips. He considered all Palaeozoic blocks as autochthonous or sub-autochthonous. Against the hypothesis of a north-to-south-moving nappe, he advanced the following important fact: the Stephanian of the Las Paules and Gotarta blocks is relatively thick, although it is absent below the Permo-Triassic overlying the axial zone immediately north of these blocks, where the roots of the nappe would be expected.

The views of Misch were accepted by Jacob (1935) and later by the Spanish geologists Almela and Rios (1947).

More recently, de Sitter (1962, 1965) interpreted the Palaeozoic blocks of the Nogueras zone as equivalents of the North Pyrenean Satellite massifs, which implies an autochthonous origin. Nagtegaal (1962) and Seguret (1964), on the other hand, took up the old nappe theory of Dalloni, explaining the inverted Permo-Triassic and overturned Palaeozoic blocks in the Nogueras zone as anticlinal hinge-zones ("*têtes plongeantes*") with a northern origin. In the case of the Malpas-Gotarta and the Las Paules blocks, the nappe could have come from an eroded area at least 6 km north of their present position in the Nogueras zone (Seguret, 1964, 1966).

Mey and Zwart (1965) did not give a map but advanced strong tectonic as well as stratigraphic

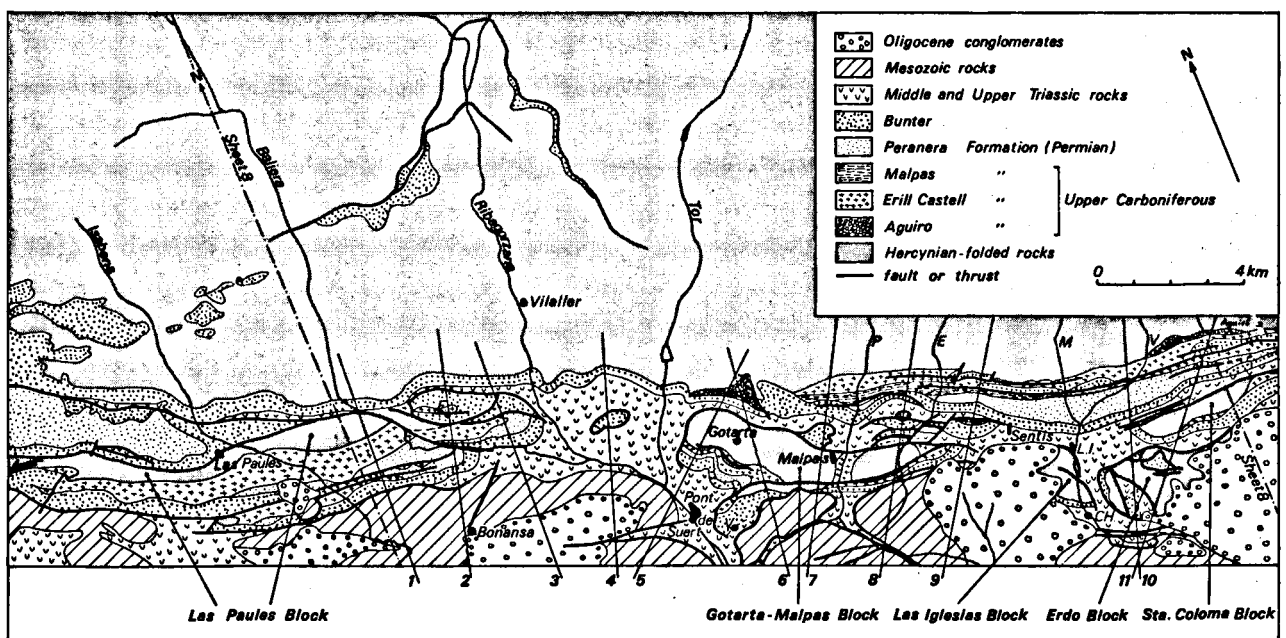


Fig. 44. Sketch-map of the Nogueras zone from west of Las Paules to Aguiró in the east.

arguments supporting an autochthonous origin of at least the Palaeozoic Las Paules block. Roger (1965) reached the same conclusion independently. He suggests that the Las Paulas-Urmella block is part of a mushroom-like fold "*in situ*".

Wennekers (1968), on the other hand, sees in the Nogueras zone a complex thrust structure everywhere present between the axial zone and the overlying younger Mesozoic series. According to him, the now outcropping part of the Nogueras zone is purely an erosion feature of an uplifted block (caused by late-tectonic block-faulting). The isoclinal recumbent folds of the Stephanian and Permo-Triassic rocks with or without accompanying Devonian/Carboniferous rocks are interpreted as coming from the north.

This brief review gives old and new conceptions put forward to explain the structure of the Nogueras zone. There are, however, a number of key arguments throwing new light on the duality of this complicated structural zone.

#### *Significant features pointing to the origin of the Palaeozoic blocks*

We prefer not to use the terms allochthonous or autochthonous for the slabs and blocks in the Nogueras zone, because hardly any of them are still in place. The two conflicting theories are better expressed by saying that some authors advocate gliding from an elevated northern area and others are convinced that faulting and thrusting within the Nogueras zone pushed these blocks upwards into the Keuper series.

With this in mind, we analysed the axial zone, the remnants of Permo-Triassic within the axial zone, the northern border of the Nogueras zone, and all slabs and blocks in the Nogueras zone itself. We soon came to the conclusion that there are a number of striking differences between the axial zone and its direct Mesozoic cover, on the one hand, and most of the slabs and blocks occurring in the Nogueras zone on the other. These differences, which can be divided into several groups, are listed below; how they pertain to each block is shown separately.

1. Difference in lithology of the Palaeozoic rocks.
2. Presence or absence of a slaty cleavage in the Palaeozoic blocks.
3. Difference in the kind of igneous bodies intrusive in Palaeozoic rocks.
4. Presence or absence of a fracture cleavage in the Bunter rocks.
5. Presence or absence of Stephano-Permian rocks between the Hercynian folded rocks and the Bunter.
6. Difference in the basal layers of the Bunter.

**A. Las Paules block.** — Of the Palaeozoic part of the Las Paules block, only the eastern extremity lies within the area of sheet 8 (see Fig. 44). However, we analysed the entire Devonian/Carboniferous, a columnar section of which is shown in Fig. 45. The upper part of this section is well-exposed along the road between the small village of Renanué and the Las Fadas Col

(sheet 7), and the lowermost part is found, albeit in a less continuous way, in the eastern part of this block. Even the most pelitic rocks of this Devonian sequence show no signs of any cleavage.

The Devonian in the axial zone north of the Las Paules block, which is everywhere cleaved, has a quite different lithology (see Fig. 3). First of all, the black massive limestones are absent in the uppermost part, as are the chert lenses and beds. The platy, sandy limestones below it, as well as the sandy limestone intercalations in the underlying mudstone sequence, are also unknown in the axial zone north of this Las Paules block. Of the lowermost part, which can be correlated with the Basibé and the Gelada Formations, no sign of any quartzite within the limestone has been found. A comparable Devonian sequence in the axial zone occurs only in the southern part of the Esera valley and further to the west, but there it is always accompanied by a strong slaty cleavage, visible even in the limestones.

On the basis of this stratigraphic evidence and the lack of a cleavage, a direct northern origin of the Las Paules block (by gliding from a northern elevated area) is very unlikely.

Another, at least equally strong argument against a northern origin is the fact that in the Las Paules block

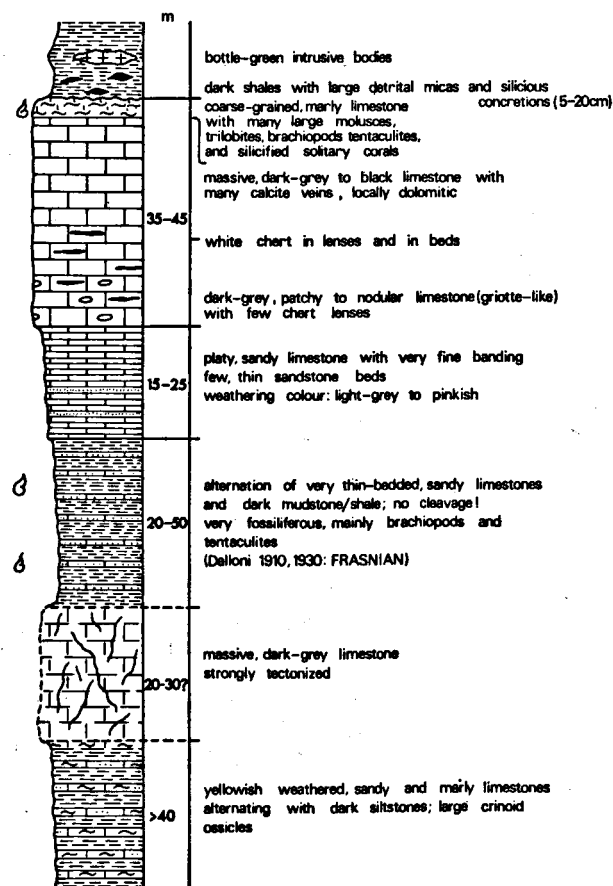


Fig. 45. Columnar section of the Hercynian-folded rock sequence in the Las Paules block.

the Hercynian sequence is unconformably overlain by Stephanian pyroclastic rocks (up to 900 m) and Permian detrital rocks (up to 500 m), whereas in the axial zone north of this block the Devonian/Carboniferous is directly overlain by the detrital deposits of the Bunter. Since we know from the area east of the Tor that in the exposed part the Bunter shows a constant overstep over Permian and Stephanian rocks towards the north, the thickest remnants of Permo-Stephanian can be expected in the Noguerras zone itself.

A fourth argument is the strong fracture cleavage always present in the lower part of the Bunter when it overlies cleaved Palaeozoic rocks, e.g. in the overthrust synclines within the axial zone and the steep wall of Bunter of the northern border of the Noguerras zone. The Permo-Triassic rocks within the Noguerras zone, on the other hand, have no fracture cleavage or only locally a very weak one. If these blocks represented anticlinal hinge zones originating from somewhere within the axial zone and having glided into their present position in the Noguerras zone, one would

expect a fracture cleavage at least as strong as the one present in the overthrust remnants of the Triassic synclines.

Less convincing is the argument that the basal layers of the Bunter synclines in the axial zone often have a coarse limestone or quartzite breccia at the base (see page 244), whereas in the Noguerras zone the Bunter starts with a coarse but well-rounded conglomerate.

A last argument against a northern origin is provided by the numerous small and large intrusive bodies, which are restricted to the Carboniferous shales in the Las Paules block. These rocks, clearly intrusive in most outcrops, are sub-volcanic with an andesitic to basaltic composition. They are undoubtedly related to the eruptive tuffs and lavas of the overlying Stephanian. Most of the intrusive dykes within the axial zone are abyssal rocks of a (quartz-)dioritic composition; moreover, they intruded in late Hercynian times.

In our opinion, therefore, there can be no doubt about the sub-autochthonous origin of the Palaeozoic Las Paules block and its surrounding Permo-Triassic and Stephanian cover. The attitude of especially the

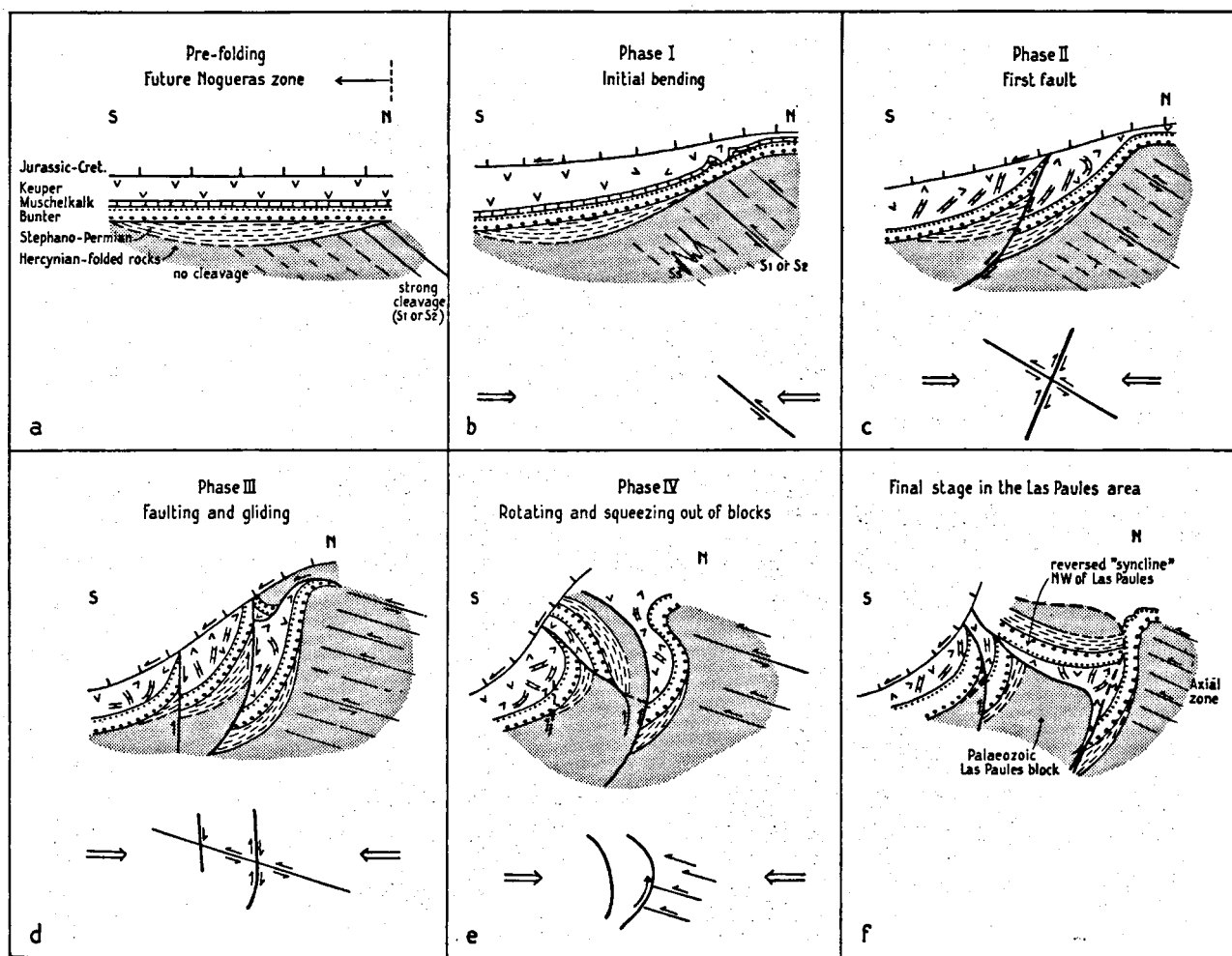


Fig. 46. Structural model of the deformational history of the Noguerras zone.

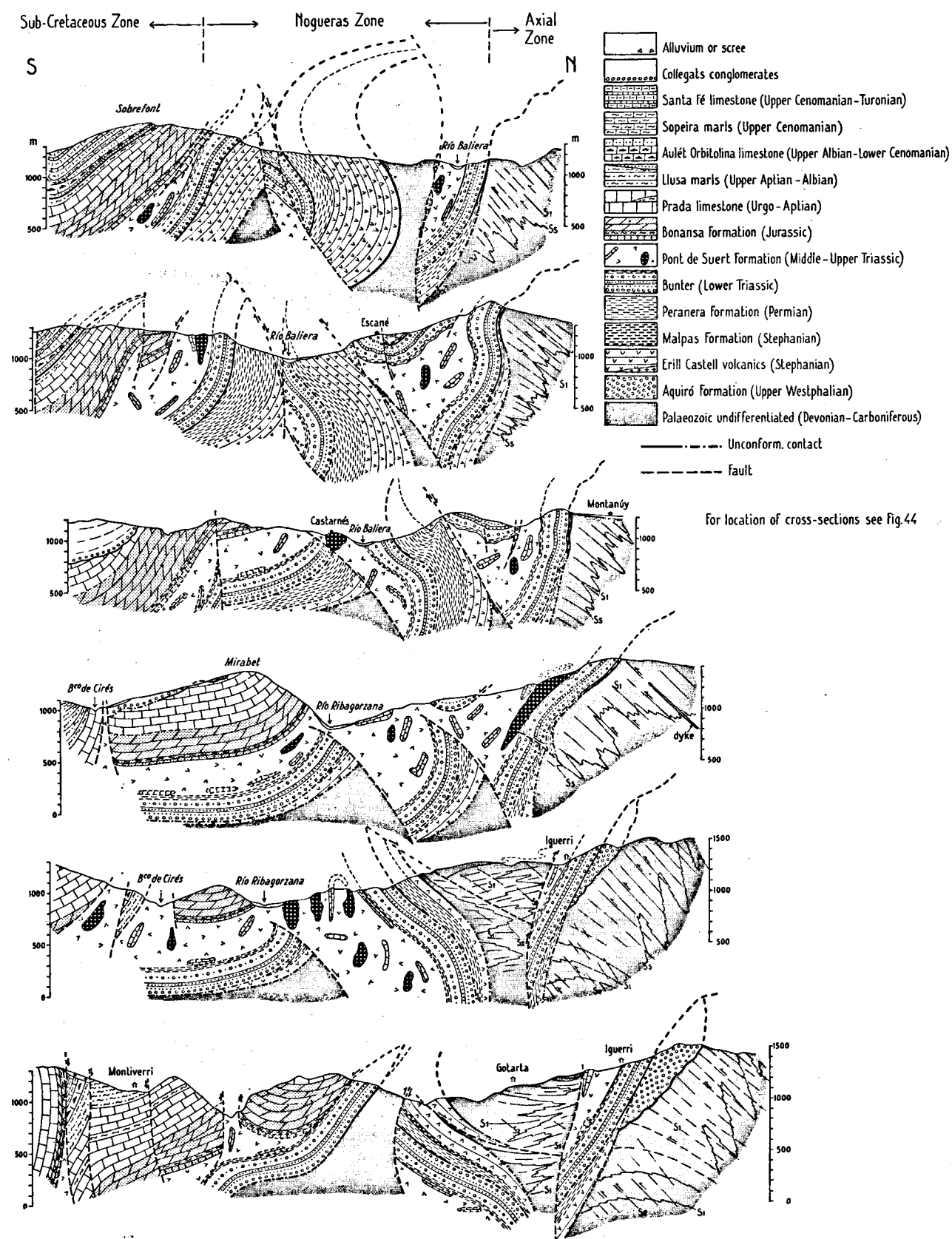


Fig. 47. Cross-sections through the western part of the Nogueras zone (scale 1 : 53,250).



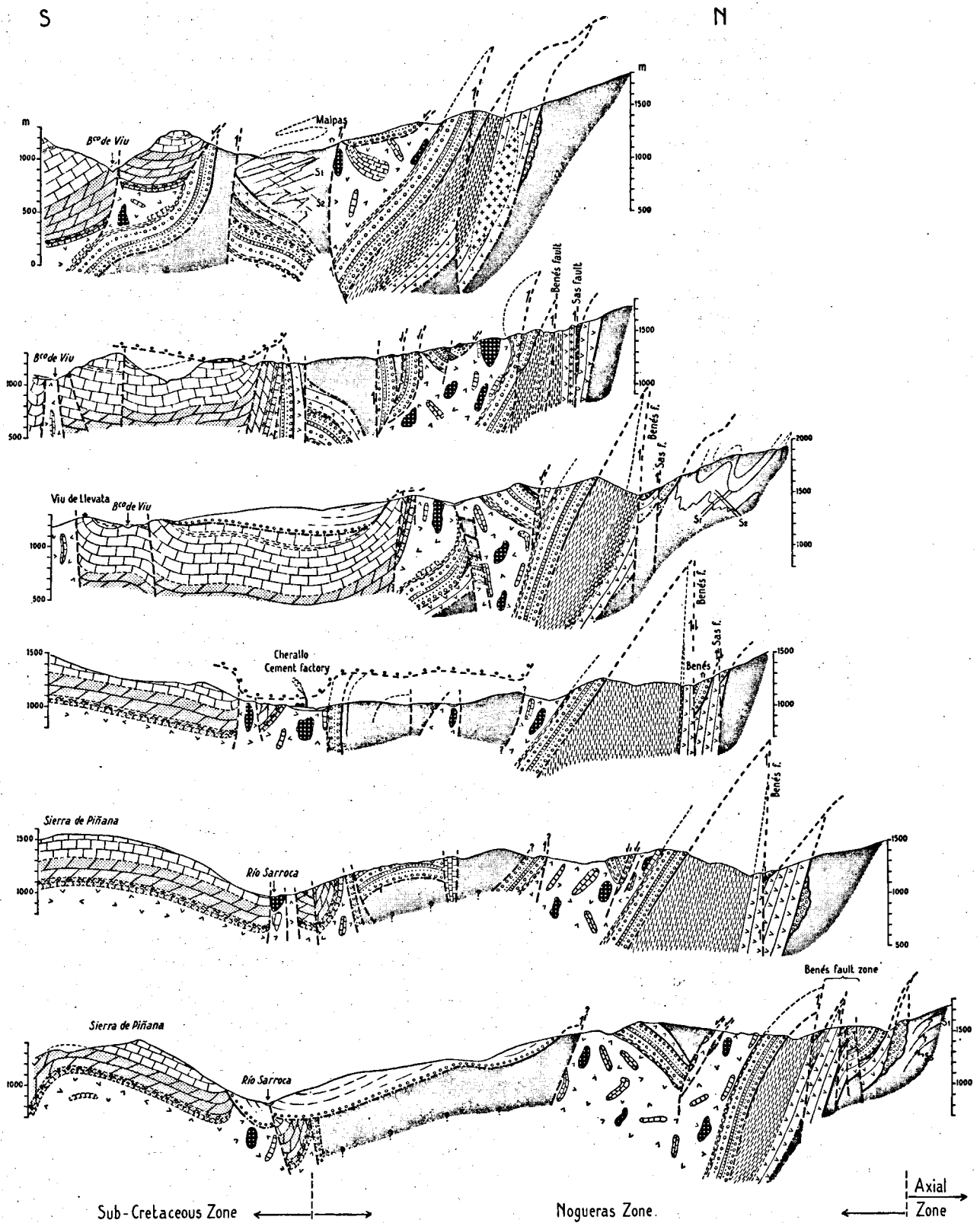


Fig. 48. Cross-sections through the eastern part of the Nogueras zone (scale 1 : 53,250).

synformal parts with a reverse stratigraphy (e.g. the large unit northwest of Las Paules and the one in the Sierra de Escan ) poses some problems, but the proposed model of deformation (see Fig. 46) is capable of explaining even these completely reversed flanks quite satisfactorily.

**B. Gotarta-Malpas block.** — The western and eastern margins of this block supply very attractive arguments for geologists who advocate gliding from an uplifted part in the axial zone. It is, however, wishful thinking to map the eastern termination as a simple reverse syncline ("*faux synclinal*") (Nagtegaal, 1962; Seguret, 1966). As shown on our map, the Gotarta-Malpas block consists of a northern and a smaller southern unit separated from each other by a sub-vertical fault.

The Palaeozoic rocks of the larger northern block are in the main comparable to the Ma anet and Fonchanina Formations, which are exposed over large areas in the axial zone to the north. As distinct from the Las Paules block, the Devonian rocks here possess a normal slaty cleavage (dipping gently towards the southwest), which shows faint local refolding. Even the overlying Permo-Triassic rocks show a faint fracture cleavage with a gentle southward attitude. So thus far there are no objections to a northern origin.

The only reasonable argument that we can put forward against a northern origin is the fact that this block is accompanied by about 100 m of Permian rocks (Peranera Formation) intercalated between the Stephanian pyroclastics and the Bunter. With respect to the northern border of the Noguerras zone, we would have to go to the Erta and Ma anet areas to find a considerable amount of Permian rocks in the same geological setting. There, however, the Palaeozoic is missing as possible source of the Gotarta-Malpas block. North of the Sas fault, where Stephanian overlies the Devonian of the axial zone, the Permian has already wedged out (see cross-sections, Figs. 47 and 48).

Unless one accepts the former presence of thick Stephano-Permian basins within the axial zone below a cover of Bunter, which we strongly doubt, the Gotarta-Malpas block must also be regarded as sub-autochthonous. Support for this view is found in the presence of several sub-volcanic intrusive bodies near the village of Gotarta, rocks also found associated with the Carboniferous of the Las Paules block (see page 275).

The Hercynian-folded rocks of the smaller southern unit are exclusively weakly-cleaved, Carboniferous shales. The post-Hercynian cover, which consists mainly of Bunter rocks and locally of some Stephanian pyroclastics, has a normal south-dipping attitude in the west, is vertical to slightly overturned in the Peranera valley, and acquires a 50–70° overturned dip in the east. A fracture cleavage, locally present in the softer Bunter rocks, lies more or less perpendicular to the bedding.

The position of this unit between the northern Gotarta-Malpas block, which in our opinion is an

upthrust unit, and the normal Jurassic-Cretaceous cover in the south, is the only argument we can advance for its probable sub-autochthonous origin. This assumption would imply, however, that the Stephano-Permian remnants below the unconformity of the Lower Triassic (Bunter) are not only wedging out towards the axial zone but also south of the axis of the Noguerras zone (see Figs. 47 and 48). One comes to the same conclusion from the construction of proper sections over the Las Paules block in the area west of sheet 8 (of course, only if one accepts the sub-autochthonous origin of this latter block).

The several, much smaller Bunter units north of the Gotarta-Malpas block, most of which are in an overturned position and are accompanied by a fracture cleavage roughly perpendicular to the bedding, all have one very important thing in common: Stephano-Permian is absent between the Bunter and the thin strip of Devonian, with only one known exception, south of the San Quiri, where less than 10 m of Peranera Formation is intercalated. In view of the fact that north and south of these units the upthrust post-Hercynian cover consists of a several hundred metres thick series of Stephano-Permian rocks, it is very unlikely that the intermediate units lack these rocks. It therefore seems plausible that these small units reached their present position by gliding down from the axial zone.

**C. Northern border of the Noguerras zone east of the Barranco de Peranera.** — The post-Hercynian rocks of the northern border of the Noguerras zone east of the Barranco de Peranera have, as everywhere else in the map-area, a constant moderate to steep southward attitude. N–S sections reveal, however, a repetition of the various lithostratigraphic units without any signs of folding. The repetition must therefore be due to faults running roughly parallel to the general strike of the strata but with a slightly steeper dip. Occasionally, the fault planes can be mapped in the field. The most important of these fractures are the Sas and the Ben s faults, both of which are sub-vertical, as can be gathered from their fault-trace.

Dip and strike measurements of the post-Hercynian strata below and above the Triassic angular unconformity provide a relatively good tool for estimating the throw along these sub-vertical faults. The maximum throw for the northern Sas fault is up to 500 m and for the Ben s fault almost 1500 m (see cross-sections, Figs. 47 and 48). Near the eastern boundary of sheet 8, the throw along the Ben s fault itself diminishes to only a few hundred metres, the total upthrow, which is still about 1500 m there, being distributed over a number of parallel faults.

**D. Las Iglesias block.** — This block, exposing mainly Devonian rocks with only in the south a thin strip of Bunter, is bounded by incompetent Keuper series in the north, east, and south, and is in the west unconformably overlain by the Oligocene conglomerates.

Except for the southern part, this unit is poorly exposed, as are the contacts with the Keuper. As far as can be seen, the exposed Devonian stratigraphy is comparable to that in the axial zone. A slaty cleavage as well as locally a secondary cleavage are developed in the more schistose rocks. There is no evidence to indicate whether this is an upthrust block or a gliding structure with a northern origin. However, if the entire structural setting of this part of the Noguerras zone is taken into account, a sub-autochthonous origin seems more likely.

**E. Erdo block.** — This block, in which the Palaeozoic rocks range from Silurian to Carboniferous and whose post-Hercynian cover consists solely of Bunter, has very badly exposed boundaries with the surrounding Keuper. In the east and southeast this block is unconformably overlain by Oligocene conglomerates. The outcrops within the unit itself are not sufficiently distinct to permit reliable interpretation of the slabs of Muschelkalk and Bunter that occur.

To return to the Palaeozoic rocks, which show a cleavage with strongly varying attitudes, some obvious differences from their occurrence in the axial zone are observed. First of all, the Fonchanina slates intercalated between the Basibé limestone and the Mañanet griotte are here developed as pink to reddish calc-schists, a facies which is known to occur in the axial zone only in and east of the Segre valley (Mey, 1967-a). Secondly, the transition from the Mañanet griotte to the Carboniferous shale sequence is marked here by the occurrence of bedded, black chert, a phenomenon otherwise unknown in this part of the axial zone.

These two facts seem to constitute sufficiently strong evidence to permit interpretation of this very irregular complicated structure as an upthrust block.

**F. Santa Coloma block.** — This structural unit is composed of mainly Devonian, Stephanian, and Bunter rocks. Its position on top of a hill (Sierra de Santa Barbara) that is everywhere surrounded by the incompetent Keuper, has long attracted the attention of geologists. Because of this peculiar position, it is hardly surprising that all workers (except Misch) have seen this block as a remnant of a gliding nappe or a detached recumbent fold with its roots near the southern margin of the axial zone. We share this opinion, since all the evidence we collected supports a northern origin. This evidence can be summarized as follows:

1. Devonian stratigraphy mainly comparable to that in the axial zone.
2. Presence of a strong slaty cleavage and secondary cleavage within the Devonian.
3. Very low reversed dip of the Bunter rocks in the south.
4. Absence of Permian, although it is very thick (up to 700 m) immediately north of and consequently below this block (see cross-sections).
5. The moderately to steeply southward-dipping fault zone bordering this block in the north in-

corporates small and large lenses and slabs of Palaeozoic and Triassic rocks, including strongly sheared bodies of ophite.

To summarize, we may say of at least this part of the Noguerras zone that the majority of the blocks and slabs of Palaeozoic, Stephano-Permian, and Bunter rocks surrounded by an incompetent matrix of the Keuper are derived from within the Noguerras zone itself and that a few others very probably have a northern origin.

#### *Structural model of the deformational history*

As the starting point for our model we take a situation such as that shown in Fig. 46a: A Stephano-Permian basin, roughly coinciding with the future Noguerras zone and unconformably overlain by Lower Triassic and younger Mesozoic rocks. The northern overstep of the Bunter towards constantly older post-Hercynian Palaeozoic formations is clearly expressed in the northern border of the Noguerras zone east of the Tor River. The overstep towards the south is also very likely, as we have shown on page 278. We further assume in our model that the intensity of the cleavage in the Hercynian basement diminishes from north to south, a tendency also expressed in the exposed axial zone (e.g. Mey, 1967-b, p. 205).

The attitude of potential planes ( $s_1$ - and  $s_2$ -cleavages and fault planes) in the Hercynian-folded basement is such that a horizontal N—S compression would be capable of pushing the fissile complex together like a strongly inclined pack of cards, resulting in a relative rise of the northern area. Since we know, however, that the northern border of the Noguerras zone is a steep flexure zone, occurring on this scale only once in the southern part of the Pyrenees, it is evident that this pack-of-cards effect must have been concentrated in a relatively narrow stretch, a mobile zone probably related to a deeper-seated major fault or thrust. The same kind of movement, but on a much smaller scale, was active in the axial zone more to the north below the remnants of the overthrust synclines (see page 271).

Because of their strongly differing competence, the individual lithostratigraphic units of the post-Hercynian cover reacted to this push-effect in a quite different manner. Whereas the pre-Keuper detrital rocks were compressed into an asymmetrical bend (Fig. 46b) and the Muschelkalk was partially broken up into a number of isolated slabs, the incompetent Keuper probably reacted as an easily deformable cushion between this bend and the more rigid Jurassic-Cretaceous roof: in the crest of this bend it is likely that the Keuper was squeezed out completely and flowed to the protected trough to the south. The relative rise of the future axial zone initiated gravitational gliding of the competent Mesozoic complex above the Keuper.

From a mechanical point of view it is feasible that further compression was considerably facilitated by an upward movement along a secondary fault plane

orientated perpendicular to the initial thrust planes, together forming a conjugate set of faults (Fig. 46c). The Sas and Benés faults are good examples of these secondary faults.

Still further compression initiated more of these faults, the blocks between these planes being pushed up and rotated (anticlock-wise in our drawings); this rotation was certainly in part stimulated by the down-gliding mass of the Jurassic-Cretaceous nappe (Fig. 46d). In this stage small blocks of Bunter with accompanying Devonian/Carboniferous rocks were carried along with the advancing Mesozoic nappe. In some parts of the Nogueras zone the outcome of this situation is still preserved.

Together with the rotation of the various blocks, the formerly southward-dipping faults were steepened and even became overturned in the higher parts, a situation which provoked the squeezing out of the most northern, upthrust unit (Fig. 46e). Once wholly surrounded by the incompetent Keuper series, further shifting, rotating, and gliding of this squeezed-out plug could occur to settle the mass in the most favourable (minimum stress) position within the over-all geological environment. Continued diapirism of the Keuper is thought to have been responsible for the later weak bending into the peculiar synformal shapes such as are encountered, for instance, northwest of Las Paules and in the Sierra de Escané (Fig. 46f).

#### SOUTHERN CRETACEOUS ZONE

We can be very brief in describing this zone, since the most important structural features already figure on the maps and cross-sections of earlier workers and need hardly any further explanation. The general folding style, as seen on our cross-sections, is governed mainly by the diapirism of the incompetent Keuper series, which intruded along most of the vertical cracks, sometimes resulting in the dragging of the adjacent Mesozoic strata. Smaller blocks bordered by such gypsum-filled cracks are often seen to have been basculated.

The folds in this zone, which are of the pure concentric type, generally measure several kilometres, although locally the more finely laminated marls show minor folds of decimetre to metre size; the latter are occasionally accompanied by a coarse axial-plane fracture cleavage (e.g. in the Ribagorzana valley where the main road crosses the Barranco de Inglada).

A peculiar feature of this area is the narrow WNW-trending zone south of the Fallada de Malpas and north of the Barranco de Inglada, in which the strata are overturned and in which north-to-south overthrusts occur. This zone probably links up with the overturned thrust zone south of the Cruz de Bonansa, which continues to show overthrusting up to the N—S-trending Turbon structure in the west (Hupé, 1954). This entire zone might be interpreted as a phenomenon comparable to the recumbent folds west of the Esera River, which are frontal features associated with the Gavarnie nappe (Misch, 1934; Seguret, 1967). There, the gravitational gliding from north to south over several kilometres is well established. In our area, where we also assume some north-to-south gliding of the Mesozoic strata on top of the Keuper (see page 279), recumbent folds were less likely to originate because the advancing nappe descended into a soft, very thick bed of Keuper where friction was negligible. Only further to the south, where thick Senonian and Eocene deposits obstructed the advancing nappe, would some overturning with accompanying upthrusts from north to south be expected.

Before closing this chapter we must add that, although the main folding of the Mesozoic strata is known to have occurred in late Eocene times, some later movements still took place in and near the cracks and zones filled with Keuper. This is shown by the local steep dips of the post-orogenic Oligocene conglomerates near these zones and the local faults associated with gypsum passing through these conglomerates. Examples of these late phenomena can be seen, for instance, in the Barranco de Cirés, west of Pont de Suert, and in the valley of the Rio de Sarroca, southeast of Cheralló.

### CHAPTER III

#### INTRUSIVE AND METAMORPHIC ROCKS

##### INTRODUCTION

The area of the present study shows a considerable variety of intrusive rocks, including a huge late-tectonic granodiorite batholith and a few smaller accompanying stocks, a large number of dykes of various types, and a few post-tectonic basic rock intrusions. The ophiolites in the Triassic of the Nogueras zone have already been discussed (page 247). The granodiorite has produced an appreciable thermal metamorphism in the host rock. In the metamorphic Bono area there seems to be a close relationship between the metamorphism, the hydrothermal lead and copper

ores, and the multitude of dykes swarming out of this area, which probably represents the metamorphic roof of a more deeply seated intrusive body.

Only a very brief description of the above-mentioned rocks will be given here; their petrological and mineralogical problems will not be discussed.

##### GRANODIORITES AND THEIR METAMORPHIC AUREOLE

The biotite-granodiorites of the Central Pyrenees of, which the Maladeta batholith is one of the largest (almost 400 sq.km), are all of the diapiric intrusive

type with sharp and clearly discordant borders. The date of their emplacement can be established on the basis of the various folding phases and with the help of porphyroblasts in the contact aureole. Rocks with late E—W cleavage folds have been hornfelsized, thus clearly indicating a late age for the contact metamorphism. The occurrence of cordierite porphyroblasts with enclosed secondary cleavage also indicates at least a very late or post-kinematic origin. That there was a compression phase (roughly N—S) subsequent to the intrusion of the Maladeta granodiorite and accompanying dykes, however, is shown by the locally overthrust attitude of the southern border of this batholith with the accompanying shear zones parallel to it, the occurrence of folded dykes, and the fact that rotated and cracked cordierite porphyroblasts have occasionally been observed in thin sections.

The large Maladeta granodiorite bounding the area to the north, consists of medium-grained, light-grey rock, usually massive in outcrop and hand-specimen and very homogeneous except for a few basic clots. This homogeneous mass is traversed in all directions by numerous thin aplites and quartz veins.

The rock is completely unorientated except for a narrow rim (under 10 m) of weakly gneissose rocks (sometimes brecciated) near the contact with the metamorphosed sediments; this contact is always very sharp and locally even faulted. A protoclastic border of this kind strongly suggests an upward movement of the granodiorite along a fault zone, a view already advanced by Zwart (1963a, p. 200).

The main constituents of the rock are quartz, plagioclase (sodic andesine 30—40 % An), microcline, and biotite. Zircon, apatite, and ore are the common accessories. The local occurrence of calcium-rich minerals (mainly hornblende) near the granodiorite border may indicate calcium assimilation from the host rock.

Plagioclase is present as idiomorphic or hyp-idiomorphic crystals, mostly with oscillatory zoning. The more basic core of the crystals has generally been altered slightly to sericite. The plagioclase forms 25 to 50 % of the rock volume. The microcline is rarely idiomorphic and occurs in smaller quantities than the plagioclase (15 to 20 %). The crystals sometimes measure up to 2 cm. The microcline often partially replaces plagioclase. Quartz occurs as an interstitial mass between the plagioclase and the biotite flakes, and forms 20 to 50 % of the rock. Biotite (5 to 15 %) occurs in more or less idiomorphic crystals, and has often been altered to chlorite and ore, sometimes to muscovite. According to the classification of Niggli (1946), this rock is a granodiorite.

North of Barruera a small separate body of rather finer-grained hornblende-bearing granodiorite is found surrounded by mainly Lower Palaeozoic rocks in the north and west. Its eastern contact with Cambro-Ordovician rocks is concealed by the wide glacial Tor valley. This stock possesses roughly the same general character as the above-described Maladeta batholith. It differs mainly by its lateral crown of north-dipping

dykes, which appear to have intruded along the planes of the slaty cleavage.

The thermal metamorphism of the Maladeta batholith has produced a contact zone of moderate width, varying in the map-area between a few hundred metres and 3 km; this zone is thinnest where the granodiorite is surrounded by Cambro-Ordovician rocks. The contact zone of the Barruera stock is nowhere more than a few hundred metres. Generally, in this aureole the argillaceous rocks have been converted into hornfels and spotted slate and the calcareous rocks into calcite marble and lime-silicate rocks.

The hornfels show an unorientated texture; the original slaty cleavage has almost completely disappeared. The matrix consists of an intergrowth of quartz, biotite, and muscovite in which small porphyroblasts of andalusite and cordierite are set; both aluminium-silicates are generally more or less strongly altered to sericite and muscovite; biotite may be chloritized. The accessory minerals are tourmaline, zircon, apatite, and ore minerals. In the spotted slates, which occur mainly on the outer border of the metamorphic aureole, the original cleavage is still pronounced; the spots consisting at present of sericite aggregates were originally andalusites or cordierites. The outer limit of the contact aureole has been drawn where the hornfels or spotted slates are no longer recognizable to the naked eye.

It is interesting to note that the Silurian black slates never become true hornfels, but remain rather friable up to the vicinity of the granodiorite. The high carbon content and the extremely low percentage of free quartz may have counteracted the metamorphic recrystallization. Chistolite seems to be the only newly formed mineral.

Where pure limestones were present in the contact aureole (e.g. the Basibé limestone) they have been converted into white marble in which the calcite has been completely recrystallized into coarse crystals. This marble often contains irregular, dirty-brown dolomite bodies (with diameters of up to several metres) which obliterate all traces of the bedding and cleavage.

The so-called "*barrégiennes*" (ridged limestones), consisting of a rapid alternation of limestones and less calcareous layers (e.g. the Rueda Formation and the Mañanet Griotte), are transformed near the contact into lime-silicate rocks. The larger minerals are: diopside, idocrase, epidote-clinozoisite, and titanite. Smaller and less frequent are: albite, microcline, basic plagioclase, actinolite, wollastonite, and a red garnet.

In general, the zone in which metamorphic minerals in impure limestones or "*barrégiennes*" can still be distinguished by the naked eye is rather narrow (up to 100 m wide).

#### DYKES AND THEIR METAMORPHIC ENVIRONMENT

The dykes encountered in the mapped part of the axial zone can be divided into several groups on the

basis of macroscopical differences and their occurrence in the field. Mineralogically, the composition of the rocks varies from dioritic to granitic.

#### *Dyke swarm of the Bono-Barruera area*

The dense belt of dykes swarming out of the metamorphic Bono area consists of hundreds of porphyrite dykes with a rather constant composition (diorite porphyrite). The more isolated dykes south of the main swarm have the same mineral composition. Most of these dykes were intruded more or less along the main cleavage plane (therefore after the main phase), but locally they sometimes follow large, continuous pre-cleavage structures, e.g. on the western slope of the Ribagorzana valley. Since these dykes are unconformably overlain by the Bunter, their intrusion can be dated as late- or post-tectonic in the Hercynian orogenic phase, but before the deposition of the Triassic.

Their thickness ranges from 30 m down to 0.5 m. Thick dykes are massive, jointed, and occasionally schistose near the contact with the host rock. Their grain-size diminishes from medium fine in the centre to rather fine near the borders. The larger dykes are flanked by a thin zone (up to 1 m) of metamorphosed rocks with roughly the same mineral association as found in the contact aureole of the granodiorite. Thin dykes are generally more finely grained, more schistose, and produce no contact metamorphism. All the dykes have a sharp contact; disturbance of the host rock was rarely observed. Due to their rather high competence and their light-grey weathering, these dykes can be distinguished in the field even from a considerable distance.

The transition zone between the dyke swarm and the metamorphic complex of Bono also contains irregular, lense-shaped bodies, several hundred metres long and 20 to 40 m thick, with the same dioritic composition and porphyritic texture. The host rock between these igneous lenses is strongly metamorphosed, but the different stratigraphical units are still recognizable.

The fresh rock of these dykes and intrusive bodies is generally white, grey, greenish, or reddish, and shows light-coloured phenocrysts of feldspar in a dense matrix. Their borders show some introduced calcium in the form of calcite veinlets. For further details concerning the mineralogical composition and texture of these dykes we refer to Mey (1967-b, pp. 212–213).

#### *Metamorphic Bono area*

The metamorphic Bono area, which represents the core of the dyke swarm, was only studied very briefly: the steep relief of this area makes detailed investigation almost impossible, and such a study would involve purely petrological problems far beyond the scope of our more general investigation.

On older maps (Dalloni, 1910, 1930; Solé Sabaris, 1956, maps of Huesca and Lerida) the Bono area is shown as a granite, but in fact about 70 % of the mass consists of (quartz-)diorite porphyrites (as irregular lenses and dykes) and the remainder of metamorphic Devonian

rocks affected by fluids and heat generated by a probably deeper intrusion. This area is thought to represent the strongly metamorphosed roof of an intrusion.

Mapping of the non-metamorphic country-rock on the eastern slope of the Ribagorzana valley revealed that all litho-stratigraphical units of the Devonian could be present in the metamorphic area. However, the metamorphism and a probable metasomatism were so intense that the original character of the calcareous and argillaceous rocks is no longer recognizable (unlike the contact aureole next to the granodiorites with hornfels, spotted slates, and marbles, whose original character and stratigraphical position can still be established). The bedding can still be deduced from the arrangement of various lime-silicate rocks, and the former cleavage trend from the alignment of intrusive dykes which, as we have shown, intruded along that plane. The fact that the metamorphic complex had already sustained a cleavage and that it is covered unconformably by the Bunter, dates the metamorphism as post-main phase but pre-Triassic. Whether the metamorphism was contemporaneous with the intrusion of the diorite dykes or post-dates them is not clear in the field. A detailed petrological investigation will certainly solve this problem. The last ascending fluids of the deeper intrusion may have caused the lead-zinc mineralization (and other ores of less importance) in this metamorphic complex.

The observed minerals in the metamorphic rocks agree fairly well with the highest temperature association in the contact aureole of the Maladeta granodiorite. In the lime-silicate rocks we found mainly: green garnet (measuring up to several centimetres), idocrase, plagioclase, pyroxene (diopside?), and wollastonite; and in the hornfels: andalusite, cordierite (strongly altered into sericite), potash feldspar, and biotite.

A small outcrop of igneous origin and now strongly altered, located in the curve of the road immediately north of the village of Bono, may represent the top part of the deeper intrusion. Adjacent to the fractures now filled with lime-silicates, the rock is bleached to a cream colour as a result of the breakdown of biotite (leaving only finely-divided magnetite) and a complete alteration of the feldspar to sericite. The considerable amount of calcite (finely dispersed and in veinlets) was most probably introduced from the surrounding carbonate rocks.

Lastly, it should be noted that it is probably not a coincidence that this metamorphic complex and the core of the dyke swarm are situated on the well-expressed bend in the general trend of the main phase cleavage (Fig. 24).

#### *Dykes of the Sierra Negra Unit*

In the Sierra Negra Unit there are many fine-grained dykes with rocks ranging in colour from light-yellow to greenish but sometimes brownish due to weathered pyrite crystals. Dark-coloured lamprophyres are rare. The rather thin dykes (20 cm up to 2 m) have generally



been intruded parallel to the main cleavage plane without causing contact metamorphism in the host rock. The contacts are sharp and non-disturbed. In the field these dykes resemble aplites, but microscopical analysis showed them to be rather similar to the above-described (quartz-)diorites but with a much finer grain and no obvious phenocrysts. They are usually strongly altered.

Locally, we encountered cleaved, folded, and boudinaged dykes constituting evidence that a compression phase occurred after their intrusion, which was certainly post-main phase, probably even later than the main intrusion of the Maladeta batholith (see Mey, 1967-b, p. 184).

Although a direct contact with the granodiorite was not observed, the dykes are probably related to the same parent rock as the granodiorite.

#### *Dykes accompanying Alpine fault zones of the Sierra Negra Unit*

The Alpine fault zones in the Sierra Negra Unit with enclosed Triassic rocks are often accompanied by fine-grained dykes with a bright white to yellowish appearance. In two outcrops the same kind of dykes were found enclosed in Triassic marls (e.g. Fig. 13); their orientation is slightly oblique to the bedding. Since no obvious fault could be observed on either side of these dykes, a post-Triassic intrusion seems rather probable.

The same kind of white dykes were also observed in the Devonian to the north, where they are occasionally situated on fault lines. In the east they seem to be connected with the Espuy line.

The thickness of these dykes is usually not more than a few metres but may locally reach 60 m, as for instance 2 km westsouthwest of the Pic del Home (coord.  $42^{\circ}34\frac{1}{2}'$   $4^{\circ}25'$ ).

Mineralogically, these dykes vary between quartz-diorite aplites, (grano-)diorite porphyries, and granite porphyries (in the legend of the geological map they are called aplites and quartz-porphyry dykes). For their mineral assemblage and texture we refer to Mey (*ibid.*, p. 214).

The main petrological distinction between these dykes and the (quartz-)diorite dykes described above is the lack of dark minerals and the higher percentage of quartz. Because they are composed mainly of quartz and very acid plagioclase, these dykes are seldomly altered.

Where these dykes occur outside the contact aureole of the Maladeta granodiorite (e.g. south of the Senet thrust), they are flanked by a few metres of slightly baked Silurian slates.

#### *Scapolitization of post-Hercynian rocks*

With respect to the metamorphism of the non-Palaeozoic (probably Triassic) rocks enclosed in the Senet fault zone and its western splays, we were unable to establish conclusively whether it was caused by the thermal influence of the accompanying aplites or by weak regional metamorphism that affected only

the post-Hercynian rocks. Dynamo metamorphism can be excluded, since the rocks show hardly any internal deformation. The original sedimentary rocks, which range from pure mudstone, marl, and limestone/dolomite to arkosic sandstone, have all undergone a slight metamorphism that caused recrystallization of the calcite and created crystalloblasts of scapolite and phlogopite. The small quantities of fine-grained quartz and plagioclase (albite-oligoclase) often present, are most probably of sedimentary origin. The scapolites occur either as tiny, white spots (0.5–5 mm) or as black, idiomorphic crystals up to 2–3 cm long. A metamorphic influence on the finely stratified, purely arkosic rocks present in the middle part of the section shown in Fig. 13 is not evident. Their mineralogical composition is mainly quartz and plagioclase (albite), with tourmaline, zircon, rutile, biotite, muscovite, sericite, and ore as common accessories.

In several thin sections of the scapolitized calcareous rocks we observed veinlets of quartz and albite cutting discordantly through the bedding. These veinlets and the fact that the above-mentioned aplites accompany the fault zones with enclosed scapolitized rocks, strongly suggest a relationship between this metamorphism and the occurrence of the aplite dykes.

#### **IGNEOUS BODIES OF THE PALAEOZOIC BLOCKS OF LAS PAULES AND GOTARTA**

Of the Hercynian-folded part of the Las Paules block, the Lower Carboniferous shale belt is characterized by a great number of small and large igneous bodies, generally lens-shaped in outcrop with the longest axis trending E–W. Where the contacts with the host rock are exposed, the surrounding shales are seen to have been pushed upwards along the generally steep borders of these bodies, which is clear evidence that these masses are intrusives. Contact metamorphism is generally negligible.

When fresh, the rock has a dense, bottle-green appearance, but the weathering colour is generally yellowish-brown. The largest of these bodies (exposed over about 5000 sq. m), which is located along the road about halfway between Noales and Castejon de Sos, is quarried and crushed to serve as road pavement.

Mineralogically, the fine-grained rock varies from a rhyodacite to a trachy-andesite; the phenocrysts are mainly lath-shaped plagioclase, a little quartz and potash feldspar, pyroxene, and probably olivine. Much calcite and chlorite is found as an alteration product of pyroxene and probably olivine.

The composition and volcanic texture of this rock make it very likely that these intrusive bodies represent the sub-volcanic part (very shallow) of the Stephanian extrusive tuffs and lavas.

Similar although much smaller igneous bodies have been found in the Gotarta block, northwest of the village of Gotarta, where they are intrusive in Devonian (Fonchanina) slates. These bodies show slightly stronger weathering but roughly the same mineralogical composition and texture.

## BASIC VOLCANIC VENTS

Evidence of a fourth and completely different type of magmatic, i.e. sub-volcanic activity, is found in rather limited areas about  $1\frac{1}{2}$  km west and southwest of the village of Castanesa (coord.  $42^{\circ}30'$  .  $4^{\circ}20'$ ) in the form of four irregular discordant masses of igneous origin with a round to elongated contour and steep walls. These bodies, which vary in diameter between 20 and 100 m, pierce through Lower Carboniferous shale and slate. Although they are discordant with respect to cleavage and bedding, the weaker parts of the igneous complex are cut by a fracture cleavage (probably Alpine) running parallel to the slaty cleavage of the host rock. Contact metamorphism is weak and restricted to a distance of several centimetres to 2 m from the contact.

The intrusive mass is an agglomerate containing a great variety of rock types in which coarse, bottle-green, porous tuff-like material predominates. The

tuff has a basaltic composition and is cemented together by carbonate locally replaced by quartz. These softer, slightly cleaved rocks contain more competent blocks varying in diameter from several centimetres up to 1 m and consisting of andesite-basaltic lava, metamorphic limestone, hornfels chips, and pieces of quartzite and diorite-porphyrity dyke. These fragments of late Hercynian dykes indicate that this basic complex must have intruded later: a Stephanian age seems the most likely. For a more complete description of these intrusive masses and their related environment we refer to Mey (*ibid.*, pp. 214—215).

The shape of these igneous bodies, their field characteristics and mineralogical composition, strongly suggest that they represent small volcanic vents of a rather explosive type of extrusion and they may be related to the Stephanian volcanism so widely encountered in the Noguera zone to the south. A much more thorough petrological investigation is required to prove this, however.

## CHAPTER IV

## GEOMORPHOLOGY

## PRE-GLACIAL RELIEF

The present relief forms of the Pyrenees are the product of nearly 40 million years of active erosion and denudation under widely differing climatological conditions. In Late Eocene times, when the Mesozoic part of the Pyrenees was folded and the Palaeozoic core pushed upwards along faults and minor gliding planes, the entire chain rose above sea-level. Its steep relief was at once attacked by fast-running waters in multi-directional gullies, but in the Southern Pyrenees the main flow was directed towards the south. This trend is indicated first by the measured scour-and-fill structures in the thick fluvial piedmont deposits (Oligocene conglomerates) indicating a major stream pattern towards the south (Nagtegaal, 1966), and secondly, by the general tendency of these coarse conglomerates to grade towards finer-grained deposits with evaporite intercalations towards the south (Misch, 1934; Birot, 1937).

In Oligocene times the relief forms were still very pronounced, but towards the Miocene, gently sloping areas with a number of persisting "Inselbergs" characterized the general landscape. The climate was warm and humid, as shown by fossilized sub-tropical plants and spores and by the thick, strongly weathered soils covering the preserved Miocene planation surfaces locally.

The most widely accepted theory concerning the Miocene planation surface in the Valle de Aran — ranging there from 2000—2500 m (Rijckborst, 1967, pp. 9—17) — is that of de Sitter (1954, 1956), who hypothesizes the doming-up in Upper Miocene times of a formerly more horizontal area characterized by

pediplains surrounded by residual mountains. The most widely known denudation level in the central part of the Pyrenean chain, i.e. the one with an elevation of 1900—2300 m (Nussbaum, 1935, 1938; García Sainz, 1940 a & b; Kleinsmiede, 1960; and Zandvliet, 1960), fits very well into this picture of a dome-shaped structure. Its regional slope dips in the direction of the present rivers, and even the Miocene water divide coincides surprisingly well with that of the present main rivers (Rijckborst, 1967, p. 17).

Owing to this morphogenetic uplift with headward erosion of the rivers and subsequent strong glacial remolding, remnants of denudation levels are scarcely recognizable in our part of the axial zone. One such remnant might be the WNW—ESE-trending zone passing from the Coll de Salinas in the west, via the Collada Gelada, the Muro to the Collado de Llevata in the east, a trough with a bottom at roughly 2000 m and sloping gently up to a level of about 2400 m.

In the granodiorite area there are many relatively flat surfaces, ranging in elevation from about 2100 m to 2400 m; the largest are those in the area of the Lagos Anglios (coord.  $42^{\circ}35\frac{1}{2}'$  .  $4^{\circ}24'$ ), the high area between the Ribagorzana and Tor rivers, the area east of the Lago Negro in the Upper Tor valley, and finally the gently southward-sloping triangle east of the Balneario de Bohi. Except for the latter, these areas have in common that the surrounding higher areas or residual mountains have steep walls, therefore causing a sharp knick in the topography. Since large glacier-basins have similar topographical features, we cannot be certain whether the gently sloping areas formed part of a pre-glacial relief or are solely the product of glacial erosion.

In the southern part of our map-area, the Cretaceous zone, the Nogueras zone, and the large triangle of Carboniferous shales in the southwestern part of the axial zone, have one important thing in common: the relief forms above 1200 m elevation are very smooth with low angles of the slopes. Below 1200 m the slopes are considerable steeper, except for the areas where the incompetent Keuper is exposed over large areas. There are, moreover, only a few residual hills higher than about 1500 m. The most prominent of these higher areas are the Sierra de Santa Barbara (about 1600 m), the Fallada de Malpas (1700 m), and in the west the area south of Bonansa, where several hills reach a level of about 1700 m. Since this gently sloping region between roughly 1200 m and 1500 m is generally situated higher than the unconformity at the base of the Oligocene conglomerates, and in other places the plane of unconformity is truncated by this plane level, it is clearly a younger erosional feature; it pre-dates the glaciations, however, since the Tor and Ribagorzana glaciers have remolded their valleys far below this planation surface. Whether this region links up with the higher Miocene planation surface of the Central Pyrenees is impossible to prove from such a relatively small area as ours, however.

#### GLACIATIONS

After the hot, moist Miocene and the dry Pliocene (Taillefer, 1951), cold dry climates prevailed in the Pyrenees. During the Quaternary period large glaciers were formed and moved downwards, guided by the larger valleys (Fig. 49). The influence of these glaciers is apparent from the frequent occurrence in the higher part of the mountain chain of U-shaped valleys, hanging tributary valleys, cirques, rock steps, glacial striae and grooves, moraines, etc. Previous authors have distinguished two or three glaciations, the first of which may have been of the Scandinavian type, leaving few traces. According to García Sainz (1940b, p. 368), the latest glaciation (Würm), which can be compared to the Alpine type, entirely modified or intensified the erosional forms of older glaciations. However, the physiographic features of the area covered by our geological map have never been studied in sufficient detail to permit the deduction of different glaciation periods from the few remaining morainal deposits. We must therefore be content for the time being with the description of the most conspicuous glacial features, disregarding their age.

In the area of our geological map only the valleys of the Ribagorzana and the Tor rivers, together with their tributaries in the granite and adjacent areas, were once occupied by moving glaciers. In the Upper Baliera and Upper Mañanet valleys the erosion forms strongly suggest the presence of much snow and ice, but there is no evidence of moving glaciers, such as U-shaped valleys and morainal deposits in the lower part of the valley. This can be explained by comparing these snow-accumulation areas with those of the Ribagorzana and Tor glaciers, the former being much

smaller, situated at a lower altitude, and more to the south. The enormous amount of debris present in the upper part of these valleys also argues against moving ice masses.

The higher the glacial erosion forms occur, the fresher they are. They have hardly been modified in the granite massif and its aureole of hornfels and marble; glacial striae, *roches moutonnées*, rock steps, and cirques with glacial lakes, are preserved predominantly in the higher part of the Maladeta area and its vicinity. The same holds for the pure U-shapes of the valleys. As nice examples of U-shaped valleys we may mention:

1. The Ribagorzana valley in the granite area.
2. The Salenca valley.
3. The Llausét valley, west and northwest of Lago de Llausét.
4. The two valleys, one on either side, of the Ribagorzana, lying more or less along the Silurian-granodiorite contact zone.
5. The Tor valley north of Barruera.
6. The upper part of the valley of the Arroyo de Bohi.
7. The San Nicolás valley.
8. The larger tributary valleys of the Tor and San Nicolás valleys.

The valleys mentioned under 2, 3, 4, 6, and 8 are hanging valleys.

Beautiful examples of large glacier-basins or *Karplatten* (Nussbaum) are provided by the flat area surrounding the Lagos Anglios and the area east of the Lago Negro, the latter in the upper Tor valley. These basins are broad, gently-sloping excavations with a flat bottom and steep walls. Via only one distinct rock step, these basins pass into the beautifully U-shaped valleys of the Barranco de Salenca and the Rio Tor, respectively. Glacial deposits are found in both the Ribagorzana and Tor valleys as far south as the northern border of the Nogueras zone, where they still occur more than 50 metres above the present valley floor. Although no distinct terminal moraine has been found in the Ribagorzana valley, the narrow V-shape of the valley at the point where the river crosses the steeply overturned wall of the Bunter, as well as the total lack of any morainal deposits downstream from this point, make it highly probable that the Ribagorzana glacier never entered the Nogueras zone. Nussbaum and García Sainz are of the same opinion. In the Tor valley the enormous number of granite boulders found on the eastern slope just north of the steep wall of Bunter up to a level of nearly 1200 m (almost 300 m above the present valley floor), may well represent the terminal moraine of the Tor glacier; no morainal deposits have been found in the Nogueras zone to the south.

Granite boulders, some of them very large, are frequently found on the valley slopes of the main glacial valleys and their most important tributaries (see also page 256). The larger concentrations of boulders are indicated on our geological map. The thickness of the original glacier can be deduced from the highest levels of these lateral moraines; a number

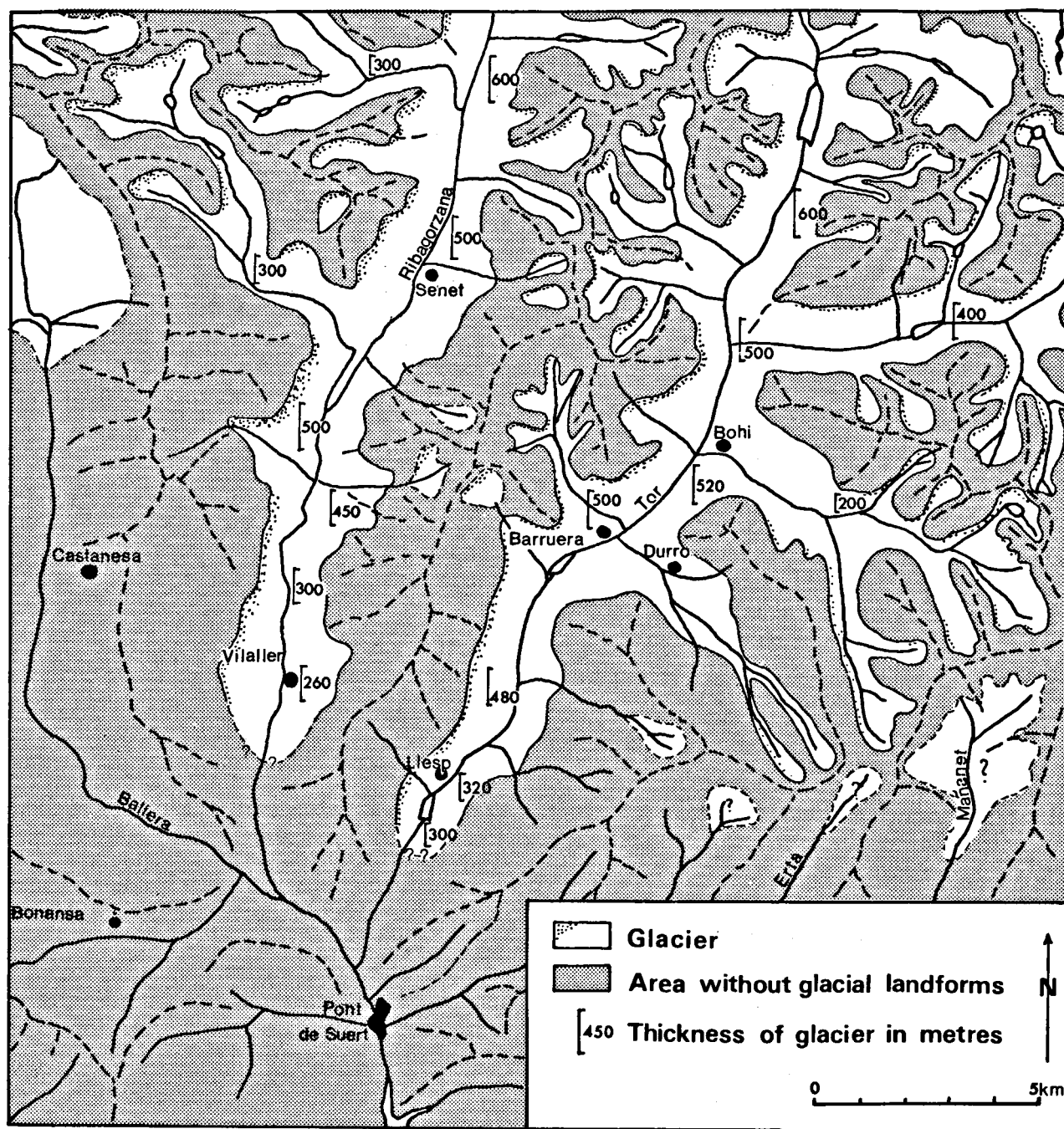


Fig. 49. Glacier extension in the Upper Ribagorzana and Tor valleys during the latest gliation.

of these measurements are shown as approximate values in Fig. 49.

#### POST-GLACIAL EROSION

After the ice melted, large and rapidly-flowing rivers re-incised their beds and removed the thin cover of moraine material. Vast cones of debris originated as soon as the pressure of the melting ice upon the walls of the glacial trough had vanished and the rock

became exposed to atmospheric disintegration. Lakes originally formed by the melting ice in the impermeable flat valley bottoms (near rock steps overdeepened by the moving glacier), were filled up with fluvio-glacial material; and in places where brooks and rivers flowed more quietly, finer material was deposited. Thresholds downstream from these filled-up lakes were cut to a certain depth that determined the river-level upstream. Good examples of this phenomenon are the threshold in the Mañanet Griotte

south of the village of Forcat in the Ribagorzana valley, and the now partly submerged threshold near Cardet in the Tor valley.

One of the most conspicuous features of post-glacial erosion is the formation of deep gorges carved in the irregularities of the glacial valley-profile by the headward erosion. These gorges occur mainly in:

1. Rock steps and thresholds in a valley, e.g. the rock step of Devonian limestone between Senet and Bono, where the gorge is 700 m long.
2. Hanging valleys, where they discharge into the deepened valley.

In the Maladeta granodiorite and its aureole of hornfels, joint systems promote the development of vast fields of dry debris. This debris resulted from mechanical denudation during freezing and thawing. Owing to the same process, the arêtes above the *Karplatten* and cirques are often sharply dentated.

#### INFLUENCE OF THE STRUCTURE ON THE PRESENT RELIEF

It is surprising to see on our geological map how independent the course of all the main rivers has remained with respect to the most prominent structural features, such as the general cleavage trend, faults, and hard limestone ridges. This is of course due to the longitudinal arching-up of post-Miocene planation surfaces in the post-Alpine morphogenetic phase, as a result of which the rivers run off the southern slope. On the cleavage trend map (Fig. 24) it is remarkable to note that the Ribagorzana River and the southern part of the Tor River more or less follow the sharp bend of the constructed cleavage traces. These bends probably correspond to fundamental fractures at greater depth, but wherever structural and stratigraphic features cross the rivers, there is no evidence whatever of a fault along the main valleys. The writer is, however, under the impression that a concentration of joints along the course of the rivers probably forms some kind of straight lineament, as described by Zandvliet for large parts of the Pallaresa River (1960, pp. 31 and 94—99).

The independence of the courses of the main rivers with respect to the structural features on the map does not hold for the smaller rivers and brooks, which often follow the general structural trend, fault-lines, and shear-zones, at least over a certain distance. The geological map shows many striking examples of this.

Not only the small rivers but the whole relief is adapted to the lithological differences of the area. Each kind of rock shows its own characteristic form of weathering. The more resistant rocks are the granodiorite, the hornfels and limestone-marble, and the quartzite, which form the highest parts of the mountain chain or, locally in a lower area, the steepest relief. The less resistant rocks are the non-metamorphic shales and slates (Silurian, Aneto, Fonchanina, and Carboniferous shale/slates) and, outside the axial zone, the gypsum and marls of the Upper Triassic and the thick marl intercalations between the massive Mesozoic limestones and dolomites. Consequently, the non-metamorphic pelitic sediments in the axial zone are characterized by smooth denudation forms and depressions, particularly where their outcrops have a considerable extension as in the broad E—W-directed band of Aneto slates in the northern part of the Baliera facies area and the triangle of Lower Carboniferous shale surrounding the village of Vilaller.

Surface creep is apparent everywhere in the schistose rock, but especially when slates form the underground. When debris is also present on the lower part of the slopes, this surface creep can result in landslides, sometimes with catastrophic consequences. Potential sites of landslides are marked by long deep clefts on the right bank of the Ribagorzana River, south of Vilaller. There, and to a much smaller extent also on the other side of the river, landslides may bring down bedrock and parts of the morainic and slope material. In August 1962, after several hours of heavy continuous rain, the Ribagorzana River washed away the foot of these landslides, resulting in a sudden mass movement several hundred metres across. The main road slid down about 50 m and the water canal of the ENHER hydroelectrical scheme was badly damaged.

#### CHAPTER V

#### ECONOMIC GEOLOGY

##### LEAD AND ZINC

The *El Cierco* lead-zinc mine southeast of the village of Bono, which is exploited by the M.I.P.S.A., is the only ore mine in the map-area still being worked. At present it employs about 100 people. The abandoned smaller mines (mainly iron and copper) indicated on the geological map were, except for the copper mine 400 m SSE of the hamlet Artiga ( $42^{\circ}30\frac{1}{2}'$ ,  $4^{\circ}25'$ ), small workings of no economic importance.

The mined galenite (PbS) and sphalerite (ZnS) occur

in a number of roughly E—W-striking quartz-calcite veins, discordant in respect of the bedding of the host rock and varying in thickness from several centimetres to about 1 m; the environment consists of metamorphic Devonian limestone cut by numerous diorite porphyrite dykes. A hydrothermal origin of the ore seems very likely, because none of the Devonian sedimentary rocks outside the metamorphic Bono area contain ore in economically exploitable quantities. It could not be determined whether the mineralizing fluids ascended along the cleavage plane, faults, or

irregular cracks in the host rock. Since these ore veins are unconformably overlain by the Bunter rocks, the mineralization must have occurred in Hercynian times.

This E—W-striking mineralization is cut by a rather straight, nearly vertical barite vein with a NNW trend; since this trend corresponds with the strike of the Alpine Bono thrust in this area, the two phenomena may be related, but this has not yet been proven.

The lead-zinc ore, which bears small amounts of silver (1 kg silver to 1 ton pure lead) is accompanied by such constituents as pyrite, chalcopyrite, and stibnite, as well as supergenic covellite, malachite, and azurite. The matrix is always calcite and quartz. The tenor of the crude ore is 6.8% lead-zinc and silver compounds.

The ore is mined in a number of horizontal galleries (about 20) situated between 300 and 600 m above the valley floor, and is carried by cable to a crushing and washing plant in the lower part of the Ribagorzana valley, about 1 km east of the village of Estét. There, the crude ore is concentrated for easier bulk transport, and the different sulphides are separated. The result is a concentrated lead ore with 66 % PbS and a zinc ore with 54 % ZnS; the actual annual production of these concentrates is about 2000 and 1400 tons, respectively. Transportation of these concentrates is done by trucks, most of the very valuable zinc being exported to France (St. Giron) via the Viella tunnel.

## COAL

Exploitation of the coal seams occurring in the narrow stretch of sedimentary Stephanian (Malpas Formation) between the villages of Erill Castell and Benés was started some centuries ago. In those early times the coal was dug in open pits and from shallow galleries. After a short decline in the first years of this century, coal mining was revived at Malpas in 1931 and even became important after 1948, when the ENHER hydraulic works started to use the coal for a new cement plant at Cheralló, forming in effect a closed economic circuit. The coal was transported directly by cable to Cheralló (distance 7 km) up to about 1958, when truck transport took over.

The average annual production of the mine over the last 15 years was about 25,000 tons, but it has recently (1965) dropped to only 22—23,000 tons. The caloric value of this coal is in excess of 7000.

Two main seams are now worked via one vertical shaft with horizontal galleries on several levels; these seams are 1.50 m and 3.50 m thick and both have a subvertical attitude. In spite of faulting and lengthy galleries, new extensions are planned. The total personnel numbers 110 at present, most of them living in the village of Malpas and few in Peranera.

## LIMESTONE

Apart from a few minor quarries providing ornamental stone and road pavement, important limestone exploitation takes place only in the large quarry located just south of Cheralló (coord.  $42^{\circ}22' . 4^{\circ}33'$ ). The

Urgo-Aptian limestone (Prada Formation) is there quarried and transported by cable to the ENHER cement factory in Cheralló. The silica for admixture is derived from a small quarry in Bunter sandstones situated two hundred metres north of the plant. The factory uses coal from the Malpas mine as fuel.

## THERMAL SOURCES

The thermal and mineral-bearing waters, which are used mainly for medical purposes, are also of economic importance. The hot springs of Caldas de Bohi (coord.  $42^{\circ}33\frac{1}{2}' . 4^{\circ}31\frac{1}{2}'$ ) in the Upper Tor valley, which pour out of joints in the granodiorite, are well known and very old. It is quite possible that the very straight upper course of the Tor River in the granodiorite is due to a major fissure in the valley bottom, along which deeper-seated waters could ascend. Since the end of last century, several baths utilizing the springs have been open to the public, and more recent times have seen the construction of two large hotels with swimming pools. These and other buildings have transformed the area into a luxurious centre of tourism called the Balneario de Bohi.

The springs can be divided into three major groups:

1. The sodium-bearing sulfurous sources with water temperatures ranging from 23 to 55° C. Number of springs: 8.
2. The calcium-bearing sulfurous sources, with temperatures varying from 33 to 38° C. Number of springs: 2.
3. The chlorine- and sodium-bearing waters, with a temperature of about 5° C, provided by one spring.

No exact data concerning the total discharge of thermal and mineral-bearing waters are available, but it certainly amounts to a few cubic metres an hour.

## HYDROPOWER

The coming of the ENHER (Empresa Nacional Hidroeléctrica Ribagorzana) to the valleys of the Ribagorzana and Tor in 1948, followed by the construction of many tunnels, canals, power stations, and dams, must not only be seen as a new link added to the energy supply of Spain and Europe, but also as the main reason for the revival and growth of several villages and towns, especially Pont de Suert, Vilaller, and Cheralló, and for the explosive development of tourism in these areas. In the first years after its arrival, the ENHER built many new roads and improved the few existing communication lines, which were in very bad condition. The building of a very modern church in Pont de Suert and of several well-equipped refuge huts in the granite area, as well as the construction of a few minor roads (e.g. to the Aiguas Tortes National Park), certainly also contributed much to the opening up of this rugged but beautiful area.

The manpower for the many construction projects was not recruited, as might have been expected, from the poverty-stricken surrounding mountain villages



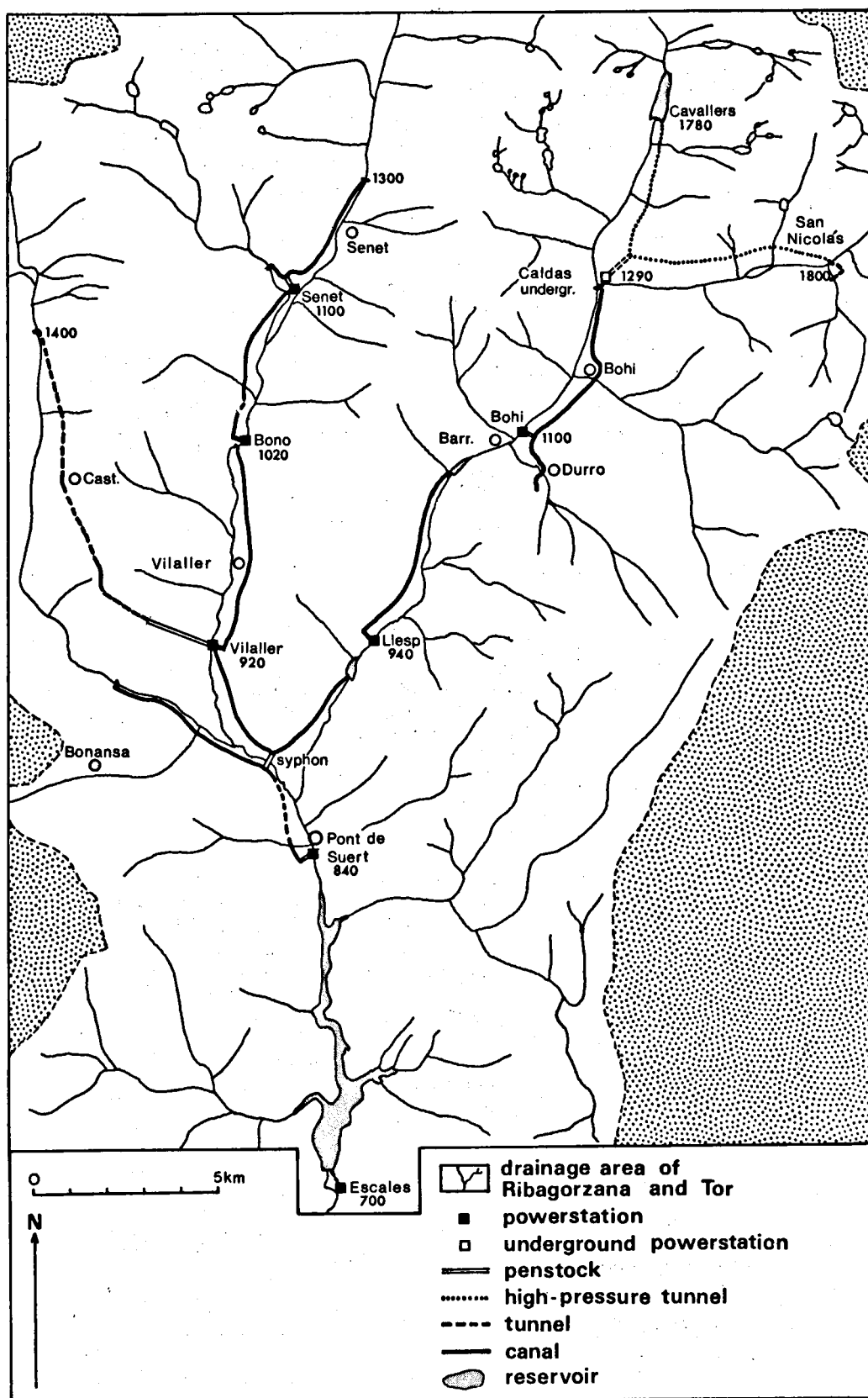


Fig. 50. Sketch-map of the hydroelectric system in the Upper Ribagorzana and Tor valleys.

already doomed to die out long before the coming of the ENHER, but mainly from still poorer areas of Spain such as Andalusia and Galicia.

The choice of the Upper Ribagorzana and Tor as a power-generating area was determined by many factors such as numerous affluents, many falls, adequate snowfall, differences in level, and the many small glacial lakes. The exploitation of these natural resources is determined by water catchment, principally on the Tor side of the basin.

The regulation of the streams and lakes was carried out by constructing a large, high buttress-dam at Cavallers (the headwater lake of the Rio Tor) and an underground power station at Caldas with a very high head (483 m), using the water from the Cavallers reservoir and that of the San Nicolás valley (Fig. 50); the water is conducted by high-pressure tunnels.

The second power station in the Tor valley is located opposite the village of Barruera (Central de Bohi) and is fed by water from the Tor River and a small canal carrying additional water from the Rio de las Foixes; total head: about 160 m.

The next step in the system is a small reservoir south of Barruera with a maximum water level at an elevation of about 1090 m. Water from this small lake is conducted into a covered canal to the village of Llesp, where it feeds the third power station in the Tor with a head of about 100 m. The water is there collected once again in a small reservoir south of Llesp (level 935 m) and brought by a partly closed canal to the Ribagorzana valley, where, together with two other branches, it feeds the power station of Pont de Suert (head about 70 m).

This Tor scheme is rather complete. The only improvement still possible is the addition of the combined granite lakes to feed a power station north or south of the Cavallers reservoir.

As Fig. 50 shows, in the Ribagorzana valley there are three more power stations north of the one mentioned near Pont de Suert. The northernmost of these, the Central de Senet, is fed by water from the Ribagorzana and from the Rio Llausét, the interception of the two rivers being affected by small weirs at a level of about

1300 m. The power station itself lies at about 1100 m, therefore working with a head of almost 200 m. The water from this plant is conducted by a half open, half closed canal to a point just south of the village of Forcat, where it feeds the power station of Bono, with a head of about 70 m.

A small weir in the Ribagorzana River immediately south of the Bono plant assures a constant inflow in the next open canal, feeding part of the turbines of the Vilaller power station (head of about 65 m). The other part is driven by water from the Baliera River via a long tunnel, providing a useful head of slightly more than 560 m. The outflowing water is conducted by a canal towards the south, where it links up with the canal from the Llesp reservoir. The crossing of the Ribagorzana River is effected by a huge syphon.

The hydraulic scheme of the Upper Ribagorzana is less complete than that of the Upper Tor, since the entire fall north of the weir of Senet is unused. It is now unlikely that further exploitation of either upper zone will take place, however, because of the expansion of Spain's nuclear power programme.

The huge reservoir behind the high Escales gravity dam (110 m) is the northernmost of a number of stream regulators. The snow melt in the late spring and the sometimes very heavy showers in the summer and fall can be stored here and distributed over a longer time span. With the power station at Escales, the total power produced in the upper section of the Ribagorzana-Tor power scheme is nearly 420,000 kwh, most of which is conducted directly to Barcelona.

In the Ribagorzana valley south of the Escales reservoir there are other power stations in Puente Montañana, Canelles, and Santa Ana, the latter two situated at the foot of high dams (cupola and arched gravity dams respectively). The Santa Ana reservoir controls the distribution and utilization of water for electric power as well as irrigation.

The Ribagorzana scheme, with a total mean annual HE production of 1,241,766,000 kw, is also integrated with that of the Lower Ebro, the whole forming a continuous year-round electricity supply.

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