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THE GEOLOGY OF THE UPPER RIBAGORZANA AND BALIERA VALLEYS, CENTRAL PYRENEES, SPAIN

BY

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ABSTRACT

In the mapped area there is a well-exposed low-grade metamorphic marine sequence from Ordovician to Lower Carboniferous, unconformably overlain by Permo-Triassic continental deposits. Determinable fossils are rare. The Ordovician consists of a quartzite/shale sequence with one marly limestone intercalation in the upper part. The Silurian is developed as a classical graptolite-bearing black shale facies with an Orthoceras limestone near the top. The Devonian rock-sequence in the north differs from that in the south. The northern or Sierra Negra facies area consists of a thin (120—250 m) alternation of mainly limestone and slate; the southern, Baliera facies area, is thicker (340—780 m), shows more individual limestone-slate units, and is moreover characterized by a conspicuous quartzite member (0—50 m) in the middle part of the Devonian. The Devonian sequence in both areas is subdivided into four or five separate formations which have been mapped individually. The Carboniferous consists in both areas of micaceous slates with a low sand content.

The continental Permo-Triassic is developed in the Germanic facies of red mud- and silstones, sandstones, and conglomerates at the base, followed by a non-fossiliferous limestone/dolomite (Muschelkalk) and gypsum-bearing, vividly coloured marls of the Keuper. The major structural movements, which probably began already in early Carboniferous times, increased in strength towards the Westphalian B. Several phases of deformation have been recognized. The first deformation produced concentric, open to tight asymmetric folds without cleavage development. Their axial planes have a general E-W to ESE-WNW trend in the north (Sierra Negra Unit), a constant NE trend in the centre (Baliera Unit), and an E-W and NW-SE trend in the south (Ribagorzana Unit). The second deformation, representing the main phase, was caused by a N-S compression and is characterized by tight to isoclinal folds with a steep northward-dipping axial plane cleavage in the north, the dip becoming more moderate in the centre and south. Fold axes and δ-lineations show a girdle distribution. A third deformational phase bent the entire structure around a NNE-trending axis coinciding with the bed of the Ribagorzana River. The Maladeta granodiorite and accompanying dykes intruded parallel to the general cleavage trend and caused a metamorphic aureole of moderate width. Near its southern border, gravity folds were formed locally. The economic occurrence of galena, exploited near the village of Bono, is related to this igneous activity. A fourth deformational phase was produced by a renewed N-S compression, causing local folding of the first or main-phase cleavage, also showing a weak secondary axial plane cleavage (fracture or crenulation cleavage), and local thrust movements along the earlier cleavage plane. This deformation might be a late Hercynian or an Alpine phase.

The Alpine orogeny initiated or rejuvenated important northward-dipping overthrusts and minor thrust movements along the main phase cleavage in the Palaeozoic, which at the same time caused asymmetric folding of the Permo-Triassic strata above the unconformity.

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

RÉSUMÉ

Dans la région étudiée, une succession de terrains d'origine marine, allant de l'Ordovicien au Carbonifère inférieur, peu métamorphique et affleurant largement, est surmontée en discordance par le Permo-Trias continental. Dans toute la série, les fossiles déterminables sont rares.

Les quelques affleurements de l'Ordovicien montrent, de bas en haut: une alternance de quartzites et de schistes, un niveau mince de calcaires marneux suivi de quelques schistes sombres, et de nouveau des quartzites. Le Silurien, dans son faciès classique pour les Pyrénées, est formé de schistes noirs graphiteux à graptolites, avec vers le sommet un niveau mince de calcaire à Orthoceras. Le Dévonien se présente différemment dans le nord et le sud de la région étudiée: au nord (Facies de Sierra Negra), il est relativement mince (120 à 250 m) et est constitué de bas en haut par une alternance de calcaires marneux et de schistes gréseux, une barre de calcaire massif, des schistes noirs et finalement des calcaires noduleux (calcaire griotte); au sud (Facies de Baliera), il est plus épais (340 à 780 m) et sa succession est la suivante: des schistes sombres, un niveau de calcaire impur, une alternance de calcaires marneux et de schistes très gréseux, un niveau très compétent de calcaires noduleux, de dolomies, de quartzites (0 à 50 m) et des calcaires bitumineux et spathiques, suivis par des schistes sombres, et enfin au sommet les griottes du Dévonien supérieur. On a donné à chaque unité lithologique cartographiable des deux régions Nord et Sud, une dénomination géographique. Le Carbonifère est formé dans les deux régions de schistes micacés, localement gréseux.

Le Permo-Trias a un faciès continental de type Germanique: on y trouve à la base des argilites rouges, des grès et des conglomérats (Buntsandstein), suivis de calcaires dolomitiques azoïques (Muschelkalk) et enfin des marnes de couleurs vives, contenant du gypse (Keuper).

Les premiers mouvements tectoniques commencent probablement déjà au début du Carbonifère et atteignent une amplitude maximale au Westphalien-B. On peut distinguer plusieurs phases de déformation:

Dans une première phase se forment des plis concentriques, plus ou moins ouverts ou fermés, légèrement asymétriques, sans clivage. Les plans axiaux de ces plis sont orientés, dans la région Nord (Unité de Sierra Negra) E-O à ESE-ONO, au centre (Unité de Baliera) NE-SO, et dans la région Sud (Unité de Ribagorzana) NO-SE et E-O.
Une deuxième phase, phase paroxymale due à une compression N-S, entraîne la formation de plis isoclinaux, dont le plan axial a un pendage de direction nord, fort dans la partie nord (50 à 80°) et plus faible dans le centre et le sud de la région étudiée (30 à 50°). Les axes des plis et les δ-linéations (intersection entre stratification et clivage) se répartissent, en projection, sur un grand cercle du diagramme de Schmidt.
Toutes ces structures ont été ensuite reprises dans une 3ème phase, dans un mouvement à grand rayon de courbure, dont l'axe, orienté NNE-SSO, coïncide avec le lit du Ribagorzana.
Le granodiorite de Maladeta et les filons l'accompagnant se sont mis en place en suivant les lignes directrices de la structure; une auréole métamorphique de quelques centaines de mètres d'épaisseur s'est formée au contact du batholite. Près de la bordure sud du batholite, des plis se sont formés par gravité. Près du village Bono, on exploite un gisement de galène, lié à cette activité magmatique.
Une quatrième phase résultant d'une compression N-S plissote localement le clivage dû à la phase principale, et fait apparaître un deuxième système de clivage axial (type clivage de fracture); de plus elle provoque localement des mouvements de chevauchement parallèle au plan de clivage principal. Ce 4ème stade de déformation peut être dû, soit à une phase tardi-hercynienne, soit à une phase alpine.
L'orogénèse alpine, a créé ou accentué dans le socle paléozoïque d'importants chevauchements et des petites failles parallèles à la direction du clivage principal hercynien, entraînant dans le Permo-Trias, au-dessus de la discordance, des plis asymétriques.
L'érosion post-Miocène, succédant les mouvements épeirogéniques alpins, a produit plusieurs plateaux d'aplatissement dont quelques restes douteux sont préservés dans notre région. Des formes glaciaires se sont développées pendant le Pleistocène et l'érosion fluviale subséquente a modifié la morphologie glaciaire.

RESUMEN

En el área mapeada hay una sucesión marina, bien expuesta, de bajo grado de metamorfismo desde el Ordovícico al Carbonífero Inferior, la cual está cubierta discordantemente por los depósitos continentales del Permo-Triásico.
Son raros los fósiles determinables. El Ordovícico consiste de una sucesión de lutitas y cuarcitas con una intercalación de caliza margosa en la parte superior. El Silúrico está desarrollado en sus clásicas facies de lutitas negras con graptolites, y una caliza con Orthoceras cerca de su techo. La sucesión de rocas devónicas en el norte es diferente de la del sur. Al norte (Facies de Sierra Negra), consiste principalmente en una delgada sucesión (120—250 m.) de calizas y pizarras; al sur (Facies de Baliera), es más potente (340—780 m.) y muestra en forma más individual las unidades caliza-pizarra y está por otra parte caracterizada por un conspícuo miembro de cuarcita (0—50 m.) en la parte media del Devónico. La sucesión Devónica, en ambas áreas está subdividida en cuatro o cinco formaciones, las que han sido mapeadas individualmente. El Carbonífero consiste en ambas áreas de pizarras micáceas con bajo contenido de arena.
El Permo-Triásico continental está desarrollado en las facies Germánicas de fangolitas, limolitas, areniscas y conglomerados rojos en la base, seguidas por dolomías y calizas (Muschelkalk) no fosilíferas y margas de vívidos colores con yeso del Keuper.
Los movimientos tectónicos principales, los que probablemente han comenzado ya en el tiempo Carbonífero, aumentan su potencia hacia el Westfaliense-B. Han sido reconocidas varias fases de deformación. La primera de ellas produjo pliegues concéntricos, ligeramente asimétricos, variando de amplios a estrechos sin desarrollo de crucero. Sus planos axiales están orientados por lo general E-W hasta ESE-WNW al norte (Unidad de Sierra Negra) y al E-W y NW-SE en el sur (Unidad de Ribagorzana). La segunda deformación, la fase principal es producida por una compresión N-S y está caracterizada en el norte por estrechos pliegues isoclinales con un crucero de plano axial buzando fuertemente hacia el norte, mientras que en la parte central y en la parte sur es más moderado (30—50°). Los ejes de plegamientos y las lineaciones muestran una distribución en anillo. La estructura entera se curva segun un eje de direccion NNE, coincidente con el Rio Ribagorzana, durante una tercera fase de deformación.
La granodiorita Maladeta y sus diques acompañantes se han intruído paralelamente a la dirección general de crucero y han producido una aureola metamórfica de moderada amplitud. Se han formado pliegues de gravedad localmente en el borde sur. La presencia de galena, en forma económica, está relacionada con la actividad ígnea y es explotada en las cercanías del pueblo Bono.
Una cuarta fase de deformación fue producida por una nueva compresión NS. Esta se evidencia por un plegamiento local del crucero de la fase principal, con un crucero de plano axial secundario y débil (tipo crucero de fractura) y cabalgamientos locales a lo largo de estos tempranos planos de crucero. Esta deformación podría ser una fase Hercínica tardía o Alpina.
La orogenía Alpina ha iniciado o rejuvenecido, en las rocas paleozoicas, importantes corrimientos inclinados hacia el norte y menores cabalgamientos a lo largo del crucero de la fase principal, la cual produjo al mismo tiempo los pliegues asimétricos de las estructuras Permo-Triásicas por encima de la discordancia.
Después del levantamiento Alpino la erosión post-Miocena produjo niveles de aplanación de los cuales remanentes dudosos están preservados en nuestra área.
Las formas glaciales del Pleistoceno fueron modificados por la erosión fluvial subsecuente.

CONTENTS

INTRODUCTION

This report describes the geology of an area of mainly Upper Palaeozoic rocks on the southern slope of the Pyrenean mountains in northern Spain. The study was performed as part of a programme of geological mapping being carried out by students of the Geological Institute of Leiden under the direction of Professor Dr. L. U. de Sitter of the Department of Structural Geology and Dr. H. J. Zwart. The main fieldwork was carried out during the summers of 1959—1961, with two additional months during the summer of 1963.

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) and Bisaurri (213), enlarged to a scale of 1 : 25,000 and for some local cases even to 1 : 10,000. Aerial photographs were available, but because of their rather poor quality and the complexity of the area they could only be used to correct the topographical maps.

The area, most of which was mapped by the author (about 250 square kilometres), is situated immediately south of the Maladeta massif, which culminates in the highest mountain of the Pyrenees: the Pico Aneto (3404 m). The main river, the Rio Noguera Ribagorzana, forms the boundary between the provinces of Huesca to the west and Lérida to the east. Owing to the great range in altitude (900—3000 metres) and the rugged relief, especially of the northern part, and the watershed between the main valleys, the area is well exposed, in the higher parts sometimes completely.

The area is thinly populated and its villages are small. The population supports itself by breeding horses and cattle (cows, sheep, and goats) as well as by a little farming, in the lower part of the main valleys (grain and potatoes). The town of Pont de Suert, situated in the Ribagorzana valley just beyond the southeastern limit of the geological map, is the centre of the district, with prosperous shopkeepers and a weekly market. It is also the local headquarters of the E.N.H.E.R., a company which produces hydro-electrical energy in the Ribagorzana valley (including the Baliera and Tor valleys). Lead mining is also of some importance in the Bono area. Other smaller mines (mainly iron and copper) dispersed throughout the entire area, were all abandoned long ago.

The climate in the northern part of this area is strongly influenced by the Atlantic Ocean, being rather cool and extremely humid; most of the time the region is blanketed by clouds. A detailed report on the weather near the watershed with the Valle de Aran (Restanca, 2000 m) over a period of 10 years has recently been published in a thesis by Rijckborst (1967), from which the following data are taken:

Average winter temperature (November—April)	—2.3° C.
Average summer temperature (May—October)	+8.2° C.
Annual precipitation (rain and snow)	1800—2500 mm
Average relative humidity	60%
Clouds:	no clouds: 30% of the year
	total cloud cover: 40% of the year.

As late as July or August, the highest parts of the area carry remnants of the winter snow cover. For further information, see Rijckborst (1967).

The areas with grass vegetation have a fresh green appearance in the summer. The lower granite and metamorphic area carries forests of beaches and pines. The climate of the southern part of the area is of a more Mediterranean type: much drier, rather warm, with a yellowish-green tone dominant in the vegetation, which consists mainly of grass, broom, and a few local pine woods. According to *Geografia de Catalunya* (1962), the annual precipitation (nearly all of it rain) varies between 700 and 1600 mm (the maxima of 1200—1600 mm are restricted to the high ridges); most of the rain falls in the months of April, May,

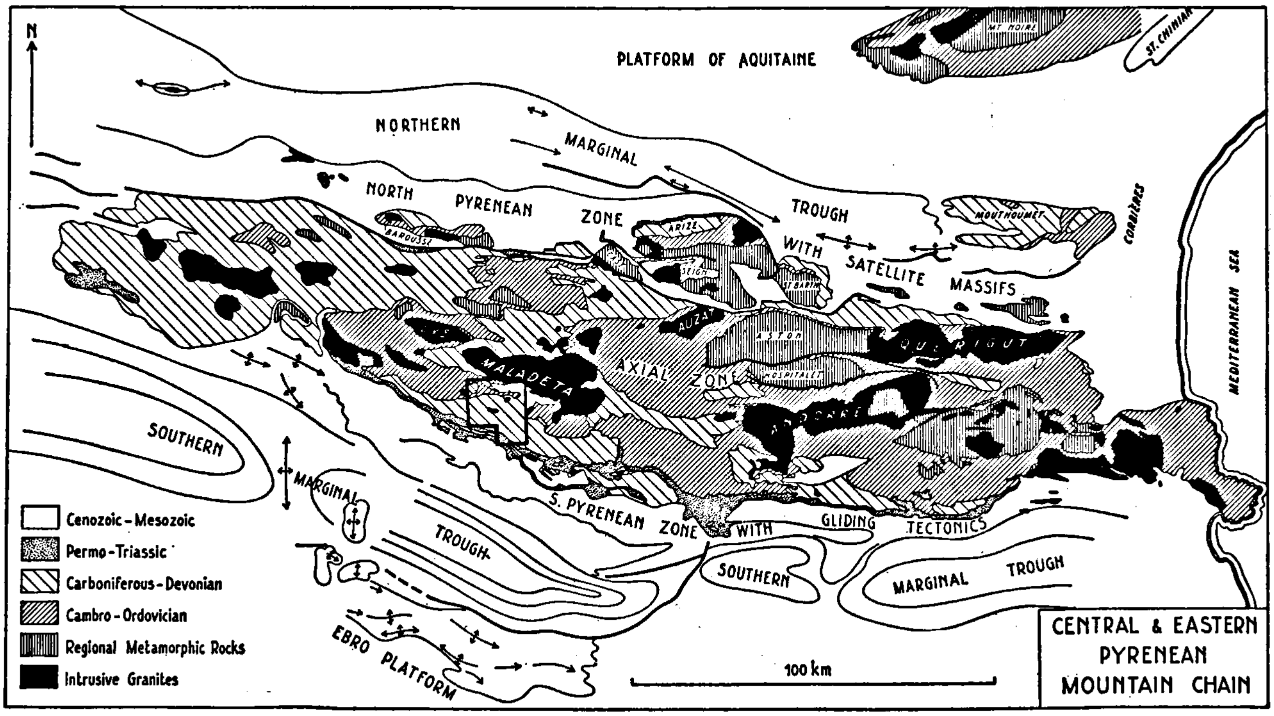

Fig. 1. Geological sketch-map of the central and eastern Pyrenees (after de Sitter, 1964), with location of the mapped area.

and June. The average temperature in the summer is about +16° C and in the winter about +7° C. Total cloudiness occurs less frequently than in the northern area.
Of the original fauna, the chamois still occurs in the highest and most isolated regions. The mountain bear and wolf, which according to many stories must have occurred in former times, seem to have vanished completely.

ACKNOWLEDGMENTS

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GEOLOGICAL SETTING

The Pyrenees proper, as a typical intercontinental orogene, constitute essentially a Hercynian structure moderately reactivated by Alpine folding, whose present form was determined by a morphogenetic uplift in Late Miocene time. The comparatively narrow Hercynian basement is fairly well exposed, sufficiently accessible, and shows only local effects of regional metamorphism.
The longitudinal arrangement of the structural zones lead to the following subdivision (de Sitter, 1956 a & b, 1964 and Fig. 1):

1. The Axial Zone, which forms the central belt, is only 30—50 km wide and consists mainly of strongly-compressed Palaeozoic rock. Some indications of Alpine influence are present.

2. A northern and southern External Zone, each separated from the axial zone by a fundamental fault. These belts, measuring 10 to 25 km in cross-section in the north and 4 to 10 km in the south, are characterized by folded Mesozoic rock surrounding domes of Palaeozoic masses, the so-called "satellite massifs". The southern external zone is called the Nogueras Zone.

3. A northern and southern sub-Pyrenean Zone, both of which are marginal with respect to the Hercynian core. These zones represent true Mesozoic basins and not post-tectonic marginal molasse basins. Their folding took place in Late Eocene time (Pyrenean folding phase) along the contact with the external zones, and died out quickly toward the continental blocks north and south of the orogene. The outer borders underwent a moderate Miocene folding, thereby forming a good illustration of the general centrifugal shift of deformational activity in the Pyrenees since the subsidence of the Devonian basin. Thus, the Pyrenees show a more or less symmetrical structure with a Hercynian core and a Mesozoic mantle passing into Mesozoic marginal basins.

CHAPTER I

STRATIGRAPHY

INTRODUCTION

The area mapped by the author is situated in the southern part of the Axial Zone, limited in the south by the northern border of the Nogueras Zone. It consists of a rather complete sequence of Palaeozoic rocks which have undergone intensive deformation (tight to isoclinal folding, thrusting, cleavage, and low-grade metamorphism). Fossils are rare in this area and strongly deformed. The stratigraphic division and mapping were based mainly on one well-characterized marker (Silurian graptolite-bearing black shale), relative age differences, descriptions of dated strata in the literature, and the author's observations of dated strata in and outside the mapped area. Consequently, the formation boundaries recorded on the map and discussed in this chapter do not coincide exactly with the ideal time-stratigraphic names sometimes applied to them. The same holds for the period names: the terms Devonian and Carboniferous rock-sequence are used here instead of group names to avoid the introduction of nomenclature differing too widely from the terminology commonly used in the literature.
The two Devonian sections shown in Fig. 5 are ideal sections assembled from various outcrops. Because of this procedure and the strong tectonical deformation, the thicknesses represented are only approximate.
The stratigraphy of this particular area has previously been discussed by Dalloni (1910, 1930), Schmidt (1931), and Solé Sabaris (1954). The explanations to the maps of the provinces of Huesca and Lérida (Alastre, Almela and Rios, 1947, 1957) give only a short summary of earlier authors and add no new data to the geology of this area.

ORDOVICIAN

The few Ordovician outcrops in this region, nearly all of which are flanked by important faults, are entirely restricted to the northern part of the mapped area (geological map 1 : 25,000).

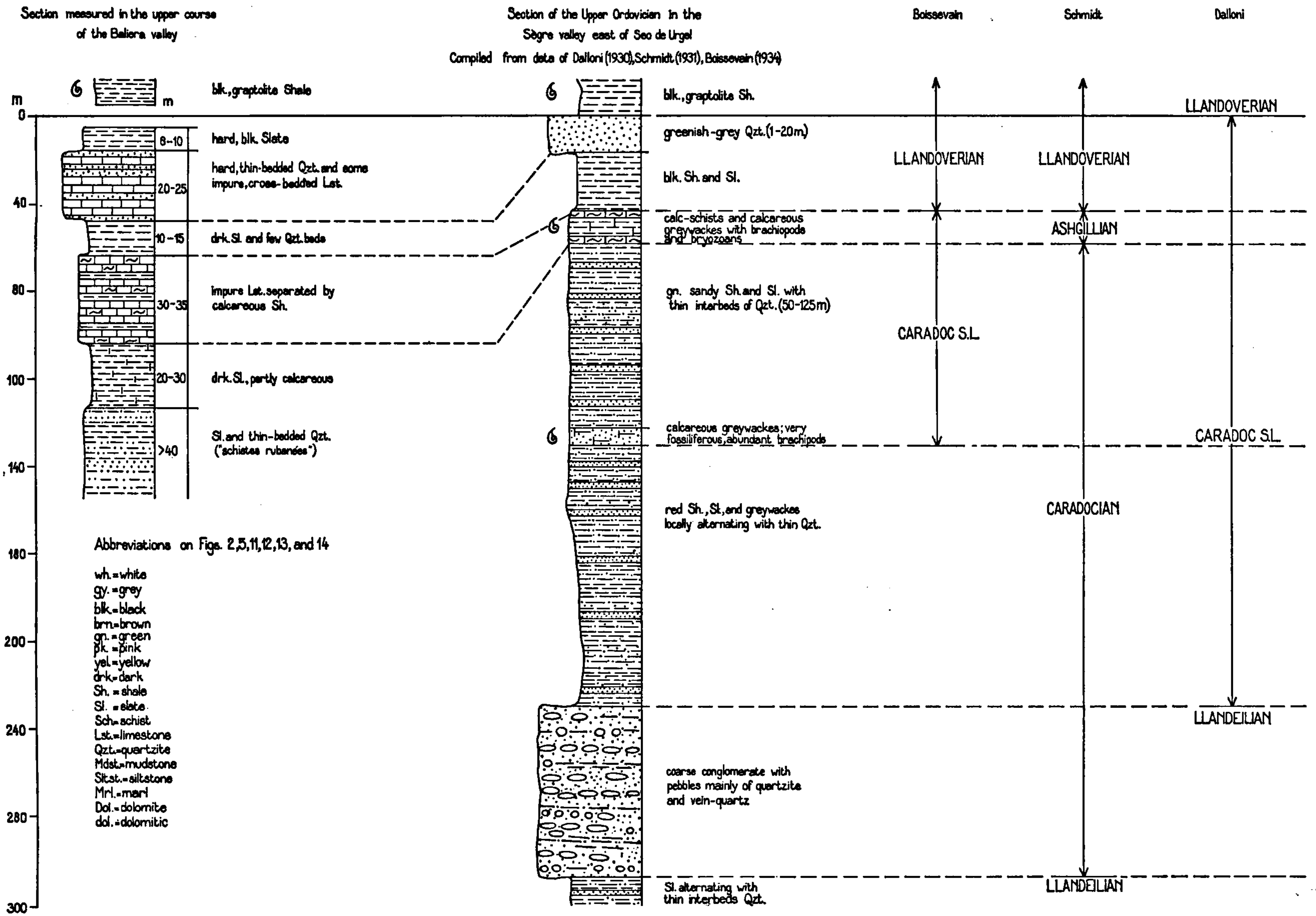


Fig. 2. Columnar section of the Upper Ordovician, taken in the Baliera valley, compared with a schematic section in the Sègre valley east of Seo de Urgel.

Below the Silurian black slates, which range from Middle Llandovery to Lower Ludow, is a formation of mainly slates and quartzites, occasionally containing some impure limestones. This formation is strongly folded, and since most of the area with Ordovician outcrops is covered by debris, the total stratigraphic sequence is difficult to reconstruct. However, the top part of the sequence can be seen in the Llausét brook and on the right bank of the Upper Baliera River just north of the confluence with the Barranco de Basibé (left hand section in Fig. 2).
Dating is difficult, because we found no fossils and these Ordovician outcrops are not mentioned in the literature. Consequently, only a lithological correlation with a dated section can be used to obtain rough indications of the age. Such a correlation is shown in Fig. 2.
The upper part of the fossiliferous Ordovician in the Sègre valley has been described by Dalloni (1930), Schmidt (1931), and Boissevain (1934). Although the time-stratigraphic division of the three authors is founded on two fossil-bearing, well-dated horizons, their interpretations show some discrepancies (Fig. 2). Comparison of the sections of the Cambro-Ordovician reported in literature (Dalloni, 1910 and 1930; Schmidt, 1931; Destombes, 1949; 1953; Zandvliet, 1960; Kleinsmiede, 1960; Clin, 1958; and Mirouse, 1962) shows that a thick horizon with conglomerates belongs to the most constant levels of the upper part of the Cambro-Ordovician. Since no conglomerate has been found in our mapped area, it may be assumed that here only the very top part of the Ordovician is present.
The lithostratigraphic units of Fig. 2 are all dark in colour (dark-grey, grey-brown, and black) except for the marly limestones, which weather to a yellowish-brown. The sequence of thin-bedded quartzites immediately below the Silurian can be followed in the field by its positive relief forms.
Although the thin-bedded quartzites and the complete profile shown in Fig. 2 are very representative for the Ordovician of this area, the impure limestones and slates without quartzite bands cannot be distinguished from any isolated outcrop of the Lower Devonian or parts of the Carboniferous. Therefore, the S-shaped, northernmost outcrop of Aneto slates in the Rio Llausét, which is bounded on the south and north by faults, could equally well be Ordovician.

SILURIAN

The Silurian is the most uniform system in the entire Pyrenees. Even far outside this mountain range, its classical facies of black bituminous shales is maintained (Cantabrian Mts., Catalan Mts., and the Montagne Noire). This uniformity is also expressed in the pelagic fauna, which consist mainly of graptolites, and show a close resemblance to those of the southern part of Central Europe and the Cantabrian Mts. In the Pyrenees the graptolites range in age from Middle Llandovery to Lower Ludlow (e.g. Dalloni, 1910—1930; Schmidt, 1931; Keizer, 1953; and Destombes, 1953).
Besides its importance as a fossiliferous guide marker between the Ordovician below it and the Devonian above it, this highly incompetent formation of black, friable slates also played an important role as lubricant during the Hercynian folding. In the northern and Central Pyrenees these slates have become the detachment horizon of disharmonic folding between the Devonian-Carboniferous above them and the Cambro-Ordovician structures below. In the southern Pyrenees this disharmony is less pronounced.
The Silurian deposits show the typical development of very fine-grained, fissile, black slates, which usually stain the fingers when handled. They are poor in quartz and not very rich in iron, but have an exceptionally high aluminum content. Usually, they are also rich in carbonaceous matter.
Graptolites are frequently found in the less disturbed parts of the Silurian. Dalloni (1910) mentions from the Sierra Negra area:

Monograptus priodon, Bronn.
„ *Sedgwickii*, Barr.
„ *spiralis*, Hisinger
„ *Becki*, Barr.
„ *convolutus*, His.
„ *Helli*, Barr.
„ *Nilsonni*, Barr.

Detailed lithological descriptions as well as the geochemical aspects of the so-called *"schistes carburés"* have already been described in great detail in the recent literature (Destombes, 1953; Zwart, 1954; Clin, 1958; Kleinsmiede, 1960; Zandvliet, 1960; de Sitter and Zwart, 1962).
Kleinsmiede (1960) gives a brief recapitulation (on page 144 of his thesis) of the most important field characteristics of these black shales, which are also valuable for the Silurian of our area. As an important extra property, I may add that in weathered outcrops these shales are often covered with a thin, white alum efflorescence. In German they are therefore termed *"Alaunschiefer"*. This phenomenon is caused by the high content of aluminum (up to 35% AL_2O_3) and sulfur in the form of pyrite and markasite, which during weathering form Al-sulphate and Alkali-sulphate. After being leached out by perculating water, the alum crystallizes at the surface of the Silurian slates.
The intensely and often chaotically contorted deformation of this highly incompetent formation makes it impossible to determine directly its thickness, which in places may well be under 100 m. In the Sierra Negra area, with its vast Silurian outcrops, the total thickness is probably more than 100 m. Since Solé Sabaris (1954) incorporated the Aneto slates into the Silurian and also underestimated the doubling-effect of isoclinal folding, his estimate of 1800—2000 m for the thickness of the Silurian is not acceptable. At present, it is generally held that the Silurian in the Central Pyrenees does not exceed a thickness of 200 m (de

Sitter, 1954a; Destombes, 1953; Kleinsmiede, 1960; Mirouse, 1962).

The top part of the slates (upper 30—50 m) carries thin-bedded, black limestones which frequently contain orthocerids, crinoid stems, and *Cardiola interrupta.* These limestones are very porous and when crushed with a hammer form a black powder which stains the fingers.

A horizon with black, elliptical, often dolomitic, fossiliferous limestone nodules, reported by many authors, was not found in the map-area.

Near the Maladeta granodiorite the limestones change from black to a dull silvery-grey and sometimes light-grey; this change in colour is accompanied by an increase in grain-size. These metamorphic limestones show an abundance of empty cubical holes, formerly occupied by pyrite crystals. The fossil content remains recognizable up to the areas with the highest metamorphism.

The black shales more or less retain their friable character even in the vicinity of the granodiorite; this is most probably due to the very low content of free quartz and the rather high content of carbonaceous matter (4—8%). The chiastolites generated by thermal metamorphism reach as much as 2 cm in length.

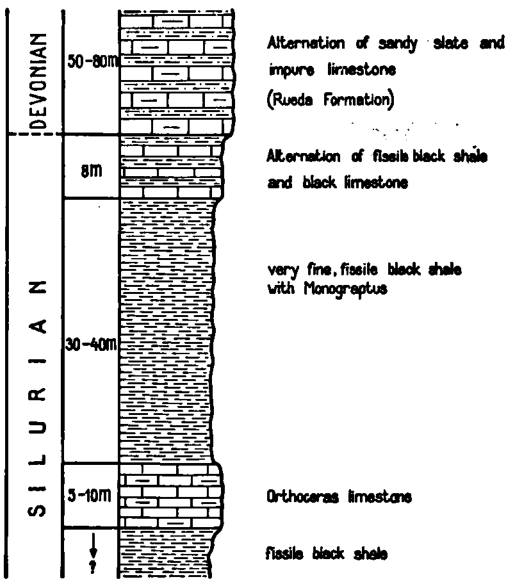


Fig. 3. Columnar section through the Upper Silurian and the lower part of the Rueda Formation.

The transition to the black Devonian Aneto slates is rather gradual, so it is often difficult to define the exact boundary. In the Sierra Negra facies area, however, the limit between the Silurian, with or without limestones, and the Rueda Formation can generally be drawn with much more accuracy. This is due to the fact that both the impure limestones and the sandy slates of the Rueda Formation differ greatly from those of the Silurian (Fig. 3). Where the Orthoceras limestone forms a continuous horizon in the field, it is shown on the map.

However, since detailed biostratigraphical evidence is lacking, this distinct boundary is in all likelihood not the exact limit between the Silurian and the Devonian but merely the limit between two formations.

DEVONIAN

Introduction

After the well-dated and uniformly-developed Silurian, the Lower Devonian does not start in the same way in all areas. There often seems to be a discontinuity between the Devonian and the Silurian, but there is no bio-stratigraphic evidence to support the assumption of a stratigraphic hiatus.

The recent literature on the geology of the Pyrenees indicates that a division into several distinct facies areas can be recognized (see review in Mey, 1967). In each of these areas the Devonian and Lower Carboniferous have a uniform and characteristic lithological development, but there is often a marked difference between them. These facies areas are separated by important fault zones or Cambro-Ordovician domes, or sometimes by a combination of both.

To establish the communication with the adjoining areas, which have been described in the literature, the stratigraphy of the Devonian and Lower Carboniferous was studied in the field over a larger area than is shown on the geological map (Fig. 4). In just this small part of the Pyrenees, three distinct areas showing different developments of the Devonian could be distinguished. These facies areas will be described in the following order:

the Sierra Negra Facies,
the Baliera Facies,
the Renanué Facies.

For the first two facies, an ideal section is shown in Fig. 5. For the description of the Renanué Facies in the soutwest (Fig. 4), we refer to Mey (1967) and Wennekers (1968).

The Sierra Negra Facies

The Sierra Negra Facies is developed in the area immediately south and west of the Maladeta granodiorite (Fig. 4). West of the Tor River it is separated from the Baliera Facies area by an important thrust zone running parallel to the general trend of the Hercynian structures in this area. Between the Baliera and Ribagorzana rivers, some Ordovician and Silurian is also involved in this boundary. East of the Tor, the southern boundary is formed by the Cambro-Ordovician Payasos Dome. For the western continuation of this facies area, we refer to Wennekers (1968).

Most of the exposed sequence of the Sierra Negra Facies was affected by the contact metamorphism of the vast Maladeta granodiorite. Although this accentuated the lithological differences between the successive formations, the original lithological properties are hard to recognize. Once the stratigraphical position of a certain formation has been established,

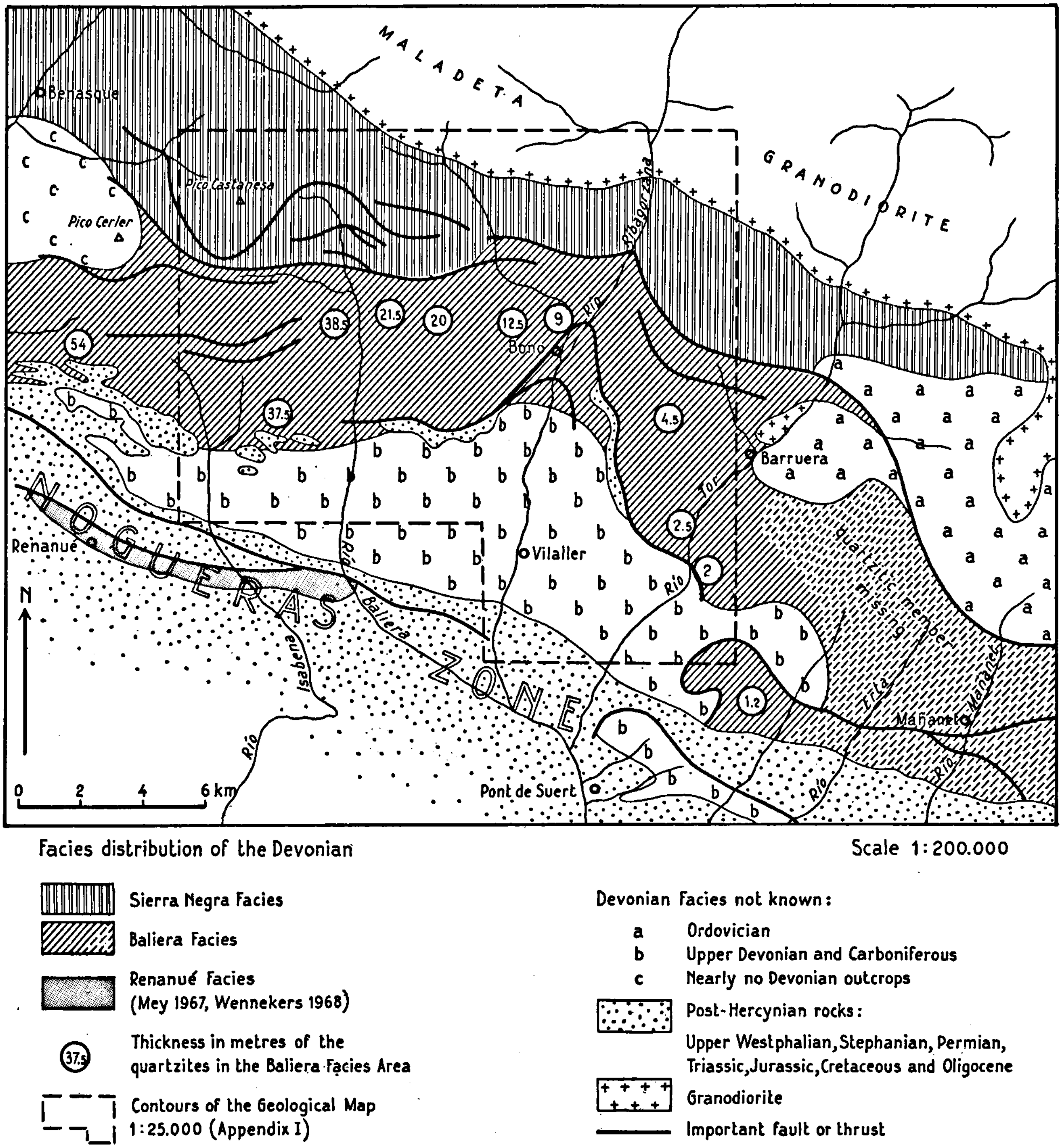


Fig. 4. Sketch-map showing the distribution of three Devonian facies south and southwest of the Maladeta granodiorite.

however, mapping presents no special difficulties. Therefore, the stratigraphy (Fig. 5a) has been worked out beyond the metamorphic aureole southwest of a line joining the Pico Castanese with the confluence of the Astos and Esera rivers.

Rueda Formation. — The Silurian, with its typical graptolite-bearing black slates and some thin limestones at the top, is followed conformably by a monotonous series of sandy slates and impure limestones. In the Montaña de Rueda (SSE of Benasque), after which this formation is named, several well-exposed complete sections of this mainly detrital series can be studied in detail.

Definition. — The Rueda Formation is defined as the rock unit in which sandy and argillaceous limestones alternate with sandy, calcareous, and slightly carbo-

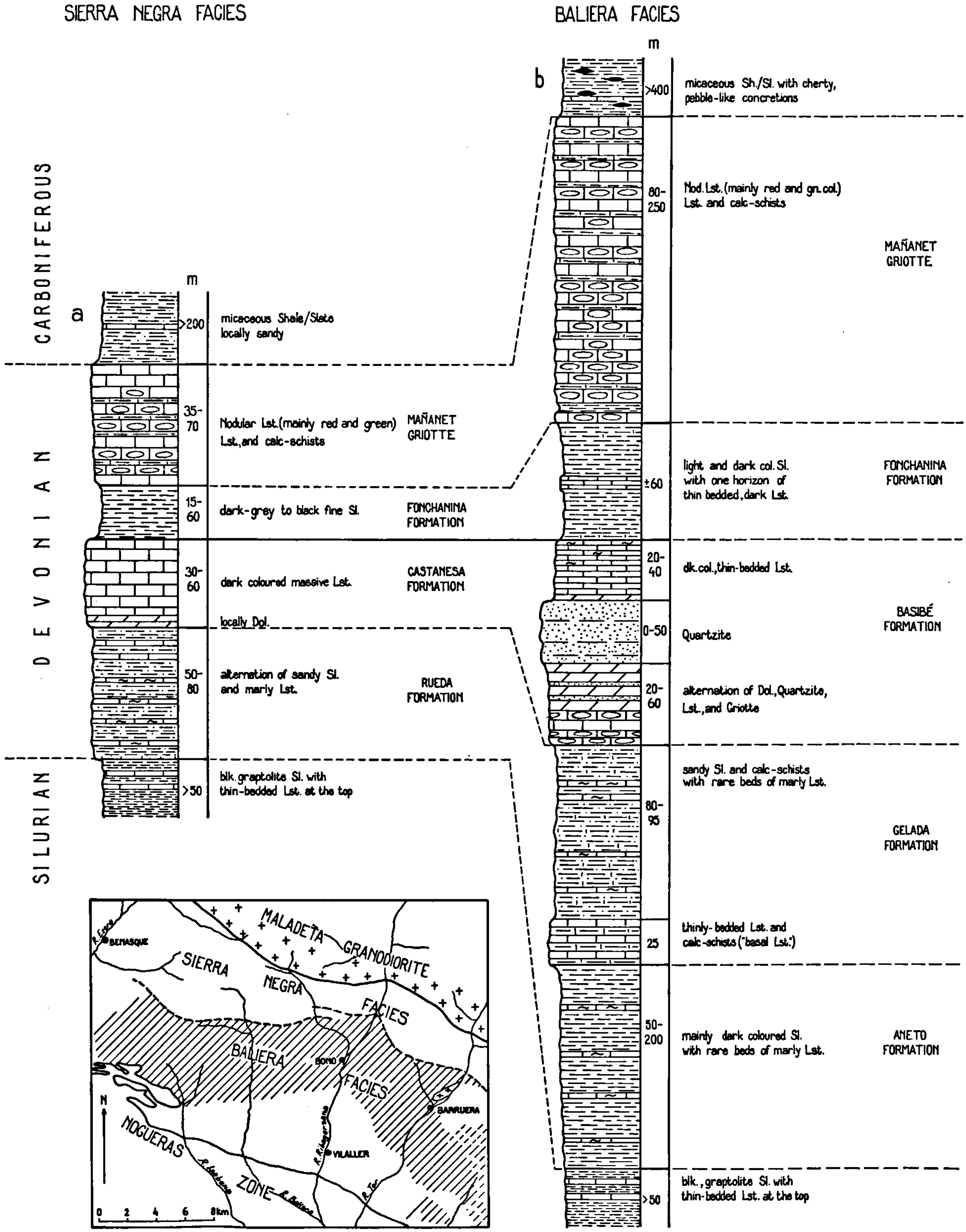


Fig. 5. Simplified lithostratigraphic Table of the Devonian Baliera and Sierra Negra Facies.

naceous shales/slates. In the lower part of the formation a few pure shales or slates may occur. The formation rests with a sharp boundary upon the typical graphite-bearing black slates of the Silurian. The top is taken at the base of the lowest pure or dolomitic limestone, which is more than 10 m thick, and is considered to belong to the Castanesa Formation.

Thickness and characteristics. — In the map-area the thickness of this formation varies between 80 and 50 m, although it may be much thinner in the strongly-flattened long limbs of large cleavage folds. The main characteristics are the dull, dirty brownish colours (due to the high iron content), the preponderence of impure rock types, and the black, irregular streaks and stains on the bedding planes of the sandy pelitic rocks. These streaks are sometimes bifurcated or branched and may strongly resemble carbonaceous plant remains (Fig. 6), thus simulating a paralic sedimentation. They are also visible on the cleavage plane, where their shape depends on the angle between the two planes. If this angle is small, the streaks are irregular and broad without a distinct orientation; if it is large, the streaks are long and very fine and show an excellent orientation parallel to the δ-lineation (intersection of cleavage and bedding). Neither microscopical analysis nor X-ray investigation gave any indications of the original composition of these streaks, but chemical analysis showed a distinctly higher percentage of carbon in the streaks than in the surrounding sediment. An organic origin therefore seems quite probable. Besides these streaks, the detrital sediments also contain many iron contaminations and detrital mica flakes measuring up to 0.5 mm.

The bedding of the various rock types can be easily distinguished from nearby, but from a distance it is hardly visible. The thickness of individual beds varies between 1 cm and about 50 cm. 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 formation contains a relatively large amount of fossils, among which the crinoid ossicles, corals *(Cyathophyllum)*, orthocerids, and brachiopods are most frequent. Some specimens were slightly deformed by the flattening of the cleavage.

With the point-counter method we established from several thin sections of the sandy pelitic rocks the following averages:

Quartz	23 —48 %	(the grain-size varying by about 60μ)
Clay	27.5—33 %	matrix
Carbonate	34.5—37 %	matrix
Muscovite	0.5— 3.5%	
Chlorite	0.5— 2.5%	
Ore	0.5— 1.5%	(mainly iron oxides)

In some cases the clay and carbonate of the matrix could not be counted separately because these minerals occurred strongly interwoven, together reaching a percentage of between 47.5 and 75.

It is interesting to note that the orientation of the thin section with respect to cleavage and bedding has a strong influence on the counts, affecting mainly the quartz-matrix (clay + carbonate) ratio. The quartz-matrix ratio of the sections perpendicular to both cleavage and bedding on the one hand, and the sections perpendicular to cleavage and parallel to the bedding on the other hand, proved to be nearly 3 to 1.

In view of the composition and grain-size of this detrital rock, the original sediment can best be described as a quartz-rich calcareous mudstone or as an extremely fine-grained calcareous quartz-wacke (Pettijohn, 1957; Carozzi, 1960). Dalloni (1910) and Schmidt (1931) describe these detrital rocks as greywackes and shales.

With decreasing distance to the granodiorite, the detrital sediments gradually become more brittle and show dark spots of metamorphic minerals. The im-

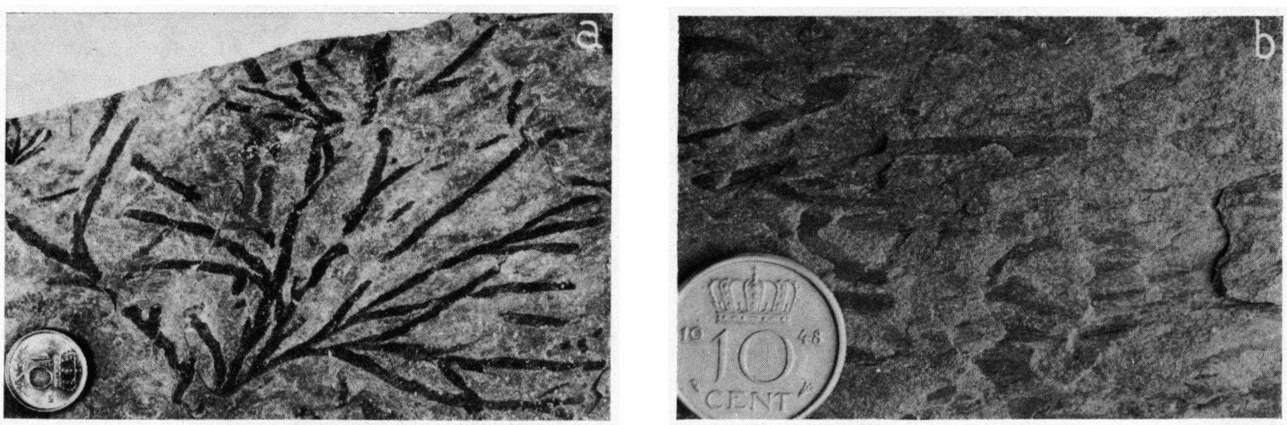


Fig. 6. Carbonaceous black streaks typical of the Rueda and Gelada Formations.
a. on the bedding plane b. on the cleavage plane (the bedding makes a low angle with the cleavage).

pure limestones tend to be recrystallized into a groundmass of pure calcite with phenocrysts of lime-silicate minerals. These mineral assemblages are discussed in more detail in Chapter III (page 212).
In the highly metamorphic contact zone in the vicinity of the granodiorite, the rocks of the Rueda Formation have been converted into hornfels, marble, and lime-silicate rocks. This formation consequently has a dull greenish-grey to dull brown appearance, but on close scrutiny the hornfels (originally detrital sediments) can be distinguished as brown and the impure limestone as ash grey. Even in this highly metamorphic state the typical black streaks mentioned above can still be recognized in the hornfels.
Selective erosion gives the rock mass the appearance of a sandwich limestone (French: *barrégienne*), the hornfelses forming the ridges and the more soluble marbles and lime-silicate rocks the recesses. This ridged appearance and the difference in colour between the hornfels and mable strongly accentuate the bedding, which is clearly visible even from a great distance, in contrast to the non-metamorphic areas described above.

Origin. — The sediment and the fossil content of this formation point to shallow marine, near-shore conditions of sedimentation, with a high influx of terrigenous material from the continent and/or eroding ridges. The rate of sedimentation was probably low.

Age. — The fossils we found were too severely deformed to permit reliable age determination. Dalloni (1910) and Schmidt (1931) determined macrofossils in the same formation further to the west and found a Lower Devonian age (Coblentzian).

Correlation. — The Rueda Formation can be correlated with the Gelada and Aneto Formations in the Baliera facies area. It can probably also be correlated with the Entecada slates from the Valle de Arán (Kleinsmiede, 1960; Mey, 1967). The Rueda Formation is more or less equivalent to what Dalloni (1910) and Schmidt (1931) call the Coblentzian, which in principle is defined as a facies of a time-stratigraphic unit but is very often applied as a litho-stratigraphic unit or formation.

The Castanesa Formation. — The massive limestone following the Rueda Formation is well exposed and complete on the Pico Castanesa (coord. 4°17½'. 42°34½'; Fig. 7), the mountain summit after which this formation is named.

Definition. — The Castanesa Formation is defined as the massive, generally dark-grey limestone (with a light-grey weathering) resting with a sharp contact upon the Rueda Formation. It is overlain, with a sharp contact, by a series of blackish shales/slates sometimes containing a few thin-bedded limestones (= Fonchanina Formation). The lowermost 4 to 8 m of this limestone may locally be dolomitic or developed as a pure dolomite, in which case the weathering colour is yellow.

Thickness and characteristics. — The total thickness of this formation varies in the map-area between 30 and 60 m, but this variation is due mainly to the difference in cleavage flattening from one place to another. The limestone is in general very fine, pure, and thick-bedded to massive; the bedding is less well pronounced than the cleavage. When bedded dolomites occur, as e.g. in the southern and central part of the Sierra Negra facies area east of the Baliera River, the bedding is the prominent feature and the cleavage is hardly developed. The top part of the limestone unit is often rather porous, and in that case the general light-grey weathering becomes a darker grey. The only fossils we found

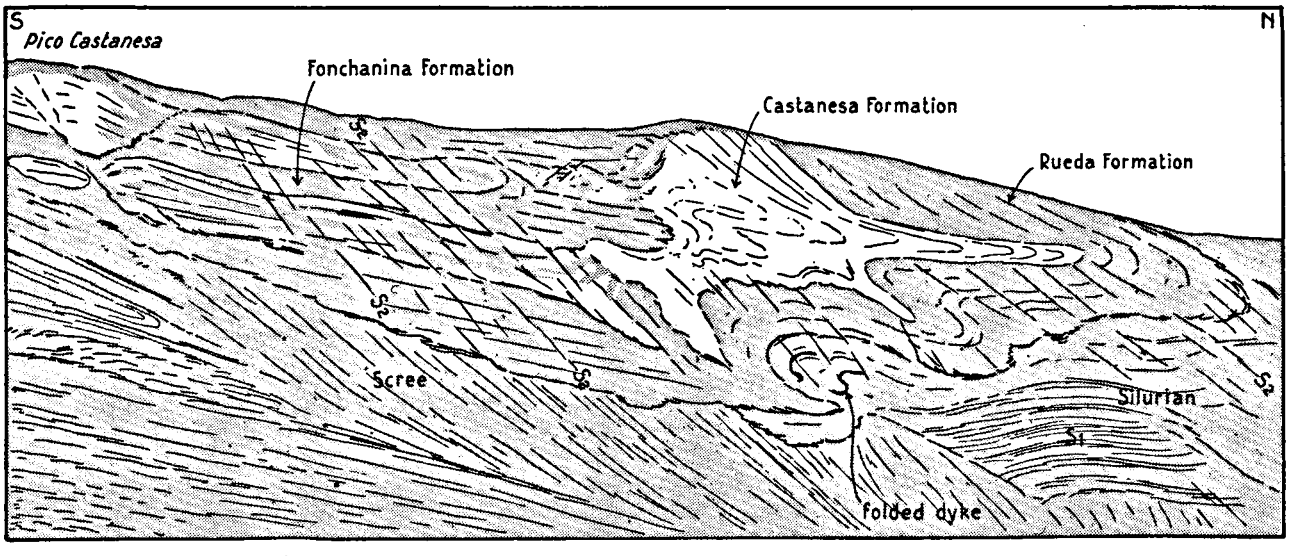


Fig. 7. Large-scale isoclinal syncline of the main phase in the Castanesa massif, slightly refolded by the s2-folding. Drawing after photograph.

are some undeterminable solitary corals and many crinoid segments. Because of its rather homogeneous calcareous composition and massive appearance, folds of this formation are seen best when viewed from a considerable distance, this formation being flanked on both sides by dark-coloured pelitic rocks.

In the contact metamorphic area the general appearance of the limestone changes very little and then because: a. the recrystallization has changed the internal colour to a lighter grey; b. the slightly marly parts protrude on the weathered surface, accentuating the bedding, and at the same time the often capricious internal folds become clearly visible on the weathered surface.

Notable and new, however, are variably sized (up to 10 m), irregular, discordant dolomite bodies which obliterate all traces of the bedding and cleavage. These bodies are orange-brown and contrast clearly with the very light-grey weathering of the marbles. They are obviously the result of secondary dolomitization, and are undoubtedly related to the complete recrystallization of the limestones by the intrusion of the granodiorite.

Karst phenomena, such as clints *(lapiés, Karren)*, dolines, and sources, occur frequently in these marbles.

Origin. — The paucity of fossils and the very homogeneous character of a rather pure, fine carbonate with hardly any pelitic material point to a quiet period of marine sedimentation under favorable climatological conditions and with little or no influx of terrigenous material from the continent or ridges.

Age. — The few corals we found could not be adequately determined. Dalloni (1910) assigns this formation to the Eifelian on the basis of fossil discoveries (mainly corals) in this limestone west of our area. On the Pico Castanesa he collected *Favosites* sp. and *Cyatocrinus.*

Correlation. — The Castanesa Formation can be correlated with the Basibé Formation of the Baliera Facies area (Fig. 5); and probably also with the limestone horizon below the turbidity sequence in the Valle de Arán (Kleinsmiede, 1960; Mey, 1967). The Castanesa limestone is most probably the equivalent of the *"calcaire à polipiers"* of Eifelian age, so well-developed in the western and northern Pyrenees (Dalloni, 1910; Mirouse, 1962; Mey, 1967).

The Fonchanina Formation. — Dark-coloured slates overly the Castanesa limestone and are in turn overlain by nodular multicoloured limestones (griotte). The same stratigraphic succession is found in the Baliera facies area (Fig. 5), where we have called these slates the Fonchanina Formation, after the type area in the neighbourhood of the village of Fonchanina (coord. 4°20½'.42°30½').

Definition. — The Fonchanina Formation is composed of a series of fine fissile dark-coloured shales/slates with lustrous cleavage planes and a few thin-bedded dark limestones, which are sometimes absent. The lower limit of these pelites is defined as the contact with the massive Castanesa limestone or, as in the case of the Baliera facies area, as the contact with the thinly-bedded, dark-coloured, and slightly bituminous limestones (upper member of the Basibé Formation). The upper limit of this formation is taken at the first nodular limestones or marly limestones measuring at least 5 m in thickness.

Thickness and characteristics. — Owing to the strong tectonic flattening of this fine pelitic material and the lack of macroscopic bedding features, the exact thickness of this formation is very difficult to estimate. In the flanks of isoclinal structures we measured thicknesses of between 15 and 35 m, but in these parts of the structure the cleavage flattening, which sometimes attains a value of 45 to 70% (page 190), is not taken into account.

Besides the main characteristics mentioned in the definition of this formation, we must add that while these slates may be black locally, they never stain the fingers as do the Silurian black slates. This formation can be distinguished from most Carboniferous slates by the absence of detrital mica flakes. Between the Fonchanina slates and most of the Aneto slates there is very little difference (page 168). To arrive at a reliable interpretation, one is forced to depend entirely on the surrounding limestones and the general tectonic picture of the specific area. No fossils except rare crinoid ossicles were found. In the metamorphic aureole these slates are brittle and spotted in the outer zone and have altered into hornfels close to the granodiorite, the colour being changed into anthracite grey, dark brown, and, locally, a rusty brown. Very often, these hornfelses are overgrown with greenish, yellowish, or reddish moss, sometimes making it difficult to see their proper colour. Even in this highly metamorphic state the difference from the Silurian slates is always clear, because metamorphic Silurian slates never become hornfels, but stay rather friable up to the vicinity of the granodiorite. Metamorphic Silurian, furthermore, always bristles with chiastolites, a mineral rarely occurring in the metamorphic Fonchanina slates.

In the metamorphic aureole it is very difficult, however, to distinguish between the hornfelses of the Fonchanina Formation and those of the Aneto Formation and between the hornfelses of the Fonchanina Formation and those of the Ordovician and Carboniferous. This distinction must there depend on the characteristics of the surrounding limestones and the general tectonic picture.

The circumstances of sedimentation deviate from those of the preceding formation only in that pure argillaceous rather than pure limestone material was deposited. From a hydrodynamic point of view, the two kinds of deposit are similar. Apparently, the biochemical conditions in the basin changed in such a way that $CaCO_3$ could no longer be deposited and the terrigenous supply became predominant.

Age and correlation. — So far, no characteristic fossils have been found in this pelitic rock sequence, and the age of this formation can only be estimated from a correlation with a dated sequence on the same lithostratigraphic level. Similar mainly pelitic sequences, locally accompanied by sandstone intercalations, lying on top of a massive limestone of more or less Middle Devonian age and below the typical griottes of the Upper Devonian, occur in most of the Devonian sections described in the literature (Mey, 1967). These rocks have been dated as Frasnian in the western Pyrenees (Mirouse, 1962) and as upper Middle Devonian to lower Upper Devonian in the Haute Garonne (Destombes, 1953). In the eastern part of the Noguras Zone this pelitic sequence is of upper Lower Devonian or lower Middle Devonian age (Ziegler, 1959; Mey, 1967).

Mañanet Griotte. — The Fonchanina slates are conformably overlain in both facies areas by mainly nodular multi-coloured limestones (griotte). This rock sequence has been defined by Roberti (pers. comm. 1967) from the Mañanet valley, some 15 km east of the Ribagorzana. The type section is located 2.5 km north of the village Mañanet, along the cattle track on the west side of the Mañanet River.

Definition. — The Mañanet Griotte is defined as the rock unit consisting mainly of nodular multi-coloured limestone (griotte) with intercalations of red and green marly limestones and shale. The lower limit is taken at the first dark-coloured fine slate without griotte intercalations; there may, however, be intercalations of thin-bedded dark limestones. The upper limit is not sharp, but is taken at the base of the first dark-coloured shales/slates, which are almost always characterized by large detrital mica flakes (Carboniferous).
The term griotte was originally applied to "marbles" of Frasnian age in the French Pyrenees (cherry- coloured spots), but in the modern literature this term is used for any nodular, vividly-coloured carbonate rock, without regard to its age. In the Pyrenees the griotte facies is almost exclusively restricted to the Upper Devonian, however (Mey, 1967).

Thickness and lithology. — The total thickness of this formation varies in the mapped part of the Sierra Negra facies area from 35 to 70 m if the tectonic flattening is not taken into account. In its normal development the griotte is very characteristic. Firstly, the rock shows a variety of colours: apart from the usually yellowish-grey to greenish-brown surface-weathering they range from white, reddish, greenish or violet-grey to rare amber tones, with either an abrupt or gradual transition. The second characteristic, from which the rock derives its name, is the nodular internal structure of the limestone. In sections taken perpendicular to the tectonic b-axis, the shapes are roundish to flat ellipsoid, comparable to the goniatite-bearing griotte limestone of the French Pyrenees (*"Marbre de Pyrénées")*. In the direction of the tectonic b-axis, rod-like structures occur.
Beside these typical nodular limestones, the formation shows many intercalations of vividly-coloured calcareous shales, pure shales, and dark-coloured limestones. The dark limestones occur more frequently in the western part of the mapped area, where they are concentrated mainly in the upper part of the formation. They attain a thickness of some 25 m in the southern part of the Esera valley (Wennekers, 1968). The bedding of the various types of rock can be easily distinguished. In the griotte the bedding planes are undulating (due to the rodding), and the transition from one bed to another may be gradual or sharp. The thickness of individual beds ranges from 1 cm to about 50 cm. The contact between the griotte and the more pelitic rocks is generally sharp. The dark limestones are regularly bedded, the thicknesses of the individual beds varying between 10 and 30 cm.

Fig. 8. Typical rodding in the nodular limestones of the Mañanet Griotte, caused by the intersection of cleavage (s1) and beddings (ss).

The slaty cleavage, which is well developed in both griotte and calc-schists but not in the dark-coloured limestones, cuts the bedding at angles of various sizes and produces the rod-like structures (Figs. 8 and 46). The individual cleavage planes are smooth and show a lustrous surface.
In the Sierra Negra facies area this formation proved to be poor in fossils; we found only the usual crinoid ossicles.
In the metamorphic state some of the highly typical characteristics, such as the nodular aspect and the vivid red colours, are entirely lost. The greenish and cream colours remain, establishing the differentiation from the other limestones. The thin pelitic layers between the calcareous nodular beds change into hornfels. The result, a sandwich limestone, is to some extent com-

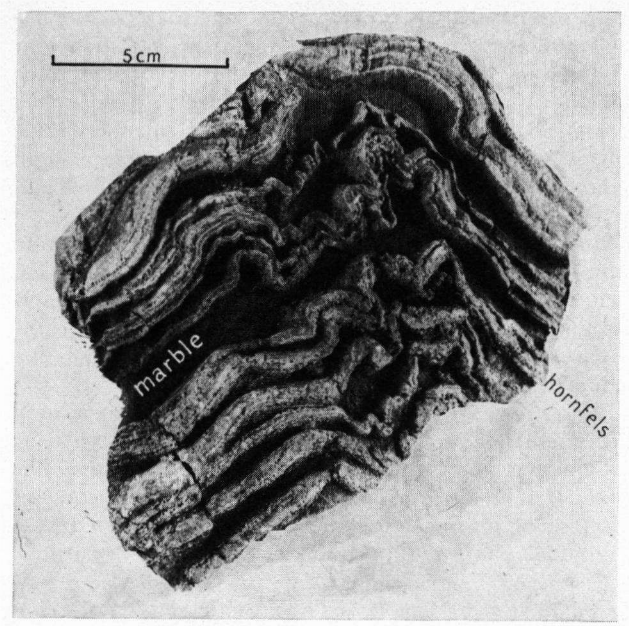


Fig. 9. Sandwich limestone *(barrégienne)* of the metamorphic Mañanet Griotte, caused by differential weathering of marble and hornfels.

parable to the metamorphic Rueda Formation, but the alternation of limestones and hornfelses is much finer in the metamorphic griotte (0.5—3 cm; Fig. 9). From far away the metamorphic griotte appears to be a dull light-grey, although distinctly darker than the Castanesa limestone and somewhat lighter than the metamorphic Rueda Formation.
The origin and conditions of sedimentation of the griotte can be studied much better in areas without strong cleavage. For these subjects, the reader is referred to Ovtracht and Fournié (1956) and the very recent publication of Nagtegaal (1967).

Age. — Neither Dalloni nor Schmidt mentions the occurrence of nodular limestones in this northern area. Since biostratigraphic data are not available, the relative age of this formation can only be inferred from correlations with dated sections in the surrounding areas. A thorough study of the literature showed that the massive multicoloured griotte limestone of the upper part of the Devonian ranges in age from Frasnian to lower Tournaisian (locally) in the Pyrenees (e.g. Dalloni, 1910, 1930; Schmidt, 1931; Destombes, 1953; Keizer, 1953; Ziegler, 1959; Mirouse, 1962; Mey, 1967).

The Baliera Facies

The characteristic properties of this facies are expressed only in the Lower and Middle Devonian, and since the southernmost sediments of this part of the Axial Zone are of Carboniferous and Upper Devonian age, the exact southern extension of this facies is not known (Fig. 4). In the Nogueras Zone, however, the Devonian already has a different development (Renanué Facies, Mey, 1967). In the north this facies area is bordered by an important fault zone (Figs. 4 and 15). The western limit very probably coincides with a N-S trending fault zone (Liri fault zone) not far to the east of the Esera River (Fig. 15). East of the Tor River the Baliera Facies gradually looses its characteristic properties of a pelitic Lower Devonian and a quartzite member in the middle part of the Devonian. The exposed sequence of the Baliera Facies is only very locally influenced by thermal metamorphism, as for example in the vicinity of the Barruera granite, near the village of Bono (coord. 4°25′.42°32′), and in the intermediate area with its many intrusive dykes. The Bono metamorphic area is, moreover, strongly silicified, so the original sedimentary rocks can no longer be recognized.
By far the greater part of the Baliera facies area, however, is not metamorphic, which is of course advantageous for accurate study of the stratigraphy and the structural history of this complicated area.

Aneto Formation. — As distinct from the Sierra Negra Facies, the Lower Devonian starts here with a thick series of slates bearing a slight resemblance to the Silurian black shales. Their Gedinnian age was first demonstrated by Dalloni (1910) in an outcrop near the Pico Cerler, about 6½ km southeast of Benasque. As the type section of this slate formation we selected the well-exposed and typical sequence in the vicinity of the village Aneto in the Ribagorzana valley (coord. 4°26′.42°33′).

Definition. — The Aneto Formation is defined as the sequence of fissile dark-coloured slates containing few intercalations of argillaceous limestones and overlying the Silurian graphite-bearing black slates. The transition between the two types of slate is often gradual, but the lower boundary of the Aneto Formation is taken at the top of the first occurrence of graphite-bearing black slates. The upper boundary of this formation is taken at the base of the lowest limestone member of the Gelada Formation. The boundary between the two formations is generally sharp.

Distribution. — In a narrow belt this slate formation can be traced from Pico Cerler in the west via Collado de Basibé, Collado de Salinas, Aneto, and Collado de Gelada, down to the Tor valley. There, northeast of the Barruera stock, a narrow syncline of these slates becomes lost between the Cambro-Ordovician Payasos Dome in the north and the Muro Dome in the south (Fig. 15). Besides this continuous belt, these slates are also exposed in the core of some anticlinal structures (see geological map). East of the Tor River the formation is much thinner and contains more limestone intercalations. Here, it becomes very difficult to separate the Aneto from the Gelada Formation, and for practical purposes it is more convenient to map both units together; this compound unit is called the Rueda Formation (see definition on page 161).

Thickness and characteristics. — Owing to the strong tectonic flattening of this incompetent rock unit and the lack of any mappable guide marker to indicate the degree of folding, it was impossible to evaluate the original thickness of this formation. Our estimate is 50 to 200 m for the western and central parts of the mapped area and 10 to 30 m for the area east of the Tor River.

The slates are fine to slightly silty, and therefore show few or no traces of bedding. Their colour is dark grey to black, but brown shades also occur locally. Even when the slates are black they never stain the fingers. In tectonically disturbed parts of these slates, as for example shear-zones, the outcrops are sometimes partly covered with a white alum efflorescence, as was also the case with the Silurian black slates (page 159); under these circumstances it is impossible to identify the kind of slate involved.

In general, however, the cleavage in the Aneto slates is developed much more regularly than in the Silurian, and this contrast and the somewhat lighter colour form the main differences between the two types of slate. The Aneto Formation also contains some impure limestones, discernable by their brownish weathering and their relatively higher competence. In small isolated outcrops, however, it is impossible to decide on purely lithological grounds whether such impure limestones in a matrix of silty slates belong to the Aneto Formation, to the upper part of the Ordovician, or to a somewhat finer level in the Gelada Formation.

Also typical of this Aneto Formation are black, spherical, or elliptical nodules with a diameter of 3 to 20 cm occurring in considerable quantities locally. Dalloni (1910) was the first to notice these nodules, and found in them the following fossils:

Orthoceras styloideum, Barr.
Pleurotomaria catelunata, de Vern. et d'Arch.
Spirifer
Tentaculites

Under the microscope these pebble-like, silicious concretions show a micro-crystalline mosaic texture; a preferred orientation of recrystallized clay minerals (cleavage) may be present locally.

Age. — In searching for fossils we came across several badly-preserved imprints of small brachiopods *(Orthis)* and incomplete trilobites *(Phacops)*, which could not be determined more exactly. Dalloni (1910) was more successful; he collected the following fossils near Pico Cerler:

Phacops aff. *plagiopthalmus,* Richt.
Orthis hipparionix, Schnur.
Rastrites linnaei, Barr.
Tentaculites acuaris, Richt.
Tentaculites typus, Richt.
Tentaculites aff. *striatus,* Richt.
Styliola clavulus? Barr.

On the basis of this fossil assemblage, he concluded a Gedinnian age. The later investigations made by Schmidt (1931) confirmed this dating. Besides some indeterminable brachiopods, he collected *Rhynchonella tarda,* Barr., a fossil which in Bohemia and the Carnic Alps is found only near the Silurian-Devonian boundary.

Mode of sedimentation. — The gradual change from the Silurian bituminous black shales with almost exclusively pelagic elements (graptolites) into the lighter-coloured and slightly silty slates of the Aneto Formation with a benthonic fauna, forms an indication that the basin floor became better aerated.

From the geological data, the rate of sedimentation of the Aneto slates can be calculated to have been about ten times that of the Silurian black slates. The fact that in the surrounding areas this slate sequence is thin or missing altogether would be explained by a relatively more rapid subsidence of the Baliera basin.

Gelada Formation. — The Aneto slates are conformably overlain by a formation nearly identical to the Rueda Formation of the Sierra Negra Facies (page 163). For correlation purposes, however, a new name had to be introduced, since there is no equivalent of the Aneto Formation in the Sierra Negra facies area, where the Rueda Formation rests directly on the Silurian black slates.

We have named the formation after the Pico Gelada (coord. 4°27'. 42°31½'), a very accessible mountain on the watershed between the Tor and Ribagorzana rivers. On the northern slope of this mountain an undisturbed and complete section of this formation is exposed; we measured a total thickness of 115 m.

Definition. — The Gelada Formation is defined as the rock unit in which sandy, calcareous, and slightly carbonaceous shales/slates alternate with few sandy and argillaceous limestones. In the lower part of the formation an impure limestone member can often be distinguished. The formation rests with a sharp boundary upon the Aneto slates. The top of the Gelada Formation is defined at the base of the lowest pure limestone, nodular limestone, or dolomite of the Basibé Formation. This boundary is always sharp but conformable.

Thickness and lithology. — 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 100 to 120 m, including the basal limestone member.

The main properties have already been described in the discussion of the Rueda Formation (page 163). It is evident, however, that in the Gelada Formation, except for the basal limestone member, the sandy pelitic rocks generally predominate over the calcareous rocks. Furthermore, the absolute sand content in the pelitic rocks seems higher than in their equivalents in the Rueda Formation.

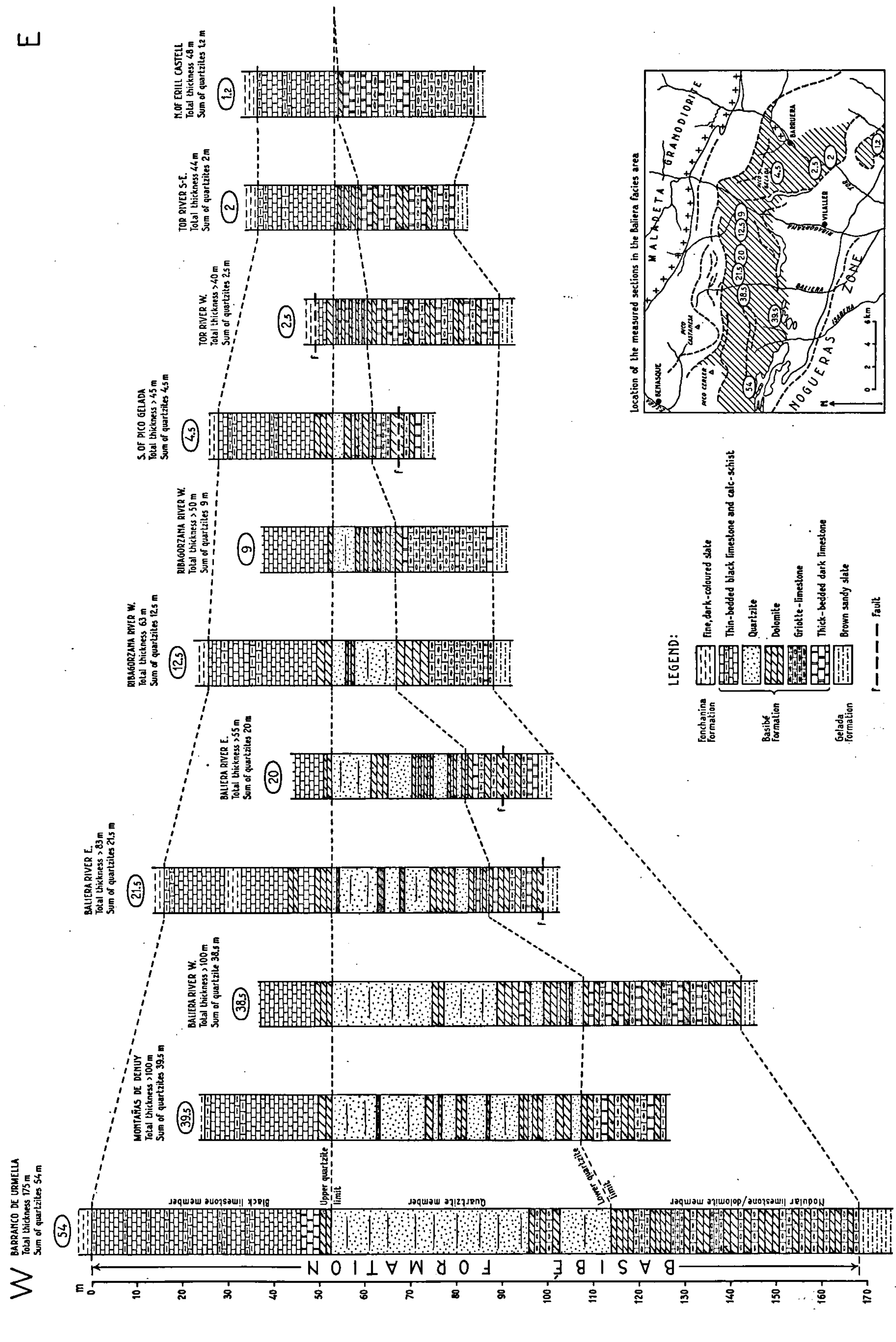


Fig. 10. Lithostratigraphic sections of the Basibé Formation, showing the decrease in thickness of its individual members from west to east.

The basal limestone member constantly present in the map-area is some 25 m thick, and consists of a thinly-bedded alternation of mainly limestones and marly limestones. Its boundary with the Aneto slates is always sharp and easy to map. Solé Sabaris (1956) took this limit for his boundary between the Silurian and the Devonian, since he considers the Aneto slates, on purely lithological grounds, to belong to the Silurian. The upper boundary of the basal limestone member with the sandy pelitic sequence is rather gradual. The individual limestone beds are dark-grey with a grey to orange-brown surface weathering. Besides an abundance of crinoid ossicles, orthocerids and brachiopods were also observed.

In the sandy pelitic sequence above the basal limestone we encountered not only the above-mentioned fossils but also strongly deformed solitary corals and traces of trilobites. Reliable determination was impossible, however. A conodont investigation yielded one doubtful Lower Devonian form (van Adrichem Boogaert, pers. comm., 1961).

From the Silurian upwards, the sedimentary succession and fossil assemblages point to a regressive cycle, with an increasing supply of coarser terrigenous material during the Gelada "time".

Age and correlation. — In this sandy, pelitic rock-sequence, Schmidt (1931) encountered the following fossils in the Esera valley near Sahun:

Phacops cf. *fecundus* Barrd.
Orthonychia hercynica Kays.
Strophomena cf. *taeniolata* SDB.
Rhynchonella (Camarotoechia) sp.
Rhipidophyllum sp.
Chaetetes roemeri Kays.
Cladochonus striatus Gieb.

He assigned a Lower Coblentzian age on the basis of this fossil assemblage. Dalloni (1910) was of the same opinion; he collected his fossils (mainly brachiopods, trilobites, and corals) in the area west of the Esera River.

On purely lithological grounds, the Gelada Formation can be correlated with the main part of the Rueda Formation. It can probably also be correlated with the Entecada slates from the Valle de Arán (Kleinsmiede, 1960; Mey, 1967). Furthermore, it is equivalent to what Dalloni (1910, 1930) and Schmidt (1931) call the Coblentzian.

Basibé Formation. — The Gelada Formation is conformably overlain by a competent unit consisting of limestones, dolomites, and quartzites, which together form a single conspicuous ridge in the field. This valuable marker group is named after the Basibé massif (coord. 4°17′.42°32½′), the dominating mountain ridge of the Upper Isabena valley.

Definition. — The Basibé Formation is defined as the rock unit whose lower part consists of a mainly nodular carbonate, which may be dolomitic, followed by a quartzite member (0—50 m) overlain by a blackish, thinly-stratified limestone. The lower boundary of this formation with either the Gelada Formation or the Rueda Formation is sharp but conformable. The Basibé Formation is overlain by dark-coloured shales/slates of the Fonchanina Formation having either a sharp contact or showing a gradual transition.

Distribution and thickness. — The Basibé Formation, with its conspicuous quartzite member in the central part, shows a gradual change from west to east in the map-area (Fig. 10). It is thickest near the Liri fault zone in the west (Fig. 15), where this formation and the entire Baliera Facies ends abruptly. There (Barronco de Urmella), we measured a total thickness of 175 m and for the quartzite member 60 m. In an easterly direction the amount of quartzites and dolomites diminishes gradually along with the total thickness of the entire formation (Fig. 10). Fluctuations in thickness in a N-S direction are of minor importance. East of the Tor River, the Basibé Formation has lost its quartzite member entirely and consists of one limestone unit whose lower part is nodular and light-coloured and upper part regularly thin-bedded and dark-coloured.

The northern limit of the quartzite occurrence and the Baliere Facies now consists of thrust and fault structures. During the period of sedimentation there was probably a more gradual transition from the Baliera into the Sierra Negra Facies.

The exact southern boundary of the sand sedimentation is not known, because in the southern part of the map-area only Upper Devonian and Carboniferous deposits are exposed. In the Palaeozoic block in the Nogueras Zone, however, the quartzite member has already disappeared (Mey, 1967; Wennekers, 1968).

Lithology. — In its most typical development the lower member of the Basibé Formation consists of an evenly-bedded alternation of mainly nodular limestones, grey spatic limestones, and dolomites. The thickness of the individual beds varies between 15 and 30 cm. Between these competent beds, thin calc-schists may occur. The general weathering colour is yellowish-brown. The internal colour of the nodular limestones is white, cream, pink, greenish, grey, or light-brown. The dolomites are also light-coloured and have a sugar-like texture. They contain many white quartz veins which, as a result of selective erosion, project above the weathered surface and can cause serious damage to hand, knees, and clothing.

The cleavage, which cuts the bedding at angles varying in size, is well developed in the calc-schists and nodular limestones and less so in the dark limestones; in the dolomites it is usually absent. The cleavage tends to have an orientation more perpendicular to the bedding with increasing competence of the individual beds.

The fossils encountered comprised crinoid ossicles, corals (sometimes silicified), both small and large brachiopods, orthocerids, and stromatopores.

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. The thickness of the individual beds varies between 10 and 30 cm. The internal colour of the quartzites is also brownish, although somewhat lighter than the weathering colour. They never show cleavage.
Higher up, the number and thickness of the quartzites increases and the amount of dolomites diminishes until they disappear altogether in the upper part of this member, in which the quartzites are thick-bedded and massive (up to 10 m). They are never cross-bedded except in the Basibé massif. Thin quartzite layers may wedge out and become thicker within a rather short distance. The internal colour of the quartzites varies between bluish-white, grey, cream, and light-brown. Their weathering colour is light-grey. Overgrowth with yellow and green moss is seen frequently. The abundance of quartz veins in the quartzites helps to distinguish these sedimentary rocks, even at a distance, from the intrusive dykes, which also have a light-grey weathering and are overgrown with the same kind of moss.
Under the microscope, the quartzites are seen to consist almost exclusively of well-sorted quartz which shows a beautiful intergrowth of the individual grains (mosaic texture). The quartz grains often show pressure shadows. As accessories we found heavy minerals and few feldspar grains.
The upper member of the Basibé Formation consists of a series of blackish, thinly-stratified limestones with a total thickness of 20 to 50 m. Thin intercalations of dark-coloured calc-schists are frequently found in the upper part, although locally they may also occur in the middle and lower parts. The cleavage is rather well developed in both the calc-schists and limestones; its planes show a lustrous surface.
The weathering of the limestones is silvery-grey to slightly bluish. Sporadically, solitary corals and crinoid ossicles are observed. Due to their constant thickness (5—10 cm), loose slabs of these limestones produce a very uniform and rather high-frequency sound when hammered upon.
In the vicinity of the quartzite member, these thin-bedded limestones are very often dolomitized and then show brownish colours. This dolomitization suggests a relationship between the occurrence of the dolomites and the quartzites. Besides this dolomitization, which is limited to stratigraphic boundaries, a secondary dolomitization occurs locally; this is related not to the bedding but only to fracture zones. In such places the bedding remains a pronounced feature but the cleavage can no longer be distinguished. This secondary dolomitization therefore differs somewhat from the one occurring in the contact metamorphic aureole (page 165).

Depositional environment. — If the stratigraphic development from the Silurian black shales up to the Basibé Formation is regarded as a whole (Fig. 5b), it becomes evident that this sequence represents a regressive cycle. The upper boundary of this cycle is difficult to define, however, because the upper member of the Basibé Formation (dark-coloured limestones) and the following Fonchanina slates can be interpreted either as lagoonal deposits, still belonging to this regressive cycle, or as the first manifestation of a new transgressive cycle.
Concerning the origin of the sand wedge (Basibé quartzite) with its accompanying dolomites, in our opinion the only explanation to be considered is that of a shifting beach-barrier, since river deposits are never regular and are usually cross-bedded and badly sorted; the accompanying dolomites are certainly difficult to explain as fluvial deposits. On the other hand, a shifting beach-border of sand derived from a river delta somewhere near the southern end of the actual Liri fault zone (Fig. 15) would readily explain the very clean and well-sorted character of the quartzite, the lack of such sedimentary structures as cross-bedding, and the rather constant development over a relative large area. Moreover, the occurrence of dolomite fits rather well in such an environment. As has already been mentioned, the upper limestone member of the Basibé Formation could therefore be interpreted as a lagoonal deposit behind a bay-mouth bar or as the first sediment of a new marine cycle showing hardly any detrital sediments from the inundated continents.
If the hypothesis of a shifting beach-barrier is valid, it is evident that the time boundaries would run straight across the lithologic boundaries.

Correlation. — The occurrence of considerable sandy deposits in the middle part of the Devonian has so far only been reported from the Valle de Arán (Kleinsmiede, 1960). The reported turbidity sequence, with local thicknesses of up to 400 m, has a littoral facies in the west. The direction of transport is from west to east. It seems very likely that this turbidity sequence is the deeper marine equivalent of the Basibé quartzite, but we do not think that the basins on either side of the Maladeta granodiorite were directly connected during the period of sand sedimentation.

Age. — The few collected fossils could not be determined satisfactorily, but Dalloni (1910) mentions the following fossils from the Montañas de Denúy, where good exposures of the Basibé Formation are present:

Favosites reticulata, Blain.
Phacops occitanicus, Trom. Gras.
Jovellania, sp.

From the Galliro (Pico Gallinero), a pyramid-shaped mountain about 8 km southeast of Benasque where both the Gelada and Basibé Formations are well-exposed Schmidt, (1931) collected:

Machairacantus cf. *bohemicus,* Barr.
Phacops cf. *fecundus,* Barr.
Orthis striatula, Schl.

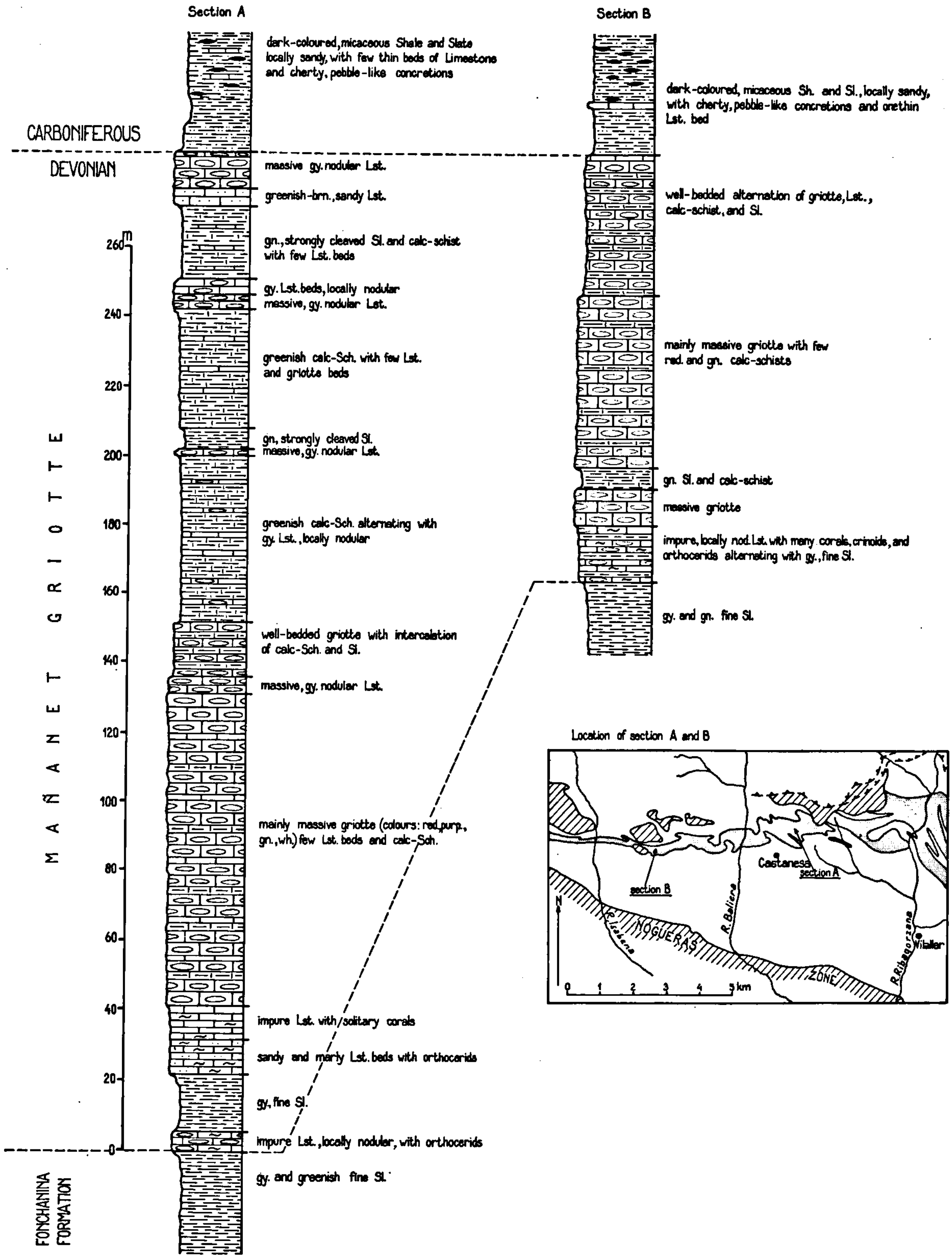


Fig. 11. Two lithostratigraphic sections of the Mañanet Griotte in the Baliera facies area.
Section A: Taken on the watershed between the Ribagorzana and Baliera rivers.
Section B: Taken on the watershed between the Baliera and Isabena rivers.

Orthis trigeri, Vern.
Orthis hamoni, Rouault.
Cyrtina heteroclyta, Defr.
Retzia andrieni, Vern.
Atrypa reticularis, L.
Zaphrentidae (three species)
Pleurodictyum cf. *selcanum,* Gieb.
Chaetetes, sp.

On the basis of this fossil assemblage he assigned a Middle Devonian age. Unfortunately, his description does not make it clear whether he collected these fossils from the upper part of the Gelada Formation or from the Basibé Formation. Both Dalloni (1910) and Schmidt (1931) assume on purely lithological grounds that the Basibé quartzite on the Pico Basibé is the base of the Carboniferous. Solé Sabaris (1954) does not mention the quartzite at all.

Fonchanina Formation. — This formation, 95 per cent of which consists of rather pure, light-grey to dark-grey slates, always forms topographical depressions owing to its incompetent nature, as for instance in the open valley of Fonchanina (coord. 4°20½'.42°30½') from which its name derives. Because of this property, mapping of this formation can be done reasonably well even in badly exposed areas.
For the definition and main characteristics of this formation, we refer to page 165. A distinctive component of this southern area is a horizon of evenly thin-bedded black limestones somewhere in the middle part of this slate formation. This limestone intercalation ranges in thickness from 5 to 10 m; it can only be mapped separately in some places.
In the slates the bedding is hardly ever discernable, so a rough estimation of their total thickness must suffice. The best guess, without taking the tectonical flattening into account, is not much over 60 m.
Except for some crinoid ossicles, no determinable fossils were found.

Mañanet Griotte. — For the definition and main characteristics of this formation, see page 166.

Thickness and lithology. — In the Baliera facies area this formation varies widely in composition and thickness. In the valleys of the Tor, Ribagorzana, and Baliera rivers, the griotte is very typical and reaches a thickness of up to 280 m. Further to the west, the typical griotte character diminishes together with the total thickness, which west of the Isabena is only some 80 m. There, the lower part of the sequence is very rich in solitary corals and crinoid stems. This decrease in thickness and facies change from east to west goes together with a decrease in the number of large folds. In the valleys of the Tor and Ribagorzana there are a number of folds measuring several hectometres on which smaller folds (measured in decimetres and metres) are superimposed. West of the Isabena, the griotte forms only a single overturned limb of a structure measuring some hectometres and having only few minor folds.
Owing to the increasing number of shale intercalations near the base and the top of this formation, its boundaries with the Fonchania slates and the Carboniferous slates are not always very sharp. In the field, however, a logical boundary can always be drawn within 3 to 6 m. The most characteristic nodular limestones occur mainly in the lower part of this griotte formation, the rest of the sequence being occupied by greenish and reddish marls, thin-bedded limestones, and, locally, some pure shales. Fig. 11 shows two typical sections of the Mañanet Griotte.
Apart from the above-mentioned corals and crinoid stems, we found many large orthocerids (up to 50 cm), mainly in the lower part of this formation. The goniatites so frequently found in the Northern Pyrenees were not observed in this region. We also sought for conodonts, but so far no reliable results can be reported. The solubility of these limestones in dilute acetid acid is very low, probably due to the internal cleavage structure. Samples of limestone from the Nogueras zone, where there is almost no cleavage, show a much better solubility and therefore a higher percentage of residual material, which implies a better chance of finding conodonts. The investigations of Ziegler (1959) and Van Adrichem Boogaert (internal reports) have demonstrated a rich conodont fauna in the griotte of the Nogueras Zone. In this zone the massive griotte below the chert horizon comprises the Frasnian, the Famennian, and locally also the Tournaisian.

CARBONIFEROUS

In both facies areas the Mañanet Griotte is conformably overlain by a mainly pelitic sequence of probably Carboniferous age. Bedded cherts or calcium phosphate nodules, which are known to represent the base of the Carboniferous in the Pyrenees, have unfortunately not been found here, although Solé Sabaris (1954) mentions about 20 to 30 m of green cherts at the base of his northern Carboniferous. Our own investigations showed, however, that his Carboniferous is undoubtedly metamorphic Silurian and his green cherts only highly metamorphic pelitic layers between green griotte nodules *("barrégienne")*. The Silurian and the Upper Devonian griottes are separated by an important fault (see geological map).
A faint indication that the lithological boundary between the nodular limestones below and the mainly pelitic sequence above represents more or less the limit between the Devonian and the Carboniferous, is given by the cherty, pebble-like concretions, varying in size from some centimetres to some decimetres, frequently seen in the lower part of these pelites.

Lithology. — The monotonous sequence of shales, siltstones, and impure sandstones without any continuous markers does not allow subdivision. The dull-toned colours range from greenish-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 beds. The bedding of the pelitic rocks is often hard to recognize; the cleavage is the most prominent feature. Individual cleavage planes are smooth and lustrous. Pencil-shales occur quite frequently.
The impure sandstones, often with a high lime content, are between 5 to 10 m thick. Due to their relatively higher competence they can be traced in the field, sometimes over considerable distances. The tectonic structure in this mainly pelitic sequence is quite complex, and it is therefore not known whether these hard markers correspond to different stratigraphic horizons or to a tectonic multiplication of the same sandstone.
Fossils are rare in these detrital sediments and are, moreover, heavily deformed by the tectonic flattening. Besides some solitary corals and small brachiopods, a very small trilobite was found, but no reliable determination was possible. Since we know, however, that the first post-tectonic conglomerate is of Middle Westphalian age (Aguiró, see below), the detrial sequence described above must be older.
The total thickness of this shale sandstone series is unknown, because it is unconformably overlain by Stephanian, Permian, or Triassic, but the exposed thickness reaches several hundred metres in places.

POST-HERCYNIAN SEDIMENTARY ROCKS

Introduction

In the central part of the Southern Pyrenees all post-Hercynian sedimentary rocks older than the Lias are almost entirely restricted to the Nogueras Zone, the oldest being found near the village of Aguiró. There, the folded Devonian is covered by a few metres of breccia and conglomeratic sandstones followed by carbonaceous shales and thin coal seams, on which rests a thick, coarse conglomerate. The coal seams have delivered a Middle to Upper Westphalian flora (Roussel, 1904; Dalloni, 1930). This outcrop near Aguiró is unique for the Pyrenees, because it is much more common for the first unconformable sediment to be of Stephanian, Permian, or Triassic age.
The Stephanian starts with a coarse conglomerate as in Aguiró or, as north of Erill-Castell, with a shaly weathered surface-deposit of older rocks in which cracks in the Devonian limestones are filled up with pebbles and similar rock types. These basal rocks are followed by the Stephanian volcanic formation: well-bedded tuffs with bombs and lapilli alternate with thin or thick andesite lava streams. Toward the top the Stephanian carries fresh-water cherts, conglomerates, sandstone beds, black shale beds, and coal seams containing a well-preserved flora of Lower Stephanian age (Dalloni, 1930).
South of Erill-Castell, the sedimentary Stephanian is followed in concordance by a mainly red series of mudstones, feldspatic sandstones with volcanic material, breccias, and irregular-bedded dolomitic limestones, which in its turn is covered unconformably by the basal conglomerate of the Buntsandstein Formation. Although this profile yielded no determinable fossils, the investigations of Vergili (1961) make it virtually certain that this red series between the Stephanian and the Buntsandstein Formation represents the Permian. This interpretation is strongly supported by Dalloni's (1930) and Schmidt's (1931) finding of an Autunien flora immediately below the basal conglomerate of the Buntsandstein in an outcrop of black carbonaceous shales located opposite the village Arcalis in the Pallaresa valley.
Owing to the lithological similarity to the Germanic Triassic, the same tripartite division into Buntsandstein, Muschelkalk, and Keuper is applied to the Triassic of the southern Pyrenees. In the latter area, however, these sediments have never yielded any determinable fossils, so that properly speaking we are using these terms as formation names.
The lower Triassic ("Buntsandstein") is developed in its classical continental facies of red mudstones, cross-bedded sandstones, and conglomerates. It rests unconformably on any of the following Palaeozoic rocks, including intrusive granites:

a. Cambro-Ordovician (SE of Sort in the Pallaresa valley)
b. Silurian (NE of Castejon de Sos in the Esera valley)
c. Devonian (near Compte in the Pallaresa valley)
d. pre-Hercynian Carboniferous (in the valleys of the Tor, Ribagorzana, and Baliera)
e. granite (N of Plan in the Cinqueta valley)
f. Stephanian (SW of Sort)
g. Permian (N of Malpas in the Peranera valley).

This coarse detrital series is followed by the typical "Muschelkalk", i.e. grey dolomite and cavernous dolomite *(cargneules)*, and the "Keuper" multicoloured marls, mudstones, and pure gypsum, carrying small and large bodies of basic rock (ophites).
In our Baliere area only the Triassic and a thin strip of probable Permian are present.

"Permian"

In the Ribagorzana valley, on the western bank about 100 m north of the Vilaller hydro-electric power station, between the basal conglomerate of the "Buntsandstein" and the pre-Hercynian Carboniferous, we found a well-exposed section of red, mainly detrital rock showing striking differences as compared to the "Buntsandstein" to the south of it (Fig. 12). The three most conspicuous properties, which are also among the characteristics of the "Permian" south of Erill-Castell (Virgili, 1961) but do not occur in the "Buntsandstein", are the following:

1. Frequently dolomitic layers (rare in the "Buntsandstein").
2. Breccias composed of angular fragments of quartz, chert, quartzite, hornfels, and slate (the "Buntsand-

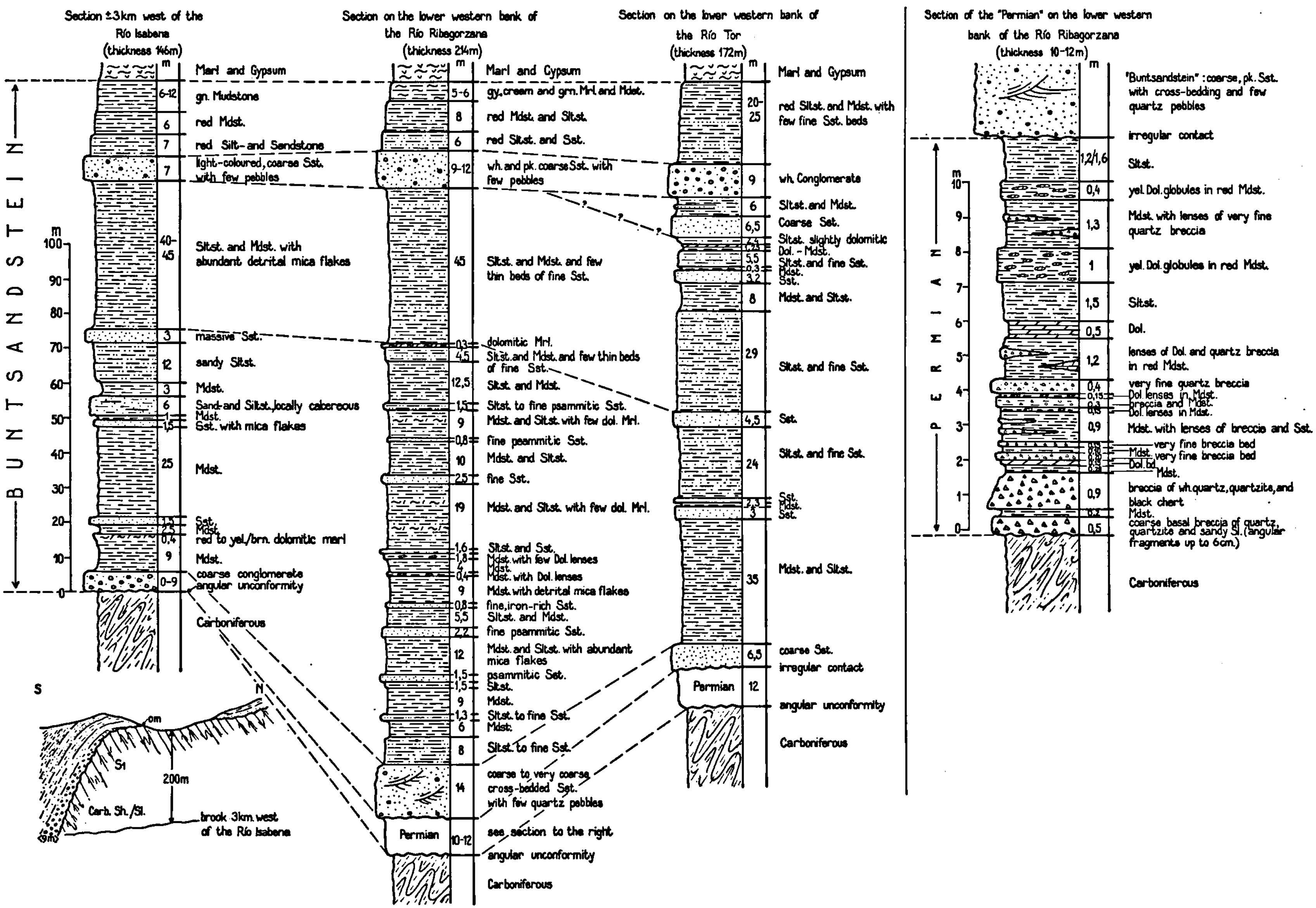


Fig. 12. Lithostratigraphic sections of the "Buntsandstein" and the "Permian", taken in the northern border of the Nogueras Zone.

stein" shows only conglomerates composed almost exclusively of well- rounded pebbles of white quartz and black chert).
3. Absence of large detrital mica flakes (abundant in the "Buntsandstein").
Because of these three properties and the slightly disconformable character of the "Buntsandstein" situated on top of this formation, this series is comparable with the "Permian" south of Erill-Castell in the Peranera valley. However, no biostratigraphical data are available to demonstrate its age.

Triassic

On the basis of their tectonic setting (Nogueras Zone, Axial Zone, and faulted facies boundary) and their lithological development, the Triassic exposures can be classified into the following three groups:

A. In the south, the northern limit of the Nogueras Zone, with a complete development of the Triassic.
B. The central synclines, with only the "Buntsandstein" present.
C. The northern Triassic, which is always related to thrust zones, represents a special facies of the probable transition from the "Muschelkalk" into the "Keuper".

A. *The Nogueras Zone.* — "Buntsandstein". — At the northern border of the Nogueras Zone, the "Buntsandstein" (together with a thin strip of probable Permian) forms one continuous band which, on our map, unconformably overlies the pre-Hercynian Carboniferous.
This formation is developed in its classical continental facies of red cross-bedded sandstones, quartz- and chert-pebble conglomerates, red mud-, and silstones. Many sedimentary structures, such as cross-bedding, wave and stream ripples, slumps, loadcasts, graded bedding, and other bottom structures, provide good evidence for determination of the top and bottom of the beds; this information is essential for solving the structures in the more complicated parts of the Nogueras Zone.
The base of this formation starts almost everywhere with a coarse-grained sandstone several metres thick and containing some small pebbles of white quartz and black chert. Only locally is a true basal conglomerate present (e.g. 2 km west of the Isabena; see Fig. 12).
The upper part of the "Buntsandstein" is also developed rather uniformly. On top of a competent, white to pinkish conglomerate layer, locally up to 12 m thick, there is a series (20—25 m) of black, red, or green mudstones containing veins of gypsum. The succeeding yellowish dolomitic marls are considered by some authors (e.g. P. Misch, 1934) to represent the Röt, the transition layers between the Buntsandstein and the Muschelkalk in Germany. When these typical top beds are missing, the reason may be tectonic in nature.
Apart from this constantly-developed base and upper part of the "Buntsandstein", the rest of this formation shows a rather arbitrary distribution of mudstones, siltstones, and sandstones (Fig. 12). The total thickness varies between 145 and 215 m in this area.
The red colour of the sediments presents a still-unsolved paleoclimatological problem. Arid deposition has been considered, but gives no explanation of their formation. Our observation that the intensity of the red colour is dependent on the grain-size (mudstones: deeply red; fine sandstones: red to pink; coarse sandstones and conglomerates: pinkish or uncoloured) may point in the direction of the conclusion of Van Houten (1948) that the red pigment (finely distributed haematite) derived from red soils of heavily-weathered limestone areas and was deposited mainly with the clay and silt fractions (low energy environment), being too fine-grained to settle in running water together with the larger grains of the sand-fraction (high energy environment).

"Muschelkalk" and "Keuper". — The incompetent character of this higher part of the Triassic undoubtedly explains the chaotic structure of the Nogueras Zone. From the "Buntsandstein" to the Lower Lias dolomite there is no continuous undisturbed profile. Almost invariably, slabs of the competent "Muschelkalk" float in a heavily distorted mass of vividly-coloured marls and gypsum. As a result, it is also impossible to establish whether all thick limestone bodies represent the same litho-stratigraphical horizon. In the present state of our knowledge it is therefore safer to group these two lithological units under one heading (Fig. 13).
The large, well-bedded limestone units, which in all probability represent the "Muschelkalk", weather to a light grey, but on fresh cuts they are blue-grey to black. Very often they are dolomitic. Near Pont de Suert, cross-bedded dolomitic limestones occur locally. Near the base and the top, the limestone mass gradually changes into yellowish, dolomitic marls which are sometimes slightly cavernous. In several undisturbed sections it can be seen that the "Muschelkalk" consists of two limestones separated by a few metres of more marly material. The total thickness, including the marl intercalation, can reach as much as 50 m. Very often, however, this marl intercalation is missing, but since this absence occurs in strongly disturbed areas, it may have a tectonic cause. The very fine grain with the sometimes millimetre-thin banding, the total lack of fossils, and the clear-cut fracturing along the bedding or joints, are the most important characteristics by which the "Muschelkalk" can be distinguished from any other limestone in the map-area.
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 gypsum member seldom contains salt and then only in very fresh outcrops, where a light salty taste betrays the presence of this

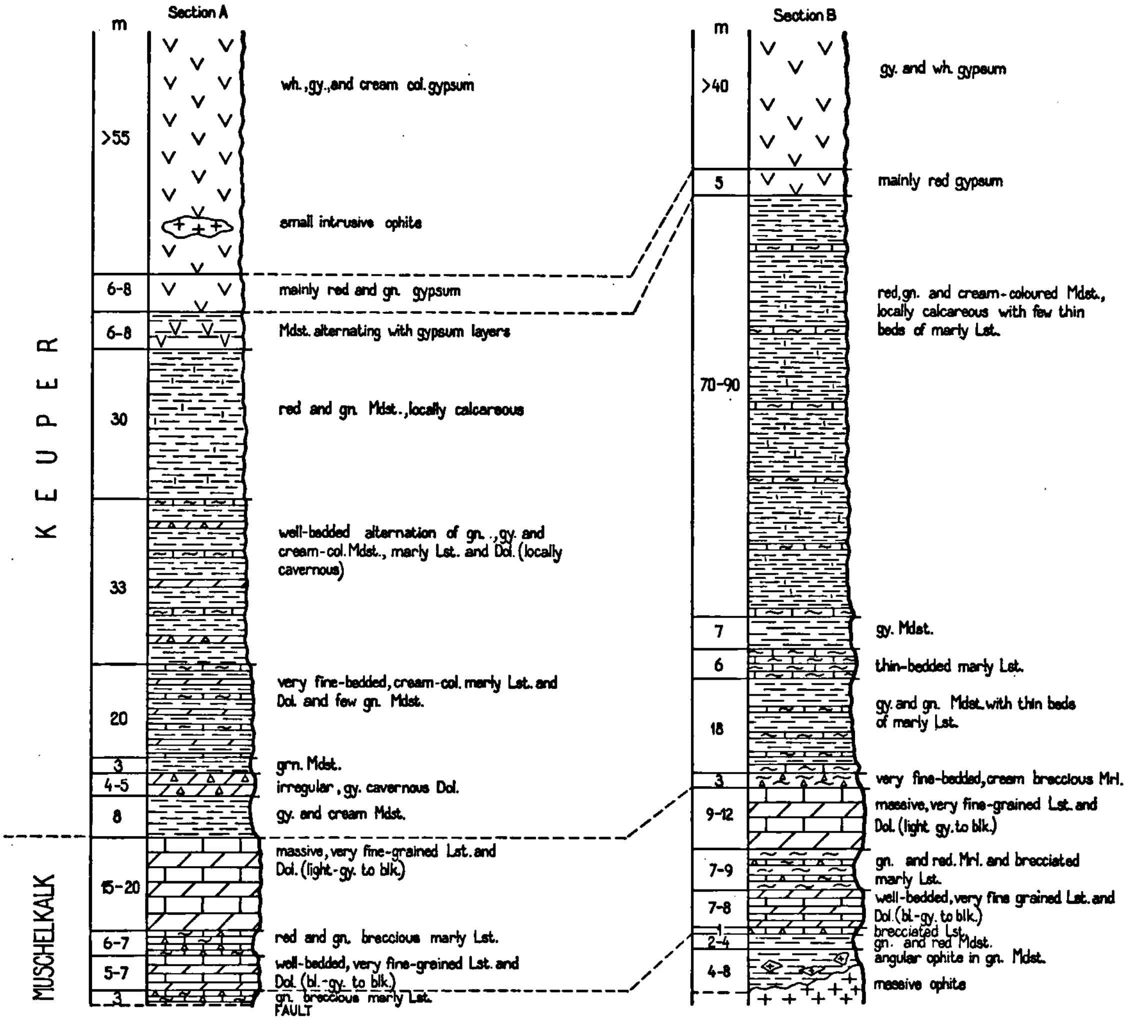


Fig. 13. Two sections of the "Muschelkalk-Keuper".
Section A: taken near the road tunnel 2 km south of Pont de Suert.
Section B: taken on the lower eastern bank of the Tor River, 800 m north of Pont de Suert.

mineral in the unweathered rock. In view of the salt springs of Gerri de la Sal, Salinas de Jaca, Salines de Sin, etc., salt must still occur in larger quantities at greater depths.

The transition into the dolomite of the Lower Lias is not present in the map-area, but according to the literature (Misch, 1934; Virgili, 1958) the upper part of the gypsum member contains yellowish cavernous dolomites *(Zellenkalk, cargneules, carneolas)*. In our area only drifting bodies of this so-called Zellenkalk have been encountered in the plastic gypsum-mudstone series. These bodies were too small to be included on the map, however.

The exposed thickness of the "Keuper" series exceeds 170 m. The two sections shown in Fig. 13 were both taken in continuous outcrops in the vicinity of Pont de Suert.

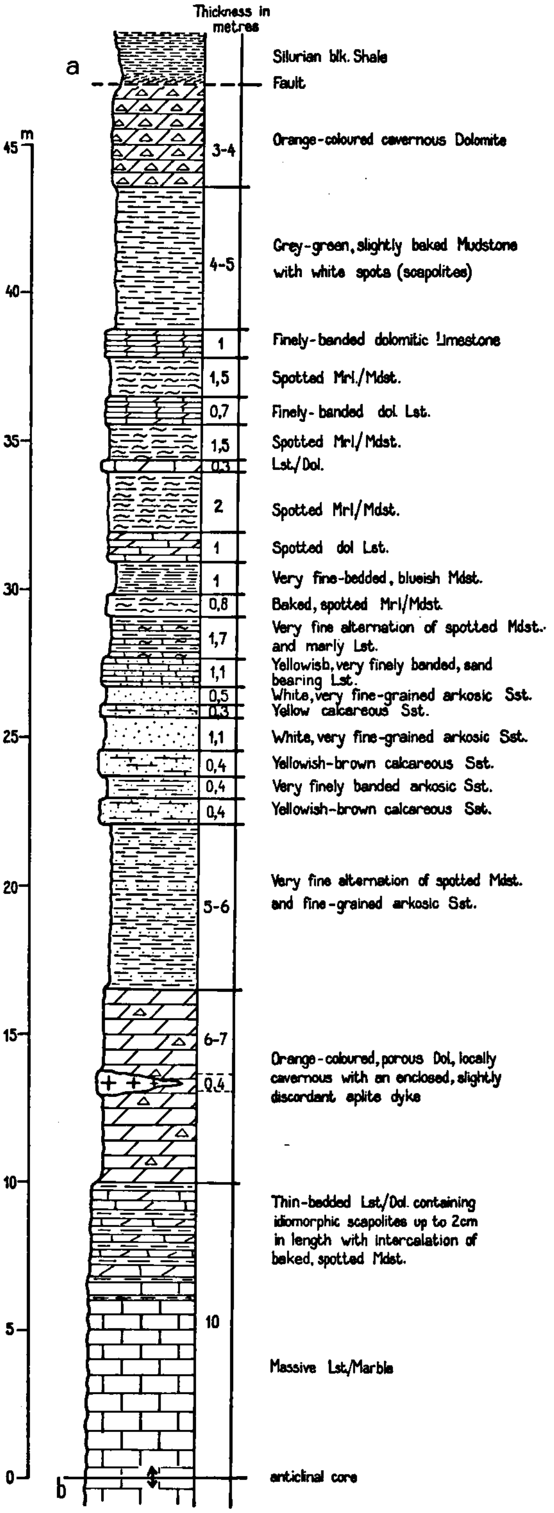
The matrix of the "Keuper" carries, besides the already-mentioned slabs of "Muschelkalk" and *Zellenkalk,* small bodies of basic rock consisting of plagioclase and clinopyroxene, with epidote and chlorite as alteration products. Owing to its ophitic texture, this basic rock is referred to in the French literature as ophite. The surrounding sediments show no signs of contact-metamorphism. These ophite bodies are dark-green to greenish-brown. A strongly-developed multi-directional joint system divides such ophite masses into numerous small polyeders, resulting in large screes at the base of the ophite. Very often, farms and hamlets are nestled close to the steep walls of ophites or steep Muschelkalk ridges.

B. *The central synclines.* — Here, only the "Buntsandstein" is present, with a thickness measuring about 70 m maximally. The basal layers of the Triassic are strongly dependent on the nature of the underlying Palaeozoic formations. The "Buntsandstein" starts:

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

At only one place (west of the Isabena, "Buntsandstein" on Fonchanina slates) was a 1 to 2 m thick conglomerate found at the base, persisting over only 300 m in the field. This dependence of the basal layers of the Triassic on the underlying rocks points to an originally more pronounced relief as compared with the Nogueras zone, where erosion and deposition during the Stephanian and Permian had already smoothed out the landscape before the deposition of the "Buntsandstein".
Above the basal layers, the "Buntsandstein" consists of a monotonous alternation of micaceous mudstones and siltstones as well as a few sandstones which, depending on their grain-size, are red, pink, or light-grey. The top and bottom of the sequence can be established only in the cross-bedded sandstones.
The synclinal structures of the central "Buntsandstein" occurrences are not deep enough to show the top of this detrital sequence. The two monoclines, one on each side of the Ribagorzana valley, which are overthrusted by Palaeozoic rocks, are probably also incomplete. However, southwest of Estét above the moraine deposits near the thrust plane we found some loose blocks of *Zellenkalk,* a strong indication that the overthrusted Palaeozoic mass had partly glided on the plastic layers of the younger Triassic.

C. *The northern zone.* — The Triassic remnants of the northern type are entirely restricted to the area of the Sierra Negra Facies. As can be seen on the geological map, these remnants are all flanked on both sides by faults, and it is therefore very difficult to reconstruct the original stratigraphic sequence.



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

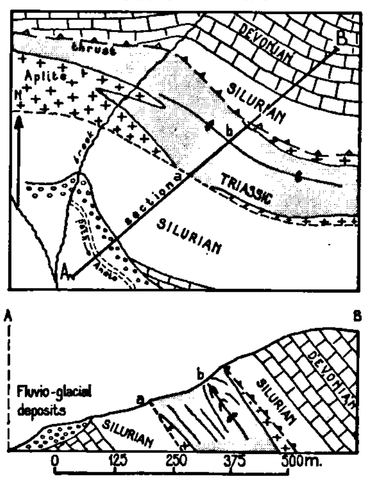


Fig. 14. Columnar section of the northern Triassic, taken 1800 m northwest of the village Aneto.

Although the shape of the large outcrop west of the Baliera River strongly suggests that there the Triassic rests unconformably on the griotte and the Carboniferous slates, we could not detect any obvious difference between the basal layers and the higher part of this Triassic outcrop. The sediments consist mainly of yellowish to orange breccias strongly resembling the cavernous dolomites *(cargneules)* encountered in the upper part of the Triassic in the Nogueras zone. Although the major part of these breccias is massive and without bedding traces, there are also patches (measuring as much as 2 to 3 m) of yellowish and sometimes grey dolomitic limestone showing an extremely fine banding. Besides this fine-bedded dolomitic limestone, we also found one massive layer (up to 4 m thick) of nearly white marble which contrasts strongly with the yellowish environment of the dolomitic breccias. Another outcrop, located near the brook coming down the southwestern side of the Pico del Home (coord. 4°25'.42°34') provided much better information about the stratigraphic sequence, however. Fig. 14 shows the detailed litho-stratigraphic section and its tectonic situation on a map and in cross-section (the section was taken in the overturned limb of the anticline). The scale of this section is ten times greater than the scale of the two "Muschelkalk"-"Keuper" sections (Fig. 13, page 177) and consequently much more detail could be recorded. But apart from this difference in scale, our northern section shows great resemblance to the transition from the "Muschelkalk" to the "Keuper".

The writer is of course fully aware of the fact that this correlation is based on purely lithological grounds and is therefore not entirely conclusive. However, the non-Palaeozoic character of these rocks combined with the fact that in Europe the Zellenkalk is known to occur exclusively in the upper part of the Triassic (in the Germanic facies), are strong arguments in favour of our hypothesis. One possible argument against a Mesozoic age for these rocks is the presence of aplitic dykes related to the outcrops of these rocks and the probably related scapolitization of the carbonates (see page 214). Scapolitization of Mesozoic calcareous sediments does occur locally in the northern Pyrenees, but not in relation to any acid rock intrusion. The latter region shows only basic intrusions, mainly lherzolites and ophites (Ravier, 1957; Avé Lallemant, 1967), which have not been found in our northern area.

Another, but less serious argument against a Mesozoic age of these carbonate rocks is the lack of "Buntsandstein" at the base of the first outcrop described above.

CHAPTER II

STRUCTURAL GEOLOGY

INTRODUCTION

The Pyrenean orogeny has been analysed by many authors. We will follow the general outlines given by de Sitter and Zwart (de Sitter 1956 a & b, 1965; Zwart 1959, 1960, 1963), incorporating our own observations.

After the epeirogenetic uplifts at the end of the Devonian, followed by a new transgression of the Dinantian, real folding started only after the deposition of the Lower Westphalian. Many successive folding phases can be distinguished but, unfortunately, so little biostratigraphic material is available for the Carboniferous that they cannot be dated separately.

In principle, there was a mainly NE-SW-striking concentric folding phase, most pronounced in the south, prior to the main, roughly E-W-striking folding phase that caused strong cleavage and in the Lower Palaeozoic locally also strong regional metamorphism (meso- and katazonal).

This main synkinematic phase was followed by a late kinematic granitization phase with still higher temperatures, accompanied by a NE-, NW- and NS-striking cross-folding and finally a new E-W-striking refolding phase. The intrusion of large and small plutons of probably palingenetic granites took place after the cross-folding, but partly before the E-W-striking refolding; these intrusive granites are therefore late-kinematic.

The present configuration of the Pyrenees is, however, due to Alpine phases. An uplift of the original Hercynian core developed along important fault zones, i.e. the north and south Pyrenean faults, probably starting immediately after the Hercynian folding but expressed most strongly during the Jurassic and Lower Cretaceous in the formation of marginal basins in two external zones on the borders of the axial zone. A pre-Cenomanian folding phase intensified the movement along these border faults of the axial zone, and started the development of the Upper Cretaceous-Eocene marginal basins outward from the Lower Cretaceous basins. Strong folding and uplift occurred in late Eocene times, followed by a strong subsidence of the Ebro block further to the south.

The Alpine tectonics are often of the overthrust type, and occasionally penetrate quite deeply into the Hercynian axial zone. There was probably more activity in the south and west than in the north and east. Due to these overthrusts and the pronounced uplift

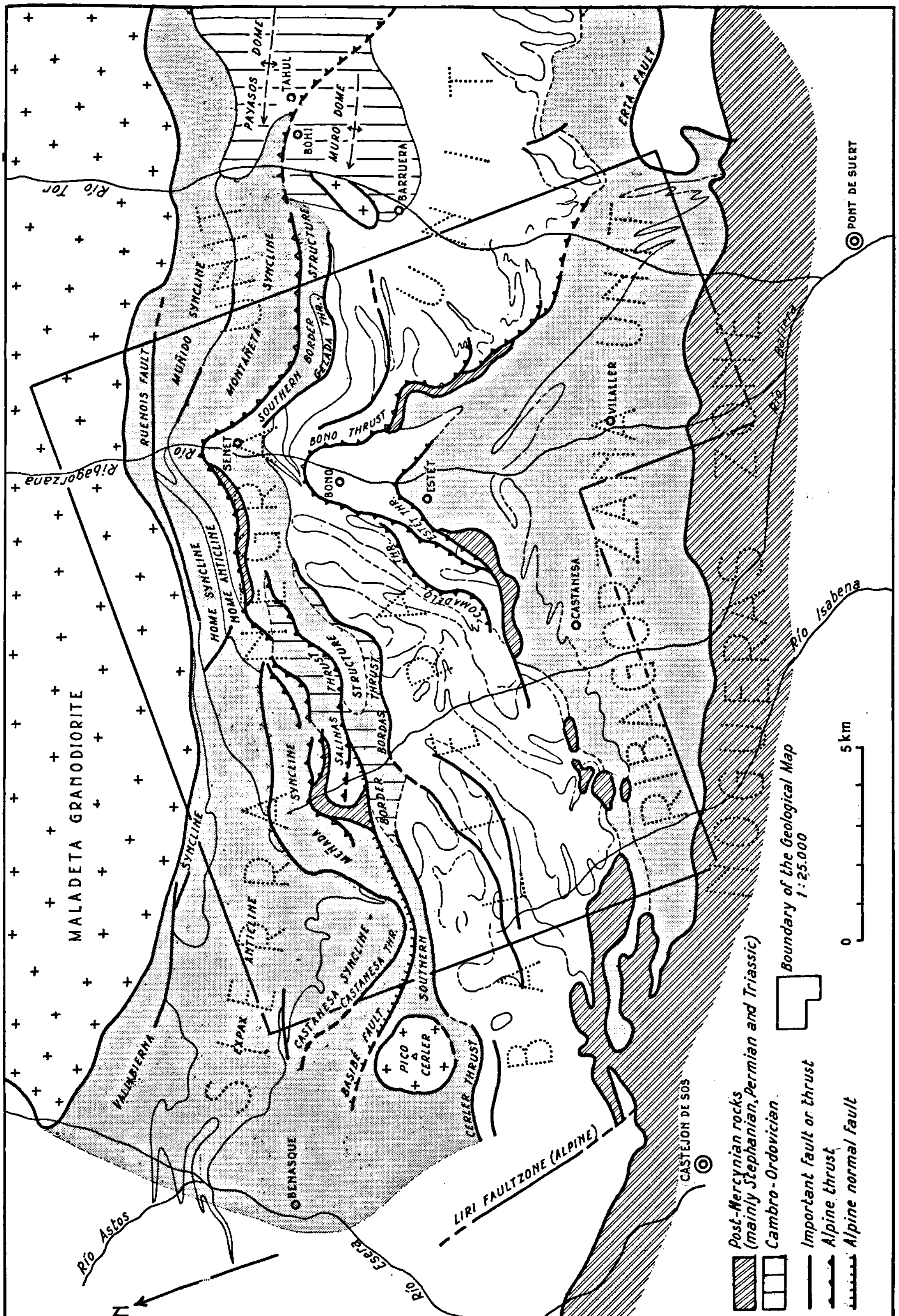


Fig. 15. Structural sketch-map of the Axial Zone of the Pyrenees south of the Maladeta granodiorite.

of the axial zone, gravity gliding also occurred locally in the south. Finally, a morphogenetic uplift in late Miocene time formed the present Pyrenean mountain chain.

The structural mapping of our area revealed only the most important of the above-mentioned folding phases. The pre-cleavage folding, which was first observed by Boschma (1963), is very pronounced in the central part of the area. The main folding, with the development of tight to nearly isoclinal cleavage folds, is clearly expressed in the entire area. Both dip and strike vary greatly from one place to another, however. The grade of metamorphism is very low (epizonal). The intensity of the cleavage, depending mainly on the degree of deformation and metamorphism and to a lesser extent on the mica content of the rock, is of the true slaty cleavage type except in the southern Carboniferous, where a coarser cleavage (incomplete slaty cleavage to fracture cleavage) dominates.

Of the refolding we could detect, in a rather limited area (Castanesa massif), only one pronounced direction (NW-SE); its shape and time relation with the intrusion of the granite and dykes suggest that this folding of the cleavage represents the late Hercynian E-W refolding.

The Alpine tectonics are rather strongly expressed in this area. In addition to several important overthrusts with a southward-directed movement of at least several hundred metres, a fracture cleavage also developed in the lower part of the "Buntsandstein". The scapolitization of the northern Triassic probably dates from the same time.

The two facies areas (Sierra Negra area and Baliera area) described in Chapter I are clearly identifiable by reason of their individual structural styles.

The Sierra Negra facies area is characterized by strong isoclinal folding with a rather steep, northward-dipping slaty cleavage that has obliterated almost all traces of former pre-cleavage structures. This area, including the E-W-striking belt of Aneto slates and the Silurian and Ordovician on the threshold of both facies areas, we shall call the *Sierra Negra Unit*.

The Baliera facies area, on the other hand is dominated by large pre-cleavage structures (NE- and NW-striking, asymmetric concentric folds) which were flattened and refolded by the main phase (cleavage folding). Since there is an obvious disharmony in folding between the Gelada, Basibé, and Fonchanina Formations on the one hand and the griotte and Carboniferous on the other hand (the Fonchanina slates acting as a lubricating intermediary), this area can be subdivided into two units:

The Baliera Unit (with large, mainly pre-main phase folds of the Gelada, Basibé, and Fonchanina Formations).

The Ribagorzana Unit (consisting of folded, E-W- and NW-SE-striking griotte and Carboniferous).

Each of these three structural units can be subdivided into large anticlines and synclines, which in their turn show many minor structures.

Symbols and nomenclature

The following symbols are used in this paper:

ss	-bedding plane
a.p.	-axial plane of the concentric pre-main phase folds
s_1	-main phase slaty cleavage
s_2	-second cleavage (either crenulation or fracture cleavage)
S_A	-coarse fracture cleavage in Lower Triassic rocks
π-ss	-pole of bedding plane
π-$s_{(1,2)}$	-pole of first and second cleavage planes
δ	-intersection of ss with s_1

The poles to the bedding planes and to the first and second cleavage planes, fold axes, and intersection lines, have been plotted on Schmidt's equal area stereographic nets. The diagrams are lower hemisphere projections. The numbers of points bear no relation to the intensity of the feature involved.

Our description of folds is based on the classification proposed by Fleuty (1964); for the different types of cleavages we followed the classification of Knill (1960).

Under the term "minor structure" is understood small structures too large for microscopical observation but too small to be indicated on a geological map.

THE SIERRA NEGRA UNIT

General remarks

This structural unit, which lies just to the south of the Maladeta granodiorite (Fig. 15), is characterized by isoclinal folding with a well-developed axial-plane cleavage (s_1) dipping to the north. The intrusion of this granodiorite formed a metamorphic aureole between 1.5 and 2.5 km wide, in which hornfelses, marbles, and spotted slates 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 to the east of the Ribagorzana, this southern boundary is less distinct, because the flattening caused by the cleavage becomes gradually less important in that direction. The most logical southern limit is the Gelada thrust, since the two patches of Cambro-Ordovician in the fault zone suggest a connection with a deeper-seated Cambro-Ordovician structure, which makes the situation comparable to the southern limit of the Sierra Negra Unit east of the Baliera River.

East of the Tor River, the function of the Sierra Negra Unit as an area of strong flattening due to isoclinal folding and slaty cleavage, is taken over by the Cambro-Ordovician Payasos Dome and the smaller Muro Dome. Structural details of both domes have been worked out by Boschma (1963, pp. 147—162). To the west of the mapped area this structural unit widens. For further detail we refer to Wennekers (1968).

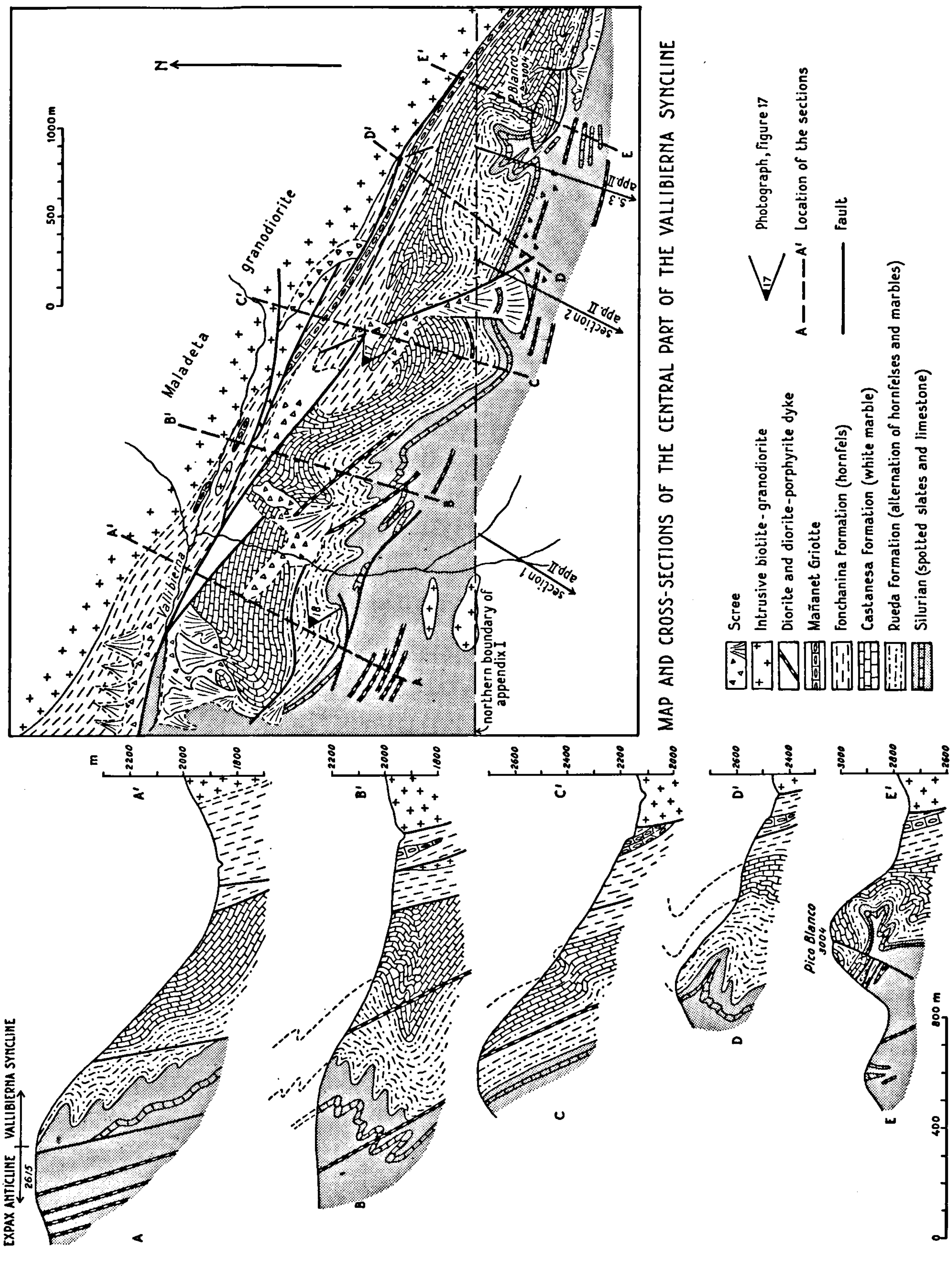


Fig. 16. Map and cross-sections of the central part of the Vallibierna Syncline.

The change in dip and strike of the main cleavage (s_1) and the bedding (ss) as well as the distribution of fold axes and cleavage-bedding lineations (δ-lineations) are shown in Fig. 19 and Appendix III (A_1—A_6). The swing in the strike of the mean cleavage follows that of the entire structure as well as the 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, more pronounced swing is present far to the south, in the Baliera and Ribagorzana units, beyond the limits of the "shouldering-aside" effect of the Maladeta batholith. This swing is therefore either an original tectonic feature, caused by a local divergent stress field during the main folding phase, or a post-main-phase (but pre-intrusion) phenomenon, possibly related to the diagonal cross-folding (NE-SW and NW-SE) so well known for many Cambro-Ordovician domes (Boschma, 1963; Zwart, 1963, 1965; Lapré, 1965; Oele, 1966).

With respect to the considerable change in the northward dip of the cleavage, which diminishes from about 80° in the north to some 45° in the south and is locally even less, we cannot at present determine whether this is an original feature or a later warping of the cleavage. However, if these flatter dips were caused by warping, this must have happened before the intrusion of the granodiorite since this batholith intruded into an already north-dipping structure, as shown by the north-dipping (65—80°) metamorphosed rocks and the contact with the granodiorite parallel to them.

The tight to isoclinal folds mapped in this area (Appendices I and II) yield typical orientation diagrams for this type of folding, with the modal cleavage and bedding planes coincident (Fig. 19, A_0-A_6). The minor fold axes observed in the field clearly have the cleavage parallel to their axial planes and consequently the δ-lineations parallel to the axes. This relationship is also clear in the diagrams (Fig. 19, A_0-A_6), which show the measured lineations and axes lying in complete girdles parallel to the modal cleavage plane. The northward-directed fold axes and δ-lineations suggest, however, that the folding preceding the cleavage development had a slightly different strike.

Folding of the cleavage (s_1) with development of a secondary crenulation cleavage (s_2) has been found locally. This 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 near the village of Aneto (coord. 4°26′. 42°33′) and its continuation to the west.

The Alpine tectonics are rather strongly developed and are represented by important overthrusts (Salinas and Senet thrusts) and one normal fault (Basibé fault).

The Sierra Negra Unit has been subdivided into a number of large anticlines and synclines, some of them separated by important thrusts or faults. These are, going from north to south:

W ← → E

N ↑

Sierra Negra Unit

Maladeta batholith

Vallibierna Syncline

Expax Anticline

Ruenois fault

Castanesa Syncline – Home Syncl. – Muñido Syncl.

Cast. thrust – Home Anticl. – (C.-Ord. Payasos Dome)

Meñada Syncline – Montañeta Syncline

Basibé fault – Salinas thrust – Senet thrust

Southern Border Structure

S

These structures will be described separately.

The Vallibierna Syncline

Although this structure, which is situated southwest of the Maladeta batholith and falls mainly outside the mapped area, has already been described by Kleinsmiede (1960, pp. 167—168), a short revision seems warranted, since the metamorphic rocks north of the Vallibierna River (sheet 4) are not of Silurian but of Upper (?) Devonian age, with a more or less synclinal shape. The so-called "Southern Anticline" of Kleinsmiede is therefore restricted to the Expax Anticline, which in the north is limited by our Vallibierna structure, several sections of which are shown in Fig. 16 and Appendix II (sections 3, 4, 5, and 6). This latter structure, which consists of metamorphic Lower, Middle, and probably even Upper Devonian, is, taken broadly, a simple syncline whose northern flank is partly truncated by the granodiorite. Both the bedding and the cleavage — the latter being visible only in the baked marly limestones — have a rather steep N to NE dip (70—80°). Smaller parasitic folds of the main phase are rare, as are the δ-lineations which at best are only poorly visible and were therefore not measured.

The contact with the Maladeta batholith, which in the central part is sometimes faulted, runs roughly parallel to the dip and strike of the surrounding metamorphosed rocks, thus being slightly overturned. This post-tectonic granodiorite intruded into a pre-existing structure, the process being facilitated by the pronounced, steep, and regular cleavage planes. Some of the steep faults near the contact might have orignated during this intrusion (faults parallel to the granodiorite contact, with a northern upthrow) but others certainly developed later (i.e. those entering the batholith with a displacement of its contact).

A phenomenon which is certainly related to the intrusion of the batholith is the rather irregular folding with a nearly horizontal axial plane, occurring locally in the higher exposed parts of the Lower and Middle Devonian (Fig. 16, sections A—E and Fig. 17). Since the flank of a main-phase fold was folded together with its cleavage, this piling-up of limestones against the granodiorite (probably accompanied by local gliding), must have developed later than the Hercynian main phase, very probably during the intrusion of the granodiorite. The same kind of "shouldering-aside" effect occurs locally, also north of the Maladeta batholith, e.g. the recumbent fold of the Pico de Viella

Fig. 17. Cascade folds in the southern flank of the Vallibierna Syncline.

(Kleinsmiede, 1960, pp. 183 and 188). Profile B and Fig. 18 show that these recumbent folds are traversed by steeply dipping intrusive dykes; these dykes have not been folded, and therefore intruded later than the piling-up of the cascade folds; consequently they are also of later date than the main intrusion of the granodiorite. Folding subsequent to the intrusion of these dykes has not been found here.

Finally, it should be noted that the granodiorite contact as well as the general dip of the main structure (apart from the local recumbent folds) dip steeply (70—80°) north to northeast. It is highly improbable that this dip could be due to later tilting caused by the upheaval of a northern block with internal gliding along "Schieferungs parallele Abschiebungsflächen" (principle advanced by Hoeppener, 1955), since the granodiorite and its aureole of hornfels and marble form one rigid mass in which hardly any internal gliding movement would be possible. Therefore, we are convinced that the general dip of 70 to 80° of cleavage and bedding in the contact aureole of the southern border of the Maladeta batholith existed before the intrusion of the granodiorite and was not the result of later tilting as has been proposed by Zandvliet (1960, pp. 100—102).

The Expax Anticline

The Expax Anticline, exposing exclusively Silurian spotted slates, is the largest continuous structure of the Sierra Negra Unit. In the Sierra Negra area, where this structure is widest (up to 5 km in cross-section), its northern flank is steep (70—80°) and dips northeast

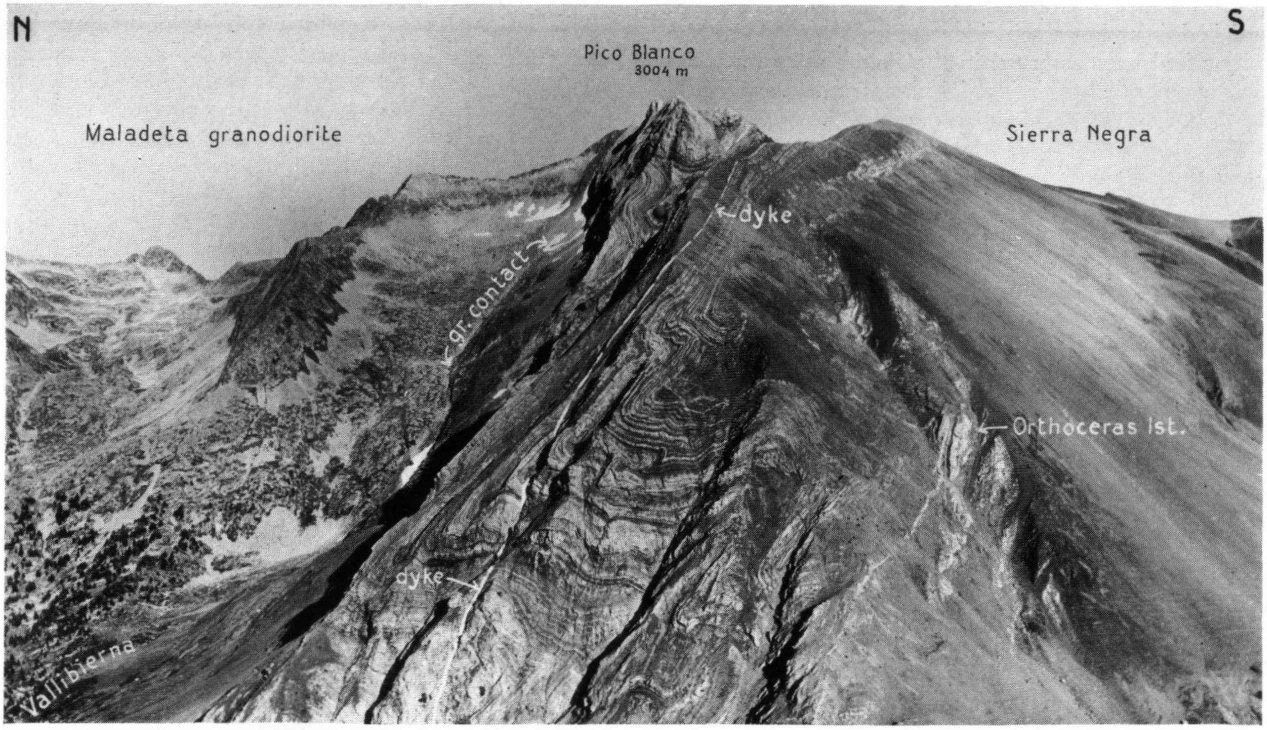


Fig. 18. Cascade folds in the southern flank of the Vallibierna Syncline traversed by an unfolded intrusive dyke.

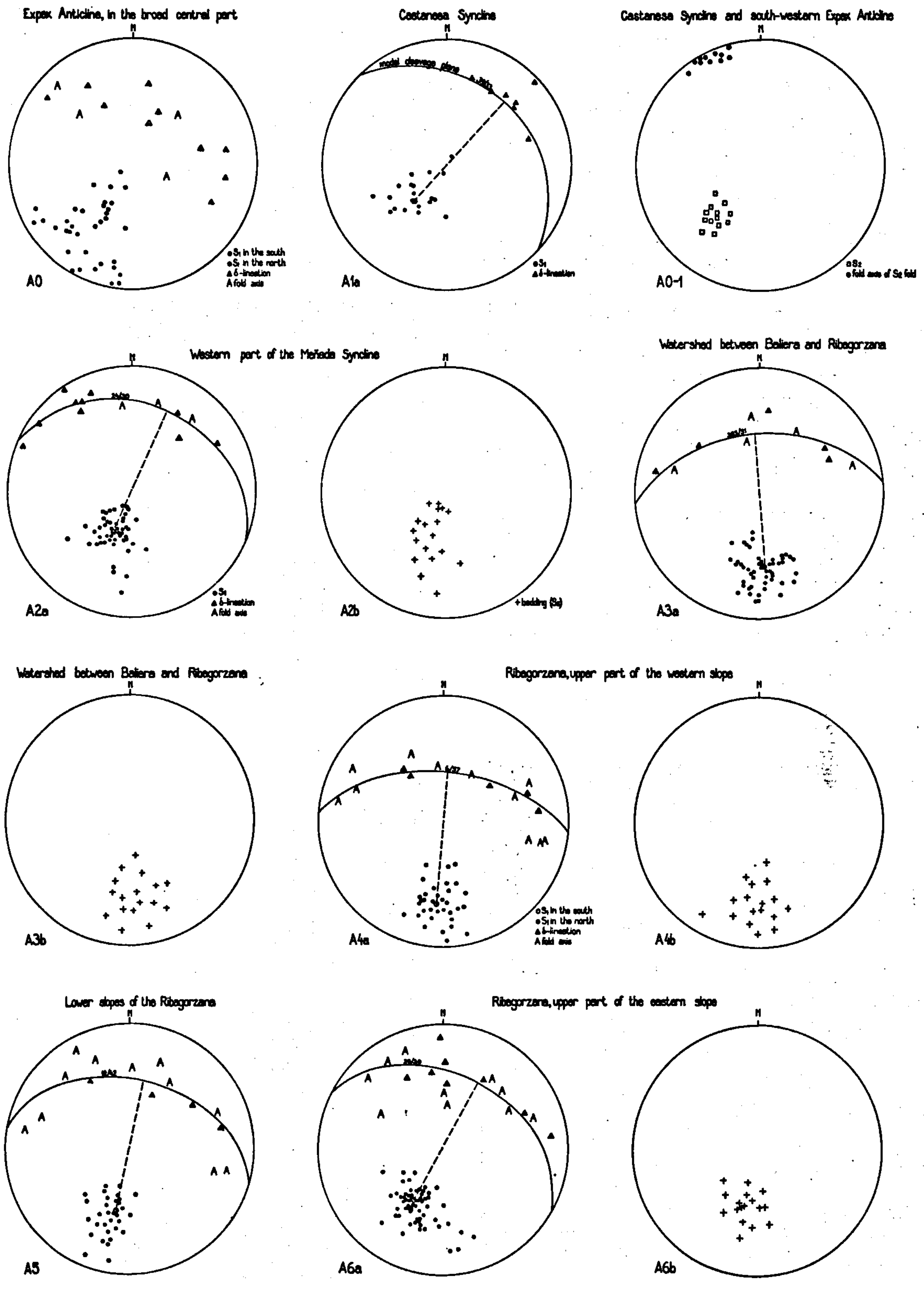


Fig. 19. Orientation diagrams of ss, s1, s2, δ-lineations, and fold axes of the Sierra Negra Unit. The numbers of the diagrams correspond to the numbers of the subareas indicated in Appendix III.

into the Vallibierna Syncline, and its southern flank is strongly overturned (15—35°), also dipping northeast into the Castanesa Syncline. In the west, near the Esera River, the Expax Anticline splits up into three or four smaller anticlines, all with axial planes dipping moderately steeply (40—60°) north to northeast. East of the Sierra de Llausét (coord. 4°20'.42°35') the anticline becomes tight to isoclinal with a steep (75—80°) northward-dipping axial plane. Its southern flank is there formed by the Home Syncline, often with a faulted contact, and east of the Lago Llausét the Vallibierna Syncline in the north is truncated by the Maladeta granodiorite. This vast batholith has a slightly overturned (75—85°) and regular contact west of the Ribagorzana, but east of this river its southern limit is rather irregular and locally faulted. In this same narrow eastern part, the nearly vertical Ruenois fault, with a northern upthrow, forms the southern limit of the Expax Anticline, which near the Tor is truncated by the Maladeta granodiorite.

Since there are no distinct mappable markers in the lower part of the Silurian, we cannot form a satisfactory idea of the actual folding style of this asymmetric dome. The only guide structures *("Leitstrukturen"*, Schwan, 1964) are small asymmetric cleavage folds in the thin-bedded Upper Silurian limestones, but unfortunately these occur only near the Devonian borders of the Expax Anticline. When the bedding is examined in thin-sections it shows very tight folding, indicating intense flattening of the rocks.

The distribution of fold axes and δ-lineations as well as the poles to the main cleavage (π-s_1) are shown in Fig. 19, diagram A_0. The rather scattered poles of lineations and fold axes suggest the presence of a folding phase prior to cleavage development.

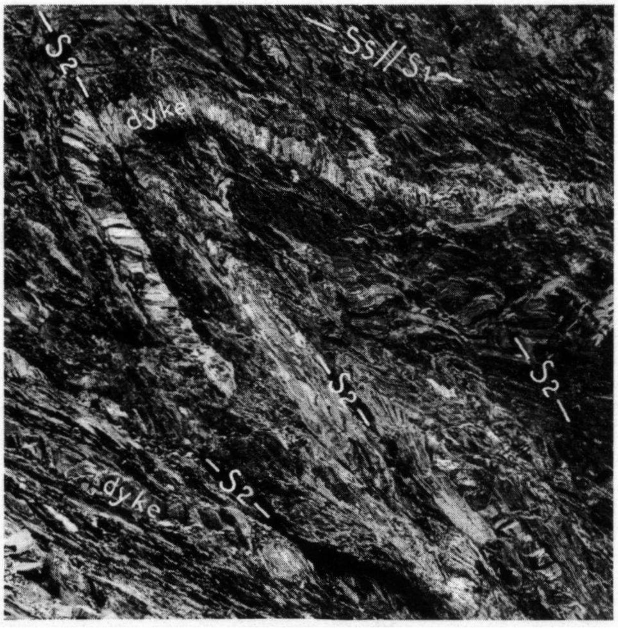


Fig. 20. Folded intrusive dyke with fracture cleavage (s_2) in the Upper Silurian limestone/slate sequence of the Castanesa massif.

In the broadest part of the Expax Anticline its southern flank shows rather strong folding of the flat, north-dipping main cleavage, causing the spread of the s_1-poles. Its fold axes have a NW-SE strike and are horizontal to slightly NW-dipping (up to 10°). The abundant intrusive dykes in this area, which in general intruded along the main cleavage plane, are also folded (Fig. 20); they are therefore younger than the main-phase cleavage but older than the refolding, which originated a new cleavage (s_2).

Since this s_2-folding occurred after the intrusion of the dykes and consequently also after the intrusion of the granodiorite stock, it must be late Hercynian (E-W refolding) or even younger (Alpine?). In the field we might call this second cleavage a fracture cleavage,

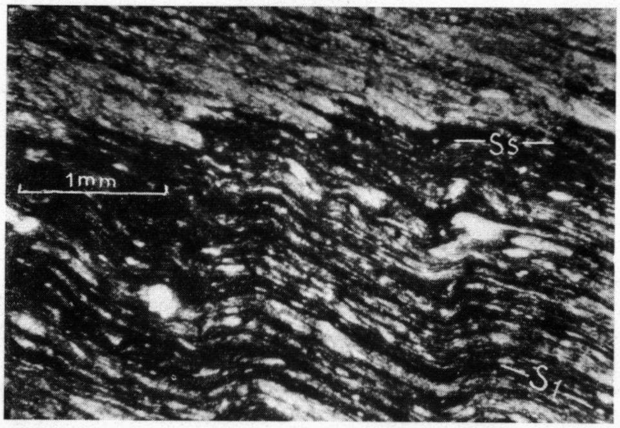


Fig. 21. Development of a crenulation cleavage (s_2) in a layered sequence (ss) with an oblique main-phase cleavage (s_1).

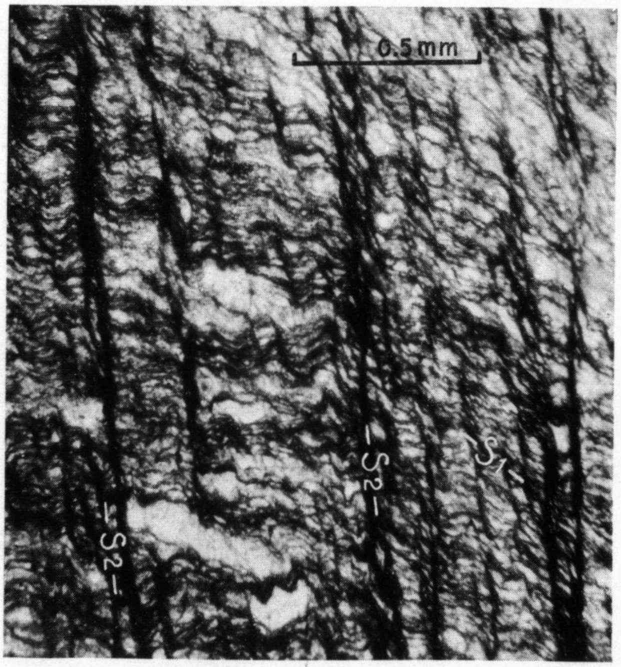


Fig. 22. Detail of thin-section, another part of which is shown in Fig. 21.

but under the microscope it is found to be a well-developed crenulation cleavage (Figs. 21 and 22).

The strike and dip of this s_2-cleavage is more or less parallel to the general strike and dip of the unfolded main phase cleavage (s_1) in this area (Fig. 19, diagrams A_0-A_{0-1}). From this it may be concluded that during both folding phases the stress field had roughly the same direction. Folding of the main cleavage could only occur where the s_1-cleavage had a different dip, in our case lower than those of the s_2-cleavage. We have, however, no evidence indicating whether this local flatter dip of the s_1-cleavage originated in this way or was caused by tilting before the time of refolding.

In the rest of the area the compression during the time of refolding resulted in a further flattening along the existing s_1-plane, accompanied by local(?) gliding along the same plane (shear cleavage with a northern upthrow). Evidence of this is provided by the often heavily deformed and rotated chiastolite crystals in the metamorphic Silurian slates (Figs. 23 and 24).

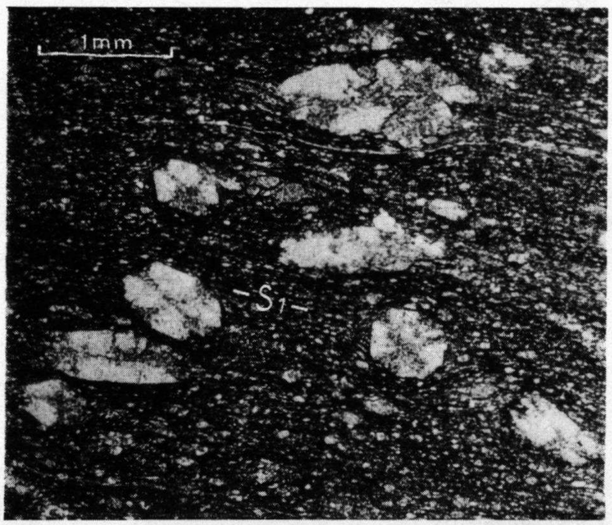


Fig. 23. Deformed chiastolite crystals in Silurian spotted slate.

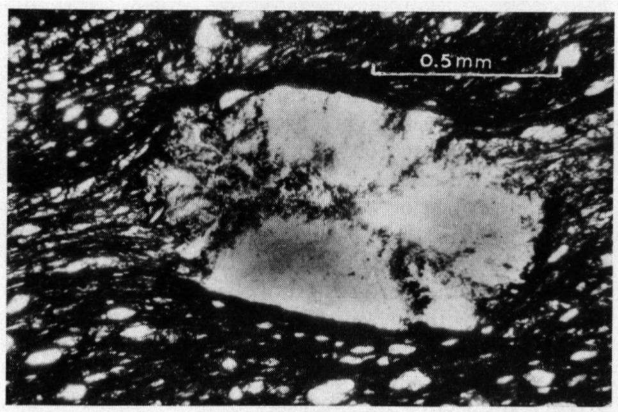


Fig. 24. Deformed and slightly rotated chiastolite crystal in Silurian spotted slate.

The Castanesa Syncline

This is a nearly horizontal to slightly northward-dipping isoclinal structure, mainly Lower and Middle Devonian, located southwest of the Expax Anticline. This structure has not been influenced by contact metamorphism. Its southern and eastern limits are formed by the flat, northward-dipping Castanesa thrust. Its extension in the westward direction will be described by Wennekers (1968). The general strike of this syncline is NW-SE, i.e. still more or less parallel to the southwestern border of the Maladeta batholith.

The isoclinal character of this structure (Fig. 7 and Appendix II, section 1), which can be distinguished from a distance, is in detail strongly refolded and intersected by many northward-dipping faults running parallel to the s_2-cleavage. Since the bedding as well as the first and second cleavages are clearly visible, the effect of the refolding can be studied much better here than in the southern flank of the Expax Anticline.

The distribution of s_1, s_2, δ-lineations, and fold axes is shown in Fig. 19, diagrams A_{1a} and A_{0-1}. The NE-dipping δ-lineations of the main phase were probably caused by a folding previous to the cleavage development.

The s_2-cleavage presents itself in the field as a rather coarse fracture cleavage very often accompanied by indications of a southward thrust movement of a few millimetres up to several centimetres per macroscopically visible plane (Fig. 25). A concentration of these planes results in real faults or thrusts, as can be seen on the Pico Castanesa (Fig. 7 and Appendix II, section 1). Microscopically, the s_2-cleavage appears as a crenulation cleavage.

Fig. 26 shows a reconstruction of the successive stages

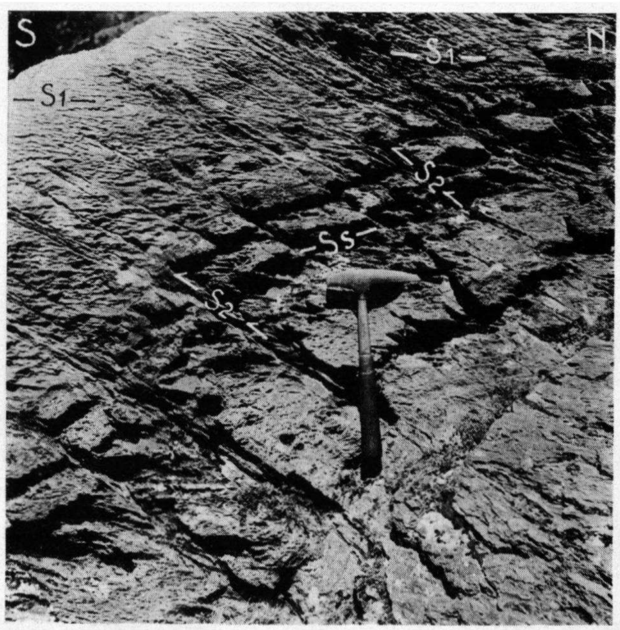


Fig. 25. Pronounced planes of the s_2-cleavage showing signs of accompanying thrust movements (northern upthrow).

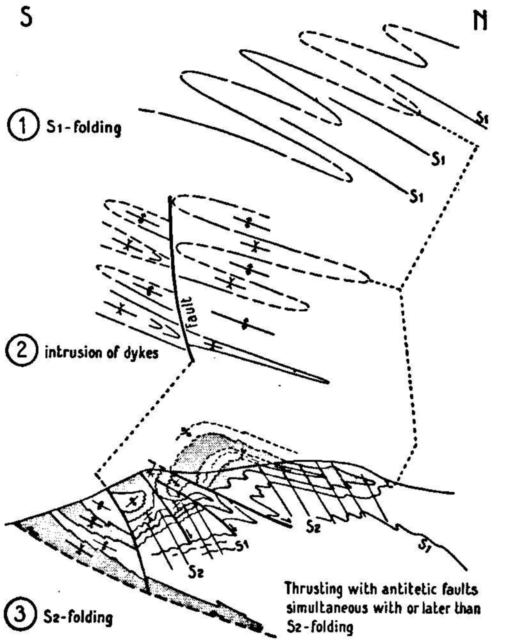


Fig. 26. Schematic representation of the structural history of the Castanesa Syncline.

of folding and faulting of the Castanesa massif. We have, however, no decisive evidence indicating whether the thrusting along the Castanesa thrust, which limits the area of s_2-folding in the south and east, caused this refolding and the accompanying smaller faults, or whether this thrust movement occurred later and independently of the refolding (Alpine?). The only certainty is that this refolding is a very late phenomenon, since the intrusive dykes are also folded (Figs. 20 and 27).

Fig. 27. Folded intrusive dyke in the Rueda Formation (Castanesa massif).

The Home Syncline

This isoclinal structure probably represents the eastern continuation of the Castanesa Syncline, but with the difference that the Home Syncline is steeper, shows no refolding, and the Upper Devonian Griotte is also exposed. Furthermore, the structure shows a rather strong influence of contact metamorphism. It is situated between the narrow part of the Expax Anticline in the north and the Home Anticline in the south. Its southwestern limit is formed by the Castanesa thrust. In the northeast, this syncline is truncated by the nearly vertical Ruenois fault.

The main phase is characterized by large isoclinal folds of hectometre dimensions (Appendix II, sections 2—9), which are accompanied by smaller folds with dimensions measured in decimetres to metres.

Beautiful examples of these smaller parasitic folds are exposed, for example, on the lower part of the western bank of the Ribagorzana valley. For the variation in cleavage and bedding and the distribution of fold axes and δ-lineations, see Fig. 19, diagrams A_3-A_4, and Appendix III, diagrams A_3 and A_4.

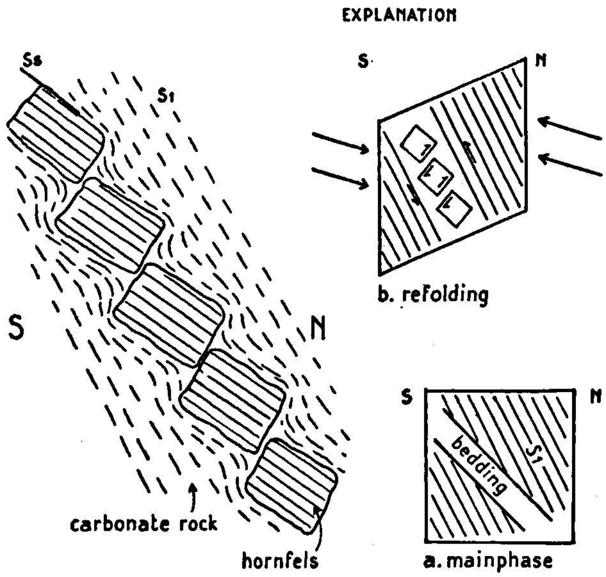


Fig. 28. Rotated hornfels chips in a plastically deformed carbonate matrix.

Although we found no folds due to refolding or a secondary cleavage, there must have been a time of compression simultaneous with or subsequent to the intrusion of the granodiorite, as shown by rotated cubic pyrite crystals and multiply-broken and rotated hornfels beds in a recrystallized carbonate matrix (Fig. 28). As in the Castanesa Syncline and the Expax Anticline, this rotation points to a relative upward movement of the northern part.

The significance of the rather thick (1—5 m) quartz veins west of the Rio Llausét is not clear. The easternmost of these quartz veins coincides with a fault, but the others seem to cross the Fonchanina and Castanesa Formations without any sign of displacement.

An interesting feature, not distinguishable on the cross-sections, is the NNW-SSE-directed vertical wrench-fault entering the Lago Llausét (coord. 4°23′. 42°35′) from the south. This fault undoubtedly originated during the main folding process, because the folding is not quite the same on both sides of the fault: east of the fault the structure is more isoclinal than in the west and lies deeper, since the Upper Devonian griotte is present here (Appendix II, sections 6 and 7). Furthermore, the number of anticlines and synclines is not the same on both sides of the fault. However, there must been renewed post-main-phase movement along this fault after the intrusion of the dykes, because these dykes are also displaced.

The Muñido Syncline

This syncline is the eastern continuation of the Home Syncline and shows more or less the same structural characteristics. These synclines are separated by the E-W striking Ruenois Fault. In the northeast and east this Muñido Syncline is truncated by the Maladeta granodiorite. West of the Tor River, the southern limit is formed by the narrow and faulted Home Anticline, and east of this river by the steep, northward-dipping flank of the Cambro-Ordovician Payasos Dome.

Besides large and small main-phase folds, comparable with those of the Home Syncline (Appendix II, sections 10 and 11), no secondary complications such as s_2-folding and recumbent folds caused by the granodiorite intrusion have been found. The variation in cleavage and bedding and the distribution of fold axes and δ-lineations are shown in Fig. 19, diagrams A_5-A_6, and Appendix III, diagrams A_5 and A_6.

The Home Anticline

This rather steep isoclinal main-phase structure exposes mainly metamorphic Silurian between the Devonian Home and Muñido Synclines in the north and the Montañeta Syncline in the south. Between the Ribagorzana and the Tor rivers, this anticline is very narrow and locally faulted. East of the Tor River, this narrow structure passes into the northern flank of the Payasos Dome. West of the watershed between the Ribagorzana and the Baliera rivers, its function as boundary between two synclines is taken over by the Castanesa thrust, which separates the Castanesa and Home Synclines to the north from the Meñada Syncline in the south.

The main cleavage of this isoclinal structure has a regular moderate northward dip (50—75°) and shows no macroscopical features of refolding. Microscopically, however, compression phenomena subsequent to the contact metamorphism were found (distorted chiastolites and rotated pyrite crystals). Fold axes and δ-lineations of the main phase were not measured in this structure. The change in strike of the the mean cleavage plane is shown on Appendix III, diagrams A_3-A_6.

The Meñada Syncline

This is a rather broad, complicated, and mainly isoclinal structure with a gently to moderately northward-dipping s_1-cleavage in the west (25—50°) becoming steeper to the east (up to 65°). Due to the vast outcrops of Carboniferous slates west of the Baliera and mainly Devonian strata east of this river, the Meñada Syncline must have a pronounced westward plunge.

In the north and east this syncline is bordered by the rather flat, mainly northward-dipping Castanesa thrust and in the southwest by the Alpine Basibé fault. East of the Baliera River, the southern limit of the Meñada Syncline is formed by a set of Alpine thrusts, probably splays of the Senet thrust to the east. A somewhat isolated wedge of the Meñada Syncline is preserved in the triangle formed by the Salinas thrust and the Basibé fault. There, the Hercynian structure is bent and tilted as a result of the Alpine deformation (Appendix II, section 3).

Due to the rather flat Castanesa thrust, the Meñada Syncline becomes very narrow near the Collado de Basibé, but widens again towards the west. Halfway between this Col and the Esera River, the Castanesa thrust ends, and one can no longer distinguish between the Castanesa and the Meñada Synclines.

West of the Baliera River, the Meñada Syncline is — apart from the flat, roughly northward-dipping Castanesa thrust — properly speaking an Alpine graben, with a downward (normal) movement along the Basibé fault (Appendix II, sections 1, 2, and 3), creating the possibility of preserving the rather incompetent Upper Triassic(?) sediments. Several hundred metres east of the Baliera, the Basibé fault is cut off by the Salinas thrust, which had a southward movement (Appendix II, sections 4—7).

The splays of the Senet thrust, bordering the Meñada Syncline in the south and cutting right through it, also had a southward movement. The variation in cleavage (s_1) and bedding (ss) orientation and the distribution of fold axes and δ-lineations are shown in Fig. 19, diagrams A_2—A_3. For the fold characteristics, see Appendix II, sections 1—6.

Distinct refolding has not been found (apart from the above-mentioned Alpine bend), but locally in the Carboniferous slates a faint crenulation of the s_1-plane suggests a compression phase subsequent to the development of the s_1-cleavage. These ripples strike E-W to WNW-ESE, i.e. roughly parallel to the strike of the s_1-cleavage.

In the slates, many knick-zones have also been observed. Most of these mainly WNW-ESE-striking flexure zones have a steep (60—80°) southward dip, in the opposite direction from that of the cleavage; the angle between the cleavage plane and the knick-zones is rather constant, measuring slightly less than 90°. The movement in the zones is downward to the south. Locally, the other component of the conjugate set has also been observed; these are usually flat planes, dipping gently north or south and with an opposite sense of movement. The origin of knick zones, their shape and relation to the Hercynian orogeny, have been explained by Hoeppener (1955), Engels (1959), Zandvliet (1960) and Boschma (1963). The dilatation of knick-zones in a horizontal sense suggests horizontal tensional stress.

Another interesting feature is the vertical NE-striking fault (probably a wrench fault) in the northernmost part of the Meñada Syncline (coord. 4°20′.42°35′). The structures on either side of this fault are not at all comparable; the fault must therefore have originated simultaneously with or later than the main folding phase. Since this fault is cut off by the Castanesa thrust, the latter must have occurred even later. This is in accordance with the time relation established in the Castanesa Syncline, where it could be demonstrated that the Castanesa thrust originated during or after the s_2-folding. Even an Alpine age could be possible.

The Montañeta Syncline

This structure, which is probably the eastern continuation of the Meñada Syncline, is purely isoclinal with a gentle eastward plunge. Its northern border is formed by the Home Anticline, which east of the Ribagorzana River is locally faulted. In the south, the Montañeta Syncline is limited by the Alpine Senet thrust. In the east, near the Tor River, it develops a westward plunge and forms the Devonian cover of the Cambro-Ordovician Payasos Dome.

West of the Ribagorzana River, this Montañeta structure consists of a single steep (60—70°) isoclinal syncline, exposing slightly metamorphic Lower and Middle Devonian rocks and, in its southern flank, some Silurian. East of the Ribagorzana, the general dip diminishes (65—40°), but the number of individual structures rapidly increases to a total of 7 synclines and 6 anticlines near the watershed with the Rio Tor. This number diminishes on the easthern slope of the Montañeta mountain (de Sitter and Zwart, 1961, pp. 46—47). The core of the synclines is mainly formed by the Castanesa limestone, except for the northernmost syncline which also comprises Upper Devonian griotte. In the anticlines the Rueda Formation is exposed. In this eastern part of the Montañeta Syncline the thermal influence of the granodiorite is only perceptible in the extreme north.

Besides these structures of hectometre dimensions, many parasitic folds measuring some metres occur, frequently with a northward directed plunge (Fig. 19, diagrams A_4—A_6). In each of the diagrams the poles of the main cleavage are closely grouped, as a result of which the northward-directed fold axes and δ-lineations suggest that a folding occurred prior to the cleavage development. The similarity between the cleavage and bedding diagrams (Fig. 19, diagrams A_{4a}—A_{4b} and A_{6a}—A_{6b}) indicates that this pre-cleavage folding must have formed only a slight angle with the later s_1-cleavage.

Folded cleavage has not been found, but the observation of broken hornfels layers in a carbonate matrix showing rotation of the individual fragments indicates that there must have been a compression phase subsequent to the intrusion of the granodiorite (see also page 188).

The Alpine influence seems to be restricted to the Senet thrust; there was no accompanying bending of the Hercynian structure, as in the case of the Meñada Syncline.

The Southern Border Structure

This mainly anticlinal structure in the southern part of the Sierra Negra Unit almost exclusively exposes strongly cleaved, pelitic rocks: Lower Devonian in the west; Cambro-Ordovician, Silurian, and Lower Devonian in the centre; and in the east, Lower and Middle Devonian. It has a sharp northern border formed by the Basibé fault in the west and the Senet thrust, with its splays, in the centre and east. Its southern border is well defined only in the west and centre by the Cerler and Bordas thrusts. East of the Ribagorzana River, we have drawn the southern boundary along the Gelada thrust because of the two patches of Cambro-Ordovician in the fault zone, although rather strong flattening by cleavage is also observed south of this thrust. In the east the Muro Dome, with its intrusive Barruera stock (fine-grained hornblende-bearing granodiorite), probably forms the structural continuation of this Southern Border Structure. In the west this structure was studied only up to the intrusive mass of the Pico Cerler (granodiorite porphyrite), which produced a contact zone of moderate width (200—400 m) only in the south. The thermal metamorphism of the Maladeta granodiorite and the Barruera stock does not influence this Southern Border Structure.

The most outstanding feature of this structural unit is the strongly developed northward-dipping slaty cleavage (s_1) generally showing only faint indications of the bedding. Some thin beds and calcite veins show very tight folding, indicating intense flattening of the rocks, which is demonstrated even better in thin sections. From these cleavage folds the amount of flattening of the rocks can be calculated by comparing the thickness of a particular bed in the hinge and in the limb (de Sitter, 1964, pp. 274—277). We established ratios varying between 4 : 1 (in pure slates) and 2 : 1 (in sandy slates and limestones), corresponding to a flattening of 50% and 30%, respectively. The amount of total flattening therefore varies between:

shortening of 90°	concentric folds + flattening due to cleavage
36%	+ 50% of (100—36%) = about 70%
and 36%	+ 30% of (100—36%) = about 55%

The change in the strike of the cleavage can be read from Appendix III and Fig. 19, diagrams A_2—A_6. The variation in the dip of the cleavage is more or less the same in sections taken perpendicular to the strike and along the strike, and ranges between 40 and 65°. Only close to the metamorphic area of Bono were lower dips observed (20—35°). Fold characteristics are shown in Appendix II, sections 1—13.

Folding of the cleavage occurred locally, mainly around E-W fold axes, causing small folds of centimetre and decimetre dimensions and crenulation of the cleavage plane (see also Boschma 1963, pp. 162—165). In areas without macroscopically visible refolding, a compression phase subsequent to the s_1 development can be inferred from boudinaged intrusive dykes (Fig.

Fig. 29. Boudinaged intrusive dyke, a phenomenon related to the s_2-deformation.

29) and in the metamorphic rim south of the Cerler stock also from rotated hornfels chips, indicating a northern upthrow. Besides these compression phenomena in the slates, a few knick-zones have also been found, indicating a later dilatation in a horizontal sense.

True Alpine phenomena are represented by the Basibé fault, the Senet thrust with its splays, and the Salinas thrust, which cut the Southern Border Structure into two individual units. The other large thrusts (Cerler, Bordas, and Gelada thrusts) may also have been active during the Alpine orogeny, but since we did not find Mesozoic rocks incorporated into the fault zone, this remains doubtful. Thin quartz veins (10—50 cm) occur locally in the Aneto slates (e.g. Collado de las Salinas and Collado de Gelada), which lie roughly parallel to the general trend of the area. These veins originated later than the s_2-folding, since they cut through the folded dyke on the Collado de las Salinas; but it is not clear whether they represent faults or merely exudations along the second cleavage plane.

THE BALIERA UNIT

General remarks

The Baliera Unit south of the Sierra Negra Unit discussed above (Fig. 15) is characterized by large, generally NE-striking pre-cleavage folds involving exclusively Lower and Middle Devonian strata (including the Fonchanina slates). These large structures are pure concentric folds without cleavage development. They were flattened and refolded by the main-phase cleavage folding, which had a roughly E-W strike. Refolding of the cleavage is seen only very locally, e.g. in a narrow rim south and southwest of the Ordovician Muro Dome, but becomes gradually stronger towards the east. The Alpine influence seems to have been restricted to the large southern border thrusts (Comadelo, Estét, and Bono thrusts) and to a few minor thrust movements along the cleavage plane.

The northern and most of the southern boundary of this structural unit are rather sharp lines formed by thrusts. In the north, these are, going from west to east: the Cerler, Bordas, Gelada, and Senet thrusts, and in the south the above-mentioned Alpine thrusts. Near the western and eastern ends of the Bono thrust, the southern limit of the Baliera Unit coincides with an imaginary line lying roughly tangentially 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 or sometimes narrow zone in the Fonchanina slates, the Upper Devonian griotte has an E-W to locally NW-SE strike. This disharmony in fold direction and folding style is most pronounced in the central and eastern part of our structural map (Fig. 15). In the area east of this map, however, the entire Devonian sequence, locally including the Carboniferous, is folded together into large NE-striking pre-cleavage structures; hence, there the structural classification into a Baliera and Ribagorzana Unit is no longer applicable.

In the western part of our structural map, mainly west of the Isabena River, this structural unit gradually looses its characteristic large NE-striking pre-cleavage folds. The area up to the Alpine Liri fault zone is dominated mainly by E-W-striking main-phase cleavage structures, although N-S-directed minor folds and δ-lineations indicate that here too the cleavage development was preceded by folding with an aberrant strike.

The presence of oblique pre-cleavage folds in this area was first detected from the distribution of δ-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 (Figs. 30 and 31).

On the geological map these folds can be recognized by the peculiar course of the lithological boundaries; these boundaries do not trend constantly E-W, as in the central part of the Axial Zone and the most of the Sierra Negra Unit, but show some kind of *"Schlingenbau"*. The cleavage, on the other hand, has a rather constant E-W to ESE-WNW trend (Appendix III, B_1—B_8).

The direction and shape of these early folds is difficult to ascertain when the later cleavage deformation is strong, as for instance in the Basibé massif (coord. 4°17′.42°32½′), where there is a complicated interference pattern of more or less equally sized folds with E-W and roughly N-S-striking axes whose plunge varies from 0 to more than 90° (Fig. 30, diagrams B_{1a} and B_{1b}). In other places, however, such as the area between the Isabena and the Ribagorzana rivers, the effect of the cleavage folding on the original oblique folds was much weaker, and 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 δ-lineations on the generally E-W striking cleavage

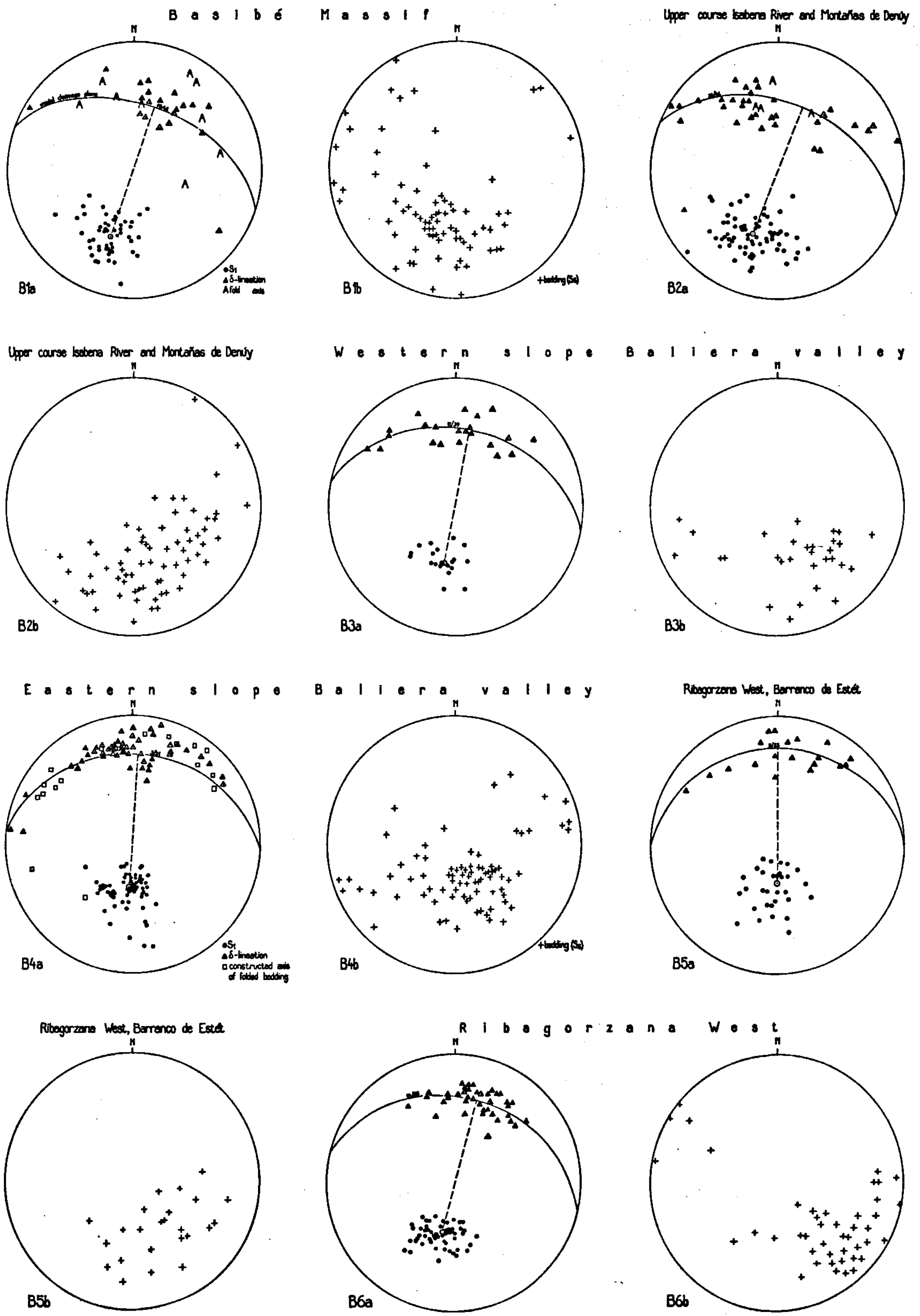


Fig. 30. Orientation diagrams of ss, s1, δ-lineations, and fold axes of the Baliera Unit west of the Ribagorzana River. The numbers of the diagrams correspond to the numbers of the subareas indicated in Appendix III.

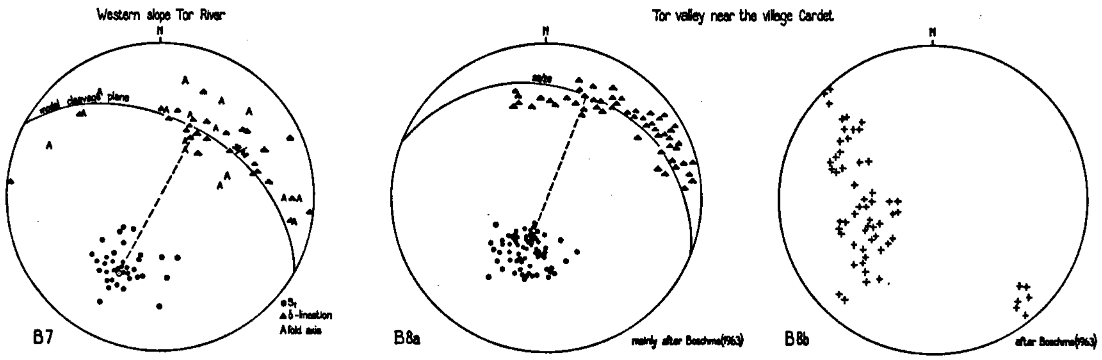


Fig. 31. Orientation diagrams of ss, s_1, δ-lineations, and fold axes of the Baliera Unit east of the Ribagorzana River.

plane are, however, clear evidence of early obliquely-striking structures (Fig. 30).

Although the pre-cleavage folds as a whole are seldom visible in the field, the relative dip of their flanks can be deduced from the shape, axial plunge, and symmetry of the superimposed cleavage folds. We noticed that on one flank cleavage folding had produced normal anticlines and synclines but on the other flank 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, δ-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) and the bedding and the distribution of fold axes and δ-lineations, as measured in the field, are shown in Fig. 30, diagrams B_1—B_6, Fig. 31, diagrams B_7—B_8, and Appendix III (B_1—B_8).

In discussing the Sierra Negra Unit we have already argued that the swing in strike of the mean cleavage must be either an original tectonic feature caused by a locally divergent stress field during the main folding phase or a post-main phase phenomenon but antedating the intrusion of the granodiorite, probably connected with the diagonal cross-folding.

The cleavage generally dips about 45° to the north, varying from 25 to 48°. We will later advance arguments making it highly probable that these rather flat dips of the cleavage originated in that position and are not the result of later warping. The argumentation for the existence of pre-cleavage folds and the proposed method of their reconstruction is, moreover, independent of this problem, since their existence is based on a difference in strike between the cleavage and the axial planes of pre-cleavage folds. Their original shape and their symmetry would, however, be slightly different if the reconstruction was based on either a steep cleavage, which came to its present position by warping (principle advanced by Hoeppener), or an original moderately-dipping cleavage. This difference becomes more pronounced as the angle between the strike of the cleavage and the axial plane of the pre-cleavage fold diminishes.

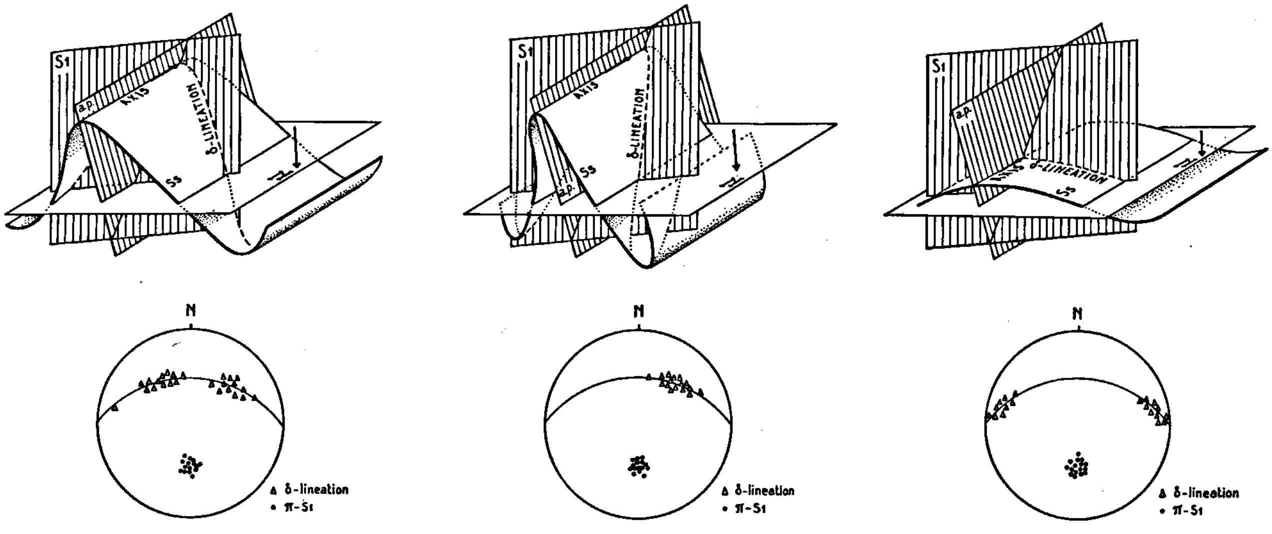
The relation between the pre-cleavage folds and the main-phase deformation

If the mainly NE-striking major structures on the geological map (Appendix I) west of the Ribagorzana are compared with the general E-W to ESE-WNW trend of the mean cleavage (Appendix III), it becomes evident that the strike of this cleavage does not at all coincide with the direction of these major structures; hence, the cleavage could not have been formed simultaneously with these major structures. In fact, the cleavage cuts right through these NE-striking, major structures and consequently must have developed later than these folds.

This is also well expressed in the field. We observed many beautiful examples of concentric, often overturned anticlines and synclines of decametre to hectometre dimensions, cut by a regular cleavage making large angles (up to 90°) with their axial planes, e.g. in the Basibé massif and the southern slope of the Rio Llausét. The theoretical relationship between bedding (ss), axial plane (a.p.), cleavage (s_1), and δ-lineation is shown in Fig. 32.

From this Figure it can be seen that the δ-lineations have a plunge roughly equal to the dip of the cleavage plane when the flanks of these pre-cleavage folds are steep. Horizontal lineations can occur only on the crest of anticlines or the trough of synclines.

When the pre-cleavage structures are very large and the outcrop of mappable units discontinuous, so that only parts of the flank of folds are visible, the ratio between the steep lineations and those with little or no plunge serves as a gauge for the original shape of the pre-cleavage structure: when there are mainly steep lineations and only a few with little or no plunge, the pre-cleavage structure must have been a fold with long, steep limbs and a narrow hinge zone, i.e. folds with a short wavelength but high amplitudes (Fig. 32b). The long, narrow NE-striking folds south of the Rio Llausét are examples of this. These very tight pre-cleavage folds cut by a cleavage with a strike



a. Close pre-cleavage fold b. Tight pre-cleavage fold c. Gentle to open pre-cleavage fold

Fig. 32. Superposition of a moderately northward-dipping cleavage on three differently shaped NE-striking pre-cleavage folds. Note relation between shape of fold and distribution of δ-lineations.

almost perpendicular to their fold axes, produce a high concentration of δ-lineations with a maximum dip in the cleavage plane (Fig. 30, diagram B_{6a}). The bedding diagram (B_{6b}) shows a concentration of bedding poles (π-ss) in the SE segment of the diagram, indicating fairly parallel limbs, which are, moreover, overturned to the SE. That the cluster of cleavage poles (π-s_1) and bedding poles in the diagram of this area show hardly any overlap, is an indication that the main-phase folding produced a cleavage that cut right through the early structure but caused few or no accompanying cleavage folds.

When, on the other hand, the flat lineations dominate, the pre-cleavage fold must have been gentle or open, with a long wavelength and low amplitude (Fig. 32c). Examples of these occur west of the Isabena River and in the western part of the Basibé massif, where the pre-main-phase folding dies out.

The above-mentioned relationship between ss, a.p., s_1, δ-lineations, and fold axes, also holds for pre-cleavage folds that were not only cut by the cleavage (under a large angle) but also refolded as a result of unequal flattening of different lithological units, e.g. the Basibé Formation, which shows almost no flattening as compared with the Gelada Formation below and the even less competent Fonchanina slates above, both of which were strongly flattened by the cleavage. These secondary folds are nearly always smaller than the pre-cleavage folds. Since the cleavage coincides well with the axial planes of these secondary folds, the δ-lineations and secondary fold axes are coincident (Fig. 33); therefore, both the δ-lineations and fold axes of these cleavage folds can be used to reconstruct the shape of the original pre-cleavage fold, as discussed above. It is obvious that under these circumstances the pre-cleavage fold was shortened in the direction of its axis by cleavage folds in the competent layers and pure flattening in the less competent rocks.

A very clear example in which several WNW-ESE-striking cleavage folds of hectometre dimensions are superimposed on a large NE-striking pre-main-phase anticline measured in kilometres can be observed in the area between the Isabena and the Montañeta de Denúy (coord. 4°18′.42°31′). This is illustrated in Fig. 34.

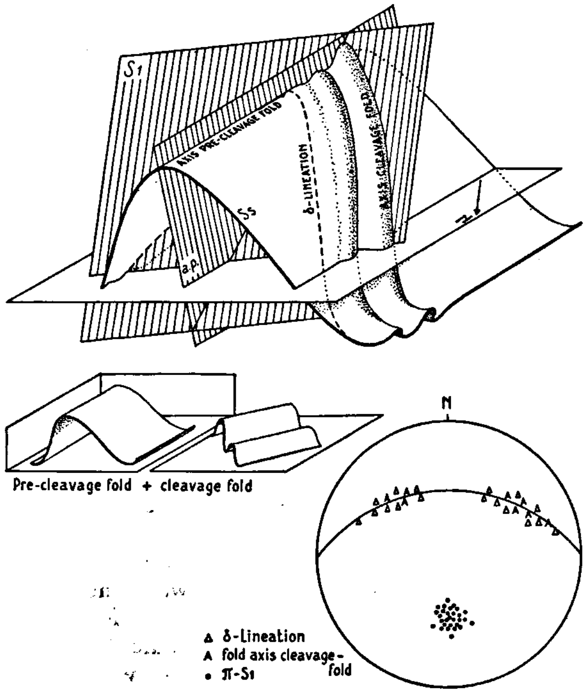


Fig. 33. Superposition of a tight cleavage fold with a moderately northward-dipping axial plane on a close NE-striking pre-cleavage fold.

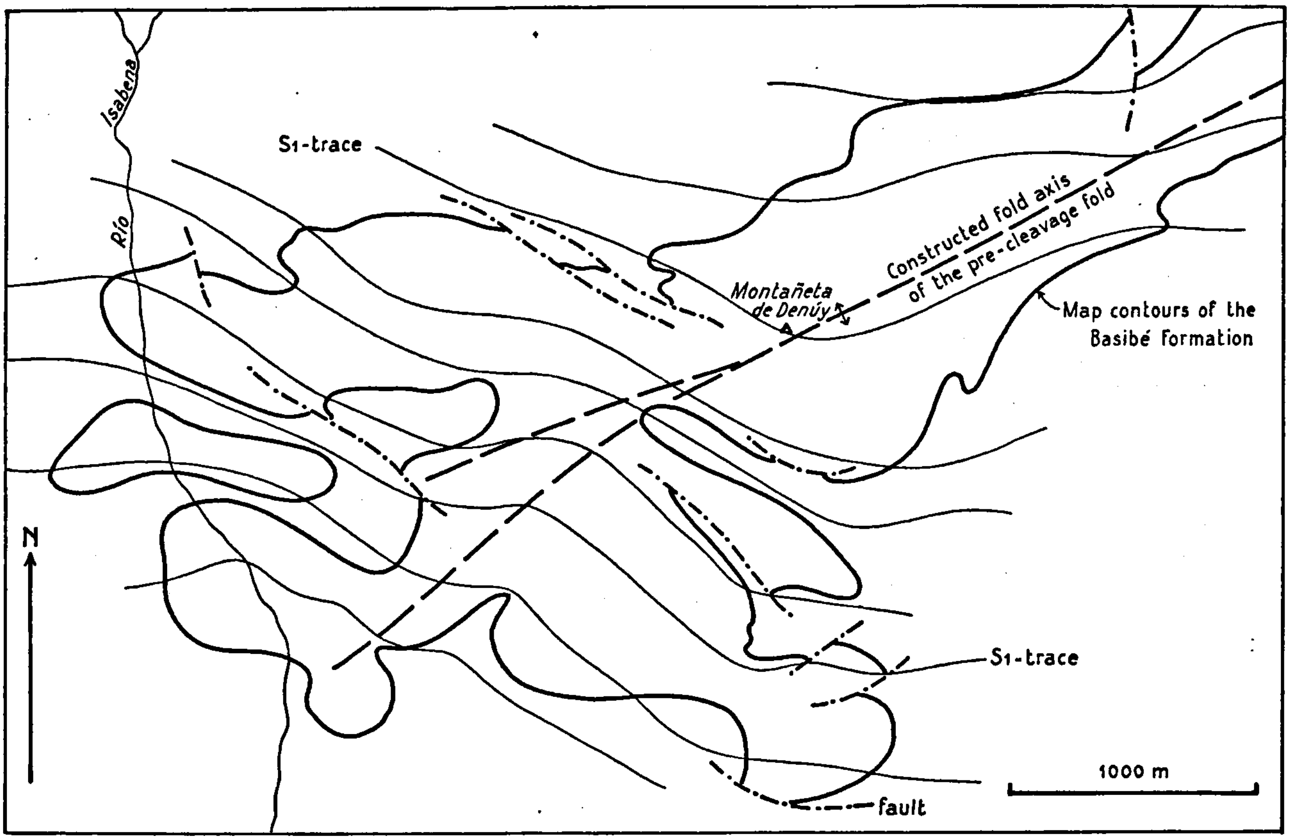


Fig. 34. Typical *"Schlingenbau"* caused by the superposition of WNW-striking cleavage folds on a roughly NE-striking pre-cleavage anticline.

The cleavage traces (intersection of the cleavage with the topography), which were constructed after the modal cleavage plane of diagram B_2 (Appendix III), coincide well with the expected axial plane traces of the secondary folds. This proves, quite apart from the field evidence, that these secondary folds are really cleavage folds.

It is interesting to note that the reclined cleavage folds on the eastern limb of this pre-main-phase anticline (south of the Mantañeta de Denúy) have overturned plunges, so that what is observed in the field as antiforms are actually synclines, and visa versa. It may therefore be concluded that this southeastern limb must have already been overturned before the development of the cleavage and accompanying folding.

This interference pattern is also well expressed on diagrams B_{2a} and B_{2b} of Fig 30. The cluster of cleavage poles (π—s_1) overlaps part of the NNE-dipping poles of the bedding (π—ss) due to the effect of the slightly isoclinal main folds, which strike WNW-ESE. The remaining bedding poles, i.e. those lying outside the cluster of cleavage poles, show rather persistant dips to the northwest. These are certainly remnants of the earlier, roughly NE-striking pre-cleavage fold. The fact that no eastward dips of the bedding were observed supports our view of an originally overturned pre-cleavage structure.

This method of reconstructing the pre-cleavage fold by subtracting the superimposed cleavage folds is of course only easily applied when the angle between the strike of the cleavage and the strike of the axial plane is more than 45° (see foregoing). If this angle is less than 45°, however, the original fold is mainly compressed into a narrower structure sometimes showing secondary folds at places where the original bedding made larger angles with the cleavage. Much depends then on the original shape of the early structure and the angle at which this structure is cut by the cleavage. In such situations it is difficult to decide whether an overturned limb of a structure was already overturned before the cleavage development or was caused by the cleavage folding.

Since the cleavage folds are generally smaller than the early structures, it is often still possible to subtract the cleavage folds of the pre-cleavage structure. But is impossible to evaluate the amount of general deformation caused by the flattening, which not only influences the shape of the pre-cleavage structure but also changes its original strike. The strike tends to adapt itself to the strike of the cleavage, except when the two directions are perpendicular to each other. The stronger the flattening and the lower the original angle between the two directions, the better the adaptation. When the extension is not only in an upward but also in a horizontal direction (both perpendicular to the direction of flattening), the adaptation is even more pronounced.

Pure flattening of a rock mass in the direction of the largest stress with accompanying extension in one or two directions perpendicular to the direction of shortening can, of course, be calculated from the deformation of known shapes, as in the case of certain fossils (e.g. Breddin, 1964; Furtak and Hellermann, 1961; Albrecht and Furtak, 1965), reduction spots, oolites (Cloos, 1947), conglomerate pebbles, etc. We could not use this method because such material is not available in our area. Less accurate is the method using concentrically folded competent layers in a matrix of flattened incompetent rocks, such as a sandstone or a quartz vein in a matrix of slate (de Sitter, 1964, pp. 276—277). The direction in which extension took place can be deduced from boudinaged beds.

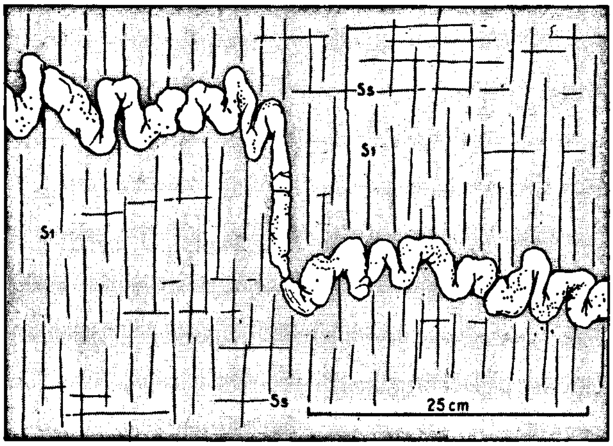


Fig. 35. Parasitically-folded calcite vein in a matrix of cleaved calc-schists. Drawing after photograph.

Using this method, we arrived at values of some 50—60% flattening for pure slate, 45% for marly limestone (Fig. 35), and 30—40% for sandy slates. Unfortunately, we were unable to calculate the amount of flattening of thick limestones, dolomites, and quartzites. The measurements show that the amount of total flattening varies from one fold to another, depending on the kind of rock involved and consequently on the dimension of the fold. In the case of our pre-cleavage folds, which involve Lower and Middle Devonian strata, the total flattening probably does not exceed 40%.

The area between the Ribagorzana and Tor rivers northwest of the village of Cardet (coord. 4°28'. 42°30') yields good examples of this kind of deformation. The large folds of the Basibé Formation shown on the geological map are former pre-cleavage structures but strongly flattened during the main-phase. This flattening caused a deviation of the original trend of the pre-cleavage structures, which were probably more NE-striking than can be evaluated from their actual trend (ENE-WSW to locally E-W). From this roughly NE-striking fold pattern the folds southwest of Cardet must be excluded. The latter are roughly NW-SE-striking and thus represent the transition to the general NW trend of the Ribagorzana Unit to the south.

The diagrams of this area between the Ribagorzana and Tor rivers (Fig. 31, diagrams B_7—B_8) show point maxima for the cleavage poles and a large spread of the δ-lineations and fold axes of minor folds, thus indicating the presence of oblique pre-cleavage structures. This is also clear from a comparison of the spread in bedding poles (diagram B_{8b}) with the point maximum of the cleavage poles.

We do not agree with the conclusion of Boschma (1963, p. 146) postulating the absence of a pre-cleavage folding in his sub-area VII (around the village of Cardet). His bedding diagram 59 implies the presence of a rather steep structure (one or several folds) with an E-W to ENE-WSW strike. His cleavage diagram 60, however, shows a point maximum of very flat, northward-dipping cleavage poles that do not at all coincide with the steep axial plane of the above-mentioned structure. Such a combination of a steep pre-cleavage structure cut by a flat cleavage with roughly the same strike also explains the presence of a cluster of δ-lineations and fold axes instead of a normal girdle (Fig. 36).

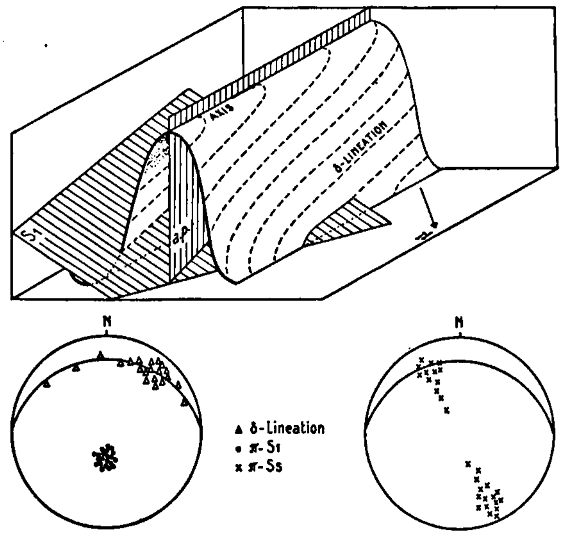


Fig. 36. Superposition of a gentle northward-dipping cleavage on a steep and tight NE-striking prc-cleavage structure.

So far, we have discussed the basic examples of the superimposed cleavage folding on large concentric folds with a different strike. But since the angle between the two fold directions does not diverge greatly from 45°, a single pre-cleavage structure often shows a combination of the above-mentioned deformations. An expression of this concurrence is found in the wide variety shown by the diagrams (Figs. 30 and 31).

This twofold deformation becomes even more complicated when the cleavage folding has created large structures comparable in size to the largest pre-cleavage folds. As far as we know, such large-scale cleavage folds occur only at the southern and northern borders of the Baliera Unit. In fact, they cause the striking disharmony in fold direction and folding style between the three major structural units. In the south we are concerned with a rather tight, inclined anticline, and in the north with an almost isoclinal, inclined syncline, both with moderately northward-dipping axial planes. The effect of these large-scale cleavage folds on the oblique-striking pre-cleavage structures can be observed best, in the gorge that the Rio Llausét has carved in the axis of this northern main-phase syncline, the northern limb of which corresponds to the southern limb of the Southern Border Structure (see page 190). The two NE-striking pre-cleavage synclines whose cores contain Fonchanina slates are refolded into antiforms, as illustrated schematically in Fig. 37. The strongly squeezed cores of these two composite folds are well exposed on the northern slope of the gorge: the eastern structure produces on the topography a so-called "eyed fold" (Ramsay, 1960) (Fig. 37); the western structure, as well as the intervening pre-cleavage anticline, is slightly overthrusted. Most of the higher part of the overturned northern limb of this main-phase syncline has been removed by erosion.

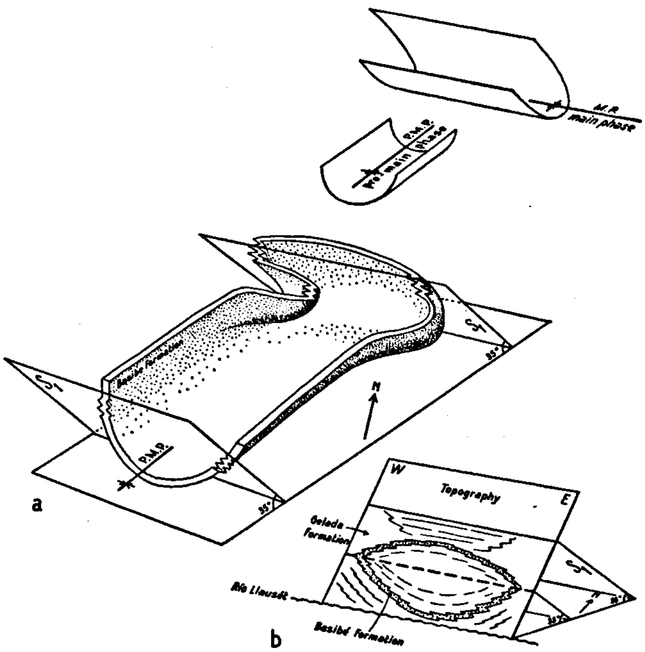


Fig. 37. Superposition of a tight cleavage fold on an equally large open pre-main-phase syncline, producing a so-called bird-eyed structure on the topography.

The Basibé massif and the area several kilometres east of the Rio Tor (outside our geological map) offer many other beautiful examples of such large-scale oblique refolding, often with both the lower and the upper overturned limbs exposed. However, the structures east of the Tor River are still more complicated, since they also underwent post-main-phase refolding and show the development of a secondary crenulation cleavage.

Now that we have shown that the NE-striking pre-cleavage structures are limited in the north and south by large main-phase folds and thrusts, the question arises of whether these structures originally extended beyond their present boundaries. Concerning the southern region, it is impossible to give a positive answer, since the Baliera and Ribagorzana Units are separated by a narrow zone of disharmony, as a result of which the NE-striking folds of the Lower and Middle Devonian of the Baliera Unit do not enter the Ribagorzana Unit with the Upper Devonian at its base. Whether this disharmony resulted solely from a difference in lithology or was caused by a former basin rim limiting the sedimentation of the Aneto slates and quartzites in the south, is not known. If the former was the case, the NE-directed folds of the Lower and Middle Devonian could theoretically extend beyond this zone of disharmony and below the Upper Devonian-Carboniferous, but in the latter case one would merely expect the NE-striking folds to die against that basin rim. For the northern region, on the other hand, it seems reasonable to suppose that the large NE-directed folds died out against the structural line (fault or flexure) limiting the Baliera basin, with its special stratigraphical development. Although pre-cleavage folding is also present north of this structural line, as shown by the northward-dipping δ-lineations and fold axes, the angle between the two fold directions was probably smaller.

However, caution is required here. It is also conceivable that due to a more severe flattening (in the horizontal direction) the adaptation of formerly oblique structures to the general cleavage trend is more complete. Persistent and rather steep lineations and axes of minor folds have been reported from other areas in the central part of the Axial Zone with a strong and steep cleavage, although no obvious oblique structures show on the geological map (Kleinsmiede, 1960; Zandvliet, 1960; Boschma, 1963). Oblique structures do occur, however, near the northern and southern borders of the Axial Zone (including the Nogueras Zone), where the intensity of the cleavage is lower. In the northern area, with a steep cleavage, large NE-striking pre-cleavage folds have so far only been reported from the Estours region (de Sitter & Zwart, 1962), but in the Southern Pyrenees, with a moderately dipping to sub-horizontal cleavage (causing extension in a horizontal direction) these NE-striking pre-cleavage folds occur very frequently (internal reports). Hence, a certain relationship between the intensity and the steepness of the cleavage on the one hand and the mappable NE trend of the pre-cleavage structures on the other hand, cannot be excluded. However, to determine whether strong flattening (in a horizontal direction) was solely responsible for the lack of NE-striking pre-cleavage structures will require much detailed map-work, supplemented by new data on the absolute values of cleavage flattening throughout the Axial Zone of the Pyrenees.

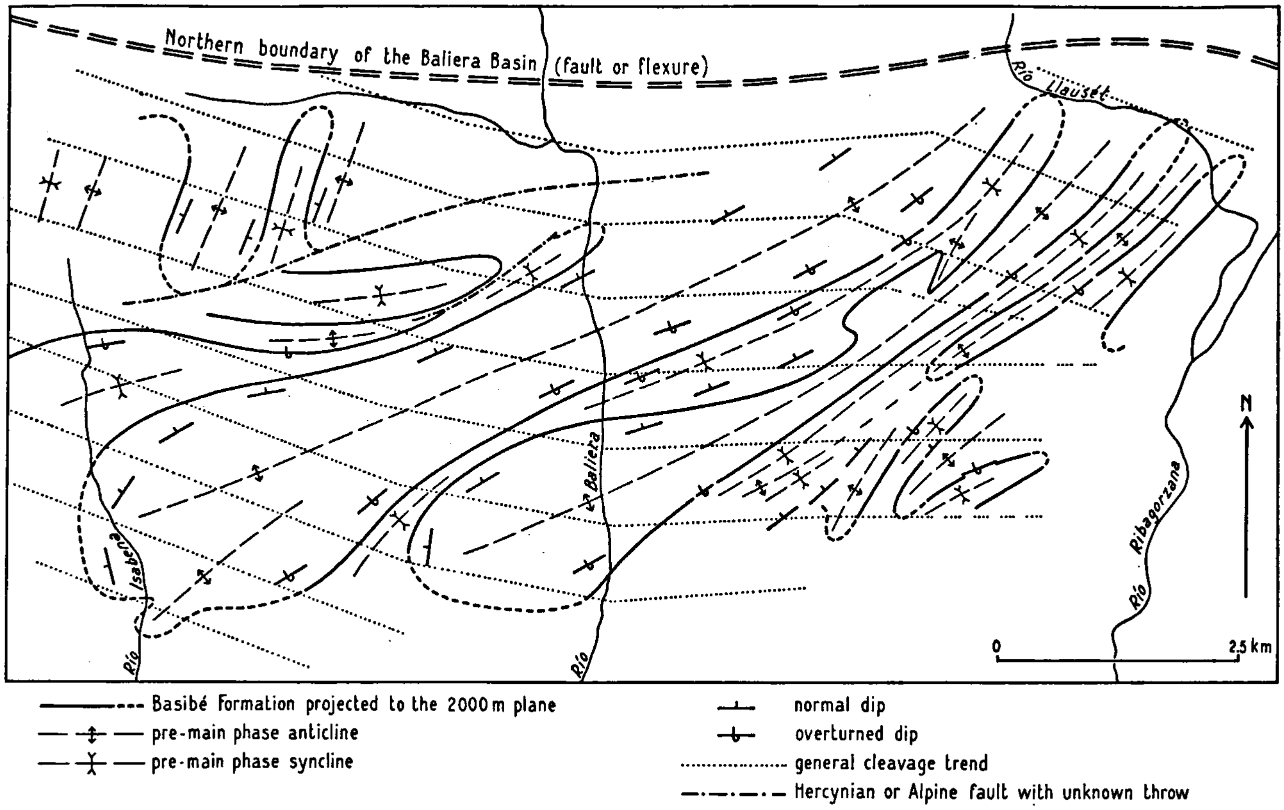


Fig. 38. Projection to the 2000 m plane of the reconstructed pre-cleavage fold pattern between the Ribagorzana and Isabena rivers.

Reconstruction of the pre-cleavage folds

For the reconstruction of the original pre-cleavage structures, all later deformations have to be subtracted. In addition to the main-phase deformations described above, we have also mentioned in the introduction the local occurrence of s_2-folding and Alpine thrust movements, which of course must also be subtracted. It is obvious that a superimposed s_2-folding on an interference pattern of two former folding phases may create structures that can no longer be distinguished as distinct types. Therefore, for our reconstruction we selected the area between the Ribagorzana and Isabena rivers, in which no Hercynian folding of the main-phase cleavage took place. The Alpine influence seems to have been restricted to a number of thrusts and minor fault movements (see page 209) which hardly influence the basic Hercynian fold pattern. Apart from these true Alpine faults, involving Triassic rocks, there are many other faults and thrusts parallel to the cleavage plane (e.g. the Bordas thrust and the one south of it) for which we do not know whether they originated during the Hercynian or during the Alpine orogeny. But since these faults cut through the cleavage folds, they too can be subtracted when their throw is known at least approximately.

For the reconstruction of the area between the Ribagorzana and Isabena rivers, we used all the available direct and indirect field data. On the basis of our geological map, horizontal stereo-photographs, and data provided by the stereograms, we constructed a simplified clay model of this area. Starting from this compound model, we subtracted all folds with E-W cleavage as axial plane, and projected the remaining structure on the 2000-metre plane (Fig. 38).

It should be noted that we did not consider the slight distortion of the original trend and the shortening of the pre-cleavage folds due to the super-imposed cleavage deformation. Our reconstruction is therefore only a very simplified picture involving many assumptions; only the distribution of folds with normal and overturned limbs can be considered more or less correct. But this basic model is adequate for the reconstruction of the much more complicated area east of our structural map in which strong s_2-folding of the s_1-cleavage occurs. An article concerning this area is in preparation.

Although we are still unable to reconstruct the pre-cleavage folds in the entire Baliera Unit, the general tendency of the folds from west to east is evident: the wide undulations west of the Isabena become toward the east narrower and more steeply inclined. The large NE-striking folds south of the Ordovician Muro Dome were probably steep and isoclinal, but without the development of a macroscopically-visible cleavage parallel to their axial planes.

The cross-sections (Appendix II), which were constructed with the help of clay models, unfortunately

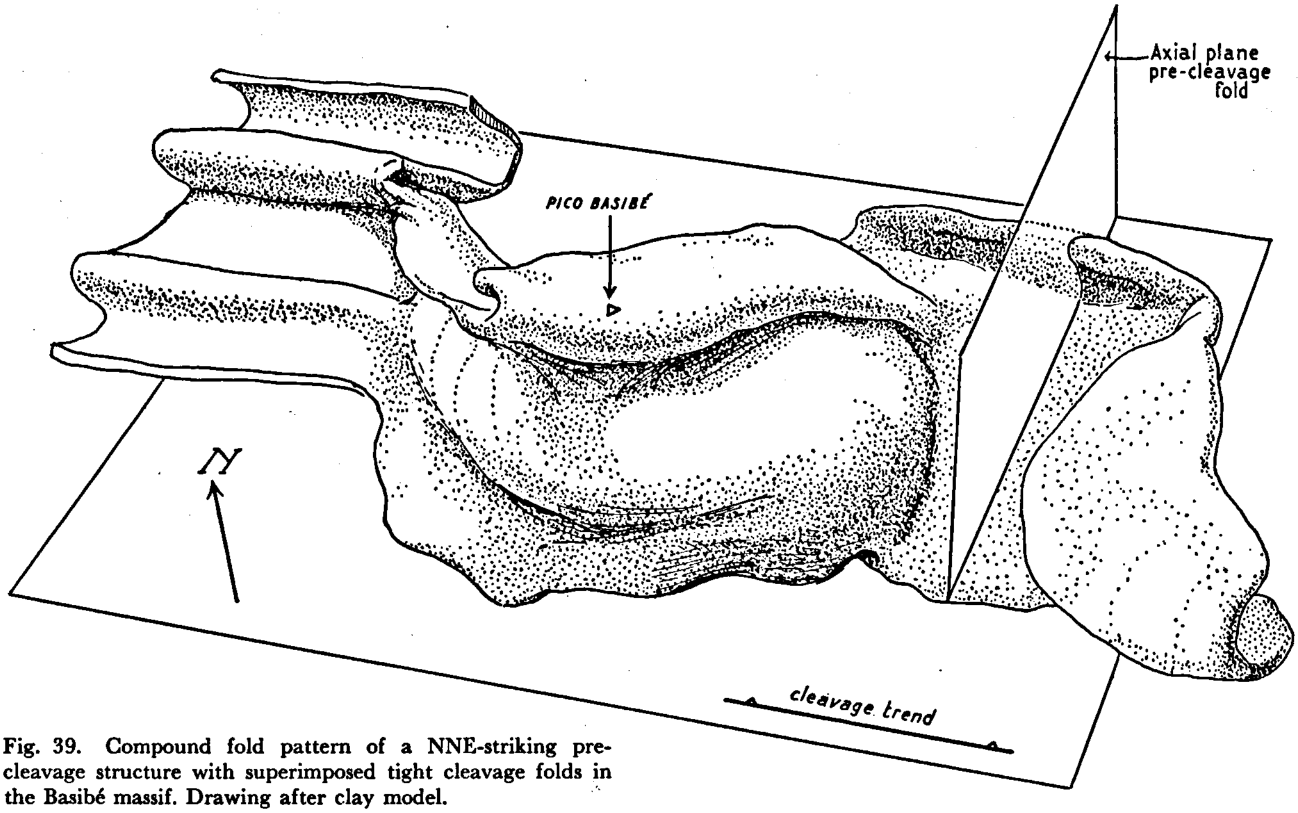


Fig. 39. Compound fold pattern of a NNE-striking pre-cleavage structure with superimposed tight cleavage folds in the Basibé massif. Drawing after clay model.

do not show the complexity of fold shapes encountered in the field. Three-dimensional drawings after models provide the only clear representations of the complicated fold pattern of this area. Two instructive examples are shown in Figs. 39 and 40, which concern the Basibé massif and the area already shown in Fig. 34, respectively.

The dip of the cleavage

In the Baliera Unit the mean dip of the cleavage varies between 25° and 48°, with an average of about 45°, in a northward direction. There are two possible explanations for these moderate dips, of which we favour the second:

1. The cleavage was originally vertical and came to its present position by warping (principle advanced by Hoeppener for the Rheinische Schiefergebirge).
2. The gentle to moderate dip of the cleavage is an original tectonic feature.

According to Hoeppener (1955), the occurrence of a non-vertical cleavage may be related to later warping. All movement would take place along normal faults parallel to the cleavage (*"Schieferungs parallele Abschiebungen"*), simultaneously causing knick-zones. The slaty cleavage of certain parts of the Central Pyrenees seems to have been affected by a similar type of deformation (e.g. Zandvliet, 1960 and Boschma, 1963). In the southern part of Zandvliet's area, which has a minimum northward dip of the cleavage of about 50°, normal faults parallel to the cleavage and abundant knick-zones both occur frequently. According to Zandvliet and Boschma, this tilting of the cleavage is a late Hercynian phenomenon and is related to an arching of the centre of the orogene.

In our area, with a cleavage dipping even less than 50°, one would expect a comparable amount of these normal faults parallel to the cleavage and many knick-zones. In slate belts striking parallel to the cleavage trend, in which the bedding cannot be traced (e.g. parts of the Sierra Negra Unit), it is rarely possible to determine in the field whether or not the cleavage plane had been a plane of movement. In the Baliera Unit, on the other hand, where a formation boundary cut by the cleavage under a large angle can be traced continuously over several kilometres, fault planes parallel to the cleavage would easily be recognizable. These faults should always cause a displacement of the bedding in the same direction (Fig. 41). Some faults parallel to the cleavage do occur, but they are seldom very persistent, have a low frequency and, moreover, show normal as well as thrust movements; knick-zones occur only sporadically. We are therefore convinced that in this part of the Pyrenees the principle of warping of the cleavage along a set of normal faults parallel to the cleavage is not applicable.

Moderately dipping cleavage is, furthermore, a rather common feature in many other orogenic belts. A

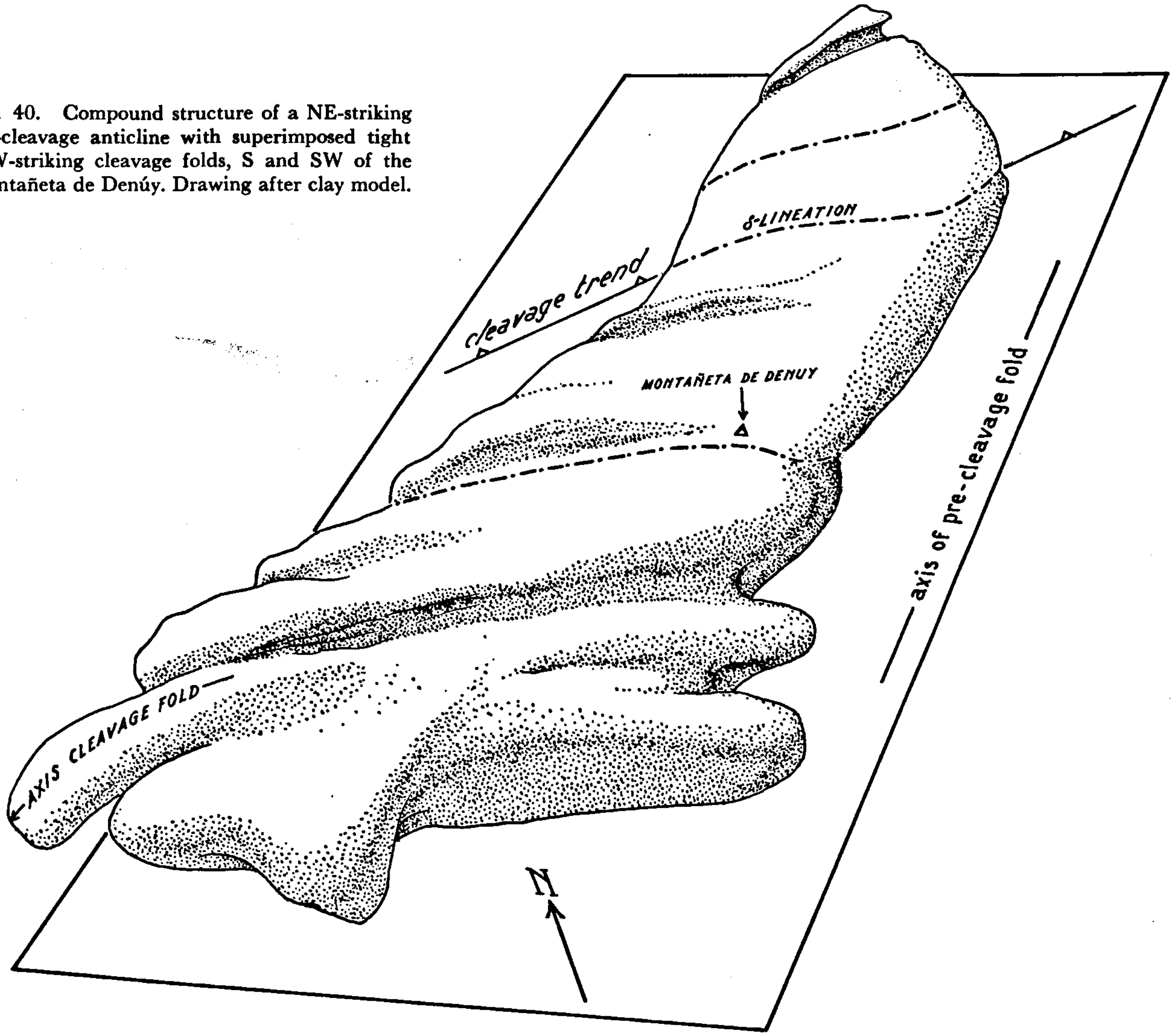
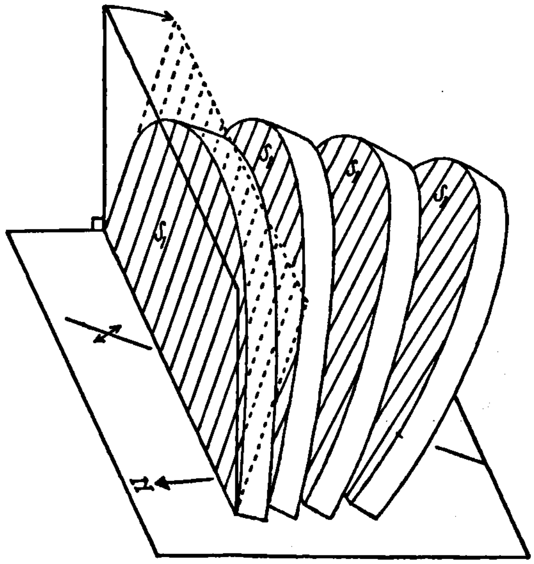


Fig. 40. Compound structure of a NE-striking pre-cleavage anticline with superimposed tight E-W-striking cleavage folds, S and SW of the Montañeta de Denúy. Drawing after clay model.



geometrical explanation of the direct origin of rather flat cleavage planes has recently been given by Breddin, Furtak and Hellermann (1964). Hellermann (1965) has given an example in the Upper Palaeozoic of Northern Germany in confirmation of their geometric explanation. Although in our area the geological circumstances differ somewhat from the example given by Hellermann, a slight modification of his explanation would make it applicable here.

An originally moderately-dipping cleavage readily explains the frequently observed asymmetric deformation of former pre-cleavage folds. We noticed that the southeastern limbs of steeply inclined NE-striking pre-cleavage folds, which are in general steeper than the northwestern limbs, show more numerous and more

Fig. 41. Warping of the cleavage along "*Schieferungs parallele Abschiebungsflächen*" causing a constantly-directed displacement of the bedding of an oblique pre-cleavage structure.

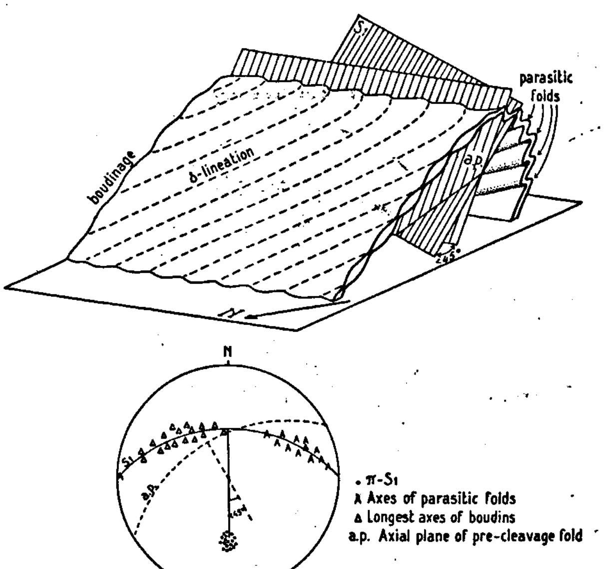


Fig. 42. Unequal deformation of differently-dipping flanks of a pre-cleavage structure in which the strike of the axial plane of the cleavage deformation makes an angle of less than 45° with the pre-cleavage trend.

severely compressed, cleavage folds than the other limb. For a steep or vertical cleavage with an oblique trend one would expect a symmetric deformation with the same amount of folding of both limbs, but a cleavage with a flatter dip can deform unequally dipping flanks in a different way. This is most obvious when the angle between the strike of the cleavage and the pre-cleavage folds is less than 45°: the southeastern limb then shows compression into cleavage folds, but the northwestern limb was flattened, which often caused boudinage in the thin competent beds within a matrix of strongly cleaved pelitic material. The total effect is comparable with a shear fold, but one with divergent fold axes and δ-lineations (Fig. 42).

We can therefore summarize by saying that the absence of *"Schieferungs parallele Abschiebungsflächen"* and the asymmetric deformation of the pre-cleavage folds both constitute arguments favouring the idea of a non-vertical origin of the cleavage in this part of the Pyrenees.

Post-main-phase phenomena

Of the post-main-phase phenomena, we shall only describe the s_2-folding and the knick-zones here; the Alpine influence, the metamorphism around the Bono area and the accompanying dyke swarm will be discussed in separate chapters.

Folding of the s_1-cleavage occurs only in the easternmost part of the mapped Baliera Unit. De Sitter (1956) and de Sitter-Zwart (1961) were the first to mention the s_2-folding in the Tor valley. More recently, Boschma (1963) published a very detailed work on the different kinds of deformation affecting the Lower Palaeozoic and locally the Lower Devonian of the Tor valley. Our own investigation revealed that this refolding of the Devonian becomes much more pronounced towards the east. It is accompanied by a very strong crenulation cleavage (s_2) with an orientation nearly equal to the unfolded main-phase cleavage (s_1) in the area to the west (Appendix III). In spite of this WNW-ESE trend we suppose this deformation phase to represent the late E-W set of refoldings rather than the earlier NW-SE set, since the fold axes are in general sub-horizontal to gently dipping. More definite proof would be provided by observations of folded dykes or rotated contact-metamorphic minerals or hornfels chips, which so far have not been found in this area. A more detailed description of this refolding and its effect on both the s_1-cleavage and the pre-cleavage structures will be the subject of a later article (Mey and Roberti, in preparation).

Knick-zones, which generally indicate a horizontal tensional stress, occur sporadically all over the Baliera Unit but are restricted to the most strongly cleaved rocks, e.g. pure shale and sandy and marly shale/slate. The flexure zones have a rather constant WNW-ESE trend and a steep southward dip, varying between 60 and 80°. The movement in these zones is downward toward the south. The other component of the conjugate set (sub-horizontal planes) was not found in this area.

THE RIBAGORZANA UNIT

General remarks

The Ribagorzana Unit, located south of the Baliera Unit, is bordered in the west and south by the Triassic, which unconformably overlies the Carboniferous shale/sandstone sequence. South of this unconformity one enters the Nogueras Zone. As we have already explained on page 191, the Ribagorzana and Baliera Units cease to be two distinct structural units east of our structural map (Fig. 15). The meaning of the Palaeozoic block of mainly Lower and Middle Devonian in this southeastern corner of Fig. 15 is not yet clearly understood. The only certainty is that the Erta fault, which limits this Palaeozoic block in the north and west, is a gently southward-dipping fault, in the south truncated by the unconformable Triassic of the Nogueras Zone.

The Ribagorzana Unit is characterized by the preponderance of schistose Upper Devonian and Carboniferous rocks with a rather constant moderate dip in a northward direction. In the field one encounters many close to tight folds of metre to decametre dimensions, having the cleavage in their axial plane. However, the many northward-directed δ-lineations and fold axes measured in the field also suggest the presence here of a pre-cleavage folding with an oblique strike. This is also confirmed by the map contours, which are often oblique to the general cleavage trend. In the field one also locally observes 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.

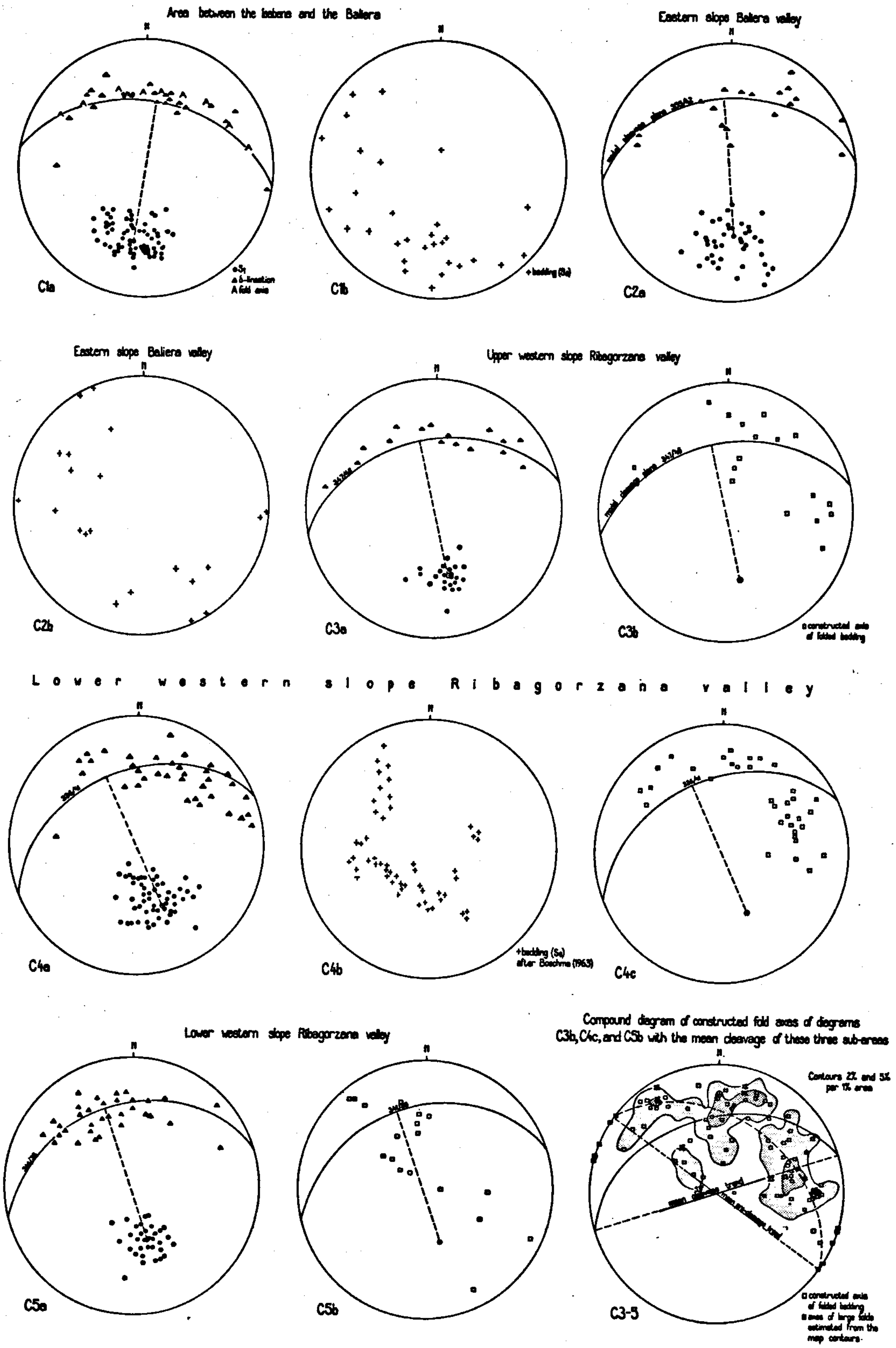


Fig. 43. Orientation diagrams of ss, s1, δ-lineations, and measured and constructed fold axes of the Ribagorzana Unit west of the Ribagorzana River. The numbers of the diagrams correspond to the numbers of the subareas indicated in Appendix III.

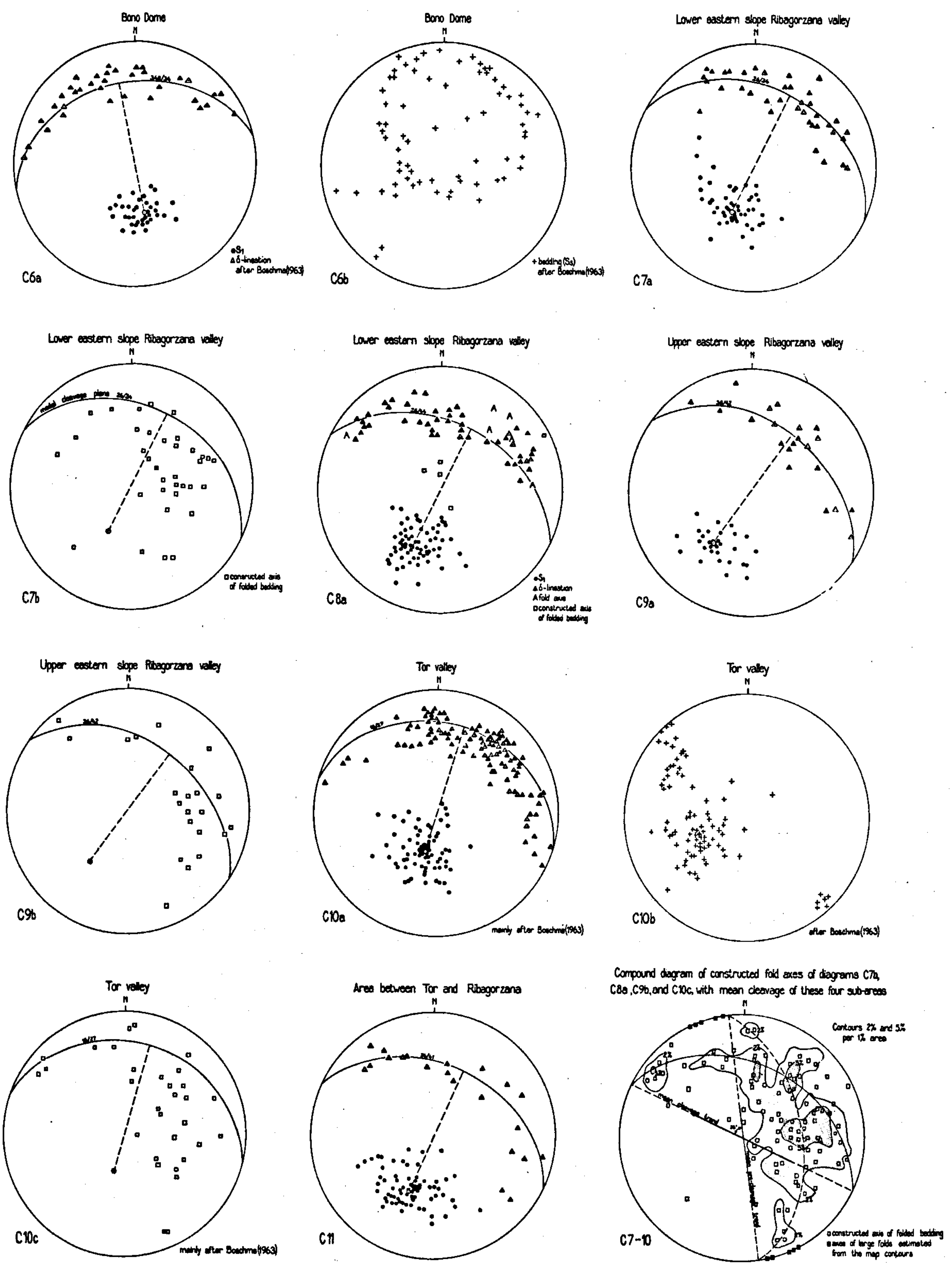


Fig. 44. Orientation diagrams of ss, s1, δ-lineations, and measured and constructed fold axes of the Ribagorzana Unit east of the Ribagorzana River. The numbers of the diagrams correspond to the numbers of the subareas indicated in Appendix III.

Doubtful secondary folding of the cleavage was encountered in one small area. Knick-zones occur more frequently than in either of the other structural units. The Alpine influence seems to have been restricted to the large overthrusts near the northern boundary of this structural unit and to a few minor thrust movements along the cleavage plane (see under Alpine Tectonics).

We can be brief with respect to the description and analysis of the minor tectonic events of this structural unit, since they have been treated extensively in Boschma's thesis (ibid, pp. 143—147).

Since the nodular limestones of the Upper Devonian and the Carboniferous shales have different fold characteristics, we will describe them separately.

The Upper Devonian griotte belt

The general E-W trend of this griotte belt changes into a peculiar NW-SE trend east of the watershed between the Ribagorzana and Baliera rivers. This change in general strike is accompanied by an increase in the number of large folds having hectometre to kilometre dimensions. If the map contours of these folds and the general cleavage trend of this area are compared (Appendix III, diagrams C_1—C_{11}), it is evident that they are not at all coincident. Whereas these folds have a general NW-SE trend, the mean cleavage strikes E-W to ENE-WSW west of the Ribagorzana and changes to a WNW-ESE trend east of this river. From the map contours of these folds between the Ribagorzana and Tor rivers it follows that the folds must be tight and sub-vertical, since the map trace of these folds is hardly affected by the topography. The cleavage of this area, on the other hand, dips moderately (40—45°) and consequently makes a large angle with the axial plane of the folds. These folds must therefore be pre-cleavage folds.

In the field this is confirmed by the orientation of the δ-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 (Figs. 43 and 44). Besides these tight cleavage folds, which occur abundantly everywhere, we observed local undulations and gentle folds of the bedding having no obvious relation to the cleavage. The axes of these gentle folds are hard to measure directly, but they can be constructed from measurements of the intersection of adjacent bedding planes on a stereographic net. This was done for a large number of adjacent bedding-plane measurements and plotted on diagrams (Fig. 43, diagrams C_{3b}, C_{4c}, C_{5b} and $C_{3\text{-}5}$; Fig. 44, diagrams C_{7b}, C_{9b}, C_{10c} and $C_{7\text{-}10}$). Although the cleavage deformation produced a large spread of these fold axes, an approximate mean original trend can still be constructed. This is shown in the compound diagrams of Fig. 43, diagrams $C_{3\text{-}5}$ and Fig. 44, diagrams $C_{7\text{-}10}$. This constructed mean strike of the pre-cleavage folds coincides rather well with the strike of the fold axes estimated from the map contours.

A striking feature of these two compound diagrams is the fact that the angles between the mean cleavage plane and the constructed original trend of the small and large folds on either side of the Ribagorzana valley is the same (53° and 56°), although the mean cleavage makes an absolute swing in strike of nearly 50° from one side of the valley to the other. This swing in strike might, of course, be an original tectonic feature caused by an unequal distribution of stresses during both the main phase and the pre-cleavage folding. In that case the constant angle between the pre-cleavage folds and the local cleavage trend is purely accidental. However, it is far more likely that the entire structure was bent only after the main phase and then around an axis coinciding with the actual course of the Ribagorzana River. A simple fault parallel to this axis must be excluded, because a competent sandstone ridge between the villages Vilaller and Ginaste crosses the river without any disturbance. Moreover, a large number of cleavage measurements on the first outcrops on either side of the river alluvia showed that the swing in strike of the cleavage is really gradual (see Appendix III).

Accepting the hypothesis of a bent structure whose influence is perceptible up to the Maladeta granodiorite (including the metamorphic Bono area in the axis of this bend), it is obvious that this movement must have happened later than the main phase but certainly before the intrusion of the granodiorite, since this batholith intruded parallel to the already bent cleavage structure (see page 183). There may very well be a relationship with the second (N-S) or third folding phase (NW-SE and NE-SW) of the tectonic scheme of deformation given by Zwart (1963).

To return to the large pre-cleavage folds mentioned above, it is striking that they have here a NW-SE trend although in the Baliera Unit only NE-SW trends have been observed. We cannot yet give a plausible explanation of this divergence, but it seems possible that the cause of this local trend is to be sought in a former basin rim of the Baliera basin.

The variations in dip and strike of the s_1-cleavage, measured fold axes, δ-lineations, bedding planes, and constructed fold axes of pre-cleavage folds, are shown in the diagrams of Figs. 43 and 44.

Fig. 45. Folded s_1-cleavage in the Mañanet Griotte, about 1 km NE of Castanesa.

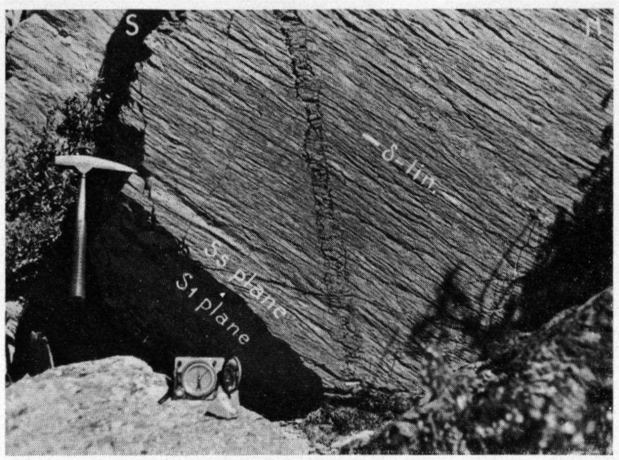


Fig. 46. True knick-zone in the Mañanet Griotte.

Minor folds of the s_1-cleavage with sub-horizontal WNW-trending axes were observed in one locality about 1 km northeast of the village Castanesa. The sub-angular and rounded close folds of the cleavage found in the griotte (Fig. 45) are suggestive of an s_2-folding comparable to the one described in the Sierra Negra Unit (page 187). A distinct crenulation cleavage is, however, absent, and therefore we cannot exclude the possibility that these folds represent knick-zones, whose hinges can be rounded in thicker foliated rocks (Hoeppener, 1955; Boschma, 1963).
Besides these rounded cleavage deformations we encountered true knick-zones, comparable in shape, strike, and dip with those seen in the Sierra Negra Unit (Fig. 46). For further details concerning this deformation, we refer to Boschma (1963, pp. 166—169).

The Carboniferous shale belt

In this monotonous sequence of mainly shale and sandy shale, only a few mappable competent beds were encountered (see Appendix I). Although these harder ridges can be followed for several hundred metres in the field, they do not give sufficient information about the major fold pattern in his area. The structural interpretation of this shale belt therefore depends almost exclusively on the general cleavage trend and on minor tectonic features, such as measured and constructed fold axes and δ-lineations.
The outstanding feature in these Carboniferous rocks is the moderately northward dipping cleavage. Its variation in strike and dip can be gathered from the diagrams (Figs. 43 and 44). Minor cleavage folds occur frequently; both axes and δ-lineations show a large spread in the cleavage plane (see diagrams), suggesting the presence here too of an oblique folding prior to the cleavage formation. This is confirmed by the constructed axes of adjacent bedding measurements, which indicate a general NW trend of the early structures. Their shape and size are unknown, however, because consistent marker beds are absent.
The cleavage folds encountered are mostly close and seldom tight. This is an indication that the compression was less severe here than more to the north, where the cleavage folds are tight to isoclinal. This difference in amount of flattening is also expressed in the fabric of cleavage samples: the cleavage in samples of sandy slates collected from the Sierra Negra Unit is rather strong (flattening of about 55%) and is, moreover, equally developed in all sections perpendicular to the cleavage plane. Samples of more or less the same rock type collected from the southern shale belt, however, show a good parallel fabric only in sections taken perpendicular to the cleavage but parallel to the bedding. Sections taken perpendicular to both cleavage and bedding show a very irregular pattern without a good parallel orientation of the micas. The recrystallization and reorientation of the original sediment fabric is therefore less complete than in the more strongly flattened Sierra Negra Unit to the north. However, it must be noted that this cleavage in the Carboniferous is nonetheless much stronger than the pure fracture cleavage locally encountered in the unconformable Triassic rocks (see under Alpine tectonics, pp. 206—211).

Fig. 47. Knick-zone-like folds of the s_1-cleavage in Carboniferous slates about 1 km NE of Castanesa.

Minor folds of the s_1-cleavage with sub-horizontal WNW-trending axes occur in the vicinity of the doubtful s_2-folding in the griotte mentioned above. The hinge zone of these folds is angular to sub-angular (Fig. 47), but here too a distinct crenulation cleavage is absent. Although these shapes differ somewhat from the true knick-zones encountered in other localities, we do not exclude the possibility that they represent the same phenomenon.
For further details concerning cleavage folds, δ-lineations, and knick-zones, as well as an analysis of these features, we refer to Boschma (1963).

SUCCESSION OF THE HERCYNIAN DEFORMATION PHASES

At the conclusion of the individual description and analysis of the three major structural units in the map-area, it may be useful to summarize the different phases and directions of deformation in chronological order as follows:

1. Pre-cleavage folding phase.
2. Main folding phase.
3. N-S deformation.
4. Intrusion of the granodiorite and accompanying metamorphism.
5. s_2-folding and thrusting.
6. Knick-zones.

1. *The pre-cleavage phase* is characterized by concentric folds without development of cleavage. This phase was active all over the map-area. In the Sierra Negra Unit the folds are probably large, with steep E-W- to WNW-ESE-striking axial planes. In the Baliera Unit, where these folds are preserved best, they are very large, NE-striking, and inclined slightly to the southeast. In the western part they are gentle to open folds and in the centre close to tight folds; in the eastern part of the area they become very tight to isoclinal. In the Ribagorzana Unit the best-preserved pre-cleavage folds occur in the centre and southeastern part of the area. The folds are steep and tight, showing a peculiar NW-SE trend. They vary in size from one metre to more than a kilometre.

2. *The main phase* is characterized everywhere by tight to isoclinal folds with an axial plane cleavage. The axial planes have a general WNW-ESE strike 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 where moderate northward-directed dips dominate. In the Ribagorzana Unit these moderate dips may locally become as low as 30°. The degree of the 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.

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. *The intrusion of the granodiorite* and its accompanying dyke swarm in this compound structure was greatly facilitated by the pronounced cleavage. The intrusion produced a metamorphic aureole of moderate width, and gravity folds were formed mainly in limestones near its southern (and northern) borders.

5. *s_2-folds* which deform the s_1-cleavage about horizontal WNW-striking axes and are clearly post-intrusional, are accompanied by fracture cleavage (crenulation cleavage). This type of deformation is clearly developed only near the Castanesa massif and its western extension. The deformation is a shear phenomenon in which the northern upthrow occurs along moderate to steep, northeasterly-dipping fracture cleavage planes. This mechanism caused small as well as large inclined shear folds, boudinage, and rotated chiastolite crystals and hornfels chips in a carbonate matrix. Some of the larger overthrusts may have originated during the same period.
This deformation phase may well represent the late E-W refolding, but an Alpine age cannot be altogether excluded.

6. *Knick-zones,* which occur sporadically throughout the area, indicate tensional stress, whereas the abovementioned 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 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 cleavage.

ALPINE TECTONICS

The Alpine tectonics can be analysed from the deformation of post-Hercynian rocks, which in the mapped area are mainly of Triassic and locally of probably Permian age. More to the east, outside our geological map, the first post-tectonic sediments are of Stephanian and, locally, of Upper Westphalian age. According to the type of deformation involved, we can subdivide the outcrops of post-Hercynian rocks in our area into three groups:

1. The northern outcrops in the Sierra Negra Unit, which are always bound to thrusts.
2. The central outcrops located near the boundary between the Baliera and Ribagorzana Units. These are mainly simple asymmetric synclines west of the Baliera River, but become overthrusted east of this river.
3. The continuous steep wall of mainly Triassic forming the northern border of the Nogueras Zone.

1. *Northern area*

The northern Triassic, which consists mainly of incompetent and slightly metamorphic marly limestone, cavernous limestone, and dolomite, is — except for one outcrop — only extant in fault zones (thrusts and one normal fault). The sediments have been strongly tectonized (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 amount of throw of the different thrusts is difficult to evaluate but must be at least several hundred metres. The extension of these thrusts and their dip can be gathered from the geological map and cross-sections (Appendices I and II).
The large patch of Triassic west of the Baliera, with its monoclinal shape, is the only outcrop in this northern part to have survived outside a fault zone. This patch is a remnant in an overthrusted Alpine graben. As we have already explained on page 189, its asymmetric anticlinal shape is most probably due to a squeezing of the entire wedge of Palaeozoic and Triassic rocks between the Basibé normal fault and the Salinas thrust (see section 3, Appendix II).
Although this wedge is limited by faults, the bending of the Triassic and Palaeozoic strata must be the result of a more or less supple deformation, however,

without the development of an obvious secondary cleavage. But as far as we know, this is the only example in our area in which the Palaeozoic reacted to the Alpine stress in a supple way. It usually reacted by shear and fault movements.

2. *Central synclines*

The central outcrops, consisting exclusively of "Buntsandstein", show an increasing grade of deformation from west to east. For a better understanding of the most compressed eastern part, we will describe the western, central, and eastern areas separately:

A. *Synclines west of the Baliera River.* — The Triassic outcrops west of the Baliera River generally have an asymmetric synclinal shape, with a horizontal to slightly northward-dipping southern flank and a steep to locally overturned northern flank (Fig. 48). These synclines are cut by a coarse fracture cleavage S_A (Figs. 49 and 50), which in the finest sediments lies roughly parallel to the axial plane of these folds. In more competent beds such as sandstones, this fracture cleavage is orientated more perpendicular to the bedding and is also more widely spaced; it may also be absent. This fracture cleavage is strongest in the steep northern limbs and in the axes of the synclines. It is much weaker in the southern limbs and, moreover, looses its importance with increasing distance from the plane of unconformity. The orientation diagrams of the bedding and the fracture cleavage of these three synclines are shown in Fig. 51. Thin sections of mud- and siltstones cut perpendicular to this fracture cleavage and parallel to the δ-lineation, show a faint orientation of fine sericite parallel to the cleavage. In coarser sediments, such parallel orientation of sericite occurs only next to macroscopically-visible fracture planes. In sections taken perpendicular to both cleavage and bedding, this parallel arrangement of sericite crystals is less regular or totally absent. These findings indicate that the reorientation of micaceous material into a cleavage plane is in a juvenile stage in which the sedimentary fabric is still very important.

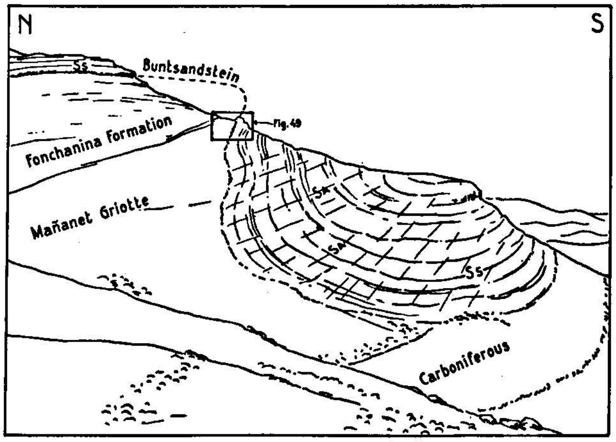


Fig. 48. Asymmetric Triassic syncline with coarse fracture cleavage (S_A) south of the Montañeta de Denúy. Drawing after photograph.

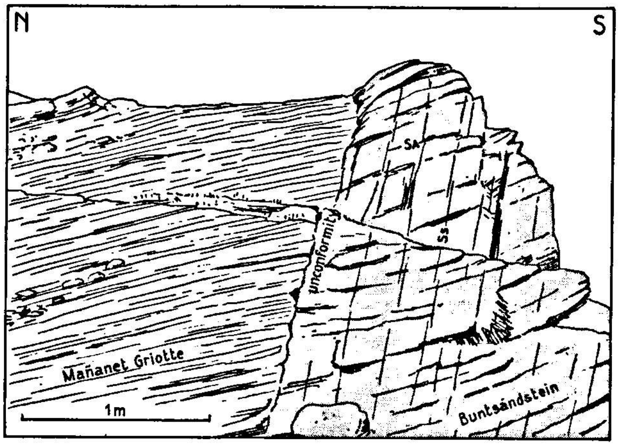


Fig. 49. Detail of Fig. 48. Fracture cleavage in the Triassic, parallel to the slaty cleavage of the Devonian griotte. Drawing after photograph.

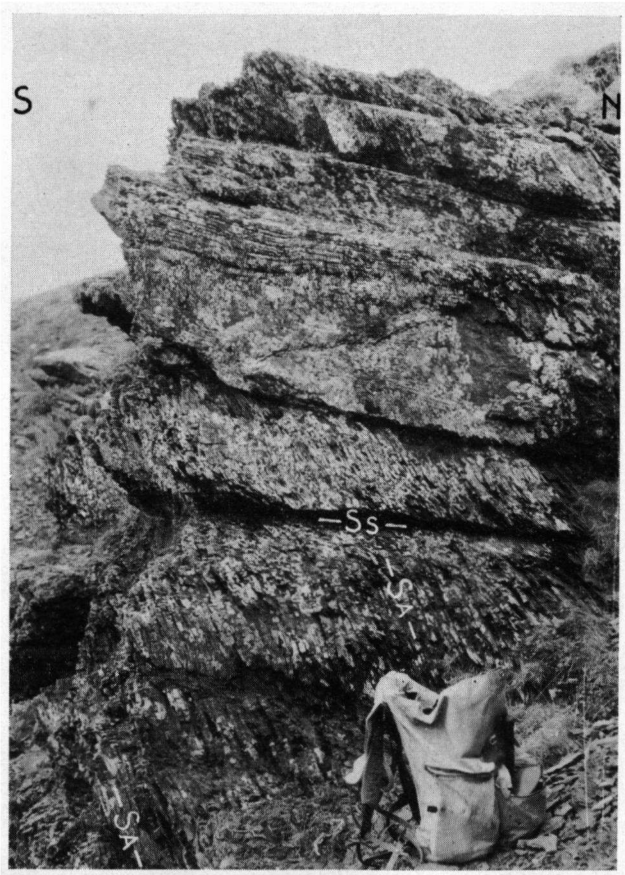


Fig. 50. Coarse fracture cleavage in siltstones of the Triassic of the Central Synclines.

These synclinal shapes and the presence of a moderately steep to rather steep fracture cleavage parallel to their axial planes, indicate a shortening of the Triassic strata in a horizontal direction. Since, as is obvious in the field, the unconformity is not a plane of movement, the Palaeozoic underground must have been shortened to the same degree as the Triassic strata. The main-

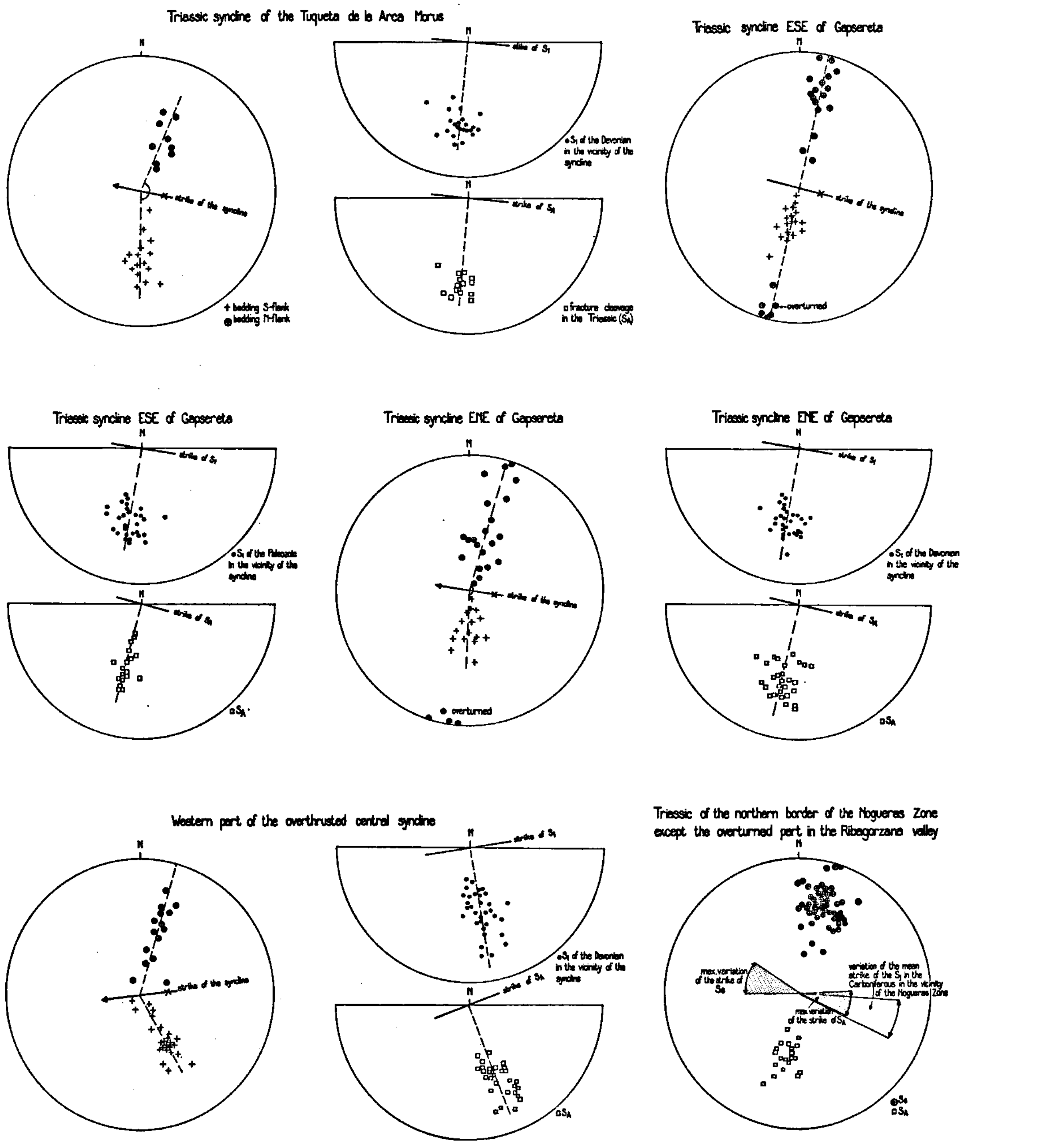


Fig. 51. Orientation diagrams of the fracture cleavage (S_A) and the bedding (ss) of the Triassic, compared with the main phase cleavage (s_1) of the Palaeozoic underground.

phase cleavage of the Palaeozoic underground has a constant dip and strike (no fanning nor refolding) below both limbs of the Triassic structures and moreover lies parallel to the fracture cleavage of the Triassic (Figs. 49 and 52). Consequently, this cleavage must either have been flattened again (Mattauer, 1964 and 1966) or used as a plane along which differential movement took place. A combination of these mechanisms is also possible. It is impossible to demonstrate a secondary flattening along a pre-existing cleavage plane where no refolding took place. But the fact that the cleavage plane served locally as a plane of movement related to the folding of the Triassic strata, can be demonstrated in several good outcrops about 2 km south of the Montañas de Denúy (coord. 4°18½′. 42°32½′). At this location we observed many examples in which a competent ridge (e.g. the Basibé Formation) penetrated like a battering-ram (moving along faults parallel to the cleavage) into the unconformable Triassic strata, causing undulations, folds, and overthrusted folds (Fig. 52, and sections 1 and 2, Appendix II). These Triassic synclines are much more regular

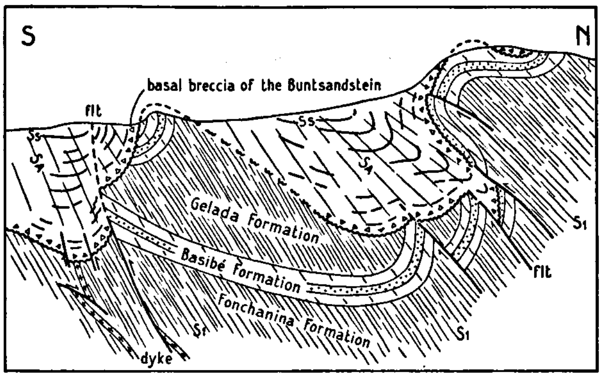


Fig. 52. Field-sketch of two Triassic synclines showing the relation between the folding above the unconformity and the faulting parallel to the s1-cleavage in the Devonian below.

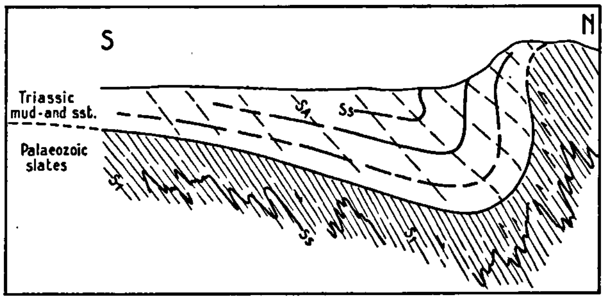


Fig. 53. Schematic representation of the development of an asymmetric Triassic syncline above schistose rocks of the Palaeozoic.

and smooth when they occur above finer and therefore better-cleaved material, such as the Fonchanina slates (Fig. 48); this may be an indication that in these well-cleaved rocks the shear movement took place along many individual cleavage planes rather than along one or two fault planes (Fig. 53).

From the foregoing considerations it follows that these Triassic folds are more shear folds than concentric folds, and that the accompanying fracture cleavage is a shear cleavage comparable to the secondary cleavage (s_2) encountered in the Castanesa massif (page 187). These folds must have been caused by a N-S-directed horizontal compression phase that shortened the Palaeozoic underground as well, although in a rather different manner: while the Alpine compression (which at least in this part of the Pyrenees has the same general direction as the Hercynian compression) reactivated such pre-existing planes as cleavage and fault planes in the Palaeozoic underground, thus causing a local upward movement of the north, the Triassic strata underwent a more supple deformation resulting in asymmetric folds.

B. *Overthrusted syncline between the Baliera and Ribagorzana rivers.* — The large outcrop of "Buntsandstein" between the Baliera and Ribagorzana rivers represents in general a NNW-dipping flank of a huge syncline which was overthrusted from the north. This thrust (Estét thrust) is rather steep near the Baliera River and has an oblique trend as compared with the Triassic syncline, whose northern limb is still extant about 1 km northeast of the village Fonchanina. Towards the east, this thrust becomes flatter and also roughly parallel to the strike of the Triassic structure (sections 4, 5, 6, and 7, Appendix II).

The general 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 the southern flanks are cut by a well-pronounced fracture cleavage with a dip to the NNW. Its characteristics in thin sections are comparable to those of the fracture cleavage of the western synclines.

For the more regular western part of this Triassic structure, orientation diagrams of the bedding, the fracture cleavage, and the main-phase cleavage of the surrounding Palaeozoic underground are shown in Fig. 51.

An explanation similar to the one given above for the western synclines could also be used to explain the shape of this structure: an initial compression resulting in an upward shear movement along cleavage planes probably formed an asymmetric syncline with an axial-plane fracture cleavage comparable to the structures west of the Baliera River. Further compression beyond the maximum point of simple folding initiated the Estét thrust, which lies roughly parallel to the general cleavage trend of this part of the Palaeozoic underground, along which further movement could take place. This thrust probably started in the axis of the syncline, but near its western end, where the thrust becomes steeper, a slight deviation of the Triassic structure occurred. Once the fault had traversed the rather competent Buntsandstein formation, further gliding was greatly facilitated by the incompetent series of the higher Triassic (e.g. marls and gypsum).

The small patch of Triassic on the Pico Comadelo (coord. 4°22′.42°31′) also has a generally northward dip and is accompanied by a rather steep, northward-dipping fracture cleavage. In the north this patch is bordered by the Alpine Comadelo thrust, which has a throw measuring at least several hundred metres.

The outcrop of Triassic in the wedge between the Bono and Estét thrusts, southwest of the village Estét, is strongly folded and also accompanied by a fracture cleavage. The relatively poor outcrops do not permit a reliable interpretation of the complicated structure.

C. *Overthrusted structure east of the Ribagorzana River.* — The elongated outcrop of "Buntsandstein" on the western 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 and 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.

In this particular case of rather constantly dipping strata with a few asymmetric minor folds and limited by a large overthrust in the east, it is hard to tell whether a reverse flank like that in the western outcrops ever existed here. But even without this knowledge it is obvious that a severe compression must have caused these structural phenomena. The maximum shortening took place in a roughly NE-SW direction. The amount of throw of the Bono thrust is certainly larger than its throw on the western side of the Ribagorzana. Estimations go as high as at least one or two kilometres.

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 Appendix III). 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 20 and 75°) 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 (Appendix III). 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.

Because of the inaccessibility of this part of the area, caused by the steep relief and local woods, only scattered measurements are available, and these are moreover insufficient to join into an orientation diagram.

3. *Northern border of the Nogueras Zone*

The "Buntsandstein" and the "Permian" of the northern border of the Nogueras Zone form one continuous WNW-striking wall with a general southward dip of some 45 to 70° (Fig. 51, last diagram). Faint undulations are found locally in this southward-dipping wall (Fig. 54). Details of the contact are shown in Fig. 55. The lower part of the strata is cut by a coarse fracture cleavage having the same strike as the bedding but a reverse (northward) dip. This fracture cleavage

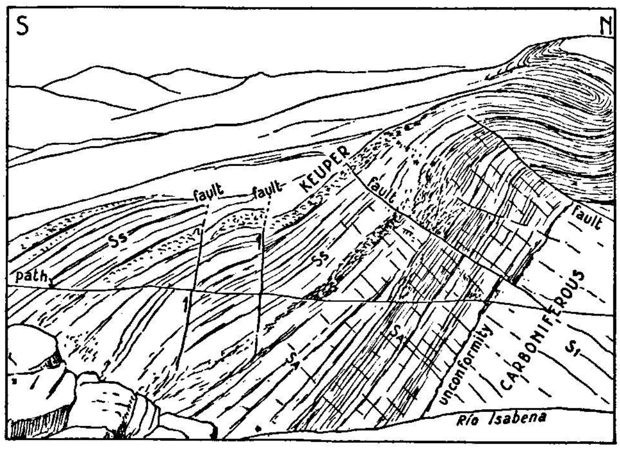


Fig. 54. Triassic of the northern border of the Nogueras Zone showing fracture cleavage, minor faults, and folds. Drawing after photograph.

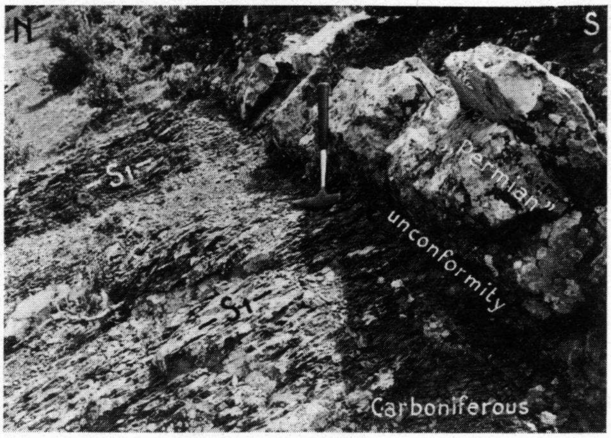


Fig. 55. Unconformity between cleaved Carboniferous siltstones and thick-bedded micro-breccia of the "Permian".

is strikingly parallel to the main-phase cleavage of the Carboniferous (Fig. 51, last diagram). Very often the plane of unconformity is displaced over one or two metres along small faults parallel to this fracture cleavage (Fig. 54). In the Carboniferous the same movement must have taken place along the pre-existing s_1-cleavage. From the unconformity upwards, this fracture cleavage becomes weaker, and it is completely absent in the top layers of the "Buntsandstein". It is therefore evident that the structures of the Triassic are strongly related to the well-cleaved Palaeozoic structure below. An upward 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 south-dipping wall of "Buntsandstein". A particular outcrop in the Isabena valley made it possible to demonstrate that the fracture cleavage of the Triassic is not only a shear phenomenon but had also caused a slight flattening of the strata. We noticed

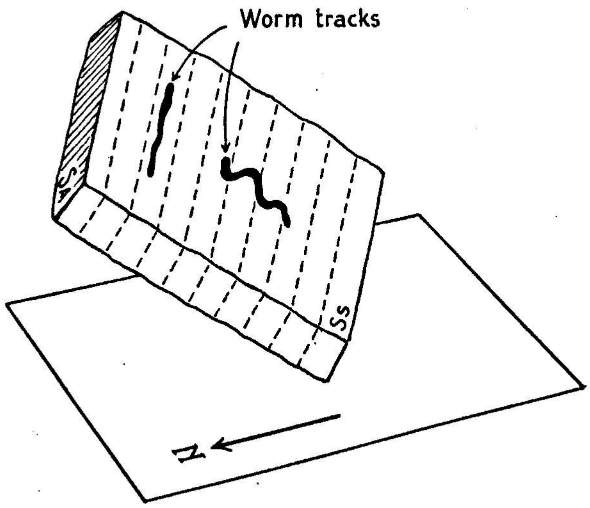


Fig. 56. Worm tracks in the Triassic, parasitically folded when orientated perpendicular to the fracture cleavage (S_A). Drawing after hand-specimen.

that worm tracks on the bedding plane orientated parallel to the cleavage were rather straight, but that these tracks were crumpled when orientated perpendicular to the cleavage (Fig. 56). From three measurements we calculated a flattening ranging between 10% and 20%.

An important deviation from the generally south-dipping wall of "Buntsandstein" is found on the lower slopes of the Ribagorzana valley, where there is a locally overturned northward dip of the Triassic strata. This divergent part 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 (Fig. 57), 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.

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 locally have also played an active role, but this cannot be proven. At the same time, the unconformable Mesozoic sediments were steepened, folded, cleaved, and overthrusted from north to south. These overthrusts may be the roots of the gliding masses encountered locally in the Nogueras Zone.

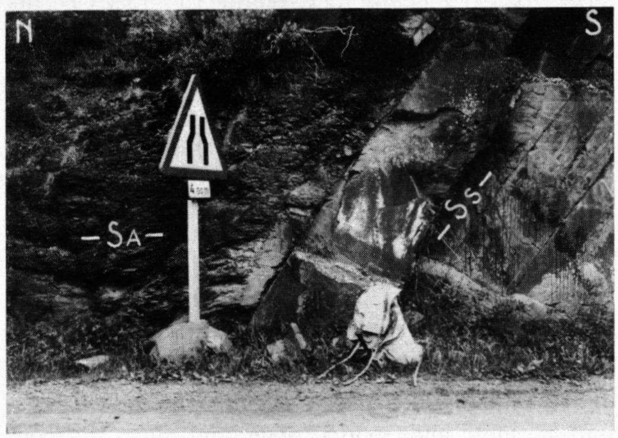


Fig. 57. Exposure of overturned Triassic strata with a sub-horizontal to slightly southward-dipping fracture cleavage in the Ribagorzana valley.

CHAPTER III

INTRUSIVE AND METAMORPHIC ROCKS; ORES

INTRODUCTION

The area of the present study shows a considerable variety of intrusive rocks, such as 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 ophites in the Triassic of the Nogueras Zone have already been discussed (page 177). 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.

THE MALADETA GRANODIORITE BATHOLITH AND ITS METAMORPHIC AUREOLE

Maladeta granodiorite

Of the Maladeta granodiorite, which is one of the largest intrusive batholiths of the Pyrenees (almost 400 sq. km), only a small part of the southern rim is found in the map-area. It is accompanied by a few smaller stocks, such as the two strongly altered bodies in the Sierra Negra mountains, the one situated northwest of the Lago Llausét, and the elongated mass about 3 km northeast of the village of Senet. As has already been discussed (pp. 183—188), the diapiric intrusion of this large body must have taken place in a late phase of the Hercynian orogeny, clearly subsequent to the main phase but very probably before the E-W refolding.

The granodiorite is a medium-grained, light-grey rock, usually massive in outcrop and hand specimen, and very homogeneous except for a few basic clots. 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 (1963).

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—

50% of the rock volume. The microcline is rarely idiomorphic and occurs in smaller quantities than the plagioclase (15—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—50% of the rock. Biotite (5—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.

The intrusive Barruera stock in the Tor valley, shown in the western part of the structural map (Fig. 15) but lying outside our geological map, is much finer grained and also bears more hornblende.

Contact aureole

The thermal methamorphism of the Maladeta granodiorite has produced a contact zone of moderate width, varying in the mapped area between 1.5 and 3 km. In this zone the argillaceous rocks have been converted into hornfels and spotted slate and the calcareous rocks into calcite marble and lime-silicate rocks.

The hornfelses 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 hornfelses 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. Chiastolite seems to be the only newly formed mineral.

Where pure limestones were present in the contact aureole (e.g. the Castanesa limestone) they have been converted to white marble in which the calcite has been completely recrystallized into coarse crystals. This marble often contains irregular, dirty-brown dolomite bodies (with diametres 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 titantite. 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 area 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 central 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 precleavage structures, e.g. on the western slope of the Ribagorzana valley. Since the dykes are covered unconformably by the "Buntsandstein", 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 porphyric texture. The host rock in 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. The phenocrysts consist mainly of:

Plagioclase: albite to andesine, in idiomorphic to hyp-idiomorphic crystals measuring 0.5 to 4 mm, mostly with oscillatory zoning, strongly altered to sericite and occasionally replaced by potash feldspar. The plagioclase forms 25—40% of the rock volume.

Biotite: in idiomorphic to hyp-idiomorphic crystals, 0.5—3 mm, usually totally altered to chlorite; partial replacement of biotite by quartz and rims of ore was occasionally observed.
Quartz: rare, but sometimes replacing biotite.
Occasionally, brown hornblende and apatite.
The matrix (up to 50% of the rock volume) generally cannot be determined on account of the intense seritization, but where this process was less intense the matrix consists mainly of plagioclase (0.05—0.01 mm), some probable potash feldspar, calcite, apatite, zircon, titanite, ore, and a very little quartz if any. The dykes have a (quartz-)dioritic composition (containing less quartz or more plagioclase than the granodiorite); they may be called (quartz-) diorites porphyrites.

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 therefore represents the strongly metamorphosed roof of an intrusion. The mapping of the non-metamorphic country-rock on the eastern slope of the Ribagorzana valley revealed that all lithostratigraphical 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 hornfelses, 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 "Buntsandstein", dates the metamorphism as post-main phase but pre-Triassic. Whether the metamorphism was contemporaneous with the intrusion of the diorite dykes or postdates 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 mineralization of the galena (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 hornfelses: 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 (Appendix III).

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.
The belt of Aneto slates west of the Collado de Basibé contains irregular lense-shaped bodies of brown rock, some 100 sq.m in outcrop and strongly altered, which have about the same dioritic composition as seen in the dykes.
Locally, we encountered cleaved, folded, and boudinaged dykes (page 186 and Fig. 29) which constitute evidence that a compression phase (E-W refolding or Alpine movements) occurred after their intrusion, which was certainly post-main phase, probably even later than the main intrusion of the granodiorite (see page 184). Although a direct relationship 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. 14); 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

Home Syncline, where they are occasionally situated on fault lines.
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. 4°25'.42°34½').
Microscopically, the rocks are usually porphyritic, occasionally non-porphyritic, unorientated, with a fine-grained matrix (0.01—0.1 mm) of mainly quartz and plagioclase with interstitial microcline. As phenocrysts are found:
Plagioclase: albite-oligoclase, in hyp-idiomorphic or xenomorphic crystals measuring 1—1.5 mm, sometimes with oscillatory zoning and only occasionally slightly altered to sericite-muscovite.
Quartz: xenomorphic crystals, up to 1 mm in diameter, usually with a wavy extinction.
Microcline: hyp-idiomorphic or xenomorphic crystals, up to 1 mm in diameter, with cross-hatched twinning and often replacing plagioclase. It is interesting to note that the phenocrysts are often deformed.
Other components are: allanite, rutile, and ore. As secondary products we found chlorite, clinozoisite, and muscovite (replacing e.g. microcline). 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). 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. 14 is not evident. Mineralogically, they are composed mainly of 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 aplite dykes.

BASIC VOLCANIC VENTS

Evidence of a third and completely different type of magmatic activity, i.e. sub-volcanic, is found in rather limited areas about 1 km east and northeast of the village of Denúy (coord. 4°20'.42°30') 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 carbonates which are 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-porphyrite dyke. These fragments of late Hercynian dykes indicate that this basic complex must have intruded later: a Stephanian age seems the most likely. Moreover, we encountered a granulite-like block containing plagioclase, scapolite, strongly-altered olivine, pyroxene, and garnets with large coronas, which may indicate a deep origin of the igneous material.
Besides these coarse tuffs with enclosed blocks of the host rock and lava, there are also more competent brecciated masses with irregular contours and a diameter of several metres. The components of this breccia are the same as those of the above-mentioned blocks; the greater degree of competence is due to a more intensive replacement of the carbonate cement by quartz.
The igneous rock forming the bulk of this intrusive mass is supposed to be a vesicular basalt with a fair amount of small xenoliths and containing phenocrysts of pyroxene and probably olivine in a groundmass of feldspar, pyroxene, and ore. A large part of these minerals has been replaced by carbonate, which also fills the vesicules, and by chloritic minerals and biotite. Locally, in the host rock neighbouring these basic, intrusive bodies, we found very thin (10—50 cm) dark-green dykes of a comparable basaltic rock con-

taining phenocrysts of pyroxene, probably olivine, probably plagioclase, and biotite, in a groundmass of pyroxene and feldspar. The rocks of these dykes are strongly altered. They are not shown on the geological map.

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. They may be related to the Stephanian volcanism so widely encountered in the Nogueras Zone to the south. A much more thorough petrological investigation is required to prove this, however.

ORES

The *El Cierco* lead mine southeast of the village of Bono, exploited by the M.I.P.S.A., is the only mine in the mapped area still being worked. The abandonned smaller mines (mainly iron and copper) indicated on the geological map were, except for the copper mine 400 m SSE of the hamlet Artiga (4°25′.42°30½′), small workings of no economic importance.

The mined galena (PbS) occurs in a number of roughly E-W-striking veins, discordant in respect of the bedding of the host rock and varying in thickness from several centimetres to 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 planes, faults, or irregular cracks in the host rock. Since these ore veins are cut by the unconformity of the "Buntsandstein", the mineralization must have occurred in Hercynian times.

The galena, which bears small amounts of silver (1 kg silver to 1 ton pure lead) is accompanied by such constituents as sphalerite, pyrite, chalcopyrite, and stibnite, as well as supergenic covellite, malachite, and azurite. The matrix is always calcite.

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 galena is mined in a number of horizontal galeries (about 20) situated between 300 and 600 m above the valley floor, and is carried by cable to the crushing and washing plant in the lower part of the Ribagorzana valley, about 1 km east of the village Estét, further transportation being done by trucks.

CHAPTER IV

GEOMORPHOLOGY

PRE-GLACIAL RELIEF

The Pyrenean (late Eocene) orogenetic phase must have been immediately followed by active erosion and denudation of the axial zone, which caused the enormous accumulation of coarse conglomerates, mostly of Oligocene age, in the southern part of the Pyrenees.

High plateaus found in the Central Pyrenees have long been considered to be well-preserved remnants of an old denudation surface. On both sides of the chain some characteristic and constant denudation levels can be distinguished, the one situated between 1900 and 2300 m being the best known (Nussbaum, 1935, 1938; García Sainz, 1940 a & b; Kleinsmiede, 1960; and Zandvliet, 1960). This surface, which is also well-developed in the Valle de Arán, has been dated as either Upper Miocene or a little older (de Sitter, 1954b, p. 276 and 1956a, p. 217). Kleinsmiede (1960), however, doubts this dating and thinks that the sequence of lignite-bearing conglomerates, sands, and clays of Upper Miocene age was downfaulted before the denudation surface was formed; in that case the surface would be post-late Miocene.

Owing to later epeirogenetic movements and headward erosion of the rivers in our area, remnants of denudation levels are scarcely recognizable. One such remnant might be the flat area in the Sierra Negra region situated at an altitude of 2500 to 2700 m, which is developed exclusively on a Silurian subsoil. Broad undulating ridges, heavily weathered, alternate with gently-dipping slopes whose lower parts have been deeply dissected by fluvial erosion. Both Zandvliet (1960) and Kleinsmiede (1960) describe some high denudation plateaus on more or less the same level.

A second, still more doubtful indication of a former aplanation surface at a lower level is to be seen in the middle part of the course of the Llausét River. Some 40 m above the upper limit of the (fluvio-)glacial deposits, lying at a level of 1860 to 1880 m, four flat ridges separated by the Llausét River and two other brooks, can be visualized as having originally constituted a nearly horizontal plane. To the west and southwest, this flat surface grades upwards into a gently-dipping slope, which at 2200 m forms the lowest point of the watershed with the Upper Baliera River (Collado de Salinas). On the eastern slope of the Ribagorzana valley no corresponding flat surface at the same level could be detected.

However, the fact that the lowest part of the watershed between the Ribagorzana and Tor rivers (Collado de Gelada: 2060 m) is at more or less the same altitude as the Collado de Salinas (2180 m) and the Collado de Basibé (2240 m) does not constitute sufficient evidence to consider them as remnants of a continuous planation surface; more reliable and conclusive data can only be provided by a study of such features over a much larger area, based on good maps and aerial photographs.

GLACIATIONS

During the Quaternary period glaciers were formed and moved downwards, guided by the more important valleys. 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, which left few traces. According to García Sáinz (1940b, p. 368), the latest glaciation, 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 such detail that different glaciation periods could be deduced from the few remaining morainal deposits. We must therefore be content for the time being with the description of the most conspicuous glacial features, irrespective of 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 area, were once occupied by moving glaciers. In the Upper Baliera and Isabena 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 becomes quite logical when we compare these snow-accumulation areas with those of the Ribagorzana and Tor glaciers, the former being much smaller and situated at a lower altitude and more to the south. The enormous amount of debris, especially in the upper part of the Baliera valley, is also an argument 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 hornfelses and marbles; 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, only part of which is shown on our geological map.
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-Granite contact zone.

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

A beautiful example of a large glacier-basin or *Karplatte* (Nussbaum) is the rather flat area on which the Lagos Anglios (coord. 4°24′.42°35½′) are situated. This basin is a broad, gently-sloping excavation with a flat bottom and steep walls. Via only one distinct rock step, this basin passes into the beautifully U-shaped valley of the Barranco de Salenca.

In the Ribagorzana valley south of the granite area, narrow parts alternate with round and wide basins. These narrowings are due to the stronger resistance to glacial erosion of limestones as compared with the slates. In the Ribagorzana valley they are, from north to south:

1. The narrowing in the valley in the Devonian limestones north of Senet (coord. 4°27′.42°34½′).
2. The gorge in the metamorphic and strongly-folded Gelada Formation between Senet and Bono (coord. 4°26′.42°33′).
3. The still-existing threshold formed by the Mañanet Griotte south of the village of Forcat (coord. 4°24′. 42°30′). This threshold is several metres higher than its basin upstream. Before this step had been completely cut by post-glacial erosion it must certainly have enclosed a lake now filled by alluvial material.
4. The Carboniferous sandstone-ridge north of Vilaller.

For the southern part of the Tor River, we may mention the narrow part of the valley in the Mañanet Griotte north of the village of Llesp (coord. 4°27′. 42°27½′).

Although glacial deposits are found in the Ribagorzana and Tor valleys as far south as Vilaller and Llesp and as high as 260 to 360 m above the present valley floor, no distinct terminal moraine has been found in either valley. The narrow V-shape of the valleys at the points where the rivers cross the Buntsandstein formation of the northern border of the Nogueras Zone, as well as the total lack of any morainal deposits downstream from these points, make it highly probable that the glaciers never entered the Nogueras Zone. Nussbaum and García Sáinz are of the same opinion.

Granite boulders, some of them very large, are frequently found on the valley slopes of the Ribagorzana and Tor rivers and their most important tributaries. The larger concentrations of boulders are indicated on our geological map. The larger morainal deposits occur:

1. East and southeast of Senet; altitude 1730 m.
2. In the middle part of the course of the Llausét River; altitude 1750 to 1880 m.
3. In the Barranco de Estét; altitude 1480 m.
4. Southeast of the Bono lead mine; altitude 1490 m.
5. In two separated outcrops north and northwest of Cardet in the Tor River, both at a level of 1580 m.
6. West and north of Iran in the Tor valley; altitude 1310 to 1330 m.

The erosional scarps of these lateral moraines show that the deposits consist of loose granite boulders embedded in argillaceous material. In the north these granite boulders have a diameter of up to 8 m, but in the southern areas they are much smaller. The top parts of these erratics are often somewhat finer and show good stratification.

From the highest level of the lateral moraine deposits the thickness of the former glacier can be deduced:

A. Ribagorzana glacier
 a. Southern granite area: 600 m.
 b. Near the village of Senet: 500 m.
 c. Near the village of Estét: 500.m
 d. Near the hamlet of Gymnaste: 300 m.
 e. Near the village of Vilaller: 260 m.

B. Tor glacier
 a. Near the village of Cardet: 500 m.
 b. Near the village of Coll: 520 m.
 c. Near the village of Llesp: 360 m.

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. A good example of this phenomenon is the threshold in the Mañanet Griotte south of the village of Forcat.

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:

a. Rock steps in a valley, e.g. the rock step of Devonian limestone between Senet and Bono, where the gorge is 700 m long.
b. Hanging valleys, where they discharge in the deepened valley, e.g. the gorge of the Llausét north and northwest of the village of Bono (800 to 900 m long) and, in the same river, the gorge southeast of the Lago de Llausét, with a length of 400 to 500 m.
c. Thresholds (see above).

Restricted to areas with vast Silurian outcrops (e.g. the Sierra Negra area and the Upper Llausét valley) are solid, sub-recent slope-breccias cemented by limonite. The long, gently-dipping slopes of the Sierra Negra are entirely covered with fine black flakes over which rain-water or melt-water still run down in broad fans to form such breccias.

In the Maladeta granodiorite and in the hornfelses of the Fonchanina Formation, 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 toothed.

INFLUENCE OF THE STRUCTURE ON THE PRESENT RELIEF

It is surprizing 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 morpho-genetic phase, as a result of which the rivers run off the southern slope.

On examining the cleavage trend map (Appendix III), it is remarkable to note that the Ribagorzana River more or less follows the sharp bend of the constructed cleavage traces. This bend probably corresponds to a fundamental fracture at greater depth, but wherever structural and stratigraphic features cross the river, there is no evidence whatever of a fault along the main valley. The writer is, however, under the impression that a concentration of joints along the course of the river probably forms some kind of straight lineament, as described by Zandvliet (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 adjusts itself 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 in the Nogueras Zone the gypsum and marls of the Upper Triassic. 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.

Surface creep occurs everywhere in the schistose rock, especially on the slopes carrying slates. 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 continuous heavy rain, the Ribagorzana River caused much damage to land, roads, bridges, and houses; two hundred metres of the main road slid down several metres, and although a new road has been constructed, part of the old, collapsed road can still be seen.

SAMENVATTING

In het goed ontsloten karteringsgebied treft men sedimentaire gesteenten aan, die in ouderdom reiken van Ordovicium tot en met Trias. In deze gehele serie komen fossielen slechts sporadisch voor en dan meestal nog sterk tektonisch vervormd. De palaeozoische gesteenten, allen van marine oorsprong, zijn zeer zwak regionaal metamorf (epizonaal), grillig verplooid en worden doorsneden door een *slaty cleavage.* De discordante Permo-Triadische gesteenten zijn fluviatiele afzettingen, die hogerop via kalken en dolomieten overgaan in evaporietgesteenten. De plooiingssteil van deze mesozoïsche gesteenten is totaal anders dan in de palaeozoïsche kern; slechts vlak boven het hercynische discordantievlak wordt locaal een grove *fracture cleavage* gevonden.
Van onder tot boven is de stratigrafische opeenvolging als volgt:

ORDOVICIUM:
— een fijne afwisseling van leien en kwartsieten,
— een niveau (30—35 m) van mergels en onzuivere kalken,
— een dun pakket (10—15 m) van donkere leien,
— aan de top harde kwartsieten met dunne zandige kalklagen (totaal 20—25 m).

SILUUR is ontwikkeld in de klassieke facies, zoals die geldt voor de Pyreneeën:
— zwart afgevende Graptoliet schalies,
— aan de top inschakelingen van zwarte, dungelaagde Orthoceras kalken.

DEVOON is in het noorden enigszins anders ontwikkeld dan in het zuiden: in het noorden, in het *Sierra Negra faciesgebied,* is het Devoon relatief dun (120—250 m) en de opeenvolging ziet er van onder tot boven als volgt uit:
— een afwisseling van dunne mergelige kalken en siltige leien (*Rueda Formatie:* 50—80 m),
— een kalkbank, locaal goed gelaagd (*Castanesa Formatie:* 30—60 m),
— donkere fijne leien (*Fonchanina Formatie:* 15—60 m),
— een afwisseling van kalken, griottes en kalkschalies (*Mañanet Griotte:* 35—70 m).

In het zuiden, in het *Baliera faciesgebied,* is het Devoon dikker (340—780 m) en meer gedifferentieerd; van onder tot boven is de opeenvolging als volgt:
— hoofdzakelijk donkere, siltige leien *(Aneto Formatie:* 50—200 m),
— een afwisseling van zandige leien, kwartswackes en mergelige kalken (*Gelada Formatie:* 105—120 m),
— een zeer competent niveau, met aan de basis griotte-kalken en dolomieten, gevolgd door dikke en dunne kwartsietlagen en afgesloten door een niveau van zeer dungelaagde, zwarte kalken *(Basibé Formatie:* 40—170 m),
— hoofdzakelijk donkere fijne leien (*Fonchanina Formatie:* ± 60 m),
— bovenin een afwisseling van kalken, griottes en kalkschalies (*Mañanet Formatie:* 80—250 m).

CARBOON bestaat uit een monotone grijze tot bruinige afwisseling van siltige leien, onzuivere zandstenen en zandige kalken; kenmerkend voor deze gesteenten zijn grote sedimentaire glimmers.
Discordant op deze palaeozoische gesteenten vindt men van onder tot boven:

PERM(?):
— rode siltstenen, breccies en knobbelige dolomitische lagen.

TRIAS:
— een rode afwisseling van conglomeraten, zandstenen, silt- en kleistenen (*Buntsandstein:* 150—200 m),
— dungelaagde dolomitische kalken (*Muschelkalk:* tot ± 50 m),
— een bonte afwisseling van mergels, cellige dolomieten, kleistenen en gipsen (*Keuper:* tot ± 150 m). Hierin komen grote en kleine lichamen van basisch gesteente voor (*ophieten*).

De eerste tektonische bewegingen maken zich merkbaar in het Midden Karboon en bereiken hun maximum tegen het einde van het Westfalien. Men kan er verschillende deformatiefasen in onderscheiden:

FASE I. — Er vormen zich concentrische plooien, die in het westen slechts flauwe welvingen zijn, maar naar het oosten toe nauwer tot bijna isoclinaal worden. Een cleavage is nooit aanwezig. In het noorden van het karteringsgebied (*Sierra Negra Unit*) zijn de assenvlakken E-W tot ESE-WNW gericht en over het algemeen steil N-hellend, in het centrum (*Baliera Unit*) zijn deze NE-SW gericht en vrij steil NW-hellend, en in het zuiden (*Ribagorzana Unit*) vindt men steile E-W en NW-SE gerichte assenvlakken.

FASE II. — Dit is de hoofdplooiingsfase der Pyreneeen. In een ongeveer N-S krachtenveld ontstaan nauwe tot isoclinale E-W plooien met een assenvlak cleavage (*slaty cleavage*). Deze heeft in het noorden een steile noordhelling (80—50°) en wordt naar het zuiden toe geleidelijk zwakker hellend (50—30°). De plooiassen en δ-lineaties (snijlijn van cleavage en gelaagdheid) hebben bij projectie op een Schmidt's net een gordelverdeling, welke samenvalt met het gemiddelde cleavage vlak.

FASE III. — De in grote trekken E-W gerichte cleavage structuur wordt om een as gebogen, die ongeveer samenvalt met de huidige loop van de Ribagorzana rivier.

INTRUSIE van de Maladeta granodioriet met de erbij horende contact metamorfose en onmiddellijk gevolgd door dioritische en granitische gangen met een porfirische textuur, die evenwijdig aan de cleavage vlakken indringen. De intrusie van de batoliet veroorzaakt locaal gravitatieplooien. Het economisch ontginbare looderts in de buurt van het dorp Bono is aan deze magmatische activiteit gebonden.

FASE IV. — In een wederom ongeveer N-S gericht krachtenveld wordt de *slaty cleavage* van de hoofdplooiingsfase locaal verplooid (b.v. in het Castanesa massief), hetgeen gepaard gaat met de ontwikkeling van een nieuwe cleavage (*crenulation cleavage*). Op andere plaatsen zijn er aanwijzingen dat het gesteente langs het oorspronkelijke cleavage vlak verder wordt afgeplat of dat langs dit vlak N-S gerichte overschuivingen optreden. Deze plooiingsfase is òf een laat-hercynische (E-W refolding) òf een Alpine plooiingsfase.

ALPINE OROGENESE veroorzaakt of accentueert in de palaeozoische kern belangrijke overschuivingen en kleine opschuivingen parallel aan het hercynische *slaty cleavage* vlak, waarbij in de Permo-Triadische gesteenten boven het discordantievlak asymmetrische plooien tot stand komen.

POST-MIOCENE EROSIE, die op de alpine morfogenetische opheffing volgt, creëert enkele aplanatie plateaus, die op hun beurt door rivieren worden ingesneden en tijdens het Pleistoceen door gletsjers worden verwijdt. Subsequente rivier-erosie en denudatie dragen zorg voor het huidige relief.

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