THE SEDIMENTARY AND STRUCTURAL HISTORY OF THE VALSURVIO DOME CANTABRIAN MOUNTAINS, SPAIN

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(Pre-issued 23 - 5 - 1962)

CONTENTS

ABSTRACT

A continuous sequence of about 1000 m of Devonian sediments has been found in the Valsurvio dome, ranging from probable Siegenian to Famennian in age. This sequence of a neritic-littoral facies shows similar characteristics to the Devonian found further west in the province of León.

North of the Valsurvio dome, a narrow E—W ridge (zone of San Martin—Camporredondo) has been uplifted by a total of 900—1000 m during two periods in the Middle and Upper Devonian. The upheaval of this zone resulted in rapid facies changes, the development of two large hiatuses in the Devonian of this zone and in the totally different facies development north (Rio Arruz area) and south (Valsurvio dome).

During Carboniferous time the San Martin-Camporredondo zone played an important role in sedimentary history. The Carboniferous is subdivided in three groups.

The Ruesga group (Pre-Curavacas folding) is developed in the Valsurvio dome as thick massive limestones of the Caliza de Montaña facies, whereas in the San Martin-Camporredondo zone predominantly clastic, reworked and highly mixed sediments of the Culm facies occur; these lithological differences being due to a subsidence of latter zone during sedimentation of the Ruesga group.

The sediments of the Yuso group (post-Curavacas and Pre-Asturian folding) are very similar to the Culm facies of the Ruesga group.

The Cea group contains at least 1200 m of coal measures, outcropping in a small E—W zone situated at the boundary of the Cantabrian mountains and the meseta of Old Castile.

At least four deformation periods have been distinguished:

1. The Curavacas folding phase which caused isoclinal E—W trending folds and altered the incompetent Devonian shales into slates. In the competent Upper Devonian rocks and limestones of the Ruesga group minor folding has played a less important role. Large isoclinal folds dip 60—40° S. These structures are most typically developed in the Valsurvio dome, but with the facies change into the zone of San Martin-Camporredondo large ESE—WNW tracing low-angle overthrusts developed. The maximum measured thrust movement to the north is about 2.5 km.
2. In the Asturian folding phase contemporaneous E—W and N—S fold directions occur. A crenulation cleavage has been developed in the already cleaved rocks, whereas in the rocks of the Yuso group a slaty cleavage developed, often sub-parallel to the bedding. A late recrystallization of small unoriented porphyroblasts of chloritoid has been observed.
3. During a Post-Stephanian (and probable Pre-Triassic) folding phase broad open E—W folds developed, which caused the updoming of the Valsurvio dome and a reorientation of first and second generation structures. In the zone of San Martin-Camporredondo this folding is presented by minor folds and crenulation cleavages, often both developed as conjugate systems.
4. Tertiary deformation. The epirogenetic upheaval of the mountainous area along a set of border faults has caused a steepening of the Cea group, the

Cretaceous and the lower part of the Tertiary conglomerates in a small E—W running zone in the south of the area.

Together with the development of the WNW—ESE Cotolorno wrench-fault, a flexure like fold has been developed in the Cea group and the Cretaceous. The influence of the Tertiary deformation apart from the wrench faults, is restricted to the southern part of the area.

INTRODUCTION

GENERAL

This study forms a part of a systematic program of geological mapping of the Paleozoic rocks of the southern slope of the Cantabric-Asturian mountains in northwest Spain, by a team of students of the Leiden University, Holland, directed by Prof. Dr. L. U. de Sitter.

The area studied is a direct continuation to the west, of that described by Kanis in his thesis "Geology of the eastern zone of the Sierra del Brezo (Palencia, Spain)".

The Cantabric-Asturian mountain chain, rising to altitudes of more than 2500 m, has an E—W trend. To the north the mountain chain is bordered by the Gulf of Biscay, whereas to the south the mountains slope towards the meseta of Old Castile, which has an altitude of 800—1000 m. There is a general tendency for progressively older sediments, to crop out at the surface from east towards the west. The eastern part, — the Cantabrian mountains, — is built up of Mesozoic sediments (Solé Sabaris 1952). In the headwater region of the Rio Pisuerga (Prov. Palencia), sediments of Carboniferous age come from below the Triassic cover (Nederlof 1959). Further to the west in the province of León, Silurian and Cambriam sediments also occur.

The present study deals with an area situated partly in the province of Palencia and partly in León, between latitudes 42° 55′30″ and 42° 46′54″ north and longitudes 1° 14′32″ and 0° 54′33″ west (App. I). The greater part of the area is drained by the Rio Rivera a confluent of the Rio Pisuerga.

To the south the area is bordered by the meseta of Old Castile or "Tierra de Campos", mainly underlain by Tertiary and Quaternary sediments. A small zone of Cretaceous and coal bearing Upper Carboniferous rocks separates the highland plateau, — altitude 1000 m —, from the Sierra del Brezo, a chain of Carboniferous limestone mountains (altitude up to 2000 m). In this zone at the foot of the mountains, a highway and railway have been constructed, which connects the district with León to one side and with Bilbao, an industrial centre and harbour at the border of the Gulf of Biscay, to the other side.

To the east, the area is bordered by a line through the Peña Redonda to the village of San Martin de los Herreros and links directly with the area mapped by Kanis (1955). To the west the boundary is formed by the interflu of the Rio Grande and the Rio Cea north of Velilla del Rio Carrion and by the Rio Carrion south of this village. The northern limit is formed by an E—W running depression, which can be followed from Valverde de la Sierra in the west via the artificial lake, "Pantano de Camporredondo", towards San Martin de los Herreros and the Rivera valley in the east. The highway Cervera de Pisuerga, via Triollo to Guardo runs partly through this depression and for the other part follows the valley of the Rio Carrion.

The name "Valsurvio dome" is derived from the village Valsurvio, which is situated in the centre of a large domelike structure. The core of the dome is formed of Devonian rocks, mainly consisting of metamorphic slates and clastic sediments, which are rather badly exposed, due to large screes and partly to a dense vegetation of heather, bilberries and scrub-oaks. The "Patremonio Forestal del Minesterio de Agricultura" is carrying out great projects to reforest the mountains with pines, as remedy to the strong erosion. The barren mountains, built up by the Carboniferous limestone, do not show any vegetation. Especially the southern slopes are dry and warm (fig. 1).

Two dams have been constructed across the Rio Carrion in the present area. One, near Camporredondo de Alba in the Upper Devonian quartz sandstones of the northern slope of the dome and a second, between Otero de Guardo and Velilla del Rio Carrion, in the same quartz sandstones but now of the southern flank of the dome structure. These reservoirs have a double purpose; the generation of electrical energy and as water reservoirs for the irrigation of the meseta in the dry season.

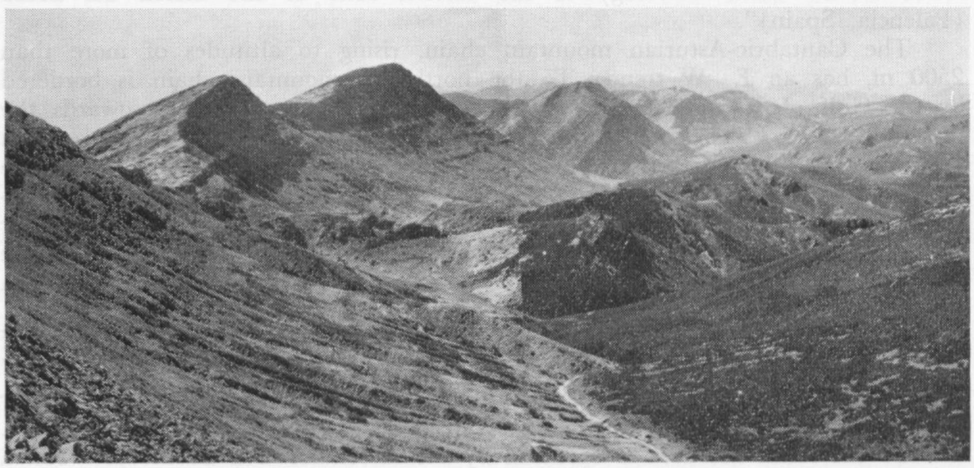


Fig. 1 The southern limb of the Valsurvio dome seen from Peña del Fraile to the WNW.

The mountainous part of the area is thinly populated; the inhabitants living from cattle-breeding. On the south side of the mountain chain, along the road Cervera-Guardo, a large number of villages occur. Here the population lives partly by agriculture (grain) and partly by coal mining.

The most important place in the area is Guardo, which will certainly expand into an important industrial centre in the near future. The coal mining is important. A large chemical industry is in construction (Union de Explosivos). Good communications by railway and highways link Guardo to all parts of Spain. Electrical energy is supplied by generating stations of the reservoirs of Camporredondo and Requejada (a reservoir in the Rio Pisuerga) and in the future will also come from the newly constructed reservoir of Compuerto.

The topographical base for the geological map was the 1 : 50.000 topographic map of the Instituto Topografico y Catastral at Madrid, sheets: 106 (Camporredondo de Alba), 132 (Castrejon de la Peña) and 105 (Riaño). A few modifications have been made. 1) The new artificial lake, the Pantano de Compuerto, has been sketched in by following the 1200 m contour line.

2) A newly constructed road, from Velilla del Rio Carrion to Camporredondo de Alba, has been drawn in, with the help of the aerial photographs. 3) The new name for the village Velilla de Guardo, as found on the topographic map, which has now been formally changed to Velilla del Rio Carrion is shown.

Use was made of the aerial photographs (Run 8839 105—098; Run 8842 016—021; Run 8870 155—161; Run 8960 179—183) of the Ministerio del Aire, placed at our disposal by the Consejo Superior de Investigaciones at Madrid (scale about 1 : 37.500).

Thanks are due to Prof. F. Stockmans (Bruxelles) and Mr R. H. Wagner (Sheffield) for determing the Carboniferous floras, to Mr A. C. van Ginkel, Mr H. A. van Adrichem Boogaert and Mr T. F. Krans for determing respectively the foraminifers, the conodonts and the Devonian spiriferids. Dr W. Struve (Frankfurt) has dated some trilobites.

Special thanks are also due to Mr J. F. Graadt van Roggen who kindly placed field data at my disposal.

I wish to express my gratitude to Mr J. F. Savage for the correction and reformulation of the English text, to Mrs E. B. M. da Costa Gomez—Heiling and Mr J. F. M. Mekel who translated the Spanish summary.

The maps were expertly drawn by Miss C. Roest.

The hospitality of the Spanish population in the area investigated made my stay during field work very pleasant. In this respect I wish to remember especially the Rodriguez family of Velilla del Rio Carrion and the Rebanal family of Santibañez de la Peña.

Furthermore I want to express my appreciation to all those other persons who have helped indirectly to prepare this paper.

PREVIOUS AUTHORS

Nearly all geological studies carried out in the present region, deal with the Upper Carboniferous coal basin, situated on the southern border of the mountain chain; or with the younger strata, cropping out further south of this coal basin. Only recently a few publications have been made about the Devonian core of the Valsurvio dome.

In the year 1856 a geological map of the province of Palencia on a scale 1 : 400.000 was published by Casiano de Prado, revised on a scale 1 : 100.000 in 1861. The Devonian core of the Valsurvio dome was joined up with the Devonian between San Martin and Ventanilla, forming one large area of Devonian rocks as far as Cervera de Pisuerga.

Little attention has been paid to the Devonian of the province of Palencia, in contrast with the intensive research carried out on the Devonian of León and the Asturian regions (a.o. Casiano de Prado 1850, de Verneuil 1850, Mallada 1875, Barrois 1882, Comte 1934—1939 and 1959). The rather uniform lithological development over a great area and the abundance of brachiopods, corals, crinoids and other fauna elements in the province of León have been responsible for this special interest. In the province of Palencia great facies changes occur, and the Devonian is rather badly exposed. Fossils are not as abundant as in the province of León and moreover paleoecological changes occur.

In 1939 Quiring published a paper, dealing with the present area. He gives only a few notes about the Devonian strata, because all fossil faunas were lost during the Spanish Civil War. His geological map of the Carrion basin differs in many details with the present map. The publication deals mainly with the Carboniferous coal basins (Carrion and Rubagón basin). Papers of the same author were published in 1955 and Dahmer and Quiring (1953), about the geology of the eastern adjacent area. Saenz Garcia (1943) has dated the Devonian quartzites, north of Velilla del Rio Carrion as Middle Devonian. Kanis (1955) described the geology of the easterly adjacent area, and gave an account of the important facies changes in the Middle Devonian. In 1960 Kullman has given a valuable contribution to the knowledge of the Devonian stratigraphy, with his study of Devonian goniatites.

Much more has been published over the Upper Carboniferous coal basin between Guardo and Cervera de Pisuerga. Oriol (1876 and 1894) wrote about the Carrion basin [1]) and after him Mallada (1892), who paid special attention to the development of the limestone conglomerates (confolites) in the basin of Valderrueda and their extension to the east in the basin of Guardo. Sanchez Lozano (1906) described the development of the Guardo basin, especially with regard to the coal mining between Guardo and Cervera. His paper of 1912 deals with a drillhole through the Cretaceous into underlying Carboniferous strata, near Vado de Cervera. In a publication by Patac (1924), some mylonites are described from the northern border of the coal basin. His conclusion, that the coal measures were deposited after the Asturian folding phase, seems correct in the light of the present investigations. The opinion of Kanis (1955), that they are pre-Asturian folding, is wrongly based on an unconformity between Stephanian A and Stephanian B sediments in the western part of the Barruelo basin, found by Wagner and Wagner—Gentis (1952). In 1934 Gomez de Llarena described the contact between the Cretaceous and the Carboniferous strata as a thrust fault. A marine fauna in the coal measures near Santibañez de la Peña has been noted by Hernandez Sampelayo (1944). The flora of the Guardo-Cervera basin, collected by the present author and determined by Wagner, was published by Wagner in 1959, in a compilation of several contributions to the Carboniferous floras from northwest Spain.

Several publications by de Sitter (1957, 1958, 1959 and 1961) are dealing with a more general picture of this part of the Cantabric-Asturian mountains.

The Cretaceous, situated south of the Guardo-Cervera coal basin, has been described by Ciry (1939). Figuerola (1953) has investigated the Cretaceous zone and its continuation below the Tertiary cover near Guardo with seismic methods. Mabesoone (1959) described the Tertiary and Quatanary sediments, deposited on the southern border of the mountain range, between the Rios Carrion and Pisuerga.

[1]) "Cuenca hullera del Rio Carrion" is the name used by Oriol (1876) for the coal basin situated between Cervera de Pisuerga and Guardo. Mallada (1892) mentioned it "Cuenca hullera de Guardo", whereas Wagner (1959) has changed it in Guardo-Cervera coal basin, which name also has been used in the present paper.

CHAPTER I

MORPHOLOGY

RELIEF

The present relief in the Cantabric-Asturian mountain range is due to vertical upheaval in Tertiary times, followed by an intensive erosion. The products of this erosion have been deposited directly south of the mountain belt, along the northern border of the meseta of Old Castile and were studied south of the present area by Mabesoone (1959). Most authors assume that the morphogenetic uplift is of Oligo-Miocene date (a.o. Richter & Teichmüller 1933, Solé Sabaris 1952), Stickel (1930) has found in the mountain range remnants of planation surfaces, uplifted in the Oligocene or Lower Miocene, which are contested by Nossin (1959, p. 377).

THE RAÑA AND PIEDMONT ALLUVIAL PLAINS

The flat surface south of Guardo is known in literature as the "Raña of Guardo" (Birot & Solé Sabaris 1954, Mabesoone 1959, Lautensach & Mayer 1961, and others). A synopsis of the investigations concerning raña deposits has been published by Oehme in 1936. In most recent studies a post-Pontian planation level has been accepted, followed by deposition of raña material in the Villafranchian.

The raña material consists of coarse quartzitic angular and subangular material of a yellow-red colour. Mabesoone considered it as a sheet-flood deposit, accumulated at the foot of a mountain range. "The climate during deposition must have been arid, with sheetfloods, ..." (Mabesoone 1959, p. 149). Lautensach & Mayer (1961, p. 174), also dating the rañas as Villafranchian, gave the following description of the climate during their deposition: "Damit wären die Rañas als Piedmont-Glacis Zeugen des ältesten Pluvial, dessen wechselfeuchtes Klima sich gegenüber dem heutigen durch höhere und intensivere Niederschläge bei gleichzeitig stärkerer Schuttanlieferung im Gebirge ausgezeichnet haben dürfte". So they consider these deposits more or less as the highest terrace, because the terrace deposits are also connected to pluvial and glacial phases.

South of Guardo, on the eastern bank of the Rio Carrion, well-rounded quartzite pebbles cover the extensive flat surface and there occurs no typical coarse subangular raña material, such as is known from the western slope of the Rio Carrion valley, near the chapel "Cristo del Amparo". Hence the surface directly southeast of Guardo has been mapped as a terrace surface. Further to the south no investigations have been carried out. The relative height of this surface corresponds with the level of the highest terrace found upstream of Velilla del Rio Carrion (cf. p. 131).

The post-Pontian planation level lies, after Mabesoone, at a height of 1200 m near the mountain range. In the Stephanian and Cretaceous border zone, which morphologically belongs more to the meseta than to the mountain range, no remains of this planation level have been found and also raña material is totally absent. In contrast with Mabesoone (1959, p. 148), we have not found raña material overlying the Cretaceous sediments on the Otero hill, west of Tarilonte.

The piedmont alluvial plains, which occur at the foot of the Sierra del Brezo, have been previously described by Kanis (1955) and Nossin (1959). These piedmont flats have been deposited by sheet-floods under a semi-arid climate, with sudden heavy showers. The angular limestone fragments have been derived from the Carboniferous limestones of the Sierra del Brezo. Mostly the fragments are 4—5 cm large and tabular in shape; however, blocks of 70—100 cm have also been measured. The limestone breccias are strongly cemented by a red calcareous clay.

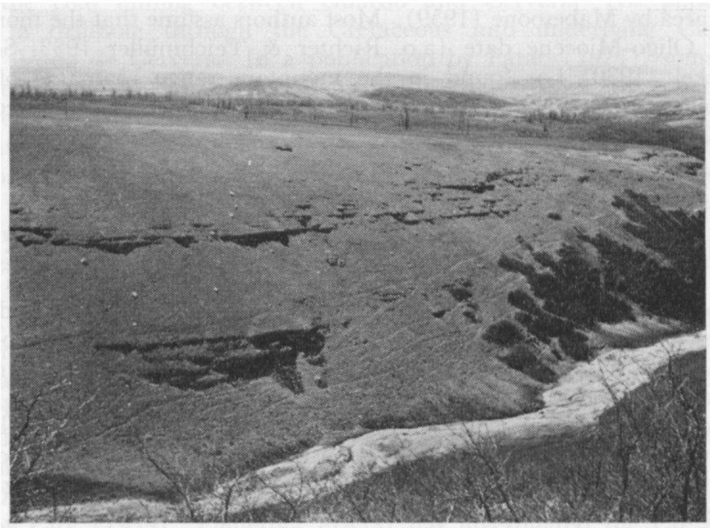


Fig. 2 View to the south over the piedmont alluvial plain situated northwest of Villanueva de la Peña.

The thickness of these sheet-flood deposits ranges up to 40 m. Northwest of Villanueva de la Peña they form a flat-lying fan, slightly inclined to the south (fig. 2). In the north the material has been deposited on an altitude of 1280—1300 m, whereas the most southerly remnant of this fan, which can be found on the western bank of the Rio de las Cuevas, between the highway and the railway from Cervera to Guardo, is situated at 1125 m. The occurrence at this low level, makes the theory that this material was deposited contemporaneously with the rañas on the post-Pontian planation level, as proposed by Nossin (1959), improbable. However, an exact dating is not yet possible.

We can agree to the view of Nossin that next to these highly cemented limestone breccias, uncemented recent or subrecent steep-angled fans also occur. Nowadays it is even possible to study their origin, when heavy thundershowers cause flash floods in the narrow canyons in the Sierra del Brezo.

CARRION TERRACES

In the valley of the Rio Carrion several river terraces have been observed. Four levels could be distinguished. Fig. 3 shows a longitudinal profile of the Rio Carrion between Triollo and Guardo, with a fall of 200 m over a distance of 27 km. Two artificial lakes have been constructed in its course. The dam of the Pantano de Camporredondo de Alba has been built in the Upper Devonian quartz sandstones north of Camporredondo de Alba, the dam of the Pantano de Compuerto in the same quartz sandstone south of the confluence of the Rio de Valcovero with the Rio Carrion.

Mabesoone (1959) has described the terraces in the lower course of the Rio Carrion, south of the mountain range. He could distinguish five levels. The terraces found in the present area correspond with the T1, T3, T4 and T5 levels of Mabesoone. The highest level, T1 terrace, occurs north of Velilla del Rio Carrion and is situated between the Rio Carrion and the Rio Grande. The gradial growth of this terrace has caused the confluence of the two above-

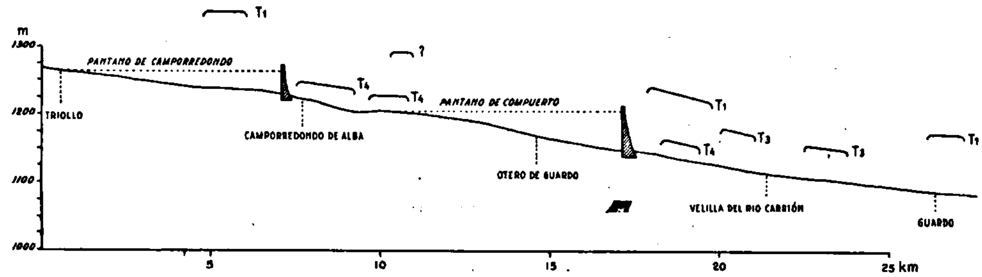


Fig. 3 Longitudinal river profile of the Rio Carrion between Triollo and Guardo.

mentioned rivers to shift in course of time to the south. Its relative height is 100—120 m and it is inclined to the south. In the southern part of the terrace, the material is locally eroded and the Devonian quartz sandstone crops out. The terrace material consists mainly of well-rounded quartzite pebbles. For their description we refer to Mabesoone (1959).

The T3 terrace level occurs north and south of Velilla del Rio Carrion. Its relative height above the river bed is 50—55 m. 1 Km south of this village, in a meander of the Rio Carrion to the west, terrace material has been accumulated. Another remnant of the T3 terrace can be found on the southerly spur of the T1 terrace north of Velilla del Rio Carrion.

The relative height of the T4 terrace level ranges from 20—35 m. The flat surface on which Camporredondo de Alba is built, belongs to the T4 terrace level, as well as the small terrace remnant somewhat further to the south. Along the T1 terrace situated north of Velilla del Rio Carrion, a small remnant of a lower level (T4) can also be distinguished, though a great deal has been removed by erosion.

The T5 terraces or Lower Terraces are situated 1—15 m above the present river bed. Sometimes two different levels are present in these T5 terraces, which, however, are not distinguished in the present paper. The T5 terraces have a large lateral extent, bearing sediments of well-rounded pebbles and boulders of mainly quartz sandstone and quartzite. A striking component of these sediments are the granite pebbles, which occur in rather great numbers.

On the higher terrace levels granite pebbles never have been found. The granite is a biotite-muscovite granite. The source rock can be found in the head-water region of the Rio Carrion and in the Arroyo las Lomas, on the southern slope of the Peña Prieta. Two glacial cirques, cut out in the granite and partly in the surrounding Carboniferous, are the valley heads of the two above mentioned rivers. Before the last glaciation the rivers probably had not yet incised the granitic body. Consequently the supply of granitic material increased highly during and after the development of these cirques and their related moraines. So we may conclude that the T5 terraces are closely connected with the glaciation of the upper course of the Rio Carrion and the Arroyo las Lomas.

In the valley of the Rio Grande only the T5 terraces have been developed.

A dating of the terraces is rather hypothetical. Most investigators assume a Quaternary age and correlate them with the glacial and pluvial phases in the mountains. Nossin (1959) and Mabesoone (1959) have shown by sediment-petrological investigations, that the lower two terrace levels were formed under periglacial conditions, whereas the terraces at higher levels do not show such features. Both authors consider the Lower Terraces, (T5 terraces) as having been built up during the last glaciation (correlated with the Würm glaciation by Spanish authors). The connection between the T5 terraces and the glacial remains on the mountain Orvillo in the present area and the Peña Prieta in the headwater region of the Rio Carrion, indicates that these cirques and moraines originated also from the last glaciation.

GLACIAL FEATURES

On the mountain Orvillo, in our area, four glacial cirques have been developed at an altitude of 1800—1900 m. The cirques are cut in the massive quartz sandstones and have all nearly vertical walls.

The cirque basin of the southernmost cirque is drained by the Arroyo de la Colina. Only one terminal moraine has been found at an altitude of about 1730 m. Outside this moraine, small remnants of fluvioglacial material occur on many places along the course of the Arroyo de la Colina. Such glacial outwashes are also exposed along the road Velilla del Rio Carrion - Otero de Guardo, near the Arroyo de la Colina, as well as south of Otero. The material shows poor sorting. The diameter of the quartz sandstone pebbles and boulders ranges from 3—120 cm. They are rounded or subrounded, but have a low sphericity. The interstitial sand is abundant.

When we follow the valley from Otero upwards to the summit of Orvillo, another cirque with moraine remnants occurs. A number of advances and recessions of the glacier caused several moraine belts, which have been partly overriden. However, two moraine ridges can be clearly distinguished. A similar picture can be observed in the two cirques which open to the north and the northeast.

The flat sloping surfaces south of Valverde de la Sierra are glacial outwash plains, connected with the cirque and the moraines on the northern slope of the mountain Orvillo. Two different levels can be distinguished in these fluvioglacial plains. Their lowest point more or less coincides with the relative altitude of the T5 terrace of the Rio Grande. This indicates that both were formed during the same glaciation. Mabesoone (1959), who investigated the river terraces of the Rios Pisuerga and Carrion along their lower course on the meseta, as well as Nossin (1959), who investigated the Rio Pisuerga terraces along their upper course, came to the conclusion, on account of sediment-petrological research, that the T5 terraces (Lower Terraces) must be ascribed to the last glacial period, the Würm glaciation. So we may deduce the same age for the lowest moraine walls, and probably for all glacial features on Orvillo mountain.

A remarkable feature is the occurrence of S-shaped ridges of quartzitic debris, which lie on the cirque floor and more or less parallel to the steep walls. Looking downwards from the cirque wall, the moraines are convex in shape,

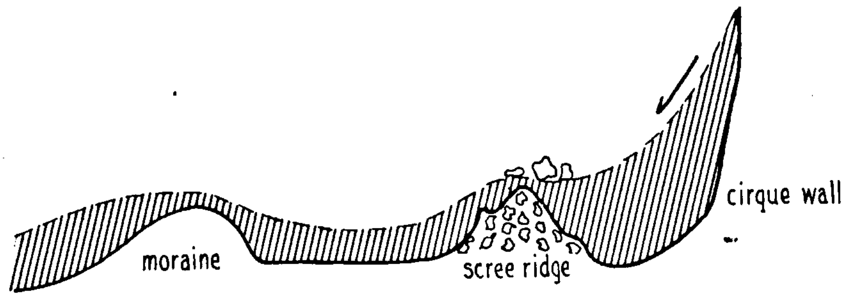


Fig. 4 Illustration of the hypothesis of the development of the irregular shaped scree ridges, found at some distance from the foot of the cirque wall. Scree material has slid over snow, accumulated against the cirque wall during the winter.

whereas these ridges are more or less concave. Between these ridges, consisting of rather angular quartz sandstone blocks and the steep cirque wall, a depression some 50 m wide occurs. In three of the four cirques such an S-shaped ridge is visible. In contradistinction to the moraines, which are regular rounded ridges with some vegetation on it, these ridges are without any vegetation and appear to have been formed recently, as irregular accumulated screes. A possible explanation for this ridges is that during the winter the snow lies thickly accumulated against the cirque wall. Rock fragments, falling down, glide along the snow slope and accumulate on its foot, which results in a scree ridge not directly situated at the foot of the cirque wall (fig. 4).

CHAPTER II

STRATIGRAPHY

Introduction

The area described in this paper is composed of Devonian, Carboniferous and Cretaceous rocks.

The Cretaceous rocks are restricted to a small zone with an E—W direction, which separates the Sierra del Brezo from the Meseta of Old Castile, the northern part of the Spanish Meseta. Ciry (1939) has described their stratigraphy in detail, so only a short lithological description of a section through the Cretaceous south of Santibañez de la Peña will be given (p. 165). For paleontological determinations and correlations with adjacent areas, we refer the reader to the work of Ciry.

A part of the Duero basin, south of the present area, filled with Tertiary sediments, has been described recently by Mabesoone (1959).

The oldest Paleozoic rocks of our region belong to the Lower Devonian. On lithological features the Devonian has been subdivided in five formations (fig. 5), from top to bottom:

5.	Camporredondo formation	— quartz sandstone and quartzite.
4.	Valcovero formation	— a shale-marl-limestone association.
3.	Hornalejo formation	— c) ferruginous sandstone, b) black slates a) quartz sandstone.
2.	Otero formation	— limestone, with local reef developments.
1.	Compuerto formation	— c) light coloured slates, b) black slates and slates with small limestone lenses, a) limestone.

The brachiopod fauna (especially the spiriferids) studied by Mr. T. F. Krans (Leiden), made some correlation with the Devonian of León possible. The total thickness of the Devonian strata is at the most some 1200 m in the south of this area, and it diminishes very rapidly to the north, near the southern shore of the Pantano de Camporredondo. A broad resemblance to the Devonian of the province of León (de Sitter 1959, Comte 1959) is obvious. The Devonian known in the upper course of the Rio Carrion and the Rio Arruz (Kullmann 1960), however, is totally different.

The Carboniferous can be subdivided into three distinct groups in the province of Palencia. These subdivisions are to be related to tectonic movements active in the region.

The Curavacas tectonic phase (Kanis 1955, Wagner 1955) has caused a strong angular unconformity. The name "Ruesga group" is proposed for Carboniferous sediments below this unconformity.

A second tectonic phase, better known in the Cantabric-Asturian mountain belt, is the Asturian folding phase, dated in general as between Westphalian D and Stephanian A. In recent literature some confusion has arised concerning the correct dating of this tectonic phase (Kanis 1955, Wagner 1959, 1960), to which we will return later.

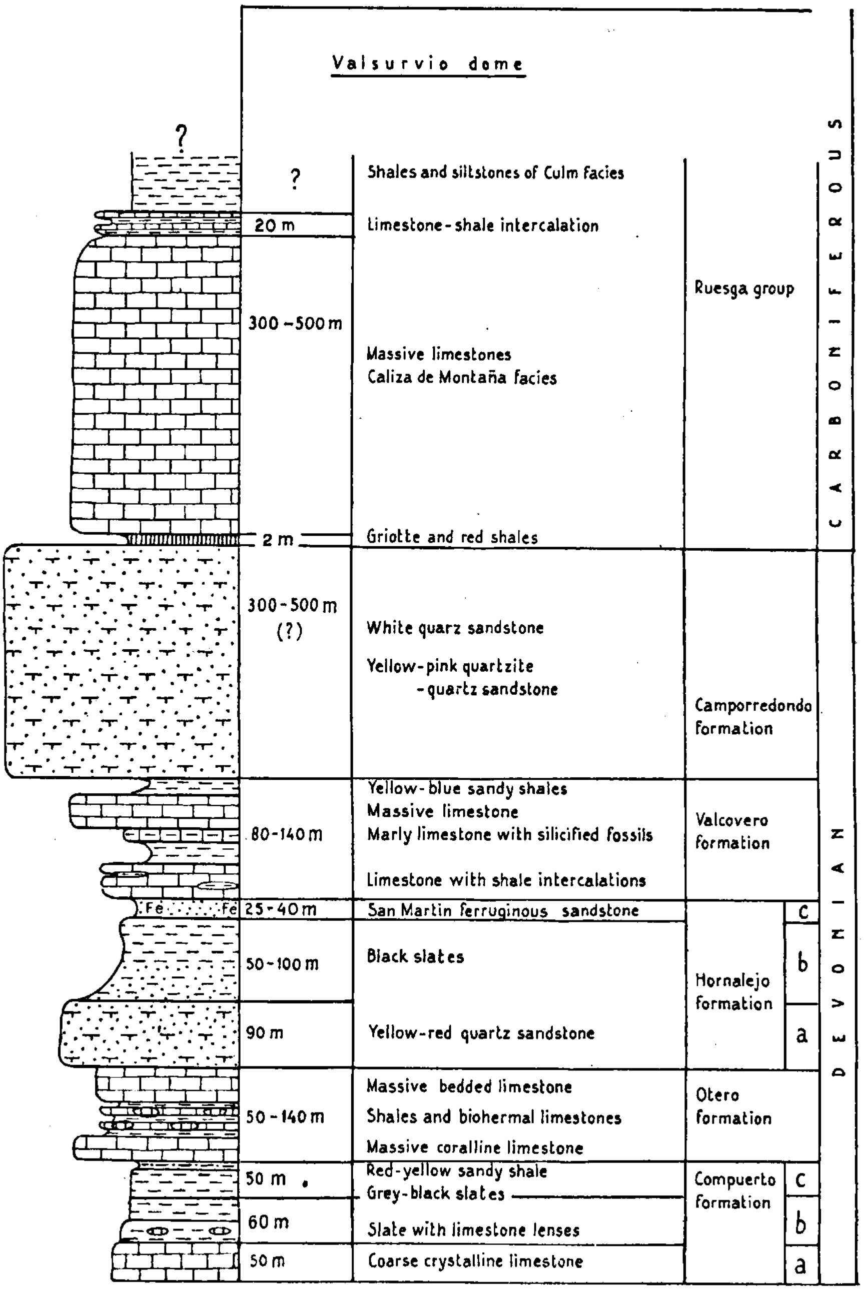


Fig. 5 Litho-stratigraphic column of the Devonian and Carboniferous Ruesga group in the Valsurvio dome.

The name "Yuso group" is proposed for the sequence of strata deposited in the time interval between Curavacas and Asturian phase, and "Cea group" for the Carboniferous sediments deposited post-Asturian phase.

Cea group	conglomerates, graywackes, shales and coal beds
——— Asturian folding phase	
Yuso group	conglomerates, shales, siltstones and limestones
——— Curavacas folding phase	
Ruesga group	Culm facies: shales, subgraywackes and conglomerates Caliza de Montaña facies: thick massive limestones Red shales and griottes (nodular limestone).

DEVONIAN

The Compuerto formation

This formation crops out in the centre of the Valsurvio dome and can be subdivided into three members from top to bottom:

c. light coloured slates and phyllites, with a chloritoid-sericite schist on top and locally a soil of strongly weathered red-yellow sandy shale.
b. sandy slates with sandstone and limestone lenses, grading upwards into black slates.
a. coarse crystalline limestones.

a. Near the mouth of the Arroyo de Abianos into the reservoir of Compuerto the following section through the limestone was measured from top to bottom:

10 m	fine grained blue stromatoporoid limestone, thick bedded,
30 m	white coarse crystallized detritic limestones, with crinoid debris and a few solitairy corals, massive thick bedded,
10 m	dark blue-black shaly limestones,
1.5 m	fine grained blue limestones, brown coloured on a weathered surface, and intercalated with dark shales.

The calcareous member *a* grades upwards into the slaty member *b* of the formation.

b. The slates are dark grey-green or black coloured and show a homogeneous texture. Sedimentary bedding is very scarce. An intensive minor folding and a strong cleavage have destroyed most sedimentary bedding indications. It is impossible to estimate the thickness of these deposits.

In thin section, small euhedral chloritoid crystals are scattered with a random orientation in a sericite-chlorite-quartz matrix. Quartz also often appears in small lenticles conforming to the general cleavage orientation. Also little granules of albite sometimes occur.

The lower part of these slates are more arenaceous. Dark shales run around light-grey sandy lenses, which are about 15 cm long and 7 cm high in cross section. Pockets of pyrite are scattered through these rocks. In a roadcut one kilometer northeast of Otero de Guardo, it appears obvious that these lenses are orientated with their longest axes in the cleavage planes and parallel with the B axis. The cleavage intensively developed in the intervening shales, does not continue straight on in the sandy lenses, but curves around these rigid segments, causing a stretching in the cleavage planes. This lensing is to consider as an advanced stage of cleavage mullions (Wilson 1953). The elongated segments are completely separated and the mutual connection between them being totally lost. Only in the sandy lenses can the original bedding be discerned in thin section. In the quartz-sericite matrix, dark coloured by a high organic content, white dots of primary and secondary quartz with interstitial sericite are orientated in the bedding plane. Indications for originally sedimentary lensing are lacking.

The limestone lenses, which occur lower in this member, are of sedimentary origin and strongly fossiliferous. However, the fossils are badly preserved, strongly recrystallized and deformed. On weathered surfaces, often light-brown coloured by iron oxydation products, the brachiopod *Uncinulus pilus* could be determined (Loc. E 41) [1]). The blue-white crystalline limestone is considered to be partly an autochthonous limestone and partly an allochthonous clastic limestone.

c. The dark slates, described above, grade upwards into light coloured slates and phyllites, with a lustrous sheen of sericite and mica on the cleavage planes. These slates have a maximum thickness of 20 m. They are overlain by a 20—30 m thick chloritoid-sericite schist. This rock is very fine grained, olive-green, hard and massive, cropping out in the country as a black resistant ridge. On the hill southeast of Otero de Guardo, between the village and the Rio Carrion, these chloritoid rocks are folded into a S-shape. The Rio Carrion follows the curve around this hill.

The chloritoid crystals only can be distinguished in thin section, they vary from 30—75 μ. X-ray investigations have shown that it is a triclinic chloritoid, however, some reflections are broad and show characteristics of the monoclinic pattern. A mixture of both polymorphs can occur (v. d. Plas 1959, Halferdahl 1961). The chloritoid is scattered through the rock in conspicuous crystals, which give a porphyroclastic effect in a sericitic matrix. The chloritoid crystals may be produced in a low grade (epizone) of regional metamorphism. Halferdahl (1961, p. 49) writes: "Stress as defined by Harker is not necessary for growth of chloritoid". In pelitic sediments which have a high content of Al and Ferro oxyde, chloritoid is one of the first new formed minerals during regional metamorphism.

On top of the chloritoid sericite schist 10—20 metres of soft red-green sandy weathered shales occur, mostly badly exposed. The character of this sediment is that of a fossil soil. The orange-red fine porous beds, which are easily pulverised in the hand, are due to laterisation and desilification of the clay by weathering. A later partial recrystallization during the low grade regional metamorphism can occur. The iron and alumina oxide content is very high, whereas the silica content is relatively low. The calcareous shales are for a part decalcified, forming a weak porous weathered rock. Internal moulds of brachiopods and corals, however, remain preserved. The rock has a yellow-brown colour due to iron oxidation products.

These rocks indicate a short local stratigraphic break, with weathering probably under tropical conditions, between the Compuerto formation and the Otero formation. The red-coloured sediments are only found on the west side of the Rio Carrion between Otero de Guardo and Camporredondo.

Soil weathering under tropic conditions, is also known from the surroundings of Portilla de Luna (prov. León), where a bauxite deposit has been discovered by Font-Altaba & Closas (1960). They describe a bauxite locality at the top of the Huergas formation (p. 8): "Having as a wall some very fragile clay slates and as a roof a bank of fossiliferous limestones".

[1]) A list of localities, mentioned in the text, are given together with their geographical coordinates on p. 227.

The brachiopods of the wall slates, determined by Villalta, are: *Uncinulus orbignyanus* Vern., *Uncinulus princeps* Barr., *Cyathocrinites pinnatus* Gold., and some corals. He concludes, that these slates belong to the Huergas formation of Comte (1959). However, *Uncinulus princeps* Barr. has been described by Comte only from the La Vid formation, whereas *Uncinulus orbignyanus* Vern. is known from the Sta. Lucia limestone, so a lower Eifelian age, proposed by Villalta, is uncertain. Also the fossil content of the overlying limestone: *Hexacrinus* sp., *Atrypa reticularis* Lin., *Fenestrella* sp. and *Fenestrellina* sp., which he correlates with the Portilla limestone, is not adequate to justify a Givetian age. Recent field studies by students of the University of Leiden have shown that the shales and limestones mentioned by Font-Altaba & Closas (1960) belong respectively to the La Vid and the Santa Lucia formation of Comte (1959).

In the crystalline limestones of the Compuerto formation *a* a few conodonts have been found (Loc. E 35), which all belong to the genus *Icriodus* Branson & Mehl (preliminary determinations by Mr. H. A. van Adrichem Boogaert, Leiden), which is known to occur through the whole Devonian.

The only determinable brachiopod, already mentioned on p. 137 is *Uncinulus pilus* Schnur., which occurs in the limestone lenses of the Compuerto formation *b*. This brachiopod attains its greatest extension in the Emsian. In León it is known from the upper part of the La Vid formation and from the lower part of the Santa Lucia formation; higher in the succession it dies out rapidly and *Uncinulus orbignyanus* takes its place (Comte 1959, p. 254). The La Vid formation consists in León of greenish shales and marls, highly fossiliferous, with fine grained limestones and dolomites at the base. A rough lithological resemblance with the Compuerto formation is apparent, although the low-grade metamorphism and intensive tectonic deformation in the Valsurvio dome has destroyed nearly all lithological characteristics of the incompetent rocks and altered them into slates and phyllites. The occurrence of *Uncinulus pilus* Schnur. supports the evidence for the correlation of the Compuerto with the La Vid formation. The strongly weathered sediments and laterisation products found on the top of the Compuerto formation, indicating a break in the sedimentation, may possibly be correlated in the future with the bauxite deposits discovered by Font-Altaba & Closas near Portilla de Luna (León).

The Otero formation

This formation consists principally of calcareous sediments, showing great lateral variations.

North of the mountain La Taramada the formation consists of thick bedded limestones, with shaly intercalations in the middle. In the valley of the Rio Carrion, directly south of Otero de Guardo — the name of the formation is derived from this village —, it consists of five distinct limestone layers separated by shales and calcareous shales. The shales are badly exposed, the here following section could be measured (thickness in metres):

	4	dark coloured sandy shales
Hornalejo form.	1.5	grey sandstone with irregular laminae of shaly material with rugose corals, bryozoans and crinoid columnals (not in situ)
	0.3	dark shales

	5	coral limestone, strongly recrystallized, the top is red coloured by a high content of hematite. Fossils: corals, bryozoans, brachiopods, all indeterminable
	4	shales
	40	grey coral limestone
	10	shales
Otero form.	20	coral limestone, which has locally a reefoid character
	8	shales
	20	coral limestone
	15	calcareous shales
	8	red-dark-blue shaly limestone, recrystallized with silicified rugose corals, bryozoans and crinoid columnals
Compuerto form.	——	

In the valley of the Arroyo de Abianos, between this small river and the road Otero-Camporredondo, small limestone reefs are developed, almost entirely made up of stromatoporoids and corals. A cleavage hase been developed in this reef like limestone, which has intensified the knotty aspect of the coral and stromatoporoid knolls. When we follow this limestone higher up the hill in the direction of Otero, no bioherms occur, but the limestone is developed as a biostrome. The lower limestone layers are strongly fossiliferous, rugose as well as tabulate corals and stromatoporoids are silicified. These layers have a faint red colour and are often more shaly than the higher ones.

In the direction of Camporredondo the thickness of the limestone suddenly diminishes to zero. Near Camporredondo we find this limestone again, but strongly reduced. This is due to local erosion in the northern part of the present area, to which we will return later (chapter III).

North of Valsurvio, the Otero limestone crops out in the core of the Hornalejo anticline, as a grey - blue, well-bedded coral limestone.

Unfortunately, fossil determinations from the Otero limestone are not to the disposal of the author. Conodonts do not occur in the limestone, and the investigations of the strongly silicified corals and stromatoporoids have not yet been completed. The few brachiopods proved to be indeterminable, due to the strong recrystallization.

So the age of this formation could be only deduced from age determinations from the underlying and overlying strata, which are respectively of (Lower ?) Emsian and Lower Couvenian age.

The Santa Lucia limestone of Comte (1959) is of the same age as the Otero limestone. Both formations are rich in corals and stromatoporoids. An upper zone of the Santa Lucia limestone, rich in brachiopods, and known as the zone of *Spirifer Cultrijugatus* in the province of León, we have not found in the Otero limestone.

The Hornalejo formation

In contrast with the Otero limestone, this formation is built up by clastic sediments. We can subdivide it in three parts:

c ferruginous sandstone (of San Martin)
b black slates
a quartz sandstone.

a Quartz sandstone. — The Otero limestone grades upwards into dark shales, some 4 m thick, which passes into a thick sequence of quartz sandstones of grey - white and often red spotted colour.

Traces of cleavage development and sedimentary bedding are rather obscure in the sandstones. In small pelitic intercalations, however, a clear cleavage has been developed. Large new crystallized mica crystals (0.5—1 mm) in the pelitic layers are orientated in the cleavage direction. Sedimentary micas also occur, primary with their orientation in the bedding planes, but with the cleavage development they are slightly deformed and mechanically reorientated in the cleavage planes. Magnetite streaks also occur, orientated in the cleavage. The cleavage planes die out when they enter the arenaceous layers. Zircon commonly occur in small amounts.

In the Hornalejo sandstones, on the mountain north of Valcovero, cleavage

Fig. 6 Fossil preservation in the sandy shales of the Hornalejo formation. The internal mould of a *Pleurodictyum* sp. is visible, lower right (× 1.5, Loc. K 119).

has totally destroyed the sedimentary bedding and has developed a tectonic banding, which is easily mistaken for bedding.

The thickness of the Hornalejo formation *a* near Otero de Guardo is about 80—100 m. The top of the quartz sandstone is formed in many places by a ferruginous sandstone, the quartz grains being coated by limonite film. This sandstone is very rich in fossils, especially brachiopods and corals. The fossiliferous horizon is laterally constant, but its quartz content may change. The fossils near Otero are preserved in a sandy often ferruginous layer. Whereas we find them in black slates in the Hornalejo anticline (p. 143). The fossils are pyritized and beautifully preserved as internal moulds. Corals, showing their complicated internal structures, have been collected (fig. 6).

b Black slates. — The black slates of this formation show a great resemblance with the slates from the Compuerto formation and a distinction is only possible by their relation to the adjacent rocks. With exception of the lowest layers, which grade downwards into the quartz sandstone, there are no fossils preserved.

Strong first and second cleavages have been developed. Bedding can scarcely be distinguished except by means of small sandy layers, but on the whole the slates are very homogeneous.

The fine preservation of the fossils in such strongly cleaved rocks is very interesting. Whereas bedding in most places has been totally destroyed by the cleavage development, the fossils remain intact.

The black slates often show a strong chloritisation (clinochlore) in thin section. The quartz content is rather high (20—25 %). Large mica or mica-chlorite aggregates occur (300—360 μ), orientated in the first cleavage direction and deformed by later cleavages. These mica crystals are also clearly visible macroscopically on the cleavage planes. Grains of zircon and stilpnomelane are always present in small amounts. Small chloritoid crystals lie scattered with random orientations in the matrix (description on p. 137).

The intensive isoclinal cleavage folding makes the thickness of these beds difficult to measure, but they should be about 80 m thick.

c. The ferruginous sandstone of San Martin. — Kanis (1955, p. 438) has already described the iron-bearing sandstone between Ventanilla and San Martin de los Herrores. We find the same horizon again in the Devonian core of the Valsurvio dome. In most places this ferruginous sandstone has a red - brown colour. In thin section the quartz grains appear to be fairly angular, with a main diameter of 0.3 mm. They are embedded in an abundant matrix of hematite (the quartz grains do not touch each other). Euhedral crystals of magnetite also occur. A few traces of oölitic texture have been found. The maximum measured iron content is 37.12 % (Loc. Q 72), a 500 m southeast of Valsurvio (analyst: Dr. C. M. de Sitter—Koomans).

Next to these red - brown hematite-rich rocks (type A), green - black shaly sandstones occur (type B), especially developed around Valsurvio, these rocks are rich in iron-bearing silicates, especially chamosite. Thin sections show that the bulk of the rock consists of a groundmass of quartz, sericite, chamosite and to less extent of chlorites and limonite. Subangular quartz grains lie scattered in the matrix. Small ellipsoidal chamosite oölites, with a pseudo-spherulitic structure and a diameter of 0.4 mm occur scarcely. Much attention to this type of rock has been paid by Hallimond (1925) and Taylor (1949). Sometimes these oölites show an alternation of shells of fresh green chamosite, with yellow-brown partial oxidized chamosite. This indicates differences in the physico-chemical conditions during deposition. Mostly, however, it is a structureless green chamosite mud. Carozzi (1960, p. 362) notes, when clay minerals predominate over chamosite it is developed as a pale-brown to pale-greenish grey micro crystalline aggregate. A later development of octahedrons of magnetite has replaced the chamosite. Occasionally large chlorite crystals have been formed, with the crystal cleavage parallel to the outlines of the magnetite crystals. The crystallization of the magnetite and the chlorite is earlier than the rock cleavage, because the cleavage traces deform these newly-formed chlorite crystals slightly.

Along the road Otero de Guardo-Velilla del Rio Carrion, a section through the ferruginous sandstone has been measured:

	thickness in metres
Valcovero formation — Calcareous shales	
Coarse grained, red - brown hematite-rich sandstone (type A) with a clear development of cross lamination of a lenticular shape.	2

Massive red - brown hematite-rich sandstone, with a high iron content (type A) with a development of cross lamination.	0.80
Detritic hematite-rich shaly sandstone (type A) with a variable iron content.	3
Massive red - brown hematite-rich sandstone (type A) high iron content, with development of cross lamination.	5
Dark green sandstone (type B) low iron content, chamosite.	2.50
Yellow - light - brown crumbly fine sandstone, fossilhorizon: brachiopods, corals, bryozoans.	0.30
Fine grained green chamosite-rich sandstone (type B), with abundant magnetite octahedrons, to the base an increase of the iron content.	1.50
Red - brown hematite-rich, friable sandstone (type A), with a high iron content.	1.50
Fossil horizon of mainly brachiopods.	0.10
Green mudstone (type B) with poorly preserved fossils, brachiopods.	4.30
Green - white fine grained, strongly weathered chamosite-rich mudstone and chamosite oölites (type B) with lenses of limonite.	2.30
Yellow - brown friable sandstone (type B) with brachiopods, corals and bryozoans.	0.70

Hornalejo formation *b* Black slates

Briefly we can say that

Type B:

a. The lowest 14 m of the ferruginous sandstone of San Martin contain chamosite and ferric-oxides.
b. The sediments are more silty than the upper part.
c. Brachiopods, corals and bryozoans are present in several fossil horizons.
d. Chamosite has been developed partly as a chamosite mud, partly as chamosite oölites.
e. Cross laminations are only rarely developed.

Type A:

f. The iron content of the uppermost 11 m consists chiefly of hematite (30—50 % ferrous-oxide, Table 1).
g. Strong cross laminations have been developed, produced by a rapid shifting of currents. Scouring and filling have developed lenticular units, which are cross-laminated in opposite direction and show a pronounced concavity.
h. Small fossil fragments are sometimes preserved.

The sediments of type A were deposited in a paralic, unstable environment, with shifting current directions. The sedimentation of the lowest part (type B), also took place in a paralic environment, but with less reworking and deposited under rather quiet circumstances, although oölites indicate a growth in turbulent waters. Changes in the physico-chemical conditions between the different layers are certain. The Huergas formation (Comte 1959) is in broad outlines to correlate with the Hornalejo formation (cf. p. 144). However, the ferruginous sandstone occurs in the Valsurvio dome in the upper part of the formation, while in the Esla region it occurs in the lower part.

Table 1. Result of the chemical analyses of the iron content.

Locality	Fe_2O_3	FeO	type of Rock (see description)
1 km S of Otero de Guardo (Loc. E 20)	41.73	0.66	type A
½ km N of Valsurvio (Loc. Q 84)	36.79	18.26	type B
100 m SE of Valsurvio (Loc. K 216)	6.00	21.54	type B
San Martin de los Herreros (Kanis 1955)	57.98	9.25	type A

Analyst: Miss H. M. I. Bik.

The fossil faunas. — Brachiopods, collected in the Hornalejo formation, were submitted for identification first to Mr. J. A. van Hoeflaken and later collections to Mr. T. F. Krans, who especially determined the spiriferids. Both are from the Leiden University and have kindly supplied the additional lists.

The following species have been determined from the Hornalejo formation *a* (Loc. E 2, E 6, E 23, E 105, Q 78, K 205).

Spinocyrtia subcuspidatus (Schnur.)
Hysterolites (Acrospirifer) bicollinae n. sp.
Conchidium ? oehlerti Barrois
Conchidium sp.
Leptaena rhomboidalis Whalenberg
Stropheodonta piligera Sandberger
Schellwienella umbraculum Schloth.
Pleurodictyum problematicum Goldfuss
Pleurodictyum sp.
single corals
crinoid fragments

This fauna of the Hornalejo formation *a* indicates a lower Couvenian age (Krans, personal communication). *Conchidium ? oehlerti* [1]) occurs abundantly, but nearly always as internal moulds. Barrois (1886) has found this species together with *Spirifer cultrijugatus* in the Moniello and Arnao limestones in Asturias. They indicate a Lower Couvenian age. *Spinocyrtia subcuspidatus* is known from the Siegenian untill the Lower Couvenian. *Hysterolites (Acrospirifer) bicollinae* n. sp. [2]) gives also indications for a Couvenian age.

[1]) This species has been investigated by Mr J. G. Binnekamp (Leiden) who comments (personal communication): The dorsal characteristics are indicative that *Conchidium* should be separated from *Gipidula.* As the dorsal valves are very scarce and never occur together with ventral valves, the determination of the genus remains uncertain. The ventral valves may be separated into two groups: 1) with a short septum, 2) with a long septum. The first group belongs to the species *oehlerti.* Determination of the second group is uncertain.

[2]) *Hysterolites (Acrospirifer) bicollinae* n. sp. will be described in due course by Mr. J. F. Krans (Leiden). This species strongly resembles *Sprifer paradoxus,* but is a more highly developed species, and so can probably be dated as Lower Couvenian, zone de Bure (Krans, personal communication).

From the base of the Hornalejo formation *b* the following fauna has been determined (Loc. E 94, E 114, K 119, K 130):

Hysterolites (Mucrospirifer) diluvianus (Steininger)
Schellwienella umbraculum Schloth.
Leptaena rhomboidalis Whalenberg
Pleurodictyum problematicum Goldfuss
single corals
crinoid columnals
bryozoans.

The spiriferid *Hysterolites (Mucrospirifer) diluvianus* (Steininger) [1]) occurs abundant and indicates a Couvenian or Lower Givetian age.

The ferruginous sandstone of the Hornalejo formation *c* has yielded a fauna collection of rather badly preserved fossils, of which only the following species could be determined (determinations by Mr. J. A. van Hoeflaken):

Schizophoria striatula Schlotheim
Atrypa reticularis Linnée
Leptaena rhomboidalis Whalenberg
Cyrtina sp.
Stropheodonta sp.
single corals
bryozoans

These species are not diagnostic of a definite level in the Devonian.

On lithological grounds this formation bears comparison to the Huergas formation, from the Esla-Bernesga region, as described by Comte (1959). In both formations, the lower part consists of clastic sediments, which pass upwards into a more pelitic sequence. In the Valsurvio dome the top of the formation is formed by a ferruginous sandstone, which is not known from the Esla-Bernesga region.

A fauna collected in the Hornalejo formation *a*, in which *Conchidium ? oehlerti* and *Hysterolites (Acrospirifer) bicollinae* n. sp. are abundant, may be correlated in age with the base of the Huergas formation, in the north of the province of León. *Spirifer paradoxus* Schlotheim which shows great similarities with *Hysterolites (Acrospirifer) bicollinae* n. sp. (Krans, personal communication), is recorded by Comte only from the Santa Lucia limestone. However, in a recently collected fauna, from the base of the Huergas formation south of Remolina in the Esla region, several highly developed species of *Spirifer paradoxus* Schlotheim [2]) has been found.

The fauna identified from the Hornalejo formation *b*, is similar to a fauna, which occurs about 30 m below the top of the Huergas formation in the Aquasalio section in the Esla region.

However, to justify a detailed correlation much more paleontological research must be done.

[1]) Young specimen of *Hysterolites (Mucrospirifer) diluvianus* are very similar to *Spirifer elegans* (Krans, personal communication).

[2]) Determination by Dr. A. Vandercammen (Bruxelles).

The Valcovero formation

The transition from ferruginous sandstone to the shales and limestones of the Valcovero formation is mostly very gradual, as near Valsurvio and San Martin de los Herreros. It takes place by a decrease of the iron content, as the sandstone grades upwards into shales or calcareous shales of a yellow - blue colour. Sometimes the contact between ferruginous sandstone and the overlying Valcovero limestone is very sharp (as north of Otero de Guardo), suggesting a break in sedimentation.

The sediments of this formation show great lateral facies changes in the present area. Similar variations are also known from the adjacent area, investigated by Kanis (1955, p. 393). From Ventanilla in the east, to San Martin in the west, a facies change occurs from fossiliferous shales into coral limestones.

Returning to our area, we find near Valcovero a thick limestone development. The total thickness of the Valcovero formation is here about 145 m. The base has been formed by 40 m of limestones with some small shale intercalations. It is overlain by 25 m of calcareous shales with thin detrital limestone beds followed by another 25 m of marls and limestones with small shale laminae, with a rich fauna of silicified single corals. A 50 m thick bedded limestone with corals, stromatoporoids and bryozoans, overlain by a 5 m sandy shale of yellow - ochrous colour, form the top of the formation.

East of Valsurvio, the upper most 80 m of this formation are composed of dark blue shales or slates, often fossiliferous. The fauna is very homogenous, rich in specimens, but poor in species, comprising only the spiriferids *Mucrospirifer bouchardi* and *Cyrtospirifer verneuili.* The spiriferid moulds are all of the same size, indicating a good sorting. In the exposure the shells are oriented with their longer axes parallel with each other and most of them lie convex side upwards. This is the position of greatest stability under current action, as Sorby (1908) has pointed out long ago. Sometimes the shells have been washed together in ridges, and after being turned into a stable position on the bottom, they were buried. The shells lie with their longer direction parallel to the length of these ridges. These deposits probably originated as shell banks on a muddy tidal flat. Turbulent water movement did not occur, because the shells are not broken, yet currents were strong enough to select the material, to turn the shells in a stable position and to wash them on to ridges. However, the exposure is too small to develop a clear picture of the sedimentary environments. Other indications, that the top of the Valcovero formation must have been developed in shallow water and possible as beach deposits, are the occurence of some small limestone reefs and the development of a hiatus between the Valcovero formation and the Camporredondo formation more to the north.

Limestone layers are developed lower in the Valcovero formation, which have sometimes a reeflike character. Near the small river east of Valsurvio a bioherm has been found. The reef core is 3 m in diameter and consists of dolomitized limestone. Reef-building fossils could not be distinguished. A detrital and marly reef talus is only partly exposed.

Limestones are often strongly recrystallized and sometimes dolomitized. Their thickness is variable. Often two or three limestone layers may be distinguished, 10—20 m thick, separated by fossilliferous calcareous shales. The corals bryozoans, brachiopods and crinoid fragments all proved to be indeterminable, due to strong recrystallization. Downwards the limestone grades into the ferriginous sandstone (Hornalejo formation *c*).

Further to the northeast, in the southern limb of the Hornalejo anticline, two limestone layers occur. The lower yellow-coloured crinoidal limestone is 20 m thick, and contains a few corals. The upper blue-grey limestone is 10 m thick; they are separated by a few meters of shale. The highest limestone is directly overlain by the quartz sandstone of the Camporredondo formation.

On the northern limb of the Hornalejo anticline the thickness of the Valcovero formation decreases more and more. Only a few small limestone outcrops have been found between the Camporredondo and the Hornalejo formations.

To the west, near Camporredondo, the thickness of the Valcovero formation remains restricted to 7 m of limestone and marly shales with bryozoans, brachiopods, corals and crinoid fragments, overlain by 7 m of black shales with a specimen-rich fauna of *Mucrospirifer bouchardi* and *Cyrtospirifer verneuili,* in a similar occurrence to that near Valsurvio.

Following the line of outcrops of the limestone from Camporredondo to the south, the thickness increases again and as we have seen, reaches its maximum near Valcovero. 1 Km south of Valcovero, the thickness of this formation appears also reduced, but due to lack of exposures it remains doubtful.

Near San Martin and in the Devonian outcrops south of La Lastra, the formation is developed as a grey-blue coral limestone, strongly recrystallized. This massive limestone is difficult to distinguish from the Otero limestone.

The sediments of this formation have been deposited under epi-neritic environment. The calcareous beds are deposited in quiet clear shallow water; locally small bioherms have been developed. The black shales also indicate quiet conditions, although the selectively orientated brachiopod shells, sometimes situated in ridges, are indicative of current action, probably on a tidal flat. Concluding we find:

a) Thickness of the formation decreases from south to north.
b) A facies change from limestone to black shales of the upper part of the Valcovero formation occurs between Valcovero and Valsurvio.
c) Near San Martin and in the Devonian outcrops south of La Lastra and Triollo, a break in the sedimentation occurs between the Valcovero and the Camporredondo formation.

The Valcovero formation is rich in fossils. In the shale intercalations brachiopods are abundant, but rather poor in different species. In the limestones brachiopods are scarce, but corals and bryozoans occur. Several samples have been investigated for their content of conodonts, but only one locality yielded a fauna. The following species have been determined by Mr. H. A. van Adrichem Boogaert:

Icriodus sp. Branson & Mehl	Dev.		
Polygnathus linguiformis Hinde	Ems.	Couv.	Giv.
Polygnathus varca Staufer		Couv.	Giv.
Polygnathus xylus Staufer		Couv.	Giv.
Polygnathus sp. Hinde			
Prioniodidae Ulrich & Bassler			

(Loc. E 81, in the Valcovero limestone, in the valley west of the Coriscao.) A Middle Devonian age could be concluded (van Adrichem Boogaert, personal communication).

Brachiopods from shale intercalations, which occur rather high in the formation, indicate a Frasnian age.

Hysterolites (Acrospirifer) bouchardi (Murchison)
Cyrtospirifer verneuili Murchison
Leptaena rhomboidalis Whalenberg
Schellwienella umbraculum Schloth.
Stropheodonta sp.
Schizophoria striatula Schloth.
bryozoans
crinoid fragments

(Loc. E 19, E 59, Q 75, K 116, BB).
The fauna has been investigated by Mr. J. A. van Hoeflaken, with exception of the two *Spirifer* species, which are investigated by Mr. T. F. Krans.

The lower part of the formation is of Givetian age, whereas the upper part ranges into the Frasnian.

A correlation with the Portilla limestone of the Esla region is obvious.

The Camporredondo formation

This formation is composed of quite homogeneous arenaceous rocks, which can reach a thickness of several hundreds of metres. The lower part of the formation is more shaly and grey-yellow or brown coloured, whereas the upper part is formed by white quartz sandstones, which are often ferruginous and brown-pink coloured. On the top sometimes microconglomerates or coarse quartz sandstones have been developed, 1—2 m thick, as for example near Besande, on the Peña Dorada south of the Pantano de Camporredondo and in the Arroyo de Miranda.

Microscopically it is found that the rock has been built up entirely by quartz grains, which are cemented with silica. The quartz grains are closely packed. They have an average grain size from 0.4—1 mm. Secondary overgrowth has mostly diminished the original nucleus. The quartz grains mostly show a wavy extinction. Mostly the quartz grains have sutured contacts. The pore space is strongly reduced by pressure solution and secondary growth. The original quartz grains are difficult to distinguish by means of a faint dust ring. Sometimes a coating of iron oxides on the grains occurs.

Bedding planes are usually poorly developed in the massive quartz sandstones. Joint systems, often well developed, are easily mistaken for bedding planes in the field.

In a few places, south of the "hermita del Cristo Sierro" and in the road section above the dam of Compuerto, small areas of spotted slates, only 2 m thick, have been formed. In thin section it appears to be an "andalusite-chloritoid-sericite schist".

The porphyroblasts of andalusite are surrounded by a matrix of sericite, quartz, chlorite (pennine), some albite, stilpnomelane, some tourmaline and organic material, forming a lepidoblastic texture. The chloritoid crystals, developed with a random orientation, are later than the development of the slaty cleavage, for the crystals have inclusions which retain their parallelism with the prevailing rock cleavage. They are rotated by a later folding, which has caused a faint fracture cleavage and a folding of the slaty cleavage. The size of the poikiloblastic andalusite porphyroblasts range from 0.8 to 0.5 mm. They enclose much quartz and sericite. The chloritoid crystals are about 0.5 mm in size. Often they are intergrown with the andalusite crystals.

The texture of these rocks suggests a development as contact mineral of the andalusite. However, in the wide environments no intrusions have been known and any visible relation lacks. Halferdahl (1961) describes the occurrence of andalusite with chloritoid in regional metamorphic rocks. In the area investigated here stilpnomelane, chloritoid and andalusite occur in the same rock.

Determination of the age of the Camporredondo formation is not possible, due to lack of fossils. At the base of the quartz sandstone, in the easterly adjacent area, between San Martin and Ventanilla, a fauna has been found containing the spiriferids: *Hysterolites bouchardi* and *Cyrtospirifer verneuili,* also known from the Valcovero shales. Hence the lower part of the quartz sandstones belongs to the Frasnian, whereas this formation grades upwards most probably into the Famennian.

Kanis (1955) has found some badly preserved fossils in what he has named the "Frasnian orthoquartzites", near Pico Almonga. Before him Dahmer (1936) collected a fauna at the same place, which he classified as Middle Devonian.

In our area Saenz Garcia (1943 and 1944) has studied the Camporredondo quartz sandstone near the present dam of the Pantano de Compuerto. He collected a poor fauna out of yellow sandy shales, along the road, 1750 m west of the village of Valcovero and from another locality, along the old drowned road Velilla del Rio Carrion - Camporredondo, 600 m north of the dam. The following determinations[1]: *Spirifer* aff. *rousseaui* Roualt, *Atrypa reticularis* Lin. and *Favosites cervicornis* Blain, would indicate a Middle Devonian age after Saenz Garcia. A glance on the geological map, however, shows that these faunas are collected in shales which belong to the Valcovero formation. The fauna collected by the author, in the same localities (cf. p. 147) indicates a Frasnian age.

A correlation with the Esla region is difficult (cf. p. 181).

Other Devonian outcrops.

Around the Pantano de Camporredondo small outcrops of Devonian rocks occur, which are older than the Camporredondo formation. In these localities, for example on the border south of the Pantano de Camporredondo near the mouth of the Arroyo de Miranda, as well as near the mouth of the Arroyo de Valderrianes and also in the small outcrops along the northern shore of the reservoir, the stratigraphic-lithological sequence which we have built up for the Devonian in the Valsurvio dome, can hardly be used. Strong lithological as well as bioecological differences occur. This together with complicated tectonic structures and the appearence of stratigraphical hiatuses in the Middle and Upper Devonian, make it very difficult to interprete the geology of these small outcrops.

In the present paper the same formation names as we have used in the Valsurvio dome are used for these rocks. But further investigations to the north would probably show, that these rocks belong to a northern facies, so that other

[1]) Determinations by Mr. Gomez Llueca. The determination of *Spirifer Rousseaui* seems improbable, because this species is only known from the Lower Devonian in the province of León and Palencia.

formation names would have to be used. However, this can only be stated by sufficient knowledge, especially of paleontological data, of a greater area. The stratigraphic correlations of these Devonian rocks are based on age determinations of trilobite and brachiopod faunas.

Near the mouth of the Arroyo de Miranda in the Pantano de Camporredondo, a Devonian outcrop is known, which shows little lithological resemblance with the Devonian sequence known from the Valsurvio dome. This outcrop could only be investigated by low water level of the reservoir.

Directly below the Camporredondo quartz sandstone, a black shale occurs, characterized by the abundant occurrence of *Hysterolites bouchardi* and *Cyrtospirifer verneuili.* The same type of rock is known from the Valcovero formation in the Valsurvio dome. These shales overlie 10 m of crystalline, often detrital, grey limestone, with brachiopods, bryozoans, crinoid fragments and corals. The shale intercalations in the lower part of this limestone have been deposited in very shallow water. Angular limestone fragments and brachiopods were washed together in a muddy matrix, filling shallow holes in the underlying limestone, so, a contemporaneous erosion and resedimentation must have occurred.

The brachiopods, found in the limestone, indicate a Lower Givetian-Couvenian age. The following species have been determined by mr. T. F. Krans (Leiden):

Mucrospirifer diluvianus Steininger
Hysterolites aculeatus Schnur.
Cyrtina sp. Davidson, aff. *heteroclita.*
(Loc. E 78, E 79).

In Mr. Krans opinion (personal communication) the fauna gives the impression of corresponding in age with the higher part of the Hornalejo formation in the Valsurvio dome.

Below this limestone, a ferruginous sandstone occurs, a few metres thick, coloured dark red by the high content of hematite. At the base of this sandstone some brachiopods have been found. The determinations of these species are not yet completed. Field observations have shown, that the fauna certainly bears no resemblance to the faunas collected in the ferruginous sandstones from the Hornalejo formation either *a* or *c.*

The ferruginous sandstone grades downwards into a mica-rich sandy siltstone of a yellow-ochreous colour. The mica crystals are 0.5 mm in diameter and of a sedimentary origin. The mica crystals are orientated in the bedding plane and give the rock a lustruous glance. Fossil impressions are beautifully preserved. A great number of pygidea of trilobites have been collected. Dr. W. Struve [1]), who investigated the trilobite fauna, came to the conclusion, that they belong to the genus *Psychopyge* of which only the species *Psychopyge elegans* Termier & Termier is known from the Lower Devonian of Morocco. This new species appears to be somewhat younger in age. The occurrence together with *Hysterolites (Hysterolites) cf. alatiformis* (Drevermann) are indicative of a Lower Couvenian age (Krans, personal communication). Gastropods, lamellibranchs, bryozoans and the coral *Pleurodictuym problematicum* Goldfuss, also occur abundant in these siltstones.

Most of these fossils are totally unknown from the Valsurvio dome (southern

[1]) Dr. W. Struve, Senckenberg Museum, Frankfurt a. M., Germany, will describe *Psychopyge* n. sp. in due course in Senckenberger Lethaea.

facies), but belong to the fauna elements of a northern facies, on which we will return on p. 172.

The determinations have shown that these siltstones can be correlated in age with the base of the Hornalejo or the top of the Otero formation (boundary Middle-Lower Devonian) This implies, that the Middle Devonian on the southern shore of the Pantano de Camporredondo is reduced to a thickness of 15—20 m. On the northern shore of the reservoir, 1½ km south of Triollo, the Middle Devonian is even not developed at all. There the Upper Devonian strata, strongly reduced, directly overlie a limestone, which is according to the presence of *Eospirifer cf. solitarius* Krantz and *Costispirifer bischofi* (Roemer) of Siegenian age.

However, the fossiliferous material is too poorly preserved and the outcrop too restricted to justify correlations with the Devonian elsewhere.

The precise age of the Devonian limestone, which occurs in the valley of the Arroyo de Valderrianes, could not be determined. On the geological map it is shown as Valcovero limestone, mainly on lithological similarity, but it is quite possible, that it belongs partly to older formations.

CARBONIFEROUS

Ruesga group

Red shale — griotte horizon. — The oldest rocks of the Ruesga group in the studied area are the red shale — griotte deposits. The red nodular limestone, often called "griotte", always occurring together with red shales and occasionally with radiolarites, form the base of the transgressive Lower Carboniferous.

This griotte horizon has a very large extension all over the Cantabric-Asturian mountain chain and appears to be a useful marker horizon. However, thickness and composition can change over short distances and from the maximum thickness of 20 m it may rapidly diminish to zero. On the maps and sections this horizon had to be drawn on an exaggerated scale.

All variations occur between the red shales, red shaly cherts and the red cherts. These rocks, of which the shale is the most common variety, are brownish-red in colour, but also often greenish. The shales being highly incompetent, often serve as a detachment horizon. They have a soft muddy texture. With a higher silica content the rock is harder and has a dense and compact texture, breaks with sharp edges or shows a smooth conchoidal fracture. In thin section this red chert consists of a very fine grained chalcedonic silica, intimately associated with clay minerals, abundantly coloured by red iron oxide pigment, the matrix sometimes becoming practically opaque.

In this red semitranslucent matrix radiolarians are visible, circular or elleptical in outline. The inner part of these radiolarians are usually free from red iron oxide pigment, so they form clearly visible transparent areas. Their size ranges from 0.1 mm to 0.5 mm in diameter. The concentric radiolarians have radial spines with maximum length of 0.025 mm, which are preserved in many instances. The outer rim, the cast of the originally hollow skeletons and the spines consist of pure silica. The inner part contains calcite and sometimes chalcedony.

Davis (1918) described radiolarian cherts of the Franciscan group, California. Table 2 lists some partial chemical analyses of siliceous red cherts, red shales and ordinary shales after data from Davis, Clarke, Niggli and the author.

Table 2

	1	2	3	4	5
SiO_2	95.08	63.47	58.38	74.90	75.04
Fe_2O_3	2.82	14.60	4.03	5.57	4.08
FeO	—	—	2.46	0.00	0.40

1. Siliceous red chert from Red Rock Island (Davis 1918)
2. Red shales from Red Rock Island (Davis 1918)
3. Composite analyses of 78 ordinary shales (Clarke 1916)
4. Red radiolarite, Weiszhorn, Schweiz (Niggli 1952)
5. Siliceous red radiolarian shales, La Lastra, Spain.

The silica content of the siliceous radiolarian shale of La Lastra lies between that of the red chert and the red shales of the analysis of Davis (1918). This is in agreement with the texture and the lithological character of the rock. It is a shaly radiolarite and the radiolarians are only sparcely scattered through the iron-stained matrix. The silica content of the radiolarite and red shales is higher than of ordinary shales, while the iron content approach in many respects that of ordinary shales. It seems that the iron is only a pigment and has no cementing effect, because when the iron oxide is dissolved out by acids, the siliceous shales do not break down.

The green or red colour of the shales is dependent on the proportion of ferric iron to ferrous iron (Grunau 1947). The green colouring is mostly of secondary origin, because it is not related to the bedding.

In the matrix a lot of organic debris occurs, only visible in thin section, among which fragments of radiolarians, siliceous spicules of sponges and other indeterminable fragments, mostly consisting of chalcedony or quartz. Calcite occurs only to a small extent in the matrix.

Concerning the origin of these rocks we can confirm the description by Kanis (1955), that these rocks are formed in a transgressive sea under oxidizing conditions. Davis (1918) has concluded in his excellent description of Jurassic radiolarian cherts, that the cherts and their associated red shales, in California, do not represent abyssal radiolarian ooze, but represent depositions in shallow water or water of moderate depth. He discusses (p. 360) the possibility that the red shales may be deposited as red mud of terrigenous origin, deposited near the shores of tropical lands and intermixed with radiolarian oozes.

The opinion of Shackleton Campbell (Moore 1954, p. D 17) — "*Radiolarians* may be deposited in much shallower water, however, and it is even probable that some deposits consist of strand-line accumulation carried inland under special circumstances" — and of Delépine (1935) that a sudden transgression over an extended continent, covered by laterite soil, has resulted in the sedimentation of red mud enriched by radiolarians, are in agreement with the view of Davis.

The griotte of the present area is composed of an alternation of red or green shales with very fine grained limestone, developed in layers of ± 1 cm, or in layered, subangular or rounded nodules. The siliceous content of the shales is strongly variable. By extreme high siliceous content radiolarite and red chert is developed. By low siliceous content the muddy shales are very incompetent and often calcareous. The calcareous nodules and the shaly matrix have the same colour variation from red, violet to green. The rounded and subangular nodules are not formed by fragments of goniatites or other fossil debris. In some places we find in the fine grained limestone of which the nodules are made up, rare scattered radiolarians, filled with calcite.

When the limestone is layered, the layers have a thickness of 1 to 30 cm. Between these limestone layers thin films of red shaly material occur. Sometimes the shaly films run like stylolitic partings through the limestone layer. All transitions exist between the layered limestones and the griotte.

The calcareous fragments in the griotte grade from a few mm to 60 mm in diameter. The nodules often give the impression of swimming in a matrix of red shale; the shale laminae curving around the nodules.

The following development can be seen: the calcareous fragments fit into each other along an intricate line, like fragments of a jig-saw puzzle; by drifting from each other, the angles become rounded and the contacts with the shales

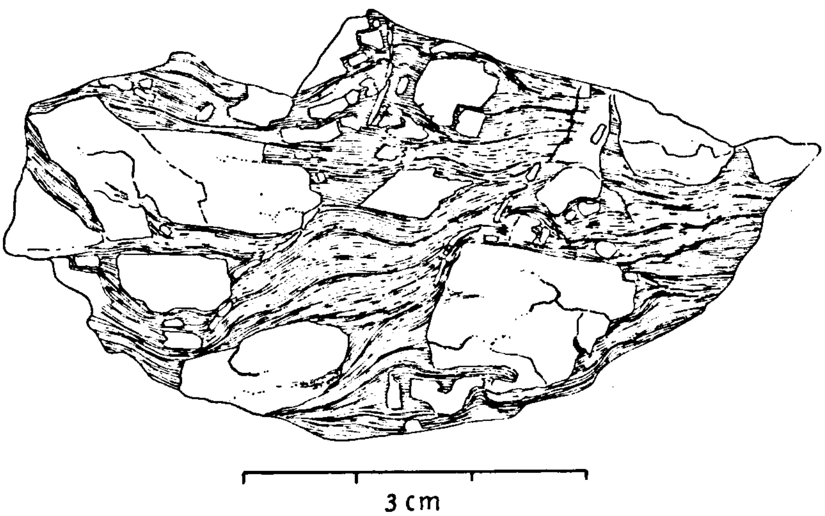


Fig. 7 Griotte; angular limestone fragments broken along irregular lines in a muddly matrix.

become straighter. Small calcareous particles along that contact line split off, sometimes remaining as a calcareous core, in other places the boundaries with the matrix become hazy and the calcareous material is mixed with the shaly matrix (fig. 7). The rounded fragments are usually ovate, flattened parallel to the bedding and generally lighter in colour than the matrix. They often keep their arrangement in rows parallel to the stratification.

The pseudo-brecciated texture seems due to a reworking of thin alternating layers of lime mud with red calcareous mud in an early diagenetic stage. The limy material was already partly consolidated, breaking along haphazard planes, due to agitation of the waters and possibly partly to desiccation; whereas the muds were unconsolidated and water-saturated, forming a flowing mass, piercing between the limestone fragments.

Indications for the development of oscillation ripple marks during sedimentation, as proposed by Kanis (1955, p. 400), have not been found. Also the nodules show no preferred orientation in the bedding plane.

The sedimentary nodules react in later deformation as competent fragments in an incompetent matrix. With a strong tectonization the shale films have been forced into the cleavage planes, which bend around the nodules and an elongation in the tectonic b axis may develop. In this way the rock will give the impression that its texture is developed by tectonic forces only.

Sometimes a rich goniatite fauna is preserved in the nodular limestone layers (known from adjacent areas). The goniatites or fragments of goniatites, more solid than the surrounding rock during reworking, will form the nodules and the shale will flow around them. However, this only occurs locally, the usual nodules are not formed by goniatites, as Dufrénoy (1833), Barrois (1882) and others have supposed.

The age of the griotte in the Cantabric-Asturian mountains is mainly based on goniatite determinations. Data of Barrois (1882), Delépine (1935, 1943), Wagner Gentis (1960) and Kullmann (1961) have shown that the griotte can be dated as Visean. The later author, especially, has made several detailed biostratigraphic sections through the Lower Carboniferous strata of localities in the provinces of León and Palencia, with special references to the goniatites. Most griotte layers are of Upper Visean age after Kullmann (1961), with exception of some localities in the Aguasalio syncline on the eastern border of the Rio Esla (prov. León), which include lower layers of Lower Visean age (*Pericyclus*-Stufe, of German authors).

In the present area no goniatites have been found. However, several griotte samples have been investigated for their content of conodonts. The following genera and species are identified from two localities, by Mr. H. A. van Adrichem Boogaert:

1) 1 km NW of Camporredondo de Alba (Loc. E 106)

Gnathodus cf. bilineatus (Roundy)	Upper Visean-Namurian
Gnathodus sp. Pander	Carboniferous

2) 125 m SSW of Triollo, along the border of the lake of Camporredondo, below the high water level (Loc. E 85)

Gnathodus texanus (Roundy)	Visean
Gnathodus bilineatus (Roundy)	Upper Visean-Namurian
Prioniodina sp. Ulrich & Bassler	Devonian-Carboniferous

The exact range for most of these conodonts is not yet known. The upper boundary of the Namurian is especially difficult to fix because of the poor development of marine Namurian in Europe. The occurrence of *Gnathodus texanus* (Roundy) together with *Gnathodus bilineatus* (Roundy), however, indicate an Upper Visean age (v. Adrichem Boogaert, personal communication), which is in agreement with the goniatite determinations from the adjacent areas.

Caliza de Montaña facies. — The thick Carboniferous limestones, which form extensive chains in the Cantabric-Asturian mountains, for example in the Sierra del Brezo, the Picos de Europa, the Covadonga mountains etc., are known in Spanish literature as "Caliza de Montaña" and named by Barrois (1882) "calcaire des canons".

In our area the massive limestone has a grey-bluish colour. In most places bedding can hardly be detected and an intensive joint pattern can be very confusing. Often the limestone is strongly recrystallized and marmorized. In several localities (north of Villanueva de la Peña, near Velilla del Rio Carrion and south of Besande) white or red marble is quarried.

Dolomitization occurs extensively. The valley between the Peña Redonda and the Peña Cotolorno consists completely of yellow-brown dolomite. To the south this dolomite zone is bordered with a straight boundary by the Cotolorno fault. This is the reason why the Cotolorno fault is clearly visible in the field and on the aerial photographs. Near the Peña Dorada the limestone is also dolomitized. The dolomitization is not related to the bedding and is to be considered as a secondary process, this in contrast with the opinion of Delépine (1935), who described only the uppermost 200 m of the "Caliza de Montaña" as Mg rich and often dolomitized.

An interesting mineralogical phenomenon is the occurrence of small bipyramidal quartz crystals, authigenic quartz, scattered through the limestone. The quartz crystals appear typical for the limestone of the Caliza de Montaña facies. De Maestre (1864) remarks: "La presencia de estos cristales es tan caracteristica en la caliza anthraxifera (Caliza de Montaña) de esta provincia, que muy bien puede suplir a la falta de fossiles". (Descripc. de Santander 1864, p. 47).

The occurrence of these quartz crystals is in our area indeed limited to the limestone of the Caliza de Montaña facies. They have never been found in the Devonian and higher Carboniferous limestones.

The dimensions of the quartz crystals are variable, but can reach a length of 2.5 cm and a diameter of 0.5 cm. They consist of euhedral hexagonal prisms, terminating at both ends in pyramids. The prism planes are horizontally striated. The colour of the crystals is bluish-grey. The quartz crystals are scattered through the limestone rock with a random orientation; rarely they show a preferred layering near joints. A preferred orientation of the quartz crystals has been found in the concentric laminae of a stromatoporoid, found in the top of the limestones. The quartz crystals have never been found in dolomite, although they do occur in strongly recrystallized limestone.

In an alternation of limestone and dolomite, on the size of a hand specimen, coloured by $CuSo_4$, the quartz crystals remain particularly restricted to the limestone.

Thin sections show that there are no indications that the quartz is recrystallized around a detritic quartz core, as often has been found in oölitic rocks or arenaceous limestones (authigenic quartz, Pettijohn 1949, p. 498, Henbest 1945 e. g.).

The secondary recrystallization of quartz has only partly replaced the carbonate rock.

A great number of carbonate inclusions form zones parallel with the crystal faces of the quartz crystal, alternating with zones where the carbonate is completely replaced, giving an impression of growth-lines. The inner part of the crystal is sometimes totally free from carbonate inclusions, at other times a great number of inclusions occur.

The crystals are euhedral and cut across all textures and structures of the calcite with a straight outline. Also the later recrystallized calcite is partly replaced (fig. 8), so post-dating the quartz crystallization. Micro-cracks in the calcite matrix run also into the quartz crystals, and have caused a knicking or slight bending of the quartz crystals.

The calcite crystals, which immediately surround the quartz metacrysts, often are crushed by the pressure of the growing quartz crystals.

Summarizing we find:

1. The above described quartz crystals are restricted in their occurrence to the limestone of the Caliza de Montaña facies.

2. They are scattered through the rock and have a random orientation.
3. Replacement of secondary recrystallized calcite by quartz, post-dates the quartz crystallization.
4. The physico-chemical conditions in the dolomite do not facilitate the crystallization of quartz.
5. The sharp outlines of the quartz crystals and the crushed calcite crystals random around are indications of a rapid crystallization.
6. The parallel zones of calcite inclusions, indicating zones of complete replacement and zones of few replacement, however, point to several stages in crystallization under various physico-chemical circumstances.
7. The source of silica remains obscure.

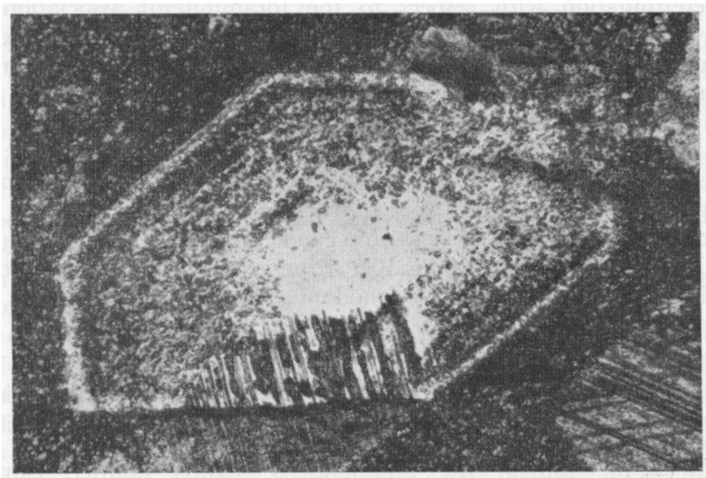


Fig. 8 Authigenic quartz crystal with zones of calcite inclusions Cal. de Montaña facies (× 24, Loc. K 9).

The base of the limestone is sometimes developed as a dark-blue well layered calcilutite. In thin section it is a very fine grained limestone with a dense and homogeneous texture. Fossils do not occur in the calcilutite, the quartz crystals mentioned above are also absent. The platy well-layered calcilutite is especially well developed in the thrust fault area of San Martin. Here, the Ruesga group has been developed mainly in the Culm facies. The limestone is reduced to a thickness of 20—40 m, with massive grey limestone at the base and the well layered calcilutite on the top, in contrast with the position of the calcilutite on other places.

The uppermost part of the limestone in the Valsurvio dome is more shaly, and the layers are often strongly detrital. Crinoid debris, coral fragments and some brachiopods, all indeterminable, can be discerned. In these layers rare foraminifers can be found.

The foraminifers from the limestone have been studied by Mr. A. C. van Ginkel (Leiden), who has given the following genetic determinations:

1. Peña Redonda (Loc. XCVII) [1]
   Pseudostaffella sp.

[1]) Locality numbers of van Ginkel.

Millerella (Eostaffella) sp.

2. N of San Martin de los Herreros (Loc. XCV) [1])
   Millerella (Millerella) sp.
   Parastaffella sp.
   Pseudostaffella sp.
3. Peñas Blancas, SW of Besande (Loc. E 7) [2])
   Parastaffella sp.
   Profusilinella ? sp.
   Bradyina sp.
   Endothyranella sp.
   Palaeotextularidae.

An age determination with respect to the foraminiferal association is not yet justified (van Ginkel, personal communication [3]). The age of the Caliza de Montaña given by Kanis (1955, p. 402) is Upper Visean — Lower Namurian. Three goniatite fragments found in the limestone, overlying the griotte near Santibañez de Resoba (just outside the mapped area), are of Upper Visean age. Kullmann (1961, p. 228) has determined these goniatites as: *Goniatites (Goniatites)* sp. ex. gr. *granosus* and *Prionoceras (Imitoceras)* cf. *globosum* Schdw. of the Go γ-zone of the Upper Visean.

In addition, a determination of some conodonts from the limestone, 1 km NW of Rebanal de las Llantas (Loc. E 101), directly overlying the griotte, confirms the age given by Kanis (1955). Conodont determinations by Mr. H. A. van Adrichem Boogaert:

Gnathodus bilineatus (Roundy)	Upper Visean-Namurian
Gnathodus gyrtyi Hass	Visean-Namurian
Gnathodus cf. *commutatus* (Branson & Mehl)	Visean-Namurian

Sometimes the top of the limestone is slightly oölitic. Limestone-shale intercalations and lateral intergrading of shales in limestones occur in several places. North of Villafria such an interbedding is well exposed. Slabs of shale up to 50 cm in diameter are found, sometimes slightly deformed, in a matrix of detrital limestone, indicating a reworking of the clay in an almost consolidated state. The shale has a light brown colour.

The thickness of the limestone is very variable. In the southern limb of the Valsurvio dome with isoclinal folded limestones, the thickness is difficult to estimate, but reaches several hundreds of meters. Kanis (1955) gives a thickness of 300 m for the "Caliza de Montaña" in the adjacent area.

In the province of León the thickness is 800 m after Delépine (1935) and Comte (1959) and 2000—2500 m near Vegacervera after Wagner (1958). In the last mentioned outcrop, however, the limestone is intensively folded in isoclines, resulting in a frequent doubling of the layers in the road section, which is easily overlooked. After Julivert (1957) the thickness of the "Caliza de Montaña" in the Asturian basin amounts to 100—300 m.

In the limestone a few indications of the occurrence of mineral veins has

[1]) Locality numbers of van Ginkel.

[2]) Locality numbers of the present author.

[3]) "All samples from the Sierra del Brezo examined on their foraminiferal content, belong to the lower part of the *Profusulinella* zone. More detailed information, regarding the correlation of these strata with the Russian marine Carboniferous will be published in due course".

been found, small traces of azurite and malachite occur throughout. In the Peña Cotolorno some veins of arsenopyrite and chalcopyrite, which run from the Camporredondo quartzite into the limestone of the Ruesga group, have been exploited on small scale a few years ago.

The Culm facies of the Ruesga group. — In contrast with the thick limestone deposition of the Caliza de Montaña facies, the Culm facies is characterized by a large detrital content. It is a thick sequence of shales, mudstones, subgraywackes and conglomerates. In the thrust fault area of San Martin they overlie a limestone strongly reduced in thickness, which is developed partly as a black well-layered calcilutite, of a thickness of 20—40 m, with at its base the griotte and red shale horizon. The calcilutite grades upwards into 1 m of calcareous shales followed by a ½ m of limestone, often coarse detrital, with a brown colour on weathered surfaces. This limestone is followed by shales grading upwards in mudstones, sandstones and subgraywackes. A thick formation with a very monotonous character. Fossils have not been found, with exception of indeterminable plant debris. All transitions between mudstones, which are sometimes calcareous, and dirty micaceous sandstones, of a brown-grey colour, occur. The sediments are badly sorted and show rapid lateral facies changes.

The isoclinal structures are difficult to recognize and the amount of deformation is not known, because the bedding is often difficult to identify and the strata are very uniform in texture. So the thickness remains unknown. A cleavage has been developed to some extent.

North of Rebanal de las Llantas a 30 m thick quartzite conglomerate is intercalated between the shales, about 100 m above the base. The quartzite pebbles and boulders are well rounded; they show a great variation in diameter, ranging from 5—40 cm. Often the boulders are broken, which is due to tectonic forces. The matrix consists of sandstone and subgraywacke. The boulders do not touch each other. Shale and subgraywacke lenses occur between the conglomerates. Washouts filled up by boulders, loading of the boulders in muddy material and imbrication of the boulders and pebbles can be used as top and bottom criteria.

In the shales, mudstones and subgraywackes it is very difficult to define top and bottom of the layers, but by a painstaking examination, small scale current bedding and foreset lamination often can be detected.

In these strata many polymict conglomerates are found, which are built up of rounded quartzite pebbles and of subrounded to subangular limestone and shale fragments. The material is of local origin, due to contemporaneous erosion and has suffered very little or no transportation before resedimentation. The limestone fragments are badly sorted. Their size ranges from 0.5 cm up to 12 cm, but especially fragments of 1—2 cm are abundant, whereas also fragments of ± 10 cm diameter occur frequently. They are elongated in the bedding plane. The quartzite pebbles range from 2—15 cm, they are well rounded and often show an elongation in the bedding plane, which is, however, less pronounced as that of the limestone fragments. Small blue shale slabs of 0.5 cm in size, lie scattered through the rock, orientated also in the bedding plane. The matrix is composed of a calcareous subgraywacke.

These sediments are originated in an unstable shelf environment. The source rock of the limestone fragments is the Carboniferous limestone of the Caliza de Montaña facies, contemporaneously deposited south and north of an E—W running zone with sediments of the Culm facies.

The subangular shale fragments indicate a reworking with little transport of older sediments, in a rather consolidated stage. The well rounded quartzite pebbles, on the contrary, indicate long transport and rather vigorous water movement. The association of quartzite pebbles and limestone and shale fragments can be easily understood, when we consider the quartzite pebbles as acquired from conglomerates, due to reworking.

Also the occurrence of well rounded dispersed quartzite pebbles and cobbles in shales and mudstones indicate a redeposition of conglomeratic material in a muddy environment. Ackermann (1951) favoured an origin involving seaward slumping and sliding of coastal clays, mixed with pebbles and other clastic material during sliding. Crowell (1957) has given a description of this type of conglomerate, which he calls a "pebbly mudstone", as a result from sliding, when unstable layers are laid down by turbidity currents on watersoaked mud.

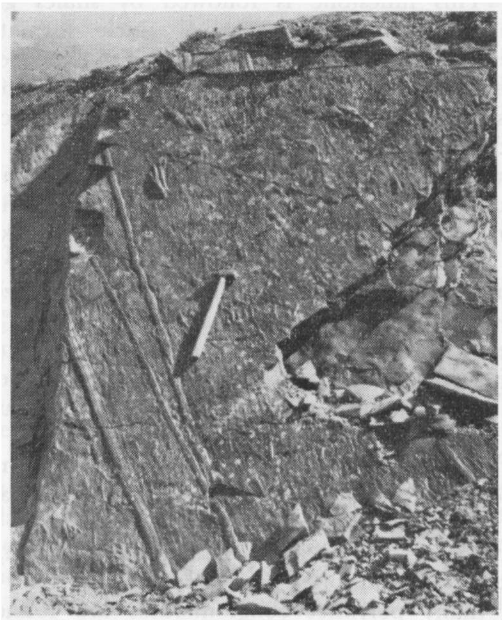


Fig. 9 Groove casts, organic trails and small load casts, Culm facies of the Ruesga group near the dam of the Pantano de Ruesga.

South of the road La Lastra - Santibañez de Resoba, between km 16 and km 19, a rather regular alternation of thin subgraywacke and shale layers has been encountered. The micaceous subgraywackes, which are often calcareous, grade upwards into more shaly material. The boundary between the shales and the overlaying subgraywackes is very sharp. The thickness of one such a sequence, from the base of a subgraywacke bed upto the base of the following subgraywacke bed, is about 10 cm. When we consider the rocks more in detail, than an evident grading does not occur. The sandy part nearly always shows a fine lamination, due to small differences in composition. Parallel as well as cross lamination occur.

A similar sequence of alternating subgraywackes and shales has been found in the adjacent easterly area, near the dam of the reservoir of Ruesga. In a small roadcut bedding planes are exposed with beautiful developed sole markings (fig. 9). Especially groove casts are well developed. They reach a length of several metres, their width is about 2—5 cm. Most of the groove casts are beset with minor ones. They lay roughly parallel. Sometimes they show a faint meandering in their traces.

The direction of the groove casts, indicating the current direction, is in agreement with the orientation of plant debris. Organic trails also frequently occur on the sole of the subgraywacke. Burrowings have not been found in these beds. Therefore the trails were originated by creeping organisms on the muddy bottom, before deposition of the overlaying subgraywacke (antecedent mud burrows).

Small scale load casts are found on the soles of the same beds. Often these small load casts occur together with fine striations. These striations are tightly crowded narrow ridges, orientated along the current direction, they are about 1 mm wide and deep. They join each other and split up again. These striations are still faintly visible upon the load casts. It is evident that the load casts are later developed than the striations. Similar markings are described by Nederlof (1959, p. 678) in relation with turbidites, as "problematic features".

There seem to be well founded reasons to consider these sediments as not deposited by turbidity currents because:

1) The lateral extension of the beds is rather small (only several hundreds of metres), which is in contrast with turbidites.
2) A clear grading has never been found. The more or less sandy beds are often laminated from bottom to top, due to an alternation of more or less clayish sand laminae. The material is rather badly sorted. [However, lamination in turbidites is also common, it has been described by many authors (van Houten 1954, ten Haaf 1959, Bouma 1961). Bouma distinguishes five intervals in a complete sequence (p. 53):

    e) Pelitic interval.
    d) Upper interval of parallel lamination.
    c) Interval of current ripple lamination.
    b) Lower interval of parallel lamination.
    a) Graded interval.

   The lowest graded interval may be absent by lack of coarse material and if the material is well sorted, current ripple lamination and convolute bedding often occur in the laminated sandstones and predominate the grading].
3) Absence of flute casts and other strong erosional features on the soles of the subgraywackes.
4) A sole is exposed, where two directions of groove casts can be distinguished, crossing each other under an angle of 60—70 degrees. The first set is disturbed by organic trails. The second set, developed in an other direction, is cross-cutting as well the burrowings as the first set. So we may conclude, that they have not been formed at one and the same moment. The first set was formed by dragging of a piece of wood or other resistent object over the soft muddy bottom. Organic trails crossing these groove casts, formed later. They are cut by a second set of groove casts, with another direction, due to changes in current direction. A rather sudden supply of coarse detritic material has burried these tracks and trails on the soft muddy bottom. Violent turbulent currents would have eroded the upper layers and destroyed the tracks and groove casts. This has never been found, so a deposition by turbidity currents is not probable. Kulick (1960) described

groove casts from the Culm in Sauerland (Germany), as developed by dragging of plant clusters, but not related to turbidity currents. Teichmüller (1960) has found them recently on the beach of Tunesia, originated by algal clusters dragging across a tidal flat, inland by a rising tide or seaward by undertow, plowing grooves in the soft muddy bottom. They show a great resemblance with the groove casts found by Kulick.

5) Convolutions do not remain restricted to turbidites. They may develop in water-saturated intraformational layers, which have a small angle of inclination, Williams (1960). The size and the intensity of contorting vary from place to place. To above and to below the contortions die out. The base and the cover of the convolute horizon are entirely undisturbed. Rich (1950) considers them as been formed after deposition of the overlying beds. A connection of convolutions with current ripple marks as proposed by ten Haaf (1956) has not been observed. The wide troughs and sharp anticlines of the convolutions can serve as top and bottom criteria, whereas the inversion of the folds indicate the direction of movement of the water saturated intraformational mass. The occurrence of these layers together with polymict conglomerates and pebbly mudstones as described above indicate sedimentation circumstances on an unstable shelf with slumping and sliding and a mixing up of the material on small scale. However, convincing indication for sedimentation by turbidity currents on extended scale has not been found.

Yuso group

This group contains the sediments which are younger than the Curavacas folding phase and older than the Asturian phase. The boundaries of the group are characterized by clearly developed angular unconformities. Kanis (1955) has described the lower part of this group (p. 405), called by him the Curavacas formation, which consists of a 500 m thick oligomict quartzite conglomerate. This conglomerate is situated with a clear angular unconformity upon older formations of the Carboniferous Ruesga group and the Devonian, near Resoba. The conglomerates are exposed in the Rio Yuso, more to the west as two thick layers, separated by shales.

In the area described here, the base of the Yuso group has been formed by pelitic sediments and in subordinate amounts of conglomerates. The sediments strongly resemble the Culm facies of the Ruesga group in lithological outlines. In the zone east of the Pantano de Camporredondo both occur and it is impossible to fix the boundary between them accurately. Sometimes, the dark blue well-bedded calcilutite of the Ruesga group grades upward into shales and subgraywackes of the Culm facies, the sedimentary transition is gradual, and there is no indication of a stratigraphical break. Elsewhere the shales and subgraywackes unconformably overlie the calcilutite and Devonian strata. Locally clear angular unconformities are visible, as for instance in the valley of the Arroyo de Valderrianes and on the northern border of the Pantano de Camporredondo. The low-angle overthrusts are also unconformable covered by shales, subgraywackes and to a small extent conglomerates of the Yuso group (W of La Lastra).

These tectonic differences were the most important features, why we have distinguished two sequences of different age. The shales and subgraywackes of the Culm facies in the east of the area are deformed and thrusted together with the limestones and griottes of the Ruesga group, whereas in the west, a sequence of shales and subgraywackes unconformably overlie these thrusted limestones and griottes.

Some small lithological differences can also be seen. In the polymict conglomerates of the Ruesga group, the limestone pebbles have been derived from the Carboniferous limestone (Caliza de Montaña facies) only, whereas the limestone pebbles of the Yuso group were derived from the Carboniferous limestones as well as from the Devonian limestones.

The conglomerates are very interesting. On the east shore of the Pantano de Camporredondo, between the mouth of the Arroyo de Miranda and the Arroyo de Valderrianes, the conglomerates are well exposed below the high water line of the reservoir and show evidences of having been formed by mud flows. The rock is composed of dispersed pebbles in a mudstone matrix. The quartzite pebbles and cobbles, well rounded, are 8—20 cm large and have an elongated shape. They show a preferred orientation of their longest axes in the current direction. The pebbles do not touch each other. Higher in the sequence, the amount of quartzite pebbles decreases, whereas the amount of limestone fragments increases. These limestone fragments are subangular and tabular in shape and vary in size from 5—30 cm. They show an imbricate arrangement which agrees with the flow direction found from the orientation of the pebbles. The packing is more compact, the pebbles and fragments often touch each other and are badly sorted. The matrix becomes more sandy. The sources of the limestone fragments are the dark blue calcilutites of the base of the Ruesga group and the Devonian crinoid-limestone, which crops out nearby in the Arroyo de Miranda. Higher in the sequence, the size of the limestone fragments decreases (0.5—5 cm) and small shale pellets occur (2—4 cm in diameter and 0.5 cm thick). These shale pellets are reworked when the clay was still in an unconsolidated stage. Their lower side is flat or slight undulated. The upper side, however, is more rounded. Thinning out in one direction they have more or less the shape of a drop, but flattened on one side. In the bedding plane they are subrounded - subangular in outline. The thinning out in one direction and the rounded nose in the other direction, visible in cross section, are indicative of a current in the latter direction, which is in agreement with observations obtained from the pebbles.

In this outcrop the layers are in an inverted flat position. This sequence is 5—8 m thick and is separated by a few metres of mudstone from an other 3 m thick layer of polimict conglomerates composed of quartzite pebbles and cobbles (8—20 cm), limestone pebbels (3—15 cm) and a few shale pellets.

The quartzite pebbles and boulders are often broken or crushed, especially when they are closely packed.

Flow lines in the matrix demonstrate the development by mud flow; a dense muddy mass sliding downwards and taking large heavy quartzite boulders with it at the bottom and lighter limestone fragments and clay pellets higher up in the mud flow.

The mountain ridge southwest of Cardaño de Abajo is wholly built up by quartzite conglomerates, several hundreds of metres thick. When we follow this ridge to the east, towards the Pantano de Camporredondo, the mighty conglomerates pass laterally into shales and polymict conglomerates, compound of well rounded quartz pebbles and huge angular limestone blocks, maximal 150 cm in size.

The pelitic sediments of the Yuso group are mostly dark coloured. Between Valverde de la Sierra and the most easterly extension of the Yuso group, south of La Lastra, a strong cleavage, with recrystallization in the cleavage planes, has been developed in the incompetent rocks. Very small euhedral chloritoid crystals (10—40 μ) can sometimes be discerned in thin sections, which are due to the low regional metamorphism, already described from the Devonian strata.

To the west these slates pass laterally into shales and mudstones. The intensity of the cleavage development decreases in this direction.

The feldspatic content of all these sediments is almost negligible. Fossils have not been found in these strata.

Cea group

This group includes the Carboniferous sediments, which are younger than the Asturian folding phase. Their name is derived from the Rio Cea area, which adjoins the present area to the west.

Thick oligomict and polymict quartzite conglomerates occur in the base of this group in the Cea area and overlie with an angular unconformity the layers of the Yuso group (Tejerina syncline and the Cueto syncline). In Cueto mountain these conglomerates pinch out rapidly to the east, and shales, siltstones and coal-beds unconformably overlie the thick massive limestone of the Peña Lampa and the Peñas Blancas. A fossil flora of the base of the sequence, determined by Prof. F. Stockmans (personal communication), indicates a high Westphalian D. age. Above the conglomerates there is the coal bearing sequence of the Cea basin, which will be described in due course by Mr. H. M. Helmig.

The Guardo-Cervera coal-basin is situated at the foot of the southern slopes of the Sierra del Brezo. It is a zone, 1 to 2 km broad, which can be followed to the east as far as Cervera de Pisuerga. The contact between the coal bearing strata of the Cea group and the limestone of the Ruesga group is a fault. West of the Rio Carrion the coal-basin widens and the coal-bearing strata bend to the north. Here the contact is an angular unconformity.

The sediments of the Guardo-Cervera coal-basin have been deposited under paralic environments. Marine influences can be deduced from the marine faunas, as found by Hernandez Sampelayo (1944), north of Santibañez de la Peña and by the author in many other localities. The maximum thickness of the strata measured along the valley north of Velilla de Tarilonte is 1100 m. From the Wealden in the south, to the limestone mountains in the north, there is one continuous sequence in a nearly vertical position. West of Santibañez de la Peña the strata are thrown into concentric folds.

All gradations between shales, mudstones and graywackes occur. The shales are mostly dark coloured, often ferruginous, with many fossil trails on the bedding planes. Moreover, a marine fauna of brachiopods, gastropods, lamellibranchs and corals is often found in these shales as well as in layers with a low sandy content. With increasing sand content the shales become lighter coloured and a tendency is observed for the marine invertebrates to be absent, whereas plant debris becomes increasingly evident. These mudstones or sandy shales are usually highly calcareous.

The graywackes and subgraywackes are poorly sorted, consisting of angular to subangular grains of quartz, micas, calcareous material, felspars and shale, slate and sandstone fragments. The rock fragments range in size upto 8 mm. The colour is predominantly gray-green. The matrix is formed by a mixture of secondary silica and sericite. Current bedding, burrowings and plant debris are often found. A first, and sometimes a second cleavage, can be observed in the slate fragments. The cleavage directions of the different fragments are so randomly disoriented, that it is obvious, that the fragments have been reworked after the development of both cleavages. These slate fragments show a resemblance to the Devonian slates of the Hornalejo and Compuerto formations.

The coal is anthracitic; analyses given by some coal mines, show that the volatiles vary between 2.5 and 11 % and the ash between 2.5 and 10 %. For detailed information concerning coal mining we refer the reader to Quiring (1939).

The three coal groups have been described by Sanchez Lozano (1906), from north to south, as "Grupo del Nord, Central and del Sur". An objection against the use of these names is, that the most northerly coal seams do not always belong to the same group. Quiring has called them, respectively "Requejada, Santibañez and Aviñante group", the names which we shall use here.

The Requejada coal group is exploited north of Villaverde de la Peña by the "Compania de Estrellaverde", previously the "Exploitation Requejada y la Positiva". Four coal seams are exploited, having a total thickness of 335 cm. In this mine, north of Villaverde de la Peña, the Requejada group is about 200 m from the fault contact with the Carboniferous limestone. To the east, north of Velilla de Tarilonte, this coal group approaches the fault contact, resulting in a strong tectonization of the coal seams, which are no longer exploitable. Further to the east only a few coal lenses accompany the fault contact. In a section near Traspeña (Appendix 3, Kanis 1955) the same group crops out again. To the west the Requejada group can be followed as far as Villafria, where it is cut off also by the border fault zone.

The Santibañez group is amongst others exploited north of Santibañez de la Peña. The following exploitable layers occur: coal seam 1 : 0.80 m, seam 2 : 0.80 m, seam 5 : 2 m and seam 7 : 1 m. Near Villanueva de Arriba this group is exploited by the Minas del Acebal and San Luis. The exploitable coal seams have a total thickness of 6.10 m. The layers have been deformed in simple concentric folds, which result in a repeated outcropping of the coal seams; of especial importance for the mining possibilities.

The Aviñante group runs parallel with the boundary of the Cea group against the Cretaceous. Its name is derived from the village of Aviñante. Four workable seams occur of which the third is about 2—3 m thick. Near Velilla de Tarilonte the total thickness of the workable layers is 4.90 m.

Between the Requejada and the Santibañez coal groups 265 m of coal-poor strata occur and between the Santibañez and the Aviñante coal groups 300 m.

The following ideal cyclothem can be discerned in these strata from bottom to top: underclay with stigmaria beds, often sandy, light gray — coal seam — sandy shales with plant fossils, light coloured — dark blue or black shales with marine invertebrates and fossil trails, often ferruginous — dark sandy shales with marine invertebrates and ironstone concretions — gradual increase of the clastic content, together with a decrease of the occurrence of marine invertebrates; plant debris, current bedding, ripple marks and burrowings occur in this unit — graywacke and subgraywacke. These graywackes are overlain with a sharp boundary by the underclay of the succeeding cyclothem.

These cyclothems have a thickness which varies from 1.5—8 m, in proportion to the amount of clastic material. Coal seams are not always developed, but stigmaria horizons indicate the vegetation levels.

From west to east in our area, the evidences of marine and brackish water conditions become more prominent, whereas the terrestrial influences, and especially the coal development diminish.

To the west, in the Cea basin, even no marine faunas have been found (Helmig, personal communication).

According to Mr. R. H. Wagner (Sheffield) and Prof. F. Stockmans (Bruxelles), who have kindly determined the fossil floras of this region, the Guardo-Cervera coal groups belong to the Westphalian D — Stephanian A. The determinations by Wagner have been published in Wagner 1959, p. 406—407. The determinations by Stockmans are listed below:

Requejada group: *Annularia stellata* (Schlotheim)
Pecopteris polymorpha Brongniart
Sphenophyllum oblongifolium Germar & Kaulfuss
Pecopteris indéterminable
Linopteris brongniarti (Gutb.)

Santibañez group: *Callipteridium gigas* (Gutb.)
Linopteris brongniarti (Gutb.)
Alethoperis grandini Brongniart
Asterotheca sp. (cf. A. arborescens)

Aviñante group: *Alethopteris grandini* Brongniart
Annularia sphenophylloides (Zenker)
Calamostachys sp.
Mariopteris carnosa Corsin
Pecopteris sp.
Linopteris brongniarti Gutb.
? *Pseudomariopteris ribeyroni* Zenker
Pecopteris polymorpha Brongniart

Determinations of the marine fauna have not been completed yet. Miss G. E. de Groot has kindly determined some poorly preserved corals, collected from the base of the Aviñante coal group north of Velilla de Tarilonte: Fam. *Aulophyllidae* (somewhat like *Dibunophyllum*); *Amygdalophylloides sp.; Pseudotimania sp.*; Fam. *Hapsiphyllidae.*

Hernandez - Sampelayo (1944) gives the following list of determinations from a fauna collected north of Santibañez de la Peña:

Natyria lyrata Phill., *Naticellas* spec., *Naticopsis neritopsidae, Bellerophon* spec., *Loxonemos* spec..

The stratigraphic value of these fossils is rather small.

CRETACEOUS AND TERTIARY

The Cretaceous crops out on the southern border of the area. It is a narrow zone, 1—2 km broad, with an E—W trend that can be followed to the west over a long distance.

For an extended litho-stratigraphic description in addition to paleontological features, we refer the reader to the valuable study by Ciry (1939), on the Cretaceous in the provinces of Burgos, Palencia, León and Santander.

The Lower Cretaceous here is of the Wealden facies. The base has been formed by a 5 m thick conglomerate, with a small angular unconformity overlying the Carboniferous strata.

From east towards the west there is an increase of detrital material and a decrease of the thickness of the layers of the Cenomanian - Turonian marls. Further west in the province of León, Ciry distinguished the Santonian immediately overlying the Wealden facies, indicating a transgression of the Upper Cretaceous from east to west (cf. Ciry 1939, fig. 74). Whether this is due to the absence of the Cenomanian, Turonian and Coniacien, or to a development of these series in a Wealden facies could not be established.

The Coniacian is built up of a hard yellow-brown limestone, strongly fossiliferous, which alternates with detrital marls. On the top we find sometimes a glauconite-rich marl.

Along the road from the station of Santibañez de la Peña to Viduerna a continuous section through the Santonian and Campanian can be observed.

The base of the Santonian consists of a white-blue limestone, very hard and microcrystalline, with some small marl intercalations. It is overlain by a grayish-white massive limestone, which is deposited as a calcareous foraminiferal mud. The foraminifers reach a dimension of 4 mm.

A red creamish Campanian limestone, strongly recrystallized overlies the Santonian deposits. In thin section it is a micro-conglomerate, of which the pebbles are recrystallized organic debris. In some of them vague structures of fragments of foraminifers, alges, echinoderms and bryozoans are visible. In the reworked material *Prealveolina tenuis* could be distinguished, which points to a reworking of Cenomanian material (J. Hofker, personal communication).

Above the calcirudite there is 5 m of sandy marl, of a light colour and rich in fossils, which grades upwards into a very hard creamish fine crystalline limestone. The total thickness of the Campanian and Maestrichtian strata is 375—325 m. Dr. Osman has determined the following foraminifers:

Lacazina elongata Munier-Chalmas (abundant)
Lacazina cf. compressa var. *galloprovincialis* Munier-Chalmas and Schlumberger
Dictyoconella complanata Henson
cf. *Dohaia planata* Henson
cf. *Lituonelloides compressus* Henson
Lacazina lamellifera Silvestri
cf. *Dictyoconella minima* Henson
cf. *Idomia iranica* Henson
cf. *Orbitolina kiliani* Prever
Orbitolina concava (Lamarck) var. *sefini* Henson
Rotalia trochidiformis.

The rock contains further species of: *Miliolidae, Cyclolina, Cuneolina, Dictyopsella, Globotruncana* (Loc. K 181, K 182, K 183, K 192). He considers this fauna as indicating a Campanian or Maestrichtian age. The total thickness of the Cretaceous strata is about 800 m.

The Cretaceous is overlain, with a slight angular unconformity, by thick conglomerates. The pebbles and boulders are derived from Cretaceous limestones, which often contain a great number of *Lacazina* individuals. For a detailed sediment-petrological description we refer to Mabesoone (1959), who described these conglomerates as the Cuevas facies (p. 50). Clear current-bedding has been found at some places in the intercalated sandstone beds, showing that the layers are overturned. An Eocene-Lower Miocene age is usually accepted by most investigators.

At several places sediments occur, which consist of sandstones, marls and conglomerates, overlying the Cretaceous rocks with a slight angular unconformity and which in their turn are unconformably overlain by the limestone conglomerates of the Cuevas facies. These sediments have been noted by Ciry (1939) from the Arroyo Pazo Brigaderos. He compares them with the "Grès de las Bodas" (deposited near Boñar, prov. León), which he dates as Upper Eocene. "L'aspect rappelle étroitement celui des grès et des sables à facies Wealdien" (Ciry 1939, p. 301). The conglomerates consist of small quartz pebbles with a sandy matrix and are easily distinguishable from the limestone boulder-conglomerates. They are clearly exposed, a 500 m NE of the station of Villaverde-Tarilonte along the small railroad, which runs from the coal mine near Velilla de Tarilonte to the station of Villaverde-Tarilonte [1]).

Ciry (1939) noted an unconformity between the "grès de las Bodas" and the Cretaceous. Mabesoone (1959), on the other hand, describes the contact as conformable. As we have already mentioned, a small angular unconformity has been found between the two sediment complexes. This is visible in the outcrop 500 m NE of the station Villaverde-Tarilonte, where the "grès de las Bodas" in the valley overlies a different Campanian limestone horizon than on the top of the hill. However on many places the unconformity is scarcely visible, because both complexes were deformed contemporaneously and turned over in a steep position.

The conglomerates of the Cuevas facies have an overturned or steep position in a 1 km broad zone. Further south they suddenly revert to the subhorizontal. This rapid decrease of dip has been interpreted by Oriol (1876) as due to an age difference between the steeply dipping and the subhorizontal strata. Ciry (1939, p. 300) agrees with this view, at least for the zone between Guardo and Cervera: "la partie inférieure, fortement plissée, des masses conglomératiques pouvait revenir à l'Eocene supérieur, tandis que les horizons supérieurs, discordants et subhorizontaux, seraient Oligocènes". Cantos Figuerola (1953) described the contact between the vertical and the horizontal conglomerates as due to a fault contact. This seems very probable, because the sudden decrease of dip occurs along a straight line, indicating a fault or a sharp flexure.

INTRUSIVES

Igneous rocks are present in the area as discordant dykes or as concordant sheets. They appear to be mainly restricted to the zone of San Martin-Camporredondo, an instable zone which played an important part in the sedimentary history of the Devonian and Carboniferous, as we will describe in chapter III.

These igneous rocks are mostly completely altered, especially those concordant sheets which show a cleavage parallel with the first cleavage in the surrounding rocks. Sometimes they are altered to such an extent, that the igneous origin of some samples could not be proved without doubt even by microscopic investigations. Both basic and acid dykes are present.

1. Quartz porphyry: Sample E 18 found 1500 m SSE of Cardaño de Abajo between limestones represents a concordant sheet of an acid volcanic rock less altered than most of these sheets. Phenocrysts of plagioclase, quartz and biotite are present. The plagioclase is almost wholly altered into sericite and carbonate. Occasionally fresh cores with An contents of 40 % are present. The plagioclase shows albite and Carlsbad twins. In wholly sericitized crystals the euhedral outlines may still be clearly visible. The quartz phenocrysts have the characteristic rounded and corroded shapes. The biotite is often euhedral, wholly chloritized and

1) Mabesoone erroneously states that this location is situated south of the station of Villaverde—Tarilonte.

may show corroded shapes. Inclusions of sagenitic rutile and some carbonate are present. The phenocrysts generally have diameters of a few mm. The groundmass is turbid. It consists chiefly of chlorite, sericite and felspar. Apatite occurs as accessory mineral.

2. North-northwest of San Martin de los Herreros, in the Arroyo de Agueras a dyke has been found, a few metres wide, which cuts across the folded and refolded limestones of the Ruesga group. It represents a rock of a spilitic composition (sample Q 108). Chlorite forms pseudomorphs after pyroxene (?) phenocrysts. The groundmass consists chiefly of albite, chlorite and carbonates. Potash felspar, probably adularia, is present in subordinate quantities. A basaltic texture is still visible; the turbid albite crystals are lath shaped and have a random orientation.
3. Porphyritic hornblende-biotite-quartz diorite. This rock has been found as a small discordant intrusive mass, near the road between Besande and Valverde, 2 km south of last mentioned village (sample Q 206). The rock has a panidiomorphic texture with euhedral hornblende, biotite and plagioclase. Unlike the biotite and plagioclase, the hornblende is still partly fresh. The biotite is chloritized and partly limonitized. The plagioclase is wholly albitized. Quartz occurs in the interstitial spaces between the euhedral crystals. Sericitic, chloritic and carbonaceous matter are present in large amounts. Apatite needles are numerous. The average grainsize is about 1 mm.

CHAPTER III

FACIES AND THE INFLUENCE OF TECTONIC EVENTS ON THE SEDIMENTARY HISTORY

Introduction

The development of the Devonian in the province of Palencia shows a great lateral facies change over a short distance in a northern direction, whereas the litho-stratigraphic development of the Devonian to the west, in the province of León, as well as in Asturias, is quite constant with exception of the Upper Devonian (Comte 1959).

We can distinguish two facies zones; a southern facies, developed in the Valsurvio dome, the Esla-Bernesga region (León) and in the region between San Martin and Ventanilla (Kanis 1955); a northern facies known from the Rio Arruz (Kullmann 1960) and the Devonian around Cardaño de Arriba (v. Veen, personal communication). These two facies areas are separated by a narrow instable zone. Vertical tectonic movements in this zone have strongly influenced the sedimentary history in the present area.

FACIES DEVELOPMENT DURING THE DEVONIAN

Sedimentary environments in the Valsurvio dome

Stratigraphic column A (La Taramada) of fig. 10 represents the most completely developed Devonian, found in this area.

The Compuerto formation. — In this formation the sediments were deposited under epi-neritic conditions. The black slates, often pyritic and without fossils, indicate sedimentation in quiet waters and in a reducing environment.

The upper surface of the Compuerto formation west of the Rio Carrion is interesting. Between Otero de Guardo and Camporredondo de Alba, strongly weathered sediments and a red-coloured fossil soil indicate a break in the sedimentation. A strong enrichment of alumina and of oxides of iron in the pelitic sediments, with crystallization during regional metamorphism, has led to a chloritoid sericite schist, a 10—20 m thick, with a very fine-grained homogeneous texture (cf. p. 137).

East of the Rio Carrion we have not found indications of soil development, so that we can assume that, although a regression took place at the end of the deposition of the Compuerto formation, the sediments only locally emerged above sea level. West of the Rio Carrion strong weathering took place, probably under warm, humid conditions, forming laterites, and decalcified sediments.

The Otero formation. — This formation has been deposited under marine conditions. A local reef-like character, as for instance found in the Arroyo de Abianos, indicates a neritic or epi-neritic environment. The biostromal limestones, are often built up of a great amount of corals and stromatoporoids. Typically brachiopods are rarely found in this formation. In thin sections the original texture of the limestone often has been destroyed by a strong recrystallization.

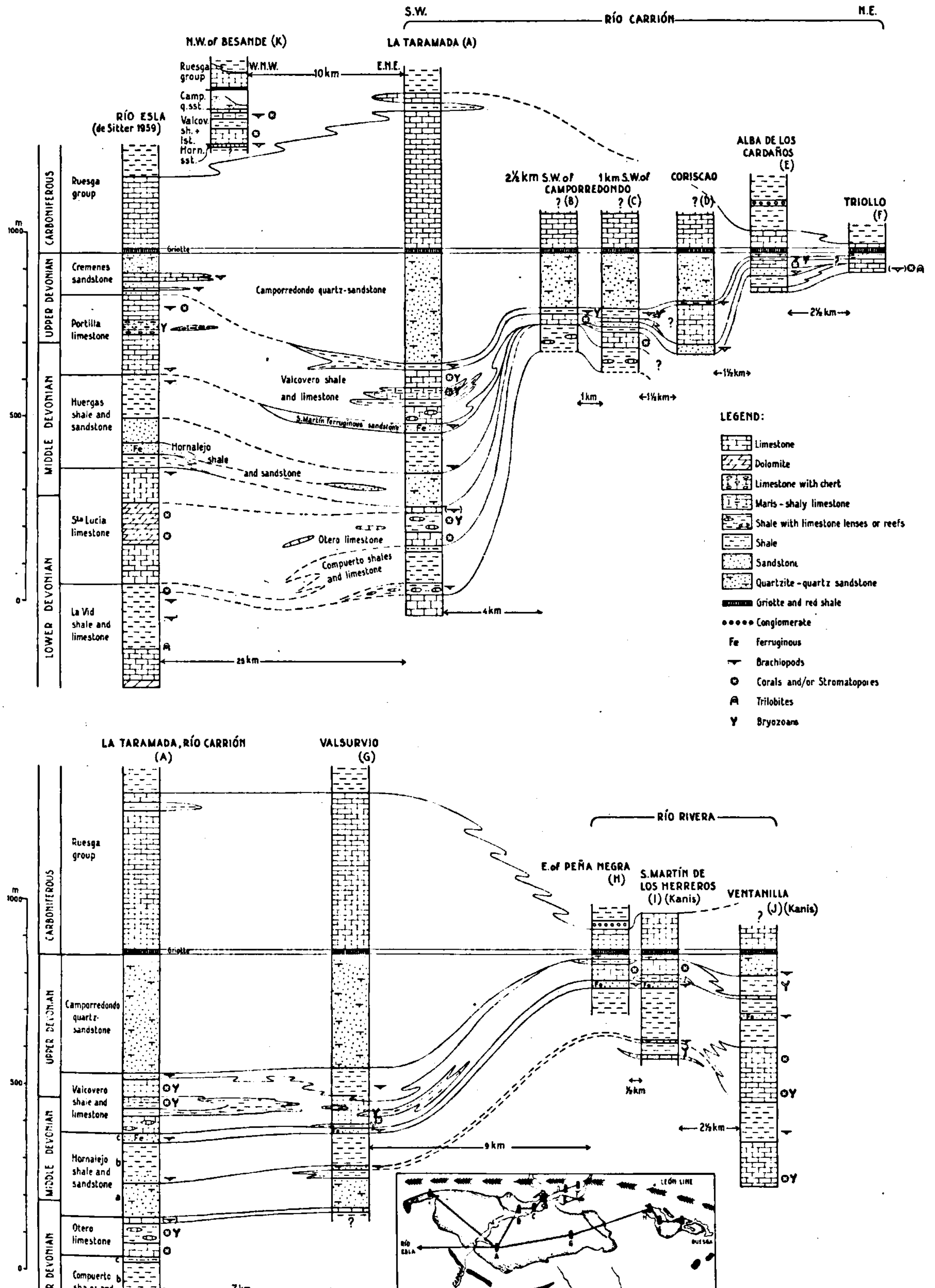


Fig. 10 Litho-stratigraphic columns of the Devonian and Carboniferous Ruesga group in the Esla region, the Valsurvio dome, the zone of San Martin-Camporredondo and near Ventanilla.

Following the limestone outcrops from the Arroyo de Abianos towards Camporredondo, their thickness diminishes locally to zero, which has been caused by a later erosion, to which we will return on p. 174. Near Camporredondo on the east side of the Rio Carrion, the Otero limestone is again fully developed.

The Hornalejo formation. — This formation is subdivided in three parts; two coarse clastic parts (*a* and *c*), separated by a 80 metres of pelitic material (*b*).

The slates have a high content of pyrite, caused by sedimentation in less agitated shallow water, which favoured the production of hydrogen sulfide and pyrite. The fossils lie badly sorted in thin layers and are very well preserved. The sapropelic bottom material afforded perfect opportunities for entombment without destruction of the fossil material. The excellent preservation of the internal moulds and the internal structures of the organisms, was due to quiet bottom circumstances, the reducing environment and the possibilities of burial by soft mud.

The upper member (*c*) of the Hornalejo formation consists of the ferruginous sandstone of San Martin, which indicates a paralic unstable environment. The oölitic hematites have been formed in agitated water; the strong development of lenticular shaped cross-lamination suggests strong shifting of currents in shallow water.

The Valcovero formation. — The environment during the deposition of this formation was again epi-neritic or littoral. The rapid lateral facies changes from coral limestone into shales (cf. p. 145) are known also in the easterly adjacent area between San Martin and Ventanilla. North of Otero de Guardo the sharp boundary between the Valcovero limestone and the underlying ferruginous sandstone indicates a small local break in the sedimentation. Near Camporredondo the Valcovero formation is reduced in thickness.

The Camporredondo formation. — This formation reaches a thickness of several hundreds of metres in the Valsurvio dome. The massive quartz sandstones indicate an enormous supply of well-sorted material. Current bedding and fore-set lamination sometimes can be observed, when small colour differences occur in thin layers. Mostly, however, the rock is very homogeneous and no layering can be discerned. Between the Valcovero and the Camporredondo formations a continuous sedimentation occurred. Sometimes a thin micro-conglomerate has been developed at the top of the Camporredondo quartz sandstone.

From the preceding we may conclude, that the Devonian in the Valsurvio dome has been deposited under neritic-littoral condition and in a rather unstable environment.

The facies of the Esla-Bernesga region

The Devonian sequence developed in the Valsurvio dome can be correlated in broad outlines, with the Devonian situated in the region of the Rios Esla-Bernesga, more to the west and described by Comte (1959).

Comte distinguished two thick limestone formations, the oldest named "Santa Lucia" limestone and a younger one "Portilla" limestone, separated by a clastic formation, the "Huergas" shales and sandstones. Below the Santa Lucia limestone there is a formation of calcareous shales with at its base a dolomite, called "La Vid" formation. On top of the Portilla limestone another thick clastic deposit has been developed, which in the Bernesga valley can be subdivided on lithological grounds in three formations; respectively "Nocedo, Fueyo and Ermitage". The Fueyo formation has a more pelitic character than the Nocedo and the Ermitage sandstones. The Fueyo shales, however, are often poorly developed, or even absent and it is difficult to distinguish the Nocedo and the Ermitage sandstones, although genetic differences occur. The Ermitage sandstone is a transgressive sandstone, resting upon various older Paleozoic formations, whereas the Nocedo sandstone is a regressive deposit.

All these Devonian formations have a great uniform lithological extension and are easily correlable with the Devonian of Asturias, described by Barrois (1882) and the Devonian of the present area, as has been pointed out already in the preceding chapter (fig. 10). Only in detail can small facies changes be made out. De Sitter (1959, 1961) has called this zone the "Leonides" on structural grounds. .Between the most easterly outcrops of the Devonian of the Esla-Bernesga area and the Devonian of the present described area, there is a distance of 10 km as the crow flies, covered by younger sediments of Upper Carboniferous age, which hide the older structures.

The most important difference between the strata of the Devonian of the Esla-Bernesga region and the Devonian of the Valsurvio dome, is that the latter has suffered a slight regional metamorphism. The folding of the Devonian of the Esla region is of a concentric type, cleavage has not been developed and a strong recrystallization, due to folding, does not occur. In the Valsurvio dome, on the contrary, intensive cleavage folding has been developed, especially in the incompetent beds. Recrystallization in the cleavage planes often has destroyed the bedding and other sedimentary characteristics. These differences, however, are of secondary origin. When we consider the primary lithological characteristics, the general similarity is striking (fig. 10, column Rio Esla and La Taramada, A).

The facies of the Devonian between San Martin and Ventanilla

The facies of the Devonian outcrop between San Martin and Ventanilla shows affinities with the Devonian of the Valsurvio dome and therefore belongs to the southern facies too. This Devonian has been studied by Kanis (1955, p. 390—398), to which we will refer for the litho-stratigraphic description.

Lateral variations in lithology occur in this zone, which is in strong contradistinction to the great lateral uniformity of the Lower and Middle Devonian in the Esla-Bernesga region.

The Camporredondo formation has been strongly reduced near San Martin. A quartz sandstone layer 1—2 m thick, overlies the Valcovero formation, with a sharp boundary. Small erosional features, on the top of the Valcovero formation, indicate a sedimentary break between the Camporredondo and the Valcovero formations.

The Valcovero formation has been developed in the same facies, as known in the Valsurvio dome. Great facies changes have been found also in the Hornalejo formation[1]) (Kanis 1955, Appendix 3, column II and III). Near San Martin the lithological sequence is identical with that of the Hornalejo formation, known from the Valsurvio dome, whereas towards Ventanilla, in easterly direction, the clastic deposits decrease in thickness and the lower part laterally

[1]) The Couvenian sequence, described by Kanis, is identical with the Hornalejo formation of the Valsurvio dome. The uppermost part of this formation ranges into the Givetian. Rocks with a Givetian age has not been mentioned by Kanis, but certainly occur, because there is a continuous sequence from the Couvenian sandstone up to the Frasnian limestone.

grades into a massive coral limestone. According to Kanis (1955, p. 392) this limestone has often a biohermal character.

An exact correlation of the Otero limestone and the Compuerto slates and limestones, with the Emsian section as described by Kanis, is not possible at the moment, with the restricted paleontological knowledge. However, general lithological resemblance certainly exists.

The facies of the Rio Arruz (northern facies)

A lithological correlation between the Devonian of the Valsurvio dome and the Devonian situated north of the Carboniferous limestone ridge of the Espiguete and the mountains near Santibañez de Resoba is completely impossible. Not only the lithology, but also the biofacies of these areas is totally different.

In the northern facies, many horizons of nodular limestones occur. When we consider a lithological sequence of the Devonian in the Arruz valley after Kullmann (1960), the differences with the sequence in the Valsurvio dome are obvious. Rio Arruz section (after Kullmann 1960).

Green nodular limestones with shale intercalations	18 m	Frasn.
		Givet?
shales with limestone layers	20 m	
grey limestones and nodular limestones	5 m	Couv.
dark shales	10—15 m	
nodular limestones and marls	55 m	Ems.
grey marls	12 m	
nodular limestones	10 m	
arenaceous slates	30 m	
dark grey limestones		Sieg.

The thickness of the strata in this section is strongly reduced, compared with that of other localities in the northern facies (unpublished report, van Hoeflaken (1955). On top of the nodular limestone of Frasnian age a thick quartzite or quartz sandstone occurs, which is overlain by another horizon of nodular limestone. Fossil determinations of trilobites and conodonts [1]) have proved that this nodular limestone was of Upper Famennian age.

In the area around Cardaño de Arriba north of the Espiguete mountain the litho-stratigraphic sequence is also totally different with that of the Valsurvio dome, as will be published in due course by J. van Veen, of the Leiden University. In the Valsurvio dome brachiopods, corals, stromatoporoids and bryozoans are the most prominent faunal elements in the Upper, Middle and top of the Lower Devonian. In the northern facies-zone on the contrary (around Cardaño de Arriba and in the Arruz valley) the fauna is characterized by goniatites, trilobites, conodonts and lamellibranchs.

INFLUENCES OF TECTONIC MOVEMENTS ON THE SEDIMENTARY DEVELOPMENT

The E—W trending ridge during the Devonian

From the preceding description of the facies distribution during the Devonian in our and adjacent areas, we may conclude that two important facies zones occur.

[1]) Preliminary determinations of the trilobites by Mr. J. A. van Hoeflaken and determinations of the conodonts by Mr. A. H. van Adrichem Boogaert indicate an Upper Famennian age.

A northern facies is represented among others by the Devonian outcrops in the Rio Arruz (Kullmann 1960) and by the Devonian near Cardaño de Arriba (Mr. J. v. Veen, personal communication).

The southern facies is characterized by the Devonian sequences in the Esla-Bernesga region, the Valsurvio dome and the San Martin-Ventanilla region. There the Devonian formations have a great lateral extension and uniformity in lithological composition, in particular in the Esla-Bernesga region. They must have been formed under neritic-epineritic environments, on a stable shelf. To the east, especially between San Martin and Ventanilla, rapid lateral facies changes occur, indicating a rather unstable, partly paralic sedimentation environment.

The northern and the southern facies are not only different in their lithological characteristics, but differ also on paleoecological grounds. To understand these great differences, the northern and southern facies must be considered as deposited in two quite distinct areas, with different sedimentary histories. There are several reasons for assuming the existence of an E—W trending ridge, running from north of San Martin in the east of the present area to Cardaño de Abajo in the west, separating the two facies zones during most of the Devonian time:

1. The existence of two Devonian facies zones, which are totally different on lithological and paleoecological grounds.
2. These two zones are separated from each other by a narrow belt. Facies changes between the northern and southern facies are restricted to this belt throughout the whole Devonian.
3. The thickness of the Devonian strata in the Valsurvio dome (southern facies) decreases suddenly to the north, over a short distance, near the Pantano de Camporredondo as well as near Besande and San Martin. (These localities are situated on the E—W trending ridge, fig. 11).
4. Several sedimentary breaks and hiatuses occur in the Devonian outcrops on this ridge, near the Pantano de Camporredondo and in the thrust fault zone of San Martin; whereas in the Valsurvio dome there is a continuous sedimentary succession from the Otero limestone untill the Camporredondo sandstone.
5. In the zone around the Pantano de Camporredondo the sedimentary history has been influenced by the southern as well as by the northern facies.
6. In the same zone, a mixing of the faunal elements of both facies zones has been found.
7. Further to the north, (north of the Espiguete and of the village of Triollo, northern facies) the thickness of the Devonian sequence increases again and appears to be a complete stratigraphic section.

The tectonic history of this ridge during the Devonian time, will be followed on the basis of the description of the sedimentary breaks, found on this ridge.

The Middle — Lower Devonian hiatus. — In the northern part of the Valsurvio dome the upheavel of the ridge is to be seen as far as 2½ km southwest of Camporredondo. The Valcovero formation, directly overlying the slates of the Compuerto formation, is strongly reduced, so that nearly the whole Middle Devonian and the upper part of the Lower Devonian is missing (compare column A and B, fig. 10). The increase of this hiatus is very abrupt and takes the place of 500 m sediments in the southern part of the Valsurvio dome. Near the village of Camporredondo, 1½ km to the northeast, as well as near Alba de los Cardaños, the Valcovero formation overlies the lower part of the Hornalejo formation *a* (column C and E, fig. 10). In these two last mentioned localities the contact between the Valcovero formation and the Hornalejo formation indicates only a stratigraphic break, whereas in the locality 2½ km southwest of Camporredondo erosional features have also been found. Consequently it is considered that, after the sedimentation of the Hornalejo formation *a*, an upheaval of the ridge occurred, resulting in a period of non-deposition and local erosion (2½ km SW of Camporredondo). In the meantime sedimentation continued in the Valsurvio dome more to the south (fig. 12a). North of the E—W ridge, in the Devonian of the Rio Arruz a continuous sedimentation — although with a reduced thickness — has been found by Kullmann (1960). No stratigraphic gap occurs in the area described by Kanis (1955) south and southeast of San Martin de los Herreros, where continuous sedimentation went on through the Middle Devonian.

Fig. 11, and the stratigraphic columns A—F of fig. 10, give a clear picture of the decrease in thickness of the Middle Devonian from the mountain La Taramada to northerly and northeasterly directions. Near Triollo the Middle Devonian is even totally absent and 10 m below the Lower Carboniferous griotte horizon a limestone with Siegenian brachiopods has been found (cf. p. 150). Also from La Taramada to the northwest, the Middle Devonian strata decrease in thickness near Besande.

The upheaval of the ridge, after the sedimentation of the Hornalejo formation *a*, has caused the reduction of these strata by local erosion and by a period of non-deposition. The uplifted zone is situated directly north of the present Valsurvio dome, including a small area SW of Camporredondo (fig. 10, column B and C). To the east this ridge must have been situated north of the Devonian outcrops between San Martin and Ventanilla, because the sedimentation of the Middle Devonian of that zone has not been influenced.

The Upper Devonian hiatus. — The Camporredondo formation shows great variations in thickness. In the Valsurvio dome, the Camporredondo quartz sandstones reach a thickness of several hundreds of metres, whereas in the thrust fault zone of San Martin and in the Devonian outcrops around the Pantano de Camporredondo the thickness decreases to some metres only. This decrease is very rapid, as may be seen between the Arroyo de Miranda and the Arroyo de Valderrianes, a distance of 2 km. Near Besande this rapid reduction of the Camporredondo quartz sandstone occurs as well (fig. 10, column K).

In the Valsurvio dome the Valcovero formation merges upwards into the Camporredondo formation and indications of continuous sedimentation are present. North and east of the Pantano de Camporredondo, as well as in the Upper Devonian around San Martin, the 2—10 m thick Camporredondo quartz sandstone lies obviously disconformably upon the limestone of the Valcovero formation. During this break in sedimentation no strong karst erosion was developed. The boundary shows only small wash-outs, filled by the overlying quartz sandstone. Hence we find that in places where the Camporredondo quartz sandstone is strongly reduced in thickness, a hiatus has been developed at its base. In places where the full thickness is present, no sedimentary break occurs (fig. 12b).

The localities of reduced thicknesses are situated on the E—W ridge. So we arrive at the conclusion, that the period of non-deposition between the Valcovero and Camporredondo formations is connected with the same ridge as in the

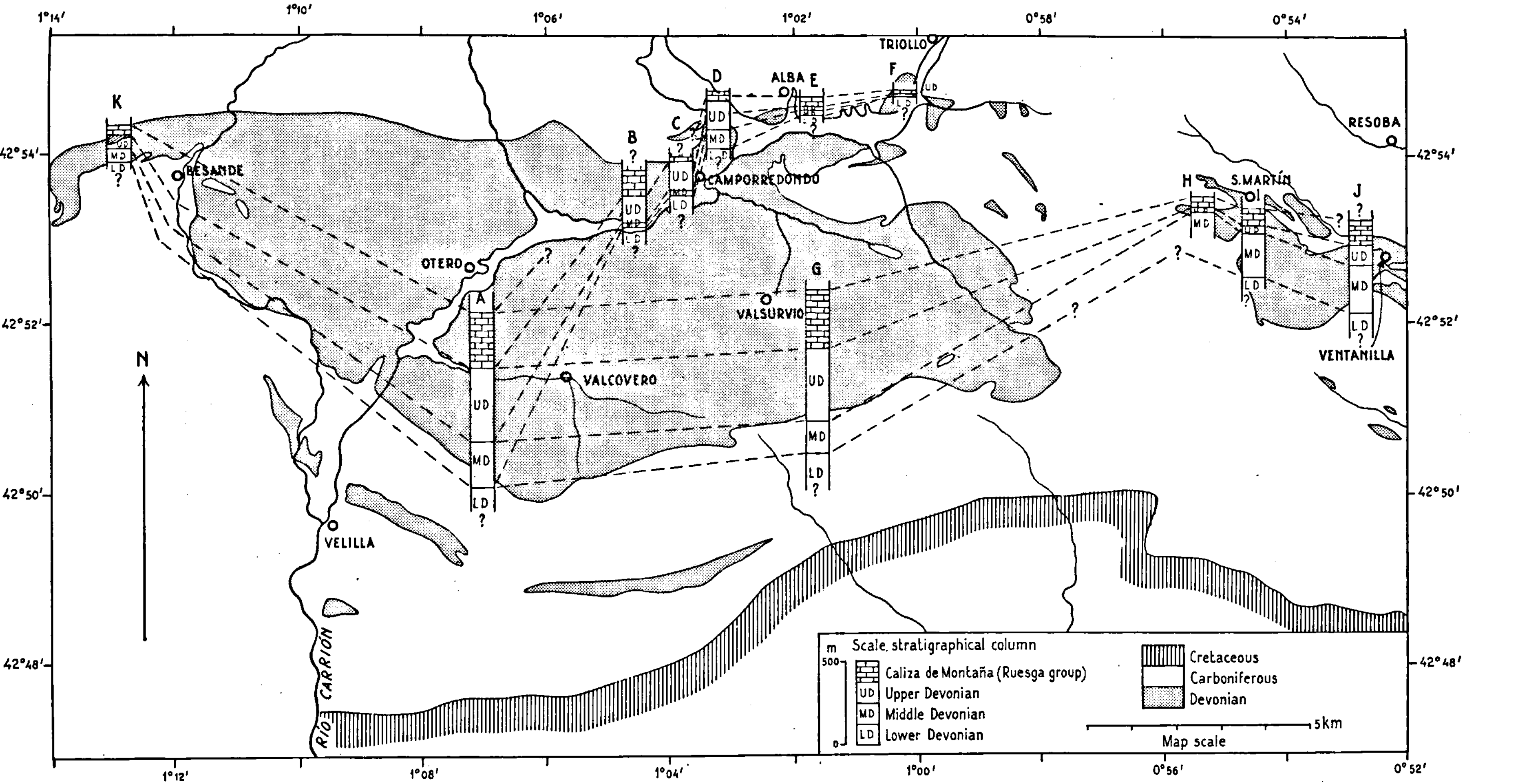


Fig. 11 Sketch map of the distribution of the thickness of the Devonian and the Caliza de Montaña. The letters refer of the detailed columns shown in Textfig. 10.

Middle Devonian, elevated again during the sedimentation of the top of the Valcovero and the lower part of the Camporredondo formations in the Valsurvio dome.

The abrupt decrease in thickness, the rapid change in facies and the sudden development of breaks in the sedimentation are indicative of tectonic movements along sharp lines. The E—W ridge has apparently been uplifted between a set of faults. In the Middle Devonian the uplifted zone was situated north of the Devonian outcrops between San Martin and Ventanilla, whereas in the Upper Devonian the ridge has been uplifted along a more southerly situated fault, including the San Martin-Ventanilla area. Near Camporredondo, on the contrary, the decrease in thickness of the Upper Devonian has been found more to the north, with respect to the change in sedimentation in the Middle Devonian. So an uplift along a more northerly situated fault during Upper Devonian must be assumed (fig. 12b). However, these are only local differences, while the general trend remains the same.

Whether the reduced Camporredondo quartz sandstone in the San Martin thrust fault area is equivalent as time-rock unit only to the uppermost layers of the Camporredondo formation in the Valsurvio dome, remains questionable. Sometimes both are developed as a micro-conglomerate or a coarse, white quartz sandstone, which is suggestive of their contemporaneity. However, one can not be sure of this correlation, because of the lack of fossils.

The hiatus on the Devonian — Carboniferous boundary. — After the deposition of the Camporredondo quartz sandstone a break in the sedimentation occurred again. This hiatus, however, is not only present in the thrust fault zone of San Martin, but also occurs in the Valsurvio dome and is known from nearly all places in the Cantabric-Asturian mountains. It includes the Tournaisian and possibly part of the Lower Visean as well as the Upper Famennian, although, in consequence of the absence of fossils, the exact extent of this hiatus remains uncertain. Delépine (1935, p. 145) writes concerning a hiatus in the Lower Carboniferous of the Montagne Noir and the Pyrenees: "Malgré une apparente concordance avec le Dévonian, sousjacent, il existe là une lacune stratigraphique importante, correspondant à toute la durée du Tournaisien". However, Ziegler (1959) and Durand Delga (1958), in respectively the Pyrenees and the Montagne Noir, have shown that beds representing the Tournaisian do occur. So it is quite possible, that further investigations will reveal the local presence of the Tournaisian in the Cantabric-Asturian mountains too.

The hiatus is due to the epirogenetic upheavel of the whole mountain ridge, whereas the hiatuses in the Middle and Upper Devonian were of rather local extent and closely connected with the E—W trending instable belt, situated north of the Valsurvio dome.

The subsidence of the E—W trending zone during the Carboniferous

The Ruesga group. — Sedimentation started again with a Visean transgression in the Carboniferous. The red shale-radiolarite and the griotte indicate, as described on p. 151, a shallow water origin. This layer has a wide horizontal extension and the stratigraphic position of this horizon is the same over long distances (Age determinations of Barrois 1882, Delépine 1935, Comte 1959, Wagner—Gentis 1960, Kullmann 1961).

The E—W ridge, which ran during the Devonian times over Valverde de la Sierra, north of the Pantano de Camporredondo and the village San Martin de los Herreros, was not predominant in the sedimentary history of the Visean griottes, as the griotte is equally developed everywhere. However, in the Namurian this E—W zone again controlled the sedimentary history of the area. The remarkable feature of this zone is now seen, in that the E—W ridge, formed during the Devonian, developed into a trough during the Carboniferous (fig. 12c).

The Ruesga group consists of two different facies, the Culm and the Caliza de Montaña facies. The latter is developed in the zone of the Valsurvio dome, whereas the Culm facies is represented in a narrow trough, situated directly north of the Valsurvio dome and in the thrust fault zone of San Martin.

The limestone in the Valsurvio dome, usually several hundreds of metres thick, are relatively clean, massive grey limestones. In thin section however, the lithology shows variations from bioclastic to oölitic. Often the strong recrystallization and dolomitization makes it impossible to identify the original sedimentary texture of the limestones.

The rock-assemblage nowhere gives signs of any great depth of sedimentation and it is supposed to have been deposited as a shelf deposit in clear waters, on a relatively gentle subsiding bottom. Many authors assume such environments for the sedimentation of Lower Carboniferous massive limestone (i. g. George 1958, Paproth 1960 and Lees 1961).

The Culm facies is composed of shales, mudstones and subgraywackes with local conglomerates, which have been deposited, probably under neritic conditions in the thrust fault zone of San Martin. In its greatest development it replaces almost completely the limestones of the Caliza de Montaña facies. At the base of the shale-subgraywacke facies, directly overlying the griotte-red shale horizon, a limestone 10—40 m thick, mostly developed as a dark-blue, well-bedded stinking calcilutite often occurs.

The wash-outs at the base of the quartzite pebble conglomerate and the imbrication of the pebbles are indicative of strong currents. The most striking feature of the pebbly mudstones or boulder beds is the association of well-rounded boulders and pebbles, with the very fine muddy matrix. The first suggests vigorous water movement and strong transport, the latter quiet conditions of sedimentation. In the present paper, sedimentation by sliding and slumping is postulated, mixing the different sediment types. Some outcrops of the polymict conglomerates also indicate such sedimentary environments.

The limestones of the Sierra del Brezo and the bioherms east of Santibañez de Resoba were deposited in a quiet and clear, shallow-water environment, whereas the clastic sediments in the thrust fault zone of San Martin are indicative of totally different sedimentary environments. The lateral passage from northern reef into southern Culm facies is rapid, implying that terrigenous material accumulated in a trough at a depth possibly greater than that of the limestones. The limestone fragments in the polymict conglomerates originated from both sides of the trough, sliding directly down the submarine slopes, are mixed up with reworked pelitic and clastic material. Density currents on small scale are possible in such environments, but no convincing indications have been found.

The lateral rapid passages from Caliza de Montaña into the Culm facies occur along more or less the same lines as the sudden thinning of the Upper Devonian quartz sandstones. This thinning was due to upheaval along faults of the E—W ridge, during the Upper Devonian. In the Namurian the faults have been reactivated, but with a contrary movement, resulting in a subsidence of that same zone (fig. 12c).

We will consider the lithological development of the Ruesga group in a greater area, as far as possible, with the present restricted knowledge of the deposits and their correlation in age.

The massive limestone development of the Sierra del Brezo, from Pico Almonga near Cervera de Pisuerga in the east, as far as Peñas Blancas in the west, is also found more to the west, in the province of León (de Sitter 1959). East of Cervera de Pisuerga the Ruesga group is developed in the Culm facies, which can be followed to the west in a narrow zone, running over Ruesga and San Martin to the Pantano de Camporredondo. North of the thrust fault zone of San Martin, we find another thick limestone development, which makes up the E—W trending mountain ridge from the mountain Santa Lucia, NW of Santibañez de Resoba, to the Espiguete, N of Cardaño de Abajo. In the massive limestone, bedding can often be distinguished from a distance. To the east these limestones develop a reeflike character, as has been described by Kanis (1955). The bioherms decrease in size from west to east. There is no perceptible age difference between these biohermal limestone and the massive limestones of the Sierra del Brezo, as proposed by Kanis. Determinations of goniatites from the limestone near Santibañez de Resoba indicate an Upper Visean age (Kullmann 1961).

A difficulty of paleogeographic description is the lack of substantiated correlations between the rocks of the two facies. Fossils are completely absent from the Culm facies, so we are never sure of their age interpretations.

The Yuso group. —In the sediments of the Yuso group, lithological differences also occur, although less pronounced and less limited to the E—W instable zone.

In the area, adjacent to the north along the Rio Carrion north of Triollo, thick conglomerates were deposited with small shale and coal intercalations containing plant remains, which indicate a terrestrial origin. These sediments are closely related to the Curavacas folding phase. The characteristics of the conglomerates point to a fluviatile origin (Pettijohn 1949). Southwards, in the zone of the Pantano de Camporredondo and Valverde de la Sierra, these conglomerates suddenly diminish in thickness and grade laterally into a thick sequence of shales, mudstones, pebbly mudstones and thin conglomerate beds, directly overlying older rocks. In these strata, only a few crinoid fragments and plant debris are found. Load casts and current ripple marks are rarely developed. Great variations in the composition of the conglomerate beds occur (see p. 161). The large angular limestone blocks of 50—150 cm in size, which are sometimes found in the polymict conglomerates, indicate almost no transport and a steep relief. They must have been formed near a limestone escarpment. The great blocks must have fallen down in a muddy environment, in which often well-rounded quartzite pebbles lay scattered. These conditions can occur on a sea-coast, near steep cliffs. The occurrence of crinoid fragments gives an indication of a marine environment.

The small shale fragments or clay pellets, which we often find in the polymict conglomerates, are due to reworking of older sediments. From many places they are known to be present in beach deposits (Allen & Nichols 1945). By sliding from a steep coast, they could be mixed up with well-rounded quartzite pebbles, angular limestone fragments and other sediments, each typical of different sedimentation environments. The reworking processes may thus generate complex sediments.

On the mountain southwest of Cardaño de Abajo the quartzite conglomerates are several hundreds of metres thick. To the east near the Pantano de Camporredondo, they grade laterally into polymict conglomerates, shales and mudstones; along the water level of the reservoir even limestone blocks, which are 150 cm large, occur. The quartzite conglomerates are also limited in extent to the west.

The sedimentological knowledge of the Yuso group is, at the moment, too limited to justify a detailed paleogeographic description.

Conclusions

A provisional analysis of the different movements in this E—W trending zone, which runs over San Martin, north of the Pantano de Camporredondo to Valverde de la Sierra, results in the following succession:

1. Existence of an E—W trending ridge during the Lower Devonian, separating a northerly facies area (Rio Arruz, Kullmann 1960) from a southerly facies area (e.g. Valsurvio dome).
2. Upheaval of this ridge above sea level in the Middle Devonian, resulting in a local erosion of the Hornalejo formation *a* and the Otero formation (for example near Camporredondo de Alba), and a period of non-deposition on the ridge during continuous sedimentation in the Valsurvio dome of the Hornalejo formation *b* and *c* and possible of the base of the Valcovero formation (fig. 12a).
3. Upheaval of the same ridge, as well as the area between San Martin and Ventanilla, in the Upper Devonian, resulting in a period of non-deposition in these places during sedimentation of the lower part of the Camporredondo formation in the Valsurvio dome (fig. 12b).
4. A subsidence of this zone with respect to the neighbouring areas in the Namurian, resulted in the sedimentation of a thick series of clastic material in a narrow trough, bordered on both sides by areas in which limestones of the Caliza de Montaña facies were deposited (fig. 12c).
5. Bottom movements of the same zone controlled to some extent the lateral facies changes in the Yuso group.

The periods of non-deposition have only been developed in this small zone. In the south, in the valley of the Sta. Eugenia, the Devonian is completely developed, whereas on the south and east border of the Pantano de Camporredondo a hiatus developed in the Upper and Middle Devonian. Two kilometres to the north, near the village of Vidreros, the Devonian is developed again in a continuous sequence, although, in another facies. The reduction in thickness of the Devonian strata from south towards the north is clearly shown in fig. 11. Near Triollo 5—10 m of Upper Devonian strata directly overlie the Lower Devonian. The rapid lateral passage from the limestones of the Caliza de Montaña facies into the mudstones and subgraywackes of the Culm facies is connected to the same zone.

A correlation with the Esla-Bernesga region (León).

More to the west, in the Devonian between the rivers Esla and Bernesga, a large hiatus has been observed (Comte 1959, de Sitter 1961), which increases rapidly in extent from south to north. Following upon this hiatus, sedimentation started again everywhere with a quartz sandstone of Upper Famennian age

(Grès de l'Ermitage, Comte 1938). The origin of this to north increasing hiatus is, according to de Sitter (1961), the tilting of a block of the Leonides. The hinge of this tilted block is an E—W trending zone, coinciding with the front

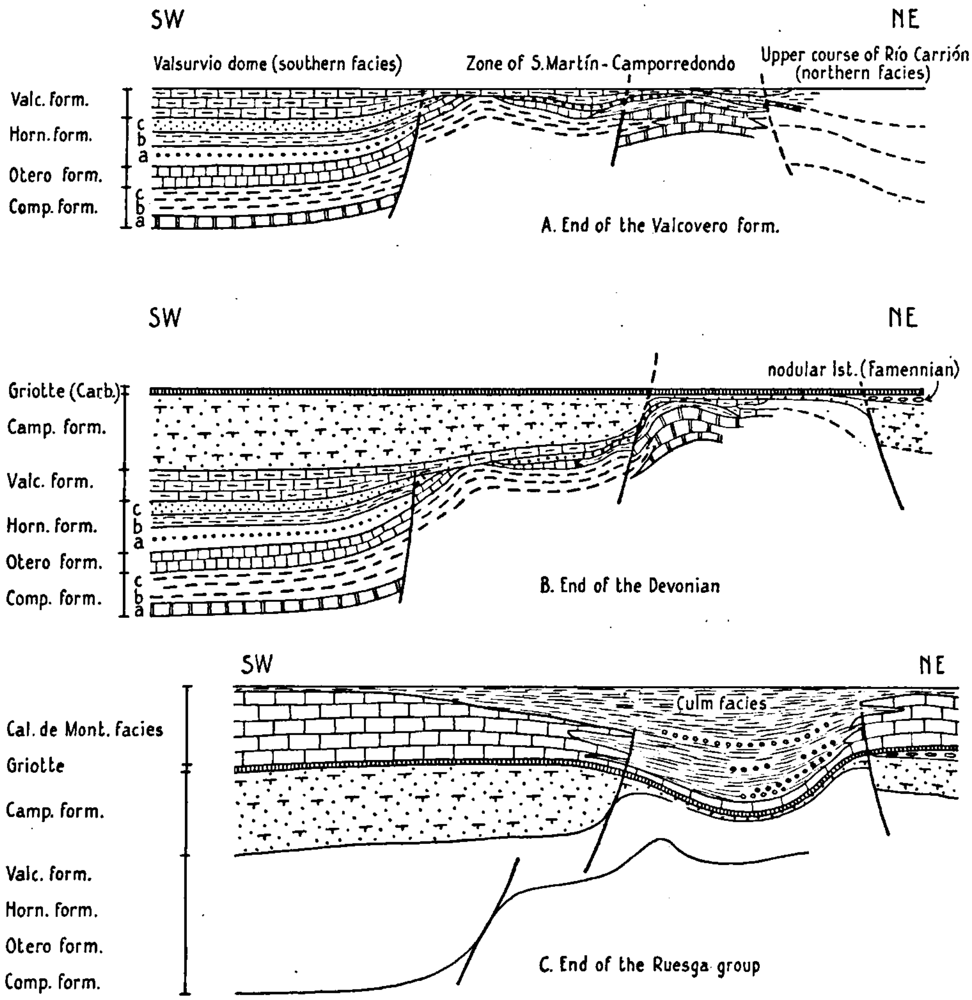


Fig. 12 Schematical lithological cross-sections (vertical scale 1 : 37000) through the Valsurvio dome (southern facies) zone of San Martin-Camporredondo and the upper course of the Rio Carrion north of Triollo (northern facies)
a. at the end of the Valcovero formation
b. at the end of the Devonian
c. at the end of the Ruesga group.

zone of the nappe structures of the Leonides. North of this zone, the hiatus has not been observed in the Devonian. The facies development of the Upper Devonian in the north, is different from the facies of the Upper Devonian south of this zone. It is a similar difference, as described from the present area. Small limnic Stephanian coal basins appear restricted to this E—W zone (de Sitter

1961). The narrow zone can be followed from Busdonga in the Rio Bernesga via Tolibia de Arriba in the Rio Curueño, to Salomon and Las Salas in the valley of the Rio Esla. In the extension to the east the instable ridge described in the preceding paragraphs is situated.

The width of this zone, influenced by these epirogenetic events, decreases from west to east. In the Devonian between the Rios Esla and Bernesga a tilting of a whole block occurred, with the E—W zone as hinge of movement, whereas north of the Valsurvio dome only a local upheaval of a rather small elongated ridge occurred during the Devonian and a subsidence of the same zone in the Carboniferous. In the present area a hiatus developed in the Middle Devonian, as well as in the Upper Devonian, both of which comprised only rather small intervals of time. In the province of León, on the contrary, only one hiatus has been found in the Upper Devonian. This hiatus increases to the north, so that even the Upper Devonian can directly overlie Cambriam strata. Whether the Upper Devonian transgressive strata are similar in age in both areas, could not be proved, because of insufficient fossil determinations in our region.

CHAPTER IV

STRUCTURE

Introduction

Three structural units can be distinguished in the present area:

1. The Valsurvio dome
2. Zone of San Martin-Camporredondo
3. Southern border zone.

In order to obtain a clear insight into the structures, they have been subdivided into eight subareas, each forming to some extent a homogeneous unit of 25—35 square kilometres in extent (indicated on the situation map of the geological map).

Lithological and structural boundaries often coincide because the folding pattern and the intensity of deformation changes with the change of rock lithological characters in many places. Often cleavage folding grades via accordion folding into concentric folding from one subarea to the other. So, correlations between the subareas can only be established with difficulty. Moreover a superimposition of different fold types occurs, which makes the picture of fold geometry in the subareas more complicated.

The analysis of the mutual relations of the rock deformations and especially of the cleavages indicates that these deformations are the result of at least four folding movements.

The structures have been divided into three groups, on the basis of their size:

1. Major folds, are mappable as individual structures, they affect the general distribution of stratigraphy in the area. These folds range from 10 m to several km in wave length.
2. Minor folds, on a scale of one single exposure, which do not affect the local distribution of stratigraphy. They may vary from 10 m to 1 cm in wave length.
3. Micro folds, which are best studied in thin sections and which are macroscopically visible as corrugations or lineations on the foliation planes. Their wave length is less than 1 cm.

Minor folds, cleavage orientations and lineations are plotted on stereographic nets, on places where a large number of measurements were available.

The lower hemisphere of the equal-area or Schmidt net has been used and contours have been prepared with the aid of a one per cent counting circle in the way described by Fairbairn (1949, p. 286). Diagrams have been left uncontoured when the number of points was too small.

The microscopic structural study is based on thin sections of orientated samples.

In this paper the definitions of Leith (1905), concerning slaty and fracture cleavages, are used.

"Slaty cleavage or flow cleavage is dependent on the parallel arrangement of the mineral constituents of the rock, which is due to recrystallization and forming of new minerals" (Leith 1905, p. 23).

"Fracture cleavage may be defined as a cleavage dependent for its existence on the development of incipient parallel fractures or actual fractures, which by subsequent welding or cementation remains planes of weakness" (Leith 1905, p. 119). The development of fracture cleavage is a mechanical process and is independent of a parallel arrangement of the mineral constituents.

Crenulation cleavage: Knill (1960) and before him many other investigators have made a subdivision into three distinct cleavage types. They distinguish fracture cleavage from strain-slip cleavage. Fracture cleavage is the result of brittle rupture, whereas strain-slip cleavage shows a parallel mineral orientation, as result of crenulations of pre-existing foliation planes. Knill (1960, p. 323) says: "Occasionally, these strain-slip cleavages show transitions into fracture cleavages.... It is apparent that both types of cleavages were developed together and this appears to be further indication of the inter-relationship of brittle and plastic conditions during folding." He suggests a descriptive term as "crenulation cleavage", which is introduced by Rickard (1961).

In the present area the transition between fracture and strain-slip cleavage is very common, due to differences of competence in the mixed sequence of rocks. Often we find both types developed in one and the same minor fold.

Knickzones (also called joint drags in English literature, Flinn 1952) are formed by a pair of cleavage planes. The finely foliated material between these planes is knicked or sharply flexured. The sharpness of the knicking is dependent of the competence of the rocks in which they are formed. Lime-silicate rocks do not show jointing as fissile slates do, but only bending occurs. Knickzones on microscopic scale often pass into crenulation cleavages and then are closely related to folding (fig. 14).

Slaty cleavage has been developed in the low-grade metamorphic slates of the Devonian Hornalejo and Compuerto formations and to a lesser extent in the shales and siltstones of the Carboniferous Ruesga and Yuso groups.

At many places in the studied area the slaty cleavage is oriented subparallel with the bedding, making an angle of only a few degrees with the bedding planes. The small angle is clearly shown in fig. 13, the white coloured sandy band representing the bedding. The slaty cleavage traces from upper left of a layer to the right below the same layer. The angle between cleavage and bedding is constant. Bedding as well as cleavage are folded again in an apparently concentric way, however, a faint axial plane fracture cleavage starts to develop.

The most important mineralogical change in slaty cleavage is the appearance of sericite and chlorite, due to recrystallization of the groundmass.

Pressure solution and internal deformation of the quartz grains have resulted in an elongated shape of the grains parallel to the cleavage plane.

Subdivision in subareas

1. Valsurvio dome

A glance at the geological map is sufficient to show the central position of the Valsurvio dome as a structural unit. The Devonian formations in the core have been strongly deformed several times. The shales and slates in particular, show an intensive cleavage development which often obscures the bedding.

A N—S structure, which runs through the village of Valsurvio, separates

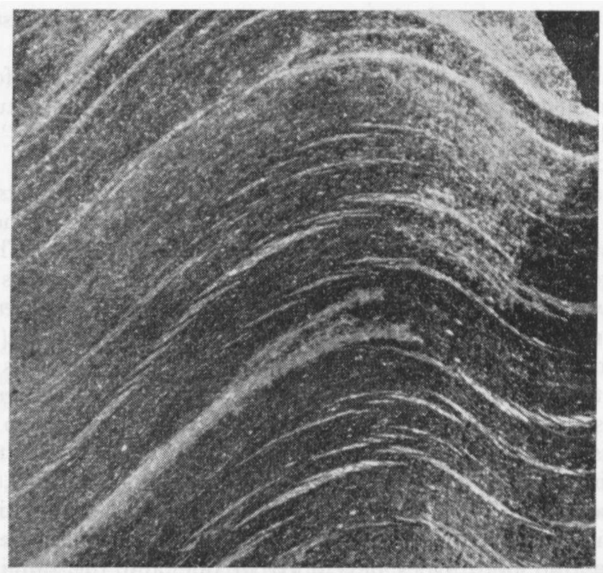


Fig. 13 Minor fold in the Hornalejo formation, refolding the slaty cleavage. A faint axial plane fracture cleavage has developed (× 24 Loc. E 100).

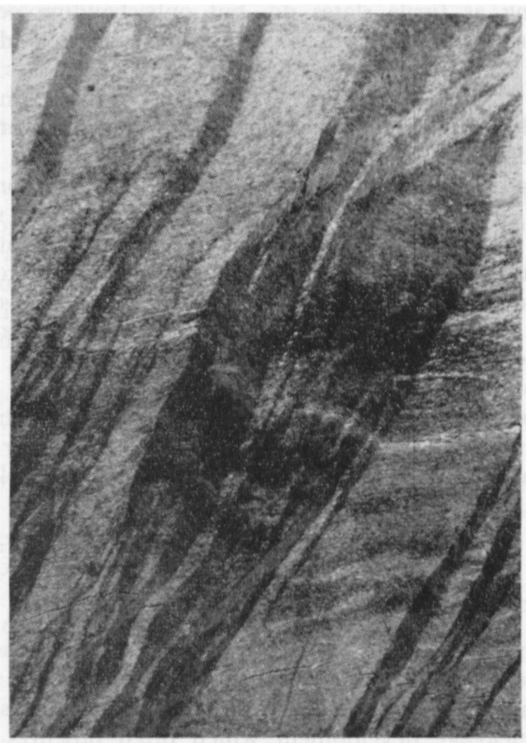


Fig. 14 Crenulation cleavage developed as an irregular bundle of microscopic knick zones, Yuso group (× 6 Loc. E. 108).

the E—W trending Otero and Hornalejo anticlines, which will be described as subareas 3 and 4 respectively.

Subarea 1 is formed by the southern limb of the Valsurvio dome, and subarea 5 by the northern limb. Both are restricted to the rocks of the Camporredondo formation and the Ruesga group. Strong recrystallization and dolomitization of the limestones of the Caliza de Montaña facies have obscured the bedding and sometimes only major folds are visible. Consequently the structural investigations are not as detailed as in subareas 3 and 4. In the limestone mountain north of Villafria (subarea 1) a crossfolding structure is visible. Notwithstanding the uniform northerly dip of the limestone and griotte layers in subarea 5 (in sharp contrast with the southerly dip in subarea 1), this area is very complicated folded and refolded.

The westerly spur of the ridge of limestone mountains (Sierra del Brezo) — amongst them the Peña Redonda, Peña Cueto, Peña del Fraile, Peña Mayor, Peña Lampa and Peñas Blancas — swings round near Besande and forms the westerly nose of the Valsurvio dome, which is treated as subarea 2. This subarea is especially interesting, because the simple structures of the Cea group here cross the older structures of the Devonian and the Carboniferous Ruesga group.

2. Zone of San Martin - Camporredondo

The northern zone of the investigated area is considered as a second structural unit. This narrow zone played a prominent role in the sedimentary history of the Devonian and Carboniferous, as we described in chapter III. The differences in lithology between these two structural units have dictated considerable variations in the types of deformation. The structures around San Martin de los Herreros are very interesting. A number of low-angle overthrusts are refolded in a complicated way. The thrust fault area of San Martin (subarea 6) trends to the west. North of the reservoir of Camporredondo the Ruesga group in which the low-angle thrusts are developed is covered unconformably by sediments of the Yuso group (subarea 7).

The Yuso group can be followed westward through Valverde de la Sierra, the Monte Viejo, forming the Tejerina syncline in the area adjacent to the west. Small patches of limestone of the Caliza de Montaña facies, sometimes with griotte, crop out and form inliers in the overlying unconformable Yuso group.

3. Southern border zone

As third principal structural unit we can consider the southerly zone of the investigated area, which coincides with the southerly border zone of the Cantabric-Asturian mountain chain. It is formed by a small zone (one of two kilometres wide) of the Upper Carboniferous Cea group, covered unconformably to the south by Cretaceous and Tertiary rocks. As the folding pattern of the Cea group and the Cretaceous is for a great part identical, they are joined in one subarea 8.

Fig. 15 can be used as geological keymap, to facilitate map-reading.

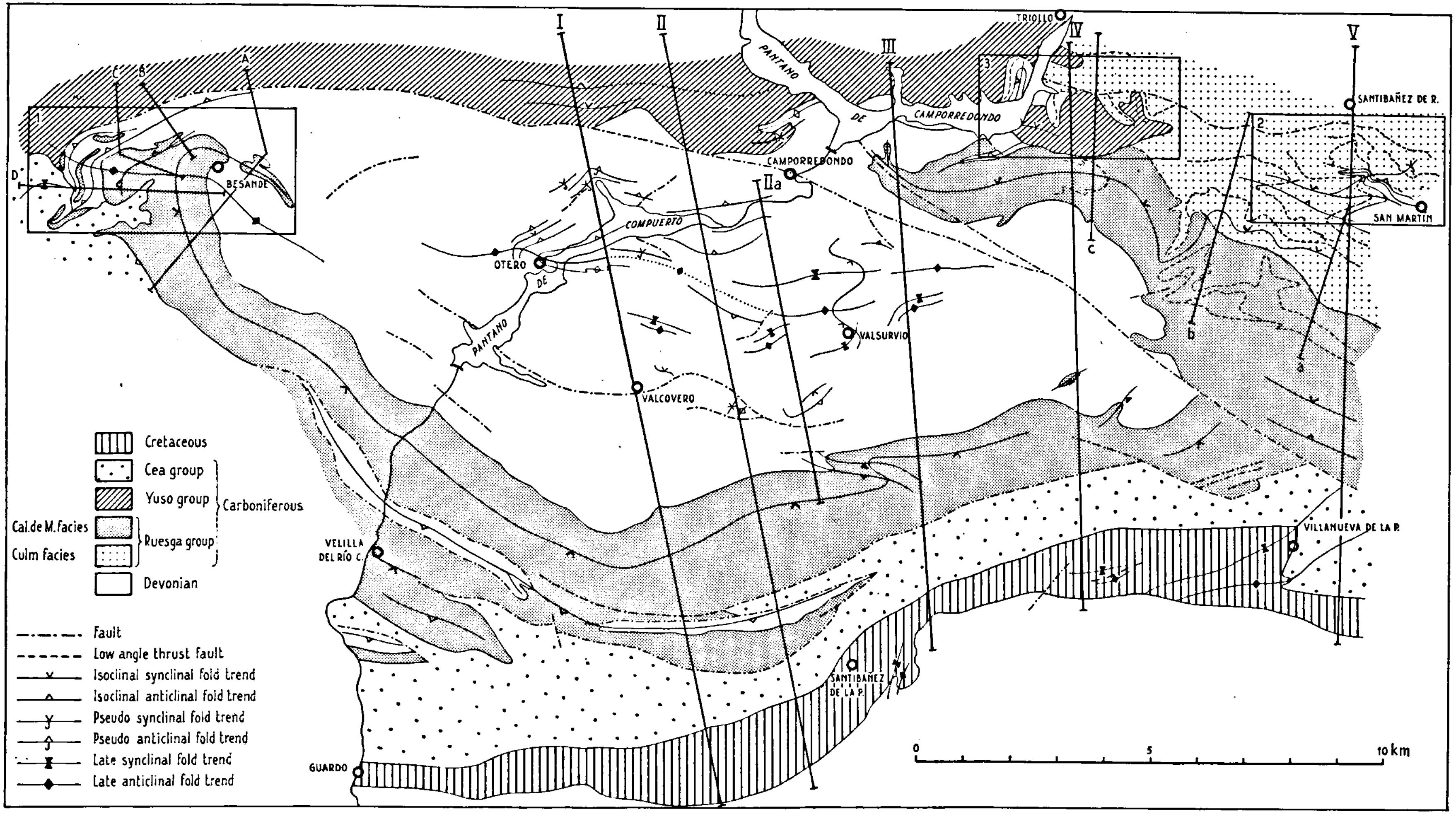


Fig. 15 Geological sketch map showing the locations of structural cross-sections and detailed maps.

VALSURVIO DOME

Subarea 1

The Ruesga group and Camporredondo formation of the southern limb

This subarea is limited to the south by the border fault-zone, a number of faults along which the Paleozoic mountains are uplifted with respect to the Duero basin in Tertiary times.

The most striking structural feature in the limestones of the Caliza de Montaña facies in this subarea is the consistent southerly dip of the bedding and axial planes of the major isoclinal folds. This is in sharp contrast with the dip of the limestones north of the Cotolorno wrench-fault, which belong to subarea 5, as well as with the dip of the limestones in the eastern zone of the Sierra del Brezo, situated north and east of the Cotolorno fault and described by Kanis (1955).

He gives the following features of the Caliza de Montaña series from the eastern zone of the Sierra del Brezo:

1. Large almost isoclinal folds.
2. Overturning to the SW in part and for the other part to the SE.
3. Structural complications where the trend curves from SW to NW.

In the western zone, discussed here, we can distinguish similar features:

1. Large almost isoclinal folds.
2. Overturning to the NW and to the NE.
3. Folds overturned to the SW occur only north and east of the Cotolorno wrench-fault and there the strike of the axial planes is more or less parallel to the fault direction.
4. SW of this fault, folds overturned to the NW occur, with an axial trend perpendicular to the fault.
5. The Cotolorno wrench-fault gives rise to a set of secondary faults and other complications near Cotolorno mountain.

The curve in the fold trends from NE—SW till NW—SE in the Caliza de Montaña south of Peña Celada, in the eastern zone of the Sierra del Brezo (Kanis 1955), is reflected in the border zone south of it, where the rocks of the Cea formation and the Cretaceous crop out. So we can conclude that this change in direction is due to a Tertiary deformation.

This is confirmed by the way in which the change in average strike of the Caliza de Montaña takes place at the Cotolorno wrench-fault southwest of the Peña Redonda. The folding of the rocks of the Cretaceous and the Cea group is connected to the development of the wrench-fault.

Further towards the west, between Peña del Fraile and Peña Mayor, a similar change in direction occurs as near Peña Celada. The relation with a curve in the Cretaceous rocks is less pronounced. This is due to an increase of the width of the intervening belt of the Cea group towards the west, caused by extra folding.

The isoclinal folds are difficult to find in the strongly recrystallized thick-bedded limestones. An intensive joint-pattern often forms open weathered-out fissures, which can easily be mistaken for bedding planes. However, from a distance structures can often be distinguished. Towards the west an anticlinal core, with griotte and Camporredondo quartz sandstones, confirms the interpretation of the structures as isoclinal folds. This Devonian core is locally concealed by slight undulations in the anticlinal fold axis. The quartz sandstones are often bounded by faults of the southern border fault-zone. We can follow the Devonian anticlinal core towards the west over the Rio Carrion up to Peña Lampa. In the valley of the Rio Carrion other folds are exposed to the south. The axial planes of these folds dip steeply (80°) to the south. In a small Devonian anticlinal core 1.5 km south of Velilla del Rio Carrion, Camporredondo quartz sandstones and Valcovero shales are exposed. In an outcrop on the western bank of the Rio Carrion of the same anticline, the Caliza de Montaña is developed as a well-bedded limestone, not so strongly recrystallized as in other places. Small minor folds are exposed, which show a very irregular folding-pattern, which is not at all characteristic for the subarea. The axial planes dip at various angles to north and south varying from a vertical position to the horizontal.

North of the village of Villafria an interesting cross-fold is visible, with a deviating axial direction. The isoclinal folds, trending WSW—ENE with axial planes dipping 60° SSE, have been refolded about axes with a WNW—ESE direction. The cross-fold axes plunge 45° ESE, in nearly vertical axial planes. Culm shales of the Ruesga group crop out where the cross-fold intersect a primary isoclinal syncline. Further towards the south the S-shaped fold has been cut off by the Tertiary border faults. Towards the northwest we can follow the S-fold in the Devonian quartz sandstones, where it clearly passes into a fault.

A summary of the structures of this subarea shows:

1. Main E—W folding in great isoclinal folds, with axial planes dipping 40°—60° S.
2. A faint WNW—ESE refolding with vertical axial planes; the plunges of the fold axes are related to the orientation of the bedding planes on which these folds are superimposed.
3. Development of the Tertiary southern border fault-zone, showing a change in direction from ENE—WSW to ESE—WNW, due to a latter warping.
4. The curves in the axial planes of the isoclinal main folds, as well as in the border faults (point 3), agree with the directions of a set of wrench-faults developed southwest of the Peña Redonda, of which the most important is the Cotolorno wrench-fault.

Subarea 2

The Besande anticline

The structures around Besande are of special importance for the understanding of the tectonic history of the area studied. This subarea is a junction of different structural units (fig. 16).

From the crest of the pass Puerto de Monte Viejo we look down to the west into the adjacent area of the valley of the Rio Cea. The E—W trending Tejerina syncline, developed in rocks of the Yuso and Cea groups, forms the northern boundary of this valley. Limestone layers of the Yuso group swing round near the road over the Puerto de Monte Viejo, forming the nose of the to the west plunging syncline. The lower part consists of shales, graywackes and conglomerates, which can be followed to the east into subarea 8.

The synclinal character clearly visible in the limestones, can no longer be distinguished in the shales and graywackes, because there are no marker horizons in this sequence and the layers show great horizontal lithological changes. Another reason for the disappearance of a clearly indicated synclinal structure further east, is the more complicated deformation of the incompetent shale beds.

The southern boundary of the Yuso group with the Devonian is formed by the vertical Sta. Eugenia fault, which runs eastwards to Camporredondo and joins the Cotolorno wrench-fault near Corbellera-Corbeñera hill. Near the Puerto de Monte Viejo the fault dies out in the unconformity plane between the Yuso group and the Devonian. The Yuso group curves together with the upper

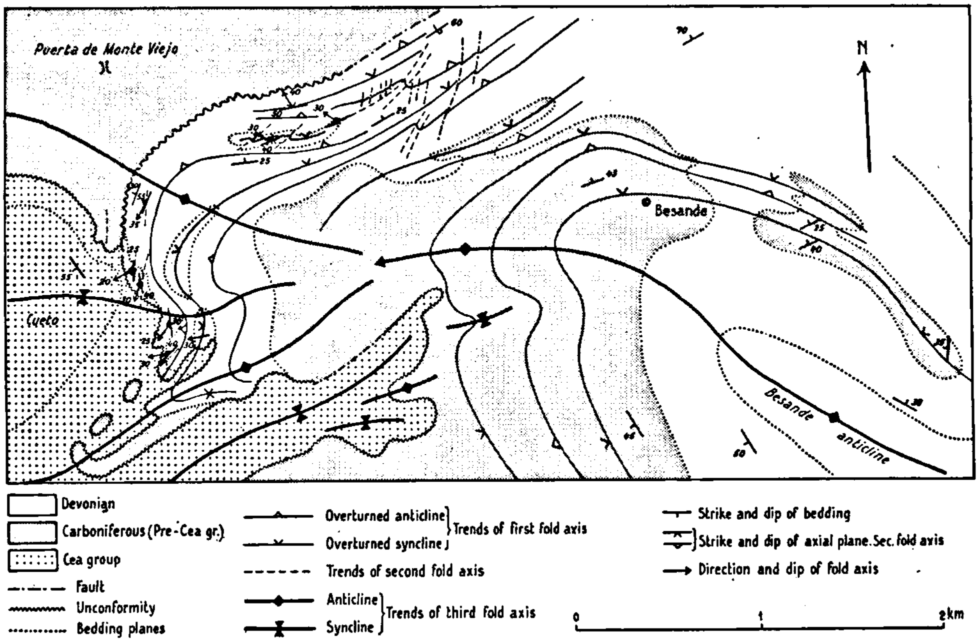


Fig. 16 Detailed structural map of the Besande subarea.
Location indicated on fig. 15.

Devonian rocks and the Caliza de Montaña to the south, under influence of a late E—W refolding. Halfway up the slope of the Cueto this structure is truncated unconformably by the strata of the Cea group. These conglomerates form an open and flat synclinal structure with an E—W trend and plunging to the west. This fold dies out towards the east where it encounters the older already folded structures of the Peñas Blancas. Between the Cueto syncline and the Tejerina syncline we find the great wide valley of Prioro, probably to be considered as an anticlinorium, but more detailed mapping will have to be done here. In our area, however, we know this anticlinal structure, which is called here the Besande anticline, forming the westerly spur of the Valsurvio dome. This anticline deforms the preseding isoclinal structures (fig. 16).

The axial trends of the isoclinal main folds of subarea 1 between the Rio Carrion and Besande, follow in broad outlines the curve of the Rio Grande. Near Besande the axial planes which dip 40° SSW, gradually flatten, so that northwest of Besande they have a dip of 20° NNW. This change of orientation of the axial planes is a result of the late E—W refolding, forming the Besande anticline (sections fig. 17). This deformation also refolds earlier folds, having axial planes making a small variable angle with the main isoclinal E—W folding. After rotating, in order to eliminate the later folding of the Besande anticline, this fold direction appears to be orientated in a NE—SW direction. This cross-

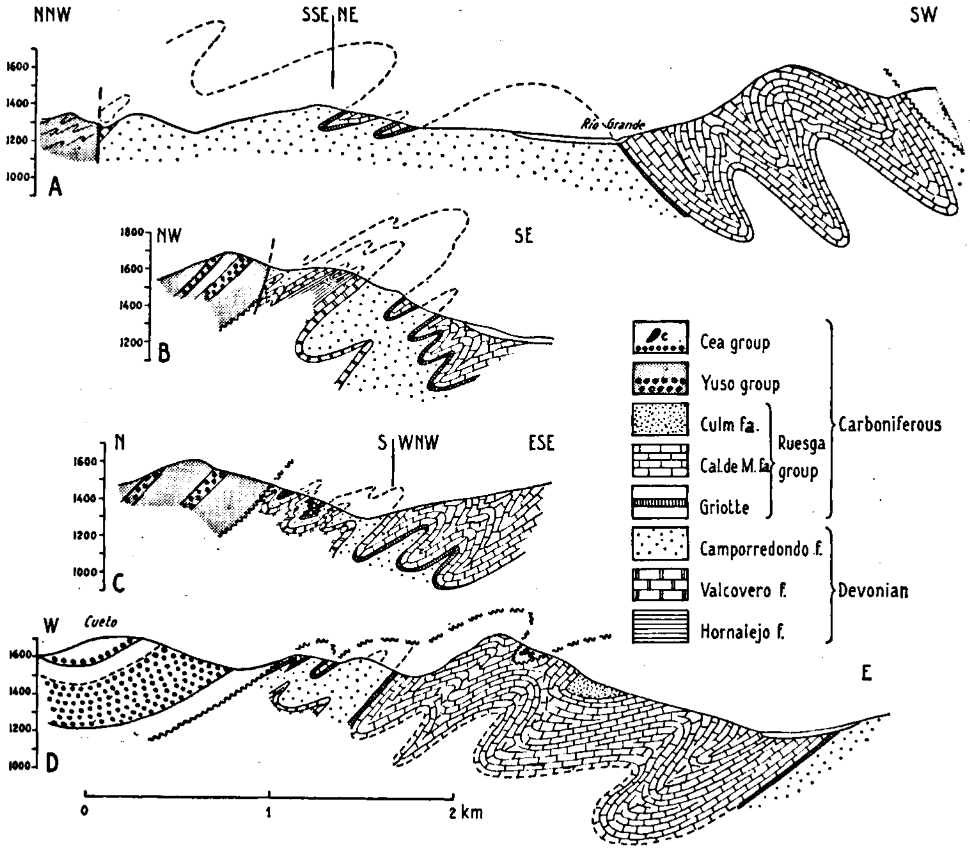


Fig. 17 Structural cross-sections through the Besande subarea. Location indicated on fig. 15.

folding deforms the main E—W structures, and has developed recumbent isoclinal folds with a coarse axial plane fracture cleavage in the limestones, fanning out at the hinges (fig. 18). The plunge of the axes of these cross-folds changes from 25—40° SW near Cueto mountain, to 30—50° NW north of the road Besande - Monte Viejo. Northwest of Besande a fault system occurs, approximately parallel with the cross-folding. Summarizing we find:

1. E—W main folding: isoclinal to north overturned large size folds
2. NE—SW cross-folding: recumbent folds, with a development of a faint axial plane cleavage, deforming the structures of the E—W main folding
3. Late E—W refolding: large open folds, with nearly vertical axial planes and plunging axes towards the west, formed after the deposition of the Çea group, deforming the preceding structures.

Subarea 3

The Otero anticline

The oldest rocks exposed in the area crop out in the core of the Otero anticline. This anticline is a wide E—W trending structure with its axial plane inclined steeply to the north. The folding picture is rather confusing, because this structure is superimposed on older E—W and N—S trending structures and therefore the formation of the Otero anticline has to be attributed to the late deformation phases. This conclusion is confirmed by the fact, that earlier cleavages and axial planes are deformed by it.

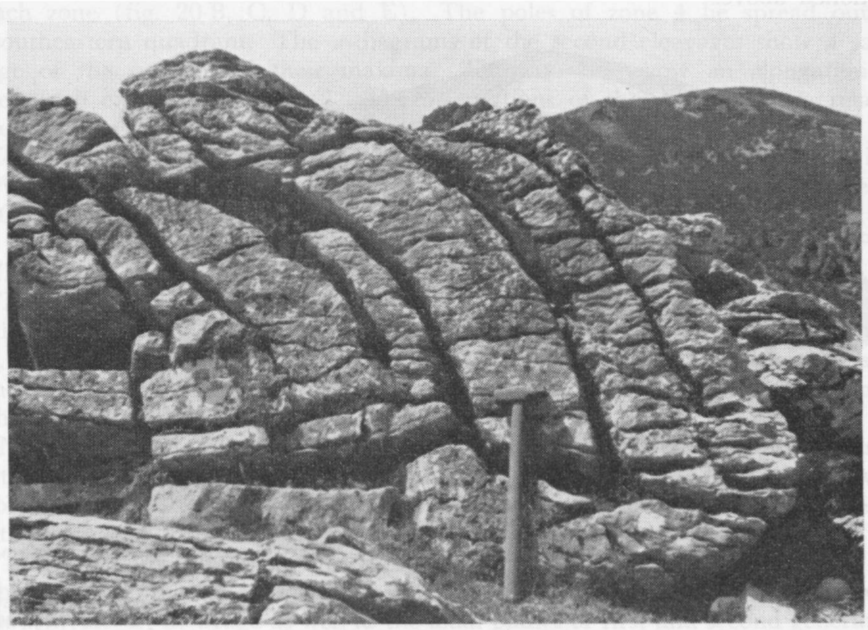


Fig. 18 Recumbent second generation fold in the limestones and griotte of the Ruesga group, 600 m E of Cueto mountain.

The gentle doming of the Valsurvio dome, visible on the map by the contours of the Caliza de Montaña and the Camporredondo formation swinging round in the west as well as in the east, is also apparent in the Middle and Lower Devonian formations. Upon closer examinations, however, several deformation phases seem to have played an important part and have caused almost inexplicable minor structures. West of Otero de Guardo the formations bend from 70° S dipping near the road Otero-Velilla, to 15° N dipping north of Otero de Guardo, forming the western nose of the Otero anticline.

In the diagrams of fig. 20 a number of poles of first and second cleavages from this subarea are plotted. For this purpose the subarea has been subdivided in five zones, each containing a homogenous part of the Otero anticline: zone 1 contains the southern limb, zones 2 and 3 the nose of the structure and zones 4 and 5 the northern limb (fig. 19). We can distinguish the following axial plane cleavages:

a. First cleavage, developed as a slaty cleavage.
b. Second cleavage, developed as a fracture or crenulation cleavage.
c. Third cleavage also developed as fracture or crenulation cleavage, deforming the first and second cleavages.

The third cleavage has not been developed clearly in many cases and can not be distinguished from the second cleavage in the field, but only in thin sections. Because of the small number of measurements of the third cleavage available, they are not included in the diagrams.

Besides micro knick zones developed as a second crenulation cleavage, we find a late generation of knick zones, cutting through all preceding structures mentioned above.

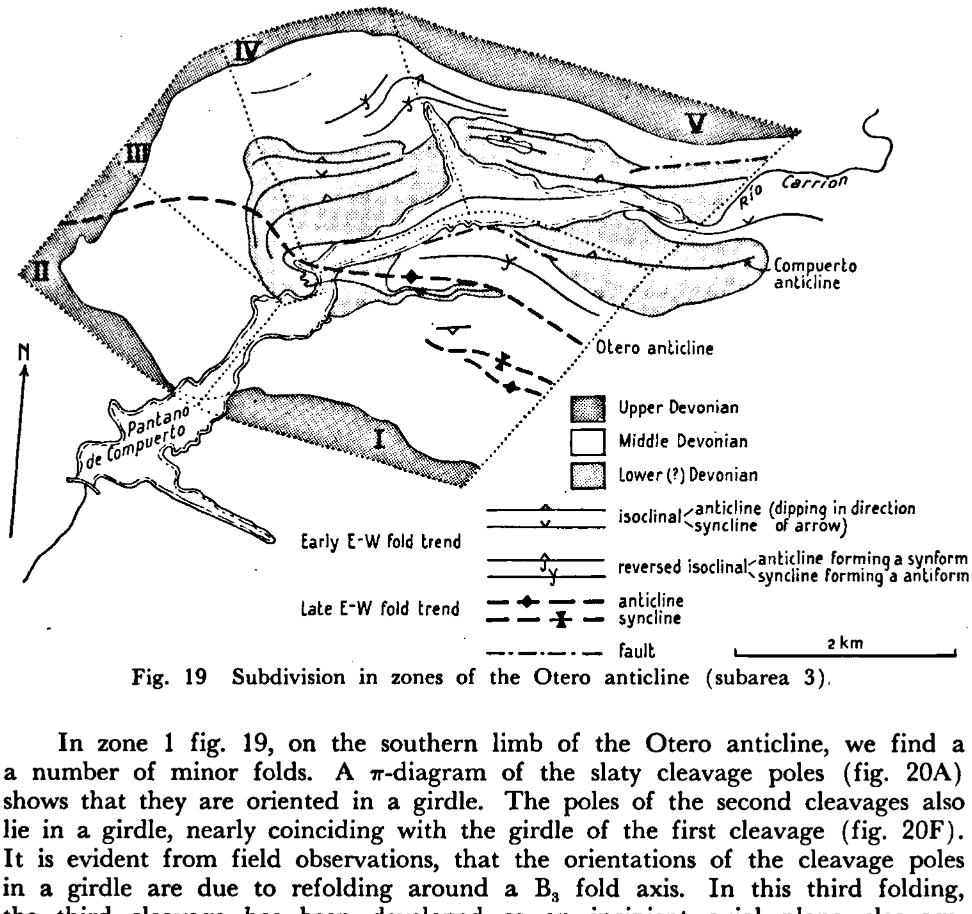


Fig. 19 Subdivision in zones of the Otero anticline (subarea 3).

In zone 1 fig. 19, on the southern limb of the Otero anticline, we find a a number of minor folds. A π-diagram of the slaty cleavage poles (fig. 20A) shows that they are oriented in a girdle. The poles of the second cleavages also lie in a girdle, nearly coinciding with the girdle of the first cleavage (fig. 20F). It is evident from field observations, that the orientations of the cleavage poles in a girdle are due to refolding around a B_3 fold axis. In this third folding, the third cleavage has been developed as an incipient axial plane cleavage. The angle between the slaty cleavage and the second cleavage planes before the refolding around B_3, is related to the difference in orientation of the girdles in diagram A and diagram F. These variations show that the second cleavages are refolded around a different axis than the first cleavages.

The axial planes of the third folds dip steeply to the north. The orientation of the axial planes of these late folds is constant; whereas the orientation of the fold axes is dependent on the earlier attitude of the bedding and cleavage planes on which this refolding is superimposed.

These third minor folds are closely related to the Otero and Besande anticlines, which form the axis of updoming of the Devonian in the Valsurvio dome. As described for subarea 2, this updoming is considered as a late E—W refolding, which has deformed the main isoclinal folds and the younger cross-folding developed in the limestones of the Caliza de Montaña facies. A resemblence between the folding history in both subarea (2 and 3) is obvious.

In the π-diagrams of the slaty cleavages for the zones 2, 3 and 5, of the Otero anticline we find that the poles are situated on elongated maxima, which form more or less great circle patterns and which have different orientations in each zone (fig. 20 B, C, D and E). The poles of zone 4 lie spread out in the southeastern quadrant. The π-diagrams of the second cleavages show a same change of the position of their maxima. They do not show an elongation in girdles, with exception of zone 1. The orientations of these maxima are related to the position of these zones on the Otero anticline (fig. 20 G, H, I and J).

Zone 1 on the southern limb has been refolded by minor folds of the late E—W refolding. The cleavages of zone 2, in the southern part of the nose of the Otero anticline (major late E—W refolding), swing from a steep S dip to a slight N dip in the nose of the Otero anticline (zone 3); then to a 30° N or NE dip in zone 4 and 5 on the northern limb of the Otero anticline.

These first and second cleavage planes are refolded by the late broad E—W anticlinal structure. The fold axes of these related isoclinal first and second folds, only make a small angle with each other, as may be occasionally seen in the field. This is also the reason why the first cleavage is less disoriented by the second folding as might be expected.

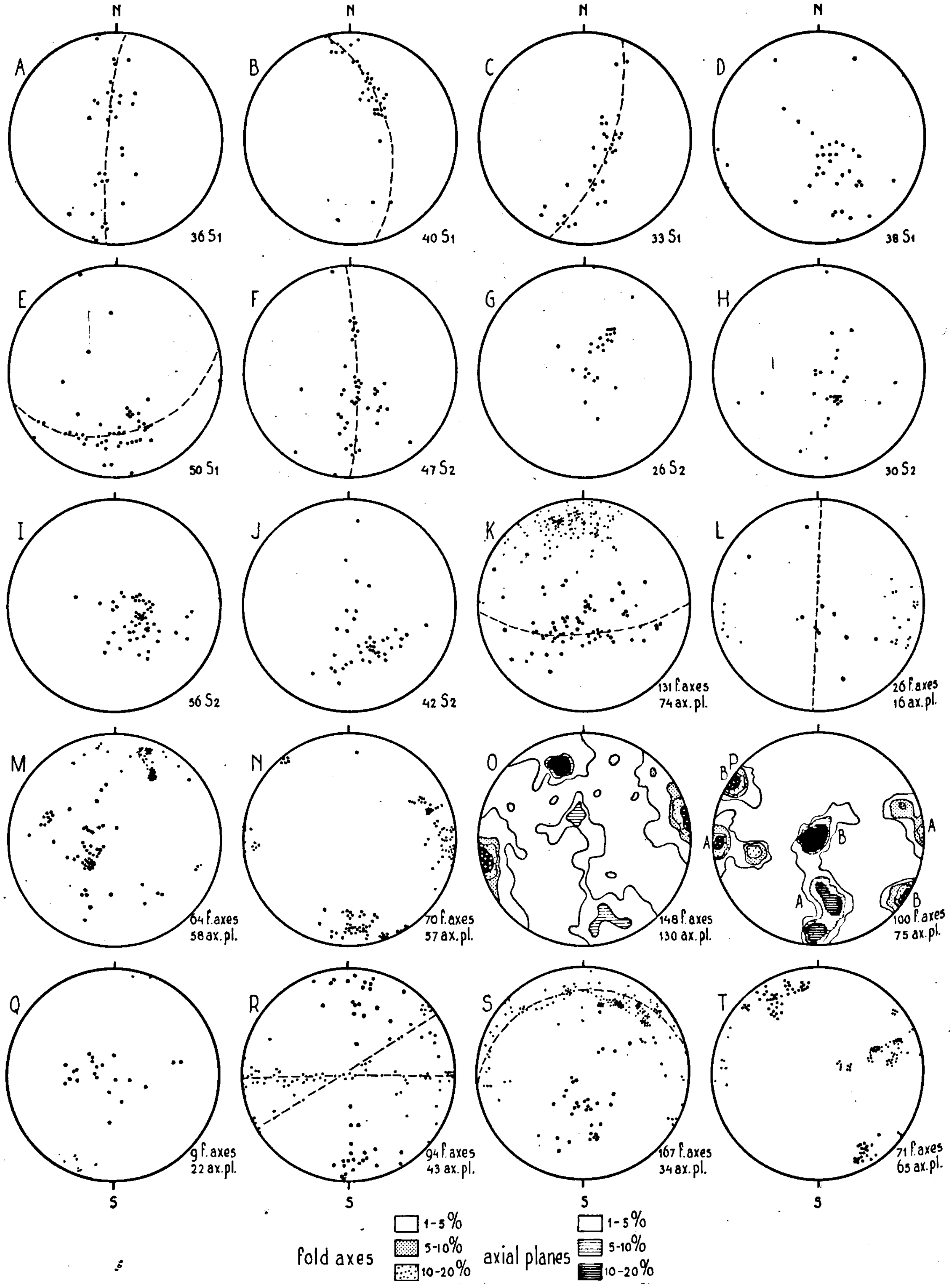
North of Otero de Guardo, in the lower course of the Arroyo Abianos, structures are clearly visible in the Otero limestone (fig. 21). These structures can be shown to be a pseudo-syncline [1]) with an antiform southeast of it. The latter has the youngest formation in the core, thus forming a pseudo-anticline.

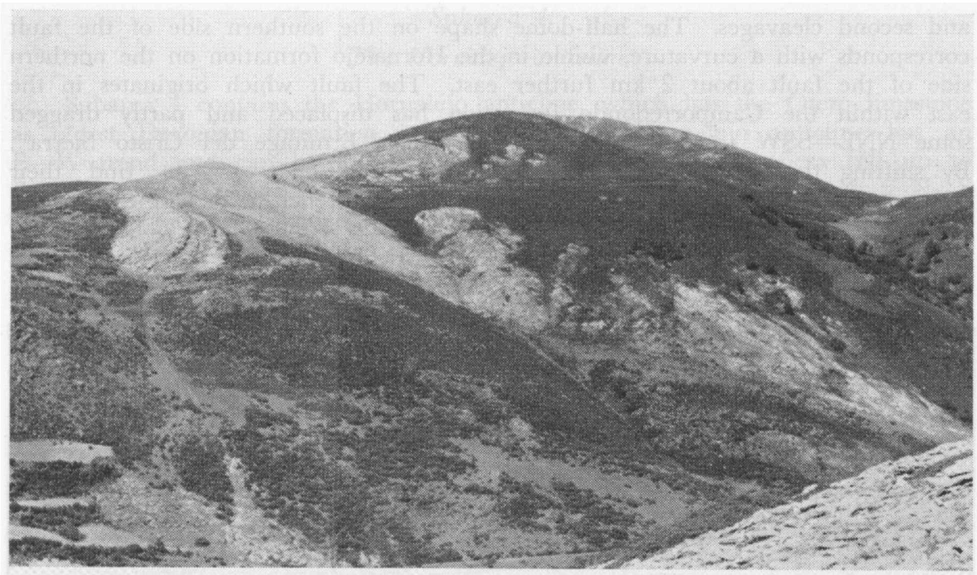
The originally south dipping second generation isoclinal folds are refolded by the Otero anticline, with the result that the noses of some of these folds are now inverted. Cleavage as well as the axial planes of first and second generation folds can sometimes be followed over the hinge of the Otero anticline changing from south to north, from a south dip throughout the horizontal to a north dip. In the limestones a coarse fracture cleavage has developed, which diverges upwards to the hinge of the anticlines.

On the eastern bank of the Arroyo Abianos, in a road cut along the highway Otero-Camporredondo, the sandy black slates of the Hornalejo formation lie in the core of a pseudo-anticline with the older Otero limestone resting on them. Notwithstanding the cleavage development a fauna of brachiopods has been collected from this exposure.

In the southern part of the subarea, in the valley of the Rio de Valcovero, the right lateral wrench-fault of Valcovero has been mapped (fig. 22) with an E—W trend and 2 km relative horizontal displacement. The bedding, as well as the cleavages of the Hornalejo slates south of the fault, dip at about 20° S. Just east and also 1 km west of Valcovero the bedding swings round in a northerly direction and is cut off by the Valcovero fault, with the result that the southern limb represents a half-dome, a form also followed by the attitudes of the first

[1]) An upside down anticlinal structure, visible in the field as a synform.


N
S
A
36 S1
B
40 S1
C
33 S1
D
38 S1
E
50 S1
F
47 S2
G
26 S2
H
30 S2
I
56 S2
J
42 S2
K
131 f. axes
74 ax. pl.
L
26 f. axes
16 ax. pl.
M
64 f. axes
58 ax. pl.
N
70 f. axes
57 ax. pl.
O
148 f. axes
130 ax. pl.
P
A
B
100 f. axes
75 ax. pl.
Q
9 f. axes
22 ax. pl.
R
94 f. axes
43 ax. pl.
S
167 f. axes
34 ax. pl.
T
71 f. axes
65 ax. pl.
fold axes
1-5 %
5-10 %
10-20 %
20-40 %
axial planes
1-5 %
5-10 %
10-20 %
20-40 %



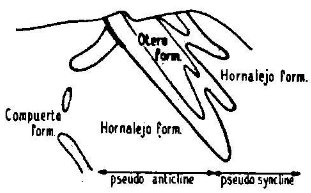


Fig. 21. Second generation folds in the Middle and Lower Devonian on the western border of the Arroyo Abianos, inverted by the late E—W refolding.

Fig. 20 Stereographic projections on a lower hemisphere Schmidt net. Contour values in percentages per 1 % area.

A—J: . indicate poles of cleavages.

K—T: . indicate fold axes and lineations,

• indicate poles of axial planes and related axial plane cleavages.

A, B, C, D and E. diagrams of the first cleavage respectively in zones I, II, III, IV and V of the Otero anticline (cf. fig. 19).

F, G, H, I and J. diagrams of the second cleavage respectively in zones I, II, III, IV and V of the Otero anticiline.

K. Minor N—S cross folding of the limestone of the Ruesga group near the Arroyo de Agueras.

L. Minor folds of the first E—W folding, at same location as K.

M. N—S cross folding, Culm facies along the road Santibañez de la Peña-La Lastra.

N. Late E—W refolding at same location as M.

O. Late E—W refolding, Ruesga group along the eastern border of the Pantano de Camporredondo.

P. A) Late E—W refolding, Yuso group along the Pantano de Camporredondo southwest of Triollo.

B) First or second generation folding, Ruesga group and Devonian along the Pantano de Camporredondo southwest of Triollo.

Q. N—S cross folding, Yuso group eastern border of the Pantano de Camporredondo.

R. Late E—W refolding at same location as Q.

S. N—S cross folding, Yuso group between Cardaño de Abajo and the Arroyo Abianos.

T. Late E—W refolding, Yuso group at same location as S.

and second cleavages. The half-dome shape on the southern side of the fault corresponds with a curvature, visible in the Hornalejo formation on the northern side of the fault about 2 km further east. The fault which originates in the east within the Camporredondo formation has displaced and partly dragged some NNE—SSW trending structures near the "Ermitage del Cristo Sierra". By shifting the northern block to its original position, we can find their continuation to the south between the Valcovero wrench-fault and a southern branch of the fault. Where the Valcovero fault is bordered on both sides by the Camporredondo formation, its trace is difficult to detect and the amount

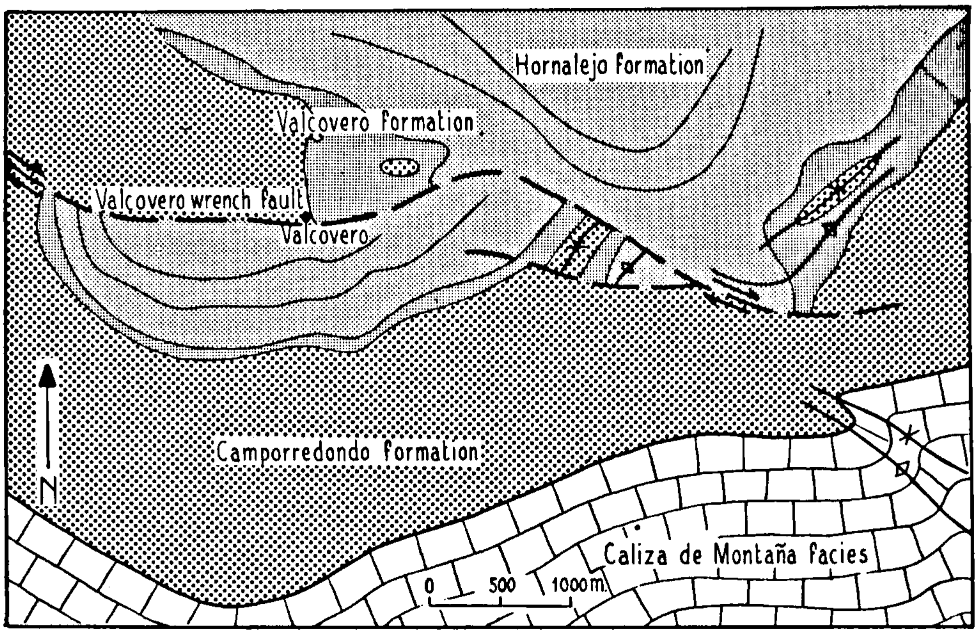


Fig. 22 The Valcovero wrench-fault, with 2 km displacement of right lateral movement.

of horizontal movement is not known. The fault zone is, however, characterized by strongly decemented and weathered quartz sandstones.

A similar zone related to the Sta. Eugenia fault also characterized by incoherent decemented sands is to be found near Camporredondo. The Valcovero wrench-fault must have been formed after the development of the Valsurvio dome, which has deformed the first and second generation cleavages and probably has the same age as the Cotolorno wrench-fault.

Summarizing we find in the subarea 3 the following succession of deformation:

1. Recumbent E—W isoclinal folding with an axial plane slaty cleavage.
2. Second isoclinal folding with an axial plane fracture or crenulation cleavage, the fold axes probably run ESE—WNW.
3. Late E—W refolding, broad open folds with a vertical or steeply north dipping axial plane, sometimes a faint axial plane cleavage has been developed. The major structure has more or less a dome shape.
4. Valcovero and other wrench-faults.

Subarea 4

The Hornalejo anticline

Subarea 4 contains the Hornalejo anticline, which has the Otero limestone as oldest Devonian formation in its core. The Hornalejo anticline has an E—W trend and can be followed from the Corbellera Corbeñera hill up to the valley of Valsurvio in the west. Here the structure interferes with a N—S structure, which as a ridge of younger rocks (Valsurvio formation) separates the older rocks of the core of the Hornalejo anticline (Otero limestone) from those of the Compuerto anticline (Compuerto formation). In detail the N—S structure is an isoclinal syncline with an axial plane dipping 55° W. This syncline,

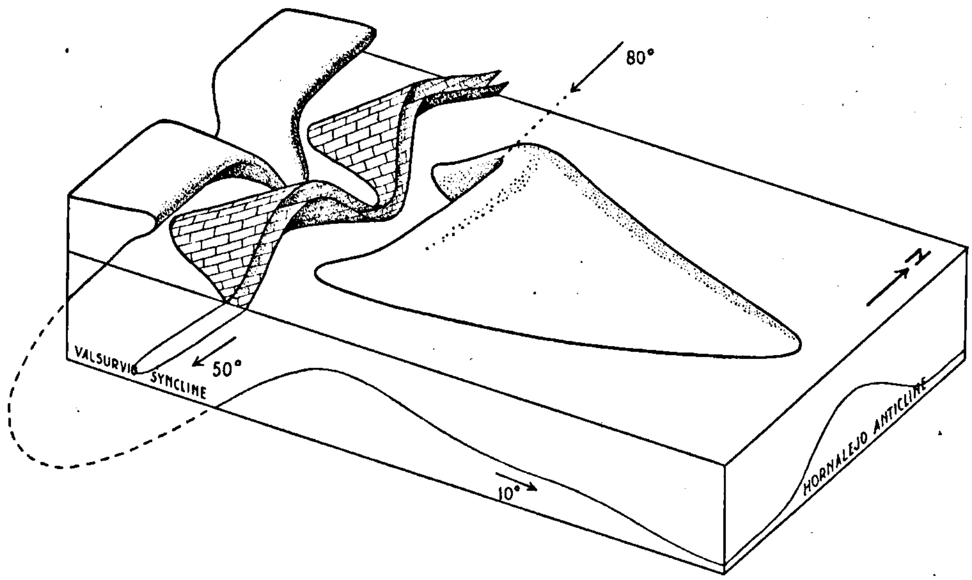


Fig. 23 Schematical block diagram of the late E—W Hornalejo anticline intersecting an earlier N—S cross fold.

which we will call the Valsurvio syncline, has a limited extent and can only be traced from Valsurvio to the north as far as the valley of the Arroyo the Sta. Eugenia. East of the isoclinal Valsurvio syncline, another N—S structure can be distinguished, developed as a broad open anticline, with a western limb dipping 50° W and an eastern limb dipping 10 ° E. Both have been strongly refolded by later E—W structures.

We will follow the axial trend of the Hornalejo anticline for a moment from east to west, to observe the development of this fold superimposed on the N—S structures. The axial plane dips about 80° N. The direction of the fold axis is more or less E—W and plunges about 10° E on the eastern limb of the broad N—S anticline. As the axial trend is traced westwards over the crest of the anticline, the plunge suddenly steepens to 80° W near the Arroyo de Valsurvio. At the same time with the changing of the plunge, the interesting feature is observable that the Hornalejo anticline passes into a steep plunging synclinal structure. On the limbs of this syncline two smaller, also steeply plunging anticlines are developed, accompanied by wrench-faults which possibly have a small vertical component of movement as well (fig. 23). Further west, where the axial trend crosses the Valsurvio syncline, the axis plunges about 50° W and the structure remains developed as a syncline. On the crest of Collada de Peña Cañales, the overturned upper limb of the Valsurvio syncline reverts to a nearly horizontal position. In consequence the E—W superimposed syncline is developed in the Hornalejo slates with a nearly horizontal fold axis. The structure continues into subarea 3 to the west.

A study of cleavage and related minor structures was hindered by the lack of good exposures. Cleavage is mainly restricted to the Hornalejo slates. Slaty cleavage developed as axial plane cleavage is often subparallel to the bedding. Two systems of axial plane fracture or crenulation cleavage can be observed. The incipient third generation cleavage is difficult to detect with the naked eye. The strike and dip of the first and second cleavages are strongly variable due to the late E—W refolding. Microscopically they are clearly developed; macroscopically we sometimes find the slaty cleavage as a plane of easiest parting and a more difficult parting along the second cleavage planes, but in other places we find the reverse. Sometimes the rock splits along both equally well. On the Collada de Peña Cañales both first and second cleavages lie in a nearly horizontal position and are deformed by the E—W syncline, which is situated in the extension of the Hornalejo anticline to the west. The third generation cleavages are related to this late refolding. If there is any relationship between the first or second generation cleavages and the N—S trending Valcovero syncline could not be found because of inadequate exposures.

The isoclinal second generation minor folds deform the slaty cleavages. As these folds have strongly compressed limbs, the angle between the first and second cleavage is very small. The fold axes and directions of lineation of these folds show only a weak preferred orientation in an east to southeast direction, horizontal or with a plunge of up to 10° E. The late E—W refolding, including the Hornalejo anticline, has developend broad folds with vertical axial planes.

Summarizing we find:

1. Isoclinal folding with a slaty cleavage developed as axial plane cleavage.
2. E—W isoclinal folding with axial plane cleavage which deforms the slaty cleavages.
3. N—S trending fold, partly isoclinal, with an axial plane dipping 55° W (Valsurvio syncline). Its relation to the first and second generation folds remains obscure.
4. Late E—W refolding, forming large open folds with vertical axial planes and varying plunges of folds axes.

Subarea 5

The Ruesga group and Camporredondo formation of the northern limb

The subarea consists of shales, graywackes, limestones and griottes of the Ruesga group and quartz sandstones of the Camporredondo formation.

In the field the griotte is clearly recognizable and is used as a stratigraphic marker horizon. The griotte exposures in this subarea give a clear picture of the complexity of folding.

In the western part of the subarea between the Pantano de Camporredondo and the Arroyo de Sta. Eugenia, several isoclinal folds occur overturned to the south. The most important one is the Dorada syncline, having limestones of the Caliza de Montaña facies in the core. It is cut obliquely by the Cotolorno wrench-fault of Tertiary age. This fault has rotated the E—W trend of the Dorada syncline into a northwesterly direction. The minor structures on the Peña Dorada, with an amplitude in the order of 10 m, have fold axes plunging 30°—40° NE—NNE. The axial planes of these folds are nearly parallel to the axial planes of the major E—W folds. An axial plane cleavage, affected by the minor folds, occurs in the limestones and griottes.

As described in chapter III great lithological differences are present between the sediments of the Valsurvio dome and the zone of San Martin - Camporredondo. The sudden decrease in thickness along a sharp line of the competent Camporredondo quartz sandstone to the north, together with the lateral facies change of the Caliza de Montaña facies into the highly incompetent beds of the Culm facies are the causes of the completely different deformation types in the two structural units.

In the southern limb of the Valsurvio dome as well as on the Peña Dorada in the northern limb, the type of deformation has been dictated by the competent Upper Devonian quartz sandstones, several hundreds of metres thick and by the massive strongly recrystallized limestones. In the limestones to north overturned isoclinal major folds have been formed (subarea 1), while the quartz sandstones have been deformed mainly in a concentric fashion, partly broken or thrusted.

In the northern limb of the Valsurvio dome, near El Otero mountain, the folding type as well as the lithological characteristics change in a northerly direction and we enter the second structural unit: the zone of San Martin - Camporredondo. The Camporredondo quartz sandstone, here 1 m thick, no longer dictated the folding style. The incompetent Culm beds dominate over the thick massive limestones. With this sudden lithological change, folding style has adapted itself. The stress which caused the large to north overturned isoclinal folds, related in size to the thickness of the competent beds, suddenly no longer encountered the resistence of these beds and with the wedging out low-angle north directed overthrusts are developed. The fold pattern is complicated by later refolding.

The most striking feature in this zone is the tectonic contact between the griotte and the underlying shales of the Culm facies. This contact is due to the low-angle thrusting of griotte and limestones over the Culm shales, in which the griotte served as a detachment horizon. Sometimes, for example near the hill of El Otero, south of Rebanal, we find a few metres of Camporredondo quartzites preserved above the thrust plane. The limestone beds lie very flat and form a dip slope. The thrust plane is exposed in the deeply eroded valleys, with below it the Culm shales, as for instance in the depression around the spring "Fuente de la Espina".

The flat-lying thrust planes, thrusted from south to north have been refolded by E—W structures and are partly overturned to the south. This superimposition of deformations results in a repeated outcropping of the thrust plane (fig. 24).

Between the section through the El Otero hill (section b) and a section 2.5 km to the west (section c) another important feature can be seen in a ridge of Caliza de Montaña, which trends approximately NNW—SSE. In this ridge isoclinal structures have NNW—SSE directed fold axes and subhorizontal axial planes and belong to the N—S cross-folding. This N—S cross-folding occurs only in a narrow zone. It is nearly impossible to make a section through this zone because of the complexity of the fold pattern.

Between Peña Cotolorno and Peña Redonda the limestone is strongly dolomitized and no structures can be distinguished. In the Peña Redondo a recumbent fold is clearly visible (Kanis, p. 427).

Summarizing we find the following succession:

1. ESE—WNW folding; ESE—WNW directed recumbent folds with horizontal or gently south dipping axial planes, which to the north pass into low-

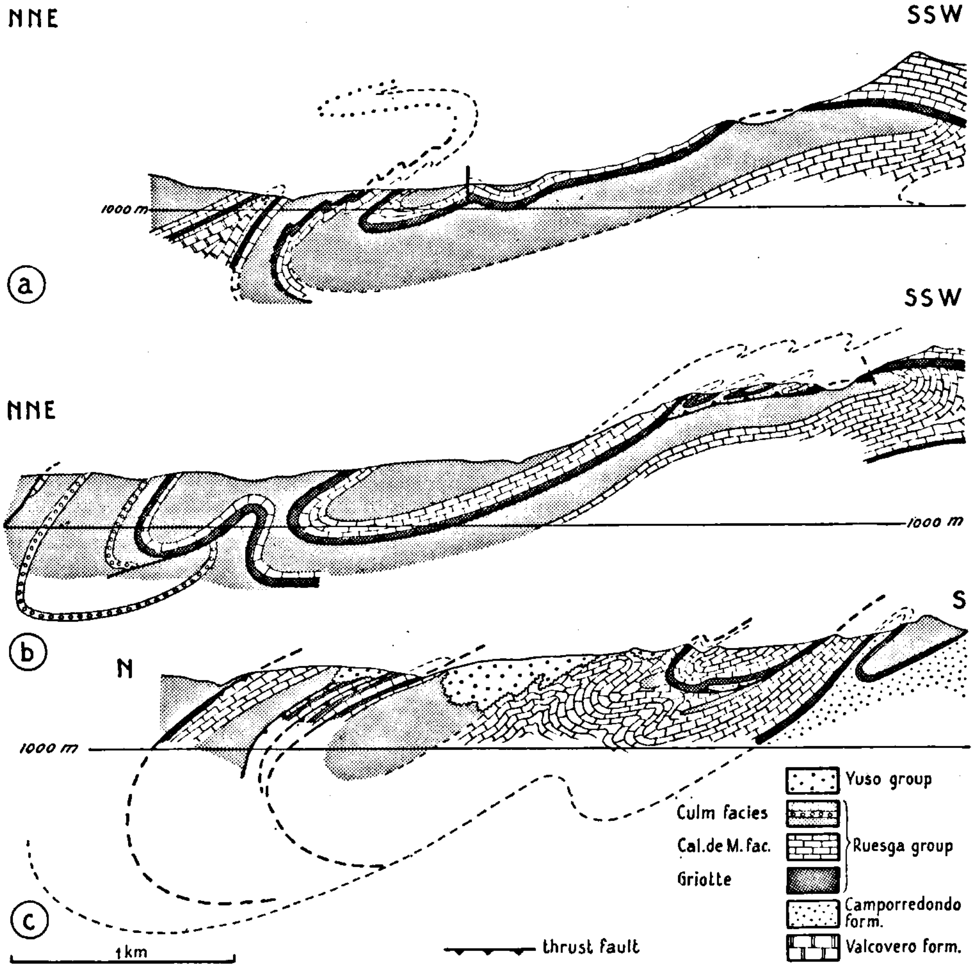


Fig. 24 N—S cross-sections through the northern limb of the Valsurvio dome (subarea 5). For location see fig. 15.

angle overthrusts, thrusted from south to north, for which Upper Visean red shales and radiolarites served as detachment horizon.

2. Second folding, with an E—W trend; the axial planes of the asymmetric folds dip about 30° N.
3. N—S cross-folding, occurring in zones. Especially developed as minor

folds with a flat lying axial plane, probably contemporaneous with the second folding.
4. Influence of the alpine Cotolorno wrench-fault, causing a slight bending in the axial planes of the folds on Peña Dorada.

ZONE OF SAN MARTIN — CAMPORREDONDO

Subarea 6

The fold complex of the thrust fault area of San Martin

The structures northwest of San Martin de los Herreros, between the Rio Rivera and the Arroyo de Agueras, are very complicated. The structural study is based on a detailed study and the preparation of a geological detail map reproduced at a scale 1 : 15.000 (app.).

The Carboniferous Caliza de Montaña as well as the Devonian Valcovero limestone are well exposed and show an intensive folding on minor as well as on macro scale. The red shale and griotte alternation of a thickness of 1—10 m is a good mappable stratigraphic defined horizon and has been drawn on the maps and on the sections in exaggerated proportions. The shales and calcareous mudstones of the Culm facies are badly exposed. The northern slope of the Peña Negra has a strong afforestation, which hides the geology a great deal. An analysis of the relation of the rocks indicates that a least two major fold movements have deformed the area. The most striking difference between the first folds and the refolding structures are the opposite orientations of their axial planes.

The first recognizable event in the tectonic history is the formation of a series of recumbent and overthrusted WNW—ESE trending major folds with axial planes dipping very gently to the south. Fig. 25 gives a section of the supposed structure just after the first folding phase, the sections of fig. 26 (see App.) the present structure after the second folding. The first folding movements were accompanied by a thrusting from south to north, with the result that parallel with the isoclinal folds a series of flat-lying overthrusts were developed. In nearly all cases the thrust faults are located in the incompetent red shale-griotte beds.

The thrust plane is difficult to detect, because with very few exceptions no mylonitization or other disturbance occur and its existence can only be concluded from the repetition of the stratigraphic sequence.

The most important major fold in this area is a recumbent anticline, the San Martin anticline (4 [1]), figs. 25 and 27), running from the Pantano de Ruesga in the east, via San Martin up to the Peña Negra in the west. This fold is described by de Sitter (1957) in the region between San Martin and Ventanilla. Together with this fold a contemporaneous low-angle overthrust (3) occurs in the southern limb. The northern limb of the recumbent San Martin anticline is strongly squeezed out at many places and shows evidences of local thrusting (near Ventanilla). These folds were primary developed as recumbent isoclinal folds, strongly overturned to the north and with the reversed northern limb strongly reduced by shearing. The flat overthrusts result in large horizontal

[1]) The numbers between brackets, noted behind first generation structures in the text, are related to the numbers of fold axes and thrust fault traces on figs. 25 and 27.

translations from the south towards the north. The present inverted position of the thrusts is due to the second deformation stage.

West of San Martin the above mentioned thrust plane (3) occurs south of the San Martin anticline. The limestone with griotte at its base and locally also small patches of crushed Camporredondo quartz sandstone are partly thrusted over the shales and conglomerates of the Culm facies and partly over the limestones of the Caliza de Montaña facies. Thrust plane (3) is succeeded to the south by a recumbent syncline (2), the Peña Negra syncline, which has in its core a thick series of the Culm facies. This syncline is followed to the south by two other thrust faults (1) described in subarea 5.

North of the San Martin anticline a tightly compressed syncline (5) occurs, followed by a series of flat overthrusts (6, 7, 8, 9, 13) and recumbent folds (10, 11, 12, 14). These first WNW—ESE trending structures are sketched in the section of fig. 25.

The E—W refolding dictates the present day geometry of the folding pattern, having deformed the earlier structures very intensively. The structural map and the sections of fig. 26 and 27 give this complicated tectonic frame-work.

The axial planes of the E—W refolding dip about 30—50° to the north. The originally south dipping strata are all overturned and now dip to the north.

The most important E—W trending cross-fold is the Rebanal syncline, which crosses the whole series of first deformation structures. We can follow this syncline over a distance of about five kilometers from west to east. Where the Rebanal syncline is superimposed on the thrust plane (1) and the Peña Negra syncline (2) in the West of the region complications occur. We can follow the Carboniferous limestone with the griotte above thrust plane (1) in an anticlinal and a synclinal bend, northwest of Rebanal, only broken by a few transverse faults. The bending must be due to the refolding by the Rebanal anticline and the Rebanal syncline, because the original flat thrust plane (1) is folded in the same way. Similarly the primary Peña Negra syncline is also refolded. When we follow the southern limb of this syncline (2) by tracing the conglomerate bed from the northwest into the bend of the secondary Rebanal syncline, this limb is soon cut off by an underthrust, which partly follows the already existing thrust plane (1). This reactivation of the thrust plane is later than the inversion of the plane, but the relative direction of movement is the same. The small anticlinal nose in the conglomerates, against the reactivated thrust plane, 300 m north - northwest of Rebanal, is not the secondary Rebanal anticline, but the inverted synclinal nose of the primary folding (Peña Negra syncline (2) now forming a pseudo-anticline or antiform).

This interpretation is confirmed by top and bottom criteria (small scale cross lamination, load casts, flame structures, wash-outs and other features visible in the conglomerates), which can only be detected by careful examination of the rocks. The series between Rebanal and the above mentioned conglomerate can be shown to have been doubled at least three times.

When we follow the conglomerate bed to the east, we are walking in the northern limb of the Peña Negra syncline (first deformation structure). This northern limb is also refolded by the Rebanal syncline — the latter has been split up to the east into two synclines — and makes a much greater S. shaped curve than the southern limb. The plunge and the direction of the superimposed fold axes is controlled by the attitude of the first deformation fold limbs, and change from a northwesterly direction with a plunge of about 40° on the southern limb to a west-northwesterly direction with a plunge of 20° to 30° on the northern limb.

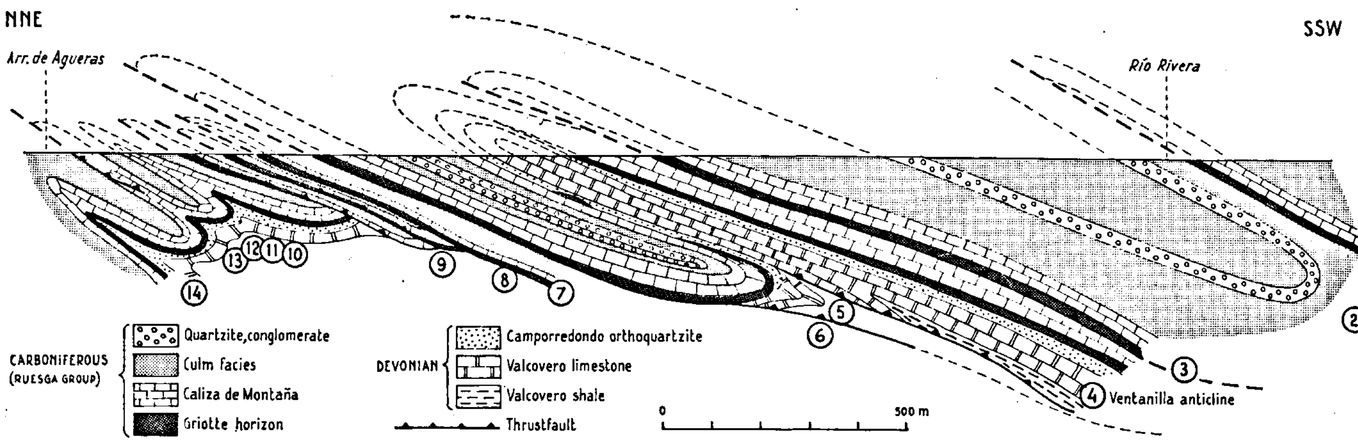


Fig. 25 Schematical reconstructed cross-section through the first generation structures in the thrust-fault zone of San Martin before refolding.

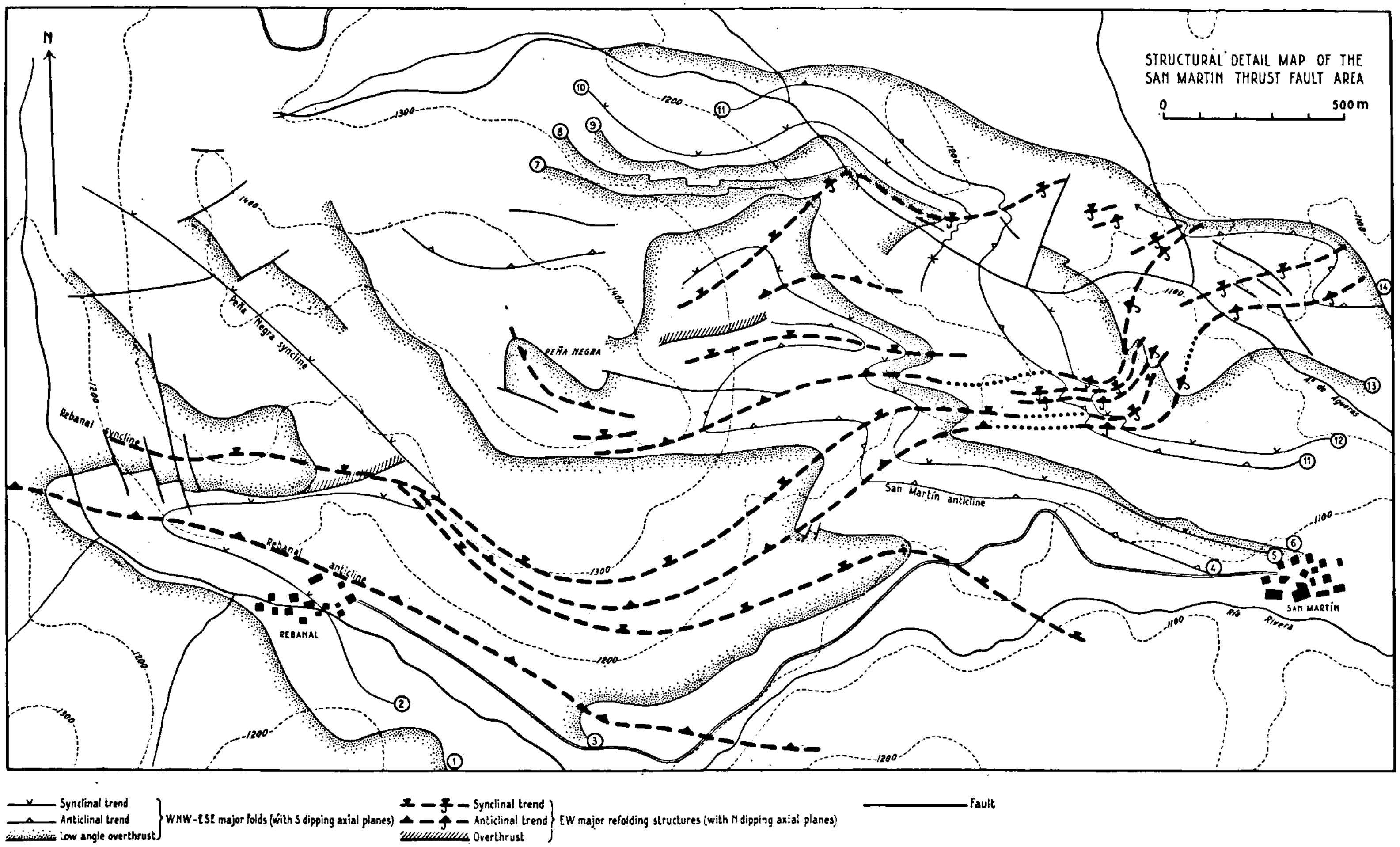


Fig. 27 Structural map of the thrust-fault zone of San Martin.

Where the road from San Martin to Rebanal turns from a southwesterly direction into a west-northwesterly direction, a small Devonian anticline core is visible on the west side of the Rio Rivera (Kanis, fig. 20). This is the easterly continuation of the Rebanal syncline. Kanis wrongly connects the San Martin anticline (primary structure) with this anticline of the E—W refolding.

North of San Martin the structures become more complicated. Tight isoclinal first fold structures are intensively refolded and the scale of the folds decreases. Following the Rebanal syncline along its axial trace to the northeast this fold becomes inverted into an antiform or pseudo-anticline near the Arroyo de Agueras. The variation in direction of the axes of these second fold structures in their parallel axial planes is due to variations in attitude of the limbs of the first generation structures on which they are superimposed.

The small limestone hills north of the Arroyo de Agueras are formed by the refolded southern limb of an anticline of the first generation (fig. 29). These limestone hills are surrounded by shales and mudstones of the Culm facies. Fig. 30 shows a pseudo-anticline refolding a low-angle thrust plane (9) of the first generation.

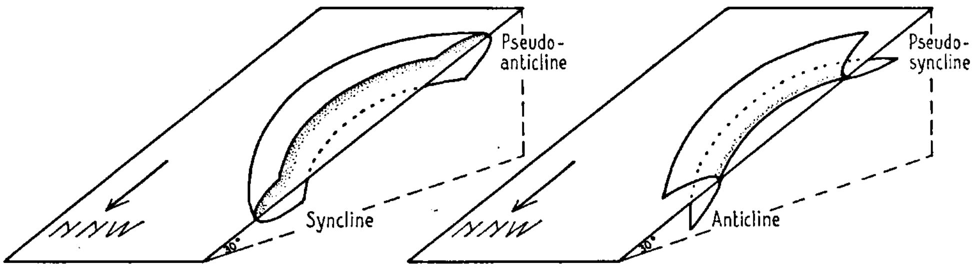


Fig. 28 Schematical diagram of the change in direction of the fold axes with respect to the dip of the axial plane, resulting in a transition from syncline into pseudo-anticline and anticline into pseudo-syuncline.

To the north and west these structures are cut off by the E—W running Rio Rivera fault (cf. Kanis 1955). Only the Peña Negra syncline can be followed as far as the watershed between the Arroyo de Valderrianes and the Rio Rivera.

Southeast of La Lastra, on the southern side of the Arroyo de Valderrianes the same thrust structures crop out again (see app. geological detail map of the thrust fault area south of Triollo and La Lastra). At least four sheets have been thrusted over each other. The E—W refolding has turned them into an inverted position. A contemporaneous E—W compression caused a slight bending of the layers into more southerly directions. This N—S direction is rather well developed in the minor folds (fig. 20, diagram M and Q). The thrust structures are cut off to the north by the Rio Rivera fault.

North of this fault a thick clastic sequence occurs. On account of a gradual transition from limestone (Ruesga group) to shales and mudstones, near the village of La Lastra, they are considered as belonging to the Culm facies. The boundaries between Ruesga and Yuso drawn on the detailed map are rather arbitrary.

In the rocks of the Culm facies (this classification is open to discussion, due to lack of fossils) along the road Santibañez de Resoba - La Lastra, small minor folds are developed which show great variations in shape, size and orientation, due to lithological differences. Diagram M (fig. 20) shows a great concentration of north trending folds with their axial planes, dipping slightly to the west. The distribution of the axial plane poles shows a more or less elongated maximum. This dispersion is due to refolding by the late E—W refolding, the folds of which are plotted on diagram N. The fold axes plunge a few degrees and the attitude of the axial planes is more or less vertical. Both maxima are somewhat elongated, due to the development of conjugate sets, especially in the crenulation cleavage. The angle between the two related sets of lineations, which form the conjugate system, is often more than 30° in one and the same hand specimen. This same phenomenon is found in the Yuso group on the eastern border of the reservoir (diagram R). A conjugate set of micro folds

Fig. 29 Refolding structures in the Ruesga group north of the Arroyo de Agueras. In the background the biohermal limestone (Caliza de Montaña facies) near Santibañez de Resoba.

of the late vertical E—W refolding is superimposed on a N—S minor fold, which gives rise to a distribution of the lineations of the micro folds on two girdles, making an angle of about 30° with each other. The axial planes of these micro folds dip steeply to the north as well as to the south.

Diagram O represents the late E—W refolding developed in the limestone and griotte of the Ruesga group on the eastern border of the Pantano de Camporredondo. Diagram P, contour values by B, represents first or second generation structures found in the Devonian and Ruesga group along the border of the reservoir southwest of Triollo. The connection of these recumbent folds with the major structure remains obscure.

Minor folds developed in the well-layered limestones near the Arroyo de Agueras are plotted in diagram K and L (fig. 20). Concentric as well as accordion folding (fig. 31) occurs and sometimes traces of incipient axial plane cleavages are visible. In the field a succession of the several minor folds is only seldom distinguishable.

Diagram L represents first generation minor folds, which could be identified by field evidence of the deformation of their axial plane cleavage by a later refolding. The distribution of the plots of diagram L is rather similar to those of the late E—W refolding (fig. 20 diagram O).

Diagram K represents second generation minor folds and shows a more or less similar distribution to diagrams M and Q. Axial planes as well as axes show a rather large spread.

Summarizing the following structural succession has been found:

1. WNW—ESE trending low-angle overthrusts and isoclinal S dipping folds.
2. E—W refolding with 30°—40° N dipping axial planes, contemporaneous with a N—S cross-folding, mainly developed as minor folds.
3. Late E—W refolding, with nearly vertical axial planes, often developed as conjugate sets.

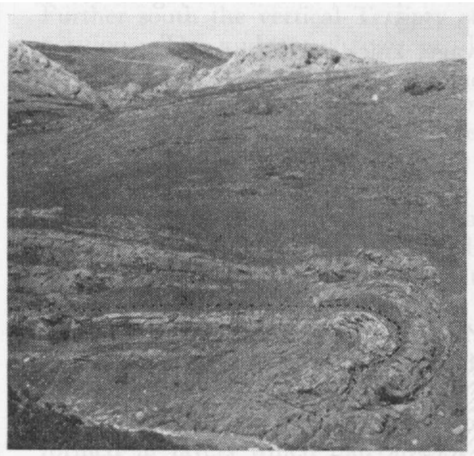


Fig. 30. Folded low-angle oberthrust forming a pseudo-anticline, north of the Arroyo de Agueras.

Fig. 31. Sharply folded minor folds of the first generation north of the Arroyo de Agueras.

Subarea 7

The deformation of the Yuso group

Major structures are difficult to detect in the Yuso group. Small outcrops of limestones and locally griotte of the Ruesga group form an E—W trending zone of inliers, which coincides with an E—W trending core of an isoclinal anticline, its axial plane dipping 30° N.

The boundary between the Yuso group and the Devonian is formed by a vertical fault, the Sta. Eugenia fault, a branch of the Cotolorno wrench-fault. Between this fault and the above mentioned anticline a synclinal curve has been found in the conglomerates on the easterly slope of the Arroyo Abianos. N—S minor folds have been formed contemporaneously with these E—W major folds. Their axial planes have similar attitudes. A contemporaneous slaty cleavage has been formed. The plotted lineations and fold axes of the minor folds are situated on a girdle. The related cleavages and axial planes are concentrated around the attitude defined by this girdle of lineations and axes (diagram S, fig. 20). So the axes change in direction in the axial plane, due to differences in attitude of the limb of the major fold on which they are formed; a similar picture to that near San Martin (p. 205). When we consider the deformation of the limestones of the Ruesga group situated as inliers in the Yuso group, a similar picture is observed.

A later deformation has been found with a direction of N 60°—70° E and nearly vertical axial planes deforming the slaty cleavage (diagram T, fig. 20). These folds are of a concentric or an accordion type. Diagram P, contour values by A, represents similar folds measured in the Yuso group SW of Triollo.

On the eastern border of the Pantano de Camporredondo an E—W trending major isoclinal syncline is formed in the Yuso group between the Arroyo de Valderrianes and the Arroyo Miranda. On the limb a few N—S minor folds are developed (diagram Q) upon which are superimposed a conjugate set of late E—W micro folds (diagram P) (see p. 215).

All transitions between crenulation cleavage, knick zones and parasitic micro folds occur. The following development could be observed (fig. 32).

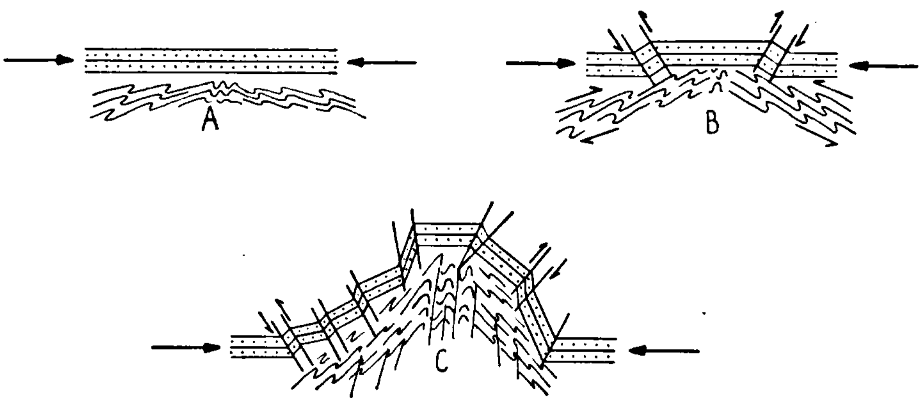


Fig. 32 Development of the parasitic micro folds in the siltstones of the Yuso group on the eastern border of the Pantano de Camporredondo (6 × reduced).

A. Slates forming micro folds, undisturbed sandy siltstone.
B. During further compression, the development of a conjugate set of knick zones in the sandy siltstone.
C. The micro folds show a tendency to grade into crenulation cleavage, whereas in the sandy siltstones a knicking has been found between pairs of joints, which resulted in an arching of the layer.

Some concordant sheets of acid intrusive rocks found in the Yuso group must have been intruded early during the Asturian folding phase, because they show a cleavage development parallel with the slaty cleavage in the surrounding rocks.

Summarizing we find:

1. E—W folding with 30—40° N dipping axial planes, contemporaneous with a N—S cross folding, development of slaty cleavage.
2. Late E—W refolding, with nearly vertical axial planes, minor folding.

THE SOUTHERN BORDER ZONE

Subarea 8

This subarea consists of sediments of the Cea group as well as Cretaceous and Tertiary rocks. To the north the subarea is bordered by a system of faults with WNW—ESE and WSW—ENE traces.

The upheaval of the Cantabric-Asturian mountains with respect to the meseta of Old Castile along this set of faults, was mainly accomplished during Oligo-Miocene time.

The coal bearing strata of the Cea group, together with the Cretaceous and some of the Tertiary conglomerates were steepened and almost entirely inverted, due to drag and thrusting to the south of the rising mountain region. Further south the vertical Tertiary conglomerates suddenly flatten to a horizontal position. The sudden change in dip along a straight line, indicates a fault movement or flexuring.

The change in direction of the border faults from WNW—ESE to WSW—ENE is also expressed in the trend of the Hercynian isoclinal folds in the limestones of the Sierra del Brezo (cf. p. 187), as well as in the strike of the Cretaceous and Tertiary layers. In addition to these faults, large wrench-faults developed, with similar directions. The dextral Cotolorno wrench-fault which runs over Cotolorno mountain in a WNW direction, is closely related to a flexure-like fold found in the Cea group and the Cretaceous near Villanueva de la Peña and so can be dated as an Alpine movement. The same age probably can be accepted for the Valcovero wrench-fault, which cut off the late E—W refolding structure of the Valsurvio dome. The small folds found in the Cretaceous near Aviñante have a similar origin as the flexure-like fold near Villanueva de la Peña. Rather steep opposite plunges of the fold axes indicate that they are formed in layers which had already steep attitudes. They are due to a change in strike of the Cretaceous zone from E—W to NE—SW. The deformation remains restricted to the Upper Cretaceous only.

The vertical or steep north dipping strata of the Cea group form a continuous sequence in the eastern part of the area. West of Santibañez de la Peña they are concentrically folded. No cleavage has ever been found in these rocks, in strong contrast with the older sediments. Slate fragments in the graywackes of the Cea group show a pre-existing first slaty and second crenulation cleavage, which indicates that both cleavages developed before reworking during Stephanian time.

The fault contacts between the Cea and Ruesga groups are accompanied by crumpling and shearing of the shale layers, brecciation and red colouration of the limestones, development of local mylonite zones and segregation of iron oxides (hematite and limonite) in veins.

Further west of the Rio Carrion the Cea group extends to the north. Here the coal measures lie unconformably upon Lower Carboniferous and Upper Devonian rocks. The southern fault boundary is not clearly recognizable in the Cea group and possibly only flexure zones and small-scale faulting developed on the surface.

The slight unconformity between the Cea group and the Cretaceous was due to the Post-Stephanian—Pre-Triassic folding phase.

STRUCTURAL DETAILS

Knick folds and conjugate fold systems

Two simultaneous asymmetrical folds with opposing senses of asymmetry form a conjugate set (Johnson 1956). Where both sets are fully developed, the result is somewhat like a box-fold.

The phenomenon of conjugate fold systems is best developed in our area in the griotte, outcropping near the place where the Arroyo de Valderrianes flows out in the Pantano de Camporredondo and is only visible at times of low water level. These folds are developed in nearly horizontal, but inverted beds.

The development of these flat-topped anticlines (fig. 35A) starts as a small initial irregularity in the bedding plane which succeeds upwards as a small asymmetric fold sometimes even overturned, usually with an axial plane dipping to the north. Upwards, a new buckle develops on the long northerly dipping limb of the fold with an opposite dipping axial plane, whereas the overturned shorter limb is no longer overturned, but dips under 70°—80° S. Higher up the freshly developed buckle on the northern limb as well as the initial fold itself get more and more the character of straight knick zones, which are broken along the knick planes only in few cases.

All transitions occur between the angular knick zones and the curved asymmetrical folds. We will call them here "knick folds".

The knick folds with the northerly dipping knick planes have sence of movement downwards towards the south, whereas those with a southerly dipping knick plane have an opposite sense of movement. Thus, the conjugate fold system widens upwards. The two related knick zones vary in distance up to 2 m apart. The warping along the knick planes vary from a few to 30 cm.

A N—S directed stress system has caused this complementary set of shear movements, whereby the knick fold which has developed out of the initial fold is indicated by a S. of synthetic, whereas the knick fold which develops on the limb of the initial fold is indicated by an A. of antithetic in fig. 35.

In fig. 33 a detailed structural picture together with some sections is given of this deformation pattern, obtained from the bedding surface, exposed in a horizontal plane.

The following features can be observed:

1. The bedding plane is nearly horizontal and only disturbed by a few knick folds.
2. The axial directions of the knick folds agree with the dominant axial directions of the late E—W refolding.
3. Two directions of knick lines (intersections of the knick planes with the bedding planes) are predominant. One in direction N 245—255 E and another in the direction N 270—275 E.
4. The knick planes of the folds trending N 245—255 E dip N and S, whereas the knick lines trending N 270—275 E are mainly related to knick folds dipping S. Knick lines often swing from one into the other direction.
5. The short limb of a knick fold dips in the opposite direction to the knick plane of its own fold.
6. The knick lines of the anticlinal conjugate fold systems mostly diverge towards the west, whereby the southerly knick line has a direction of N 245—255 E and the northerly knick line the direction of N 270—275 E.

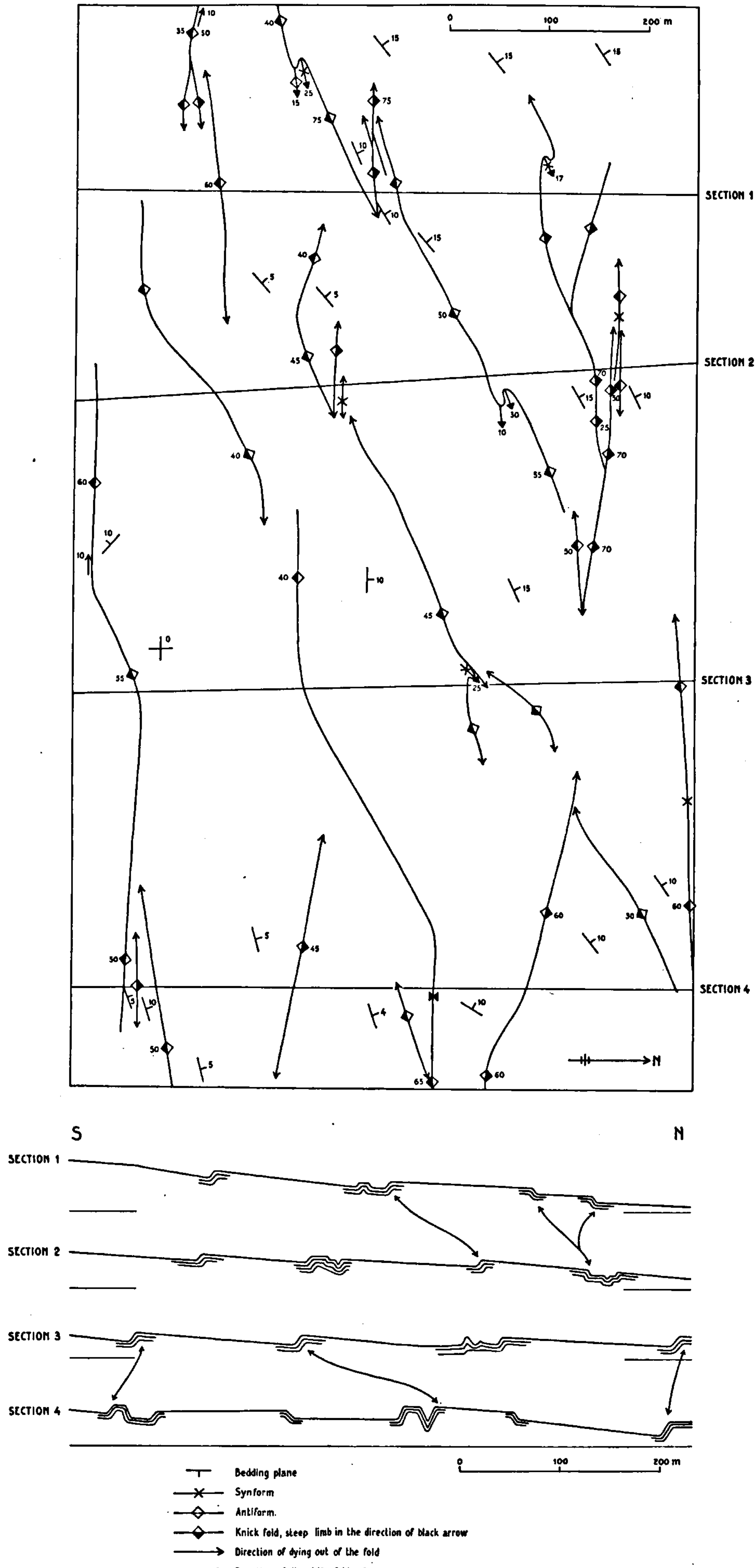


Fig. 33 Trend of the minor folds in the griotte exposure near the mouth of the Arroyo de Valderrianes in the Pantano de Camporredondo.

7. Knick folds usually have a vertical extension merely over a few layers, with a total thickness of 1—2 m. Their lateral extension in one and the same bedding plane is about 10 metres, after which they die out (indicated in fig. 33 by an open arrow; an abrupt end of the knick lines indicates that the knick line could not be followed by lack of exposure in the bedding surface).

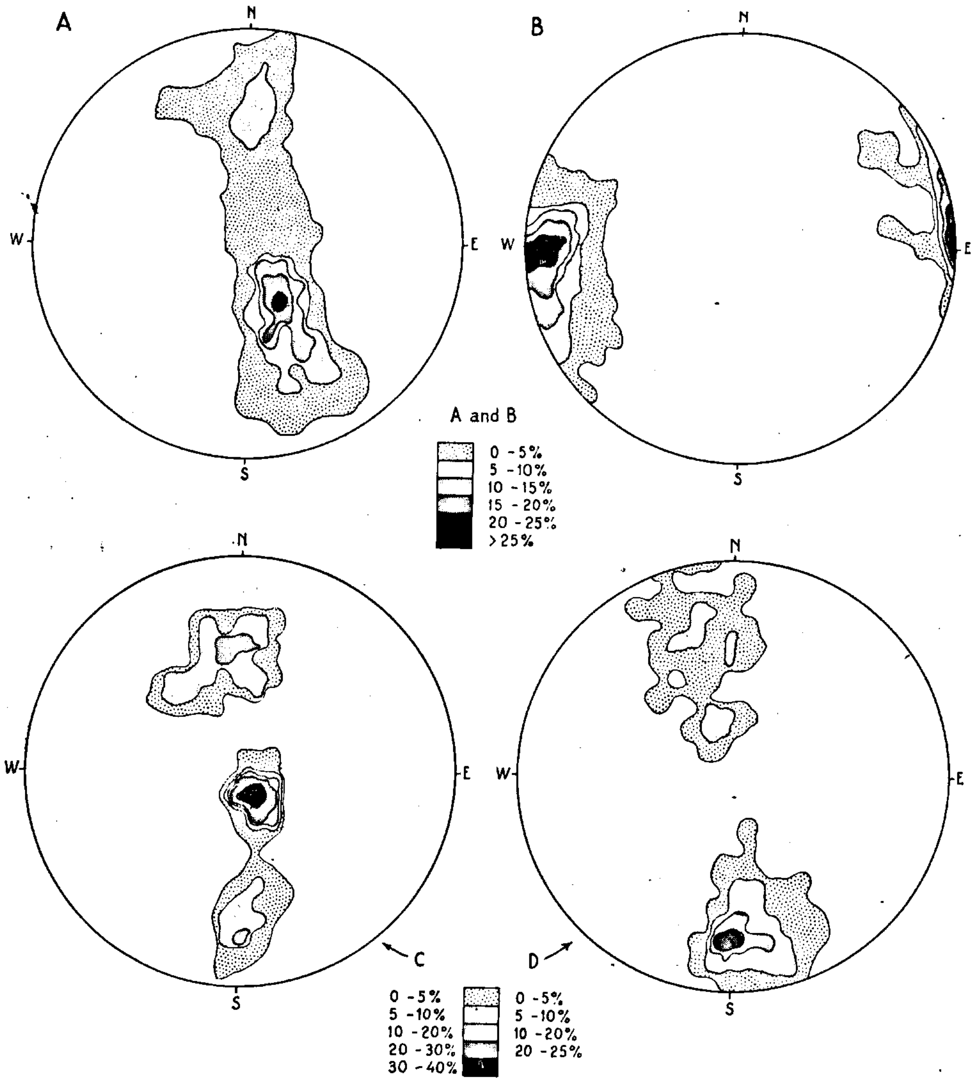


Fig. 34 Stereographic projections of minor folds of the griotte exposure near the mouth of the Arroyo de Valderrianes in the Pantano de Camporredondo.
A. poles of bedding planes of minor folds
B. fold axes of minor folds
C. poles of bedding planes of the conjugate fold systems
D. poles of the axial planes of the conjugate fold systems
Contour values in percentages per 1 % area.

8. At some places small secondary minor folds, which plunge 15—30° E, give rise to irregularities in the knick lines. A simultaneous development of the knick folds with these small folds is assumed (fig. 35 D).

The geometry of these conjugate fold systems is shown in a stereogram of bedding plane poles (fig. 34 C). It shows three fields of concentration, forming a nearly complete girdle. The concentration at the centre of the diagram represents the nearly horizontal top layers of the conjugate fold system. The smaller concentrations in the girdle represent the short steep limbs, dipping respectively north and south. The poles of the knick planes are plotted on diagram D, which shows two maxima, one with a rather high concentration, which is related to the knick folds with a downthrow to the south and another is related to the knick folds with a downthrow to the north.

The plunge and direction of the knick lines are plotted, together with fold axes of the disharmonic folding measured from the same outcrop, in diagram B. The two directions could be clearly distinguished in the conjugate fold systems, but can not be detected in the compound diagram D, which shows only an elongated maximum.

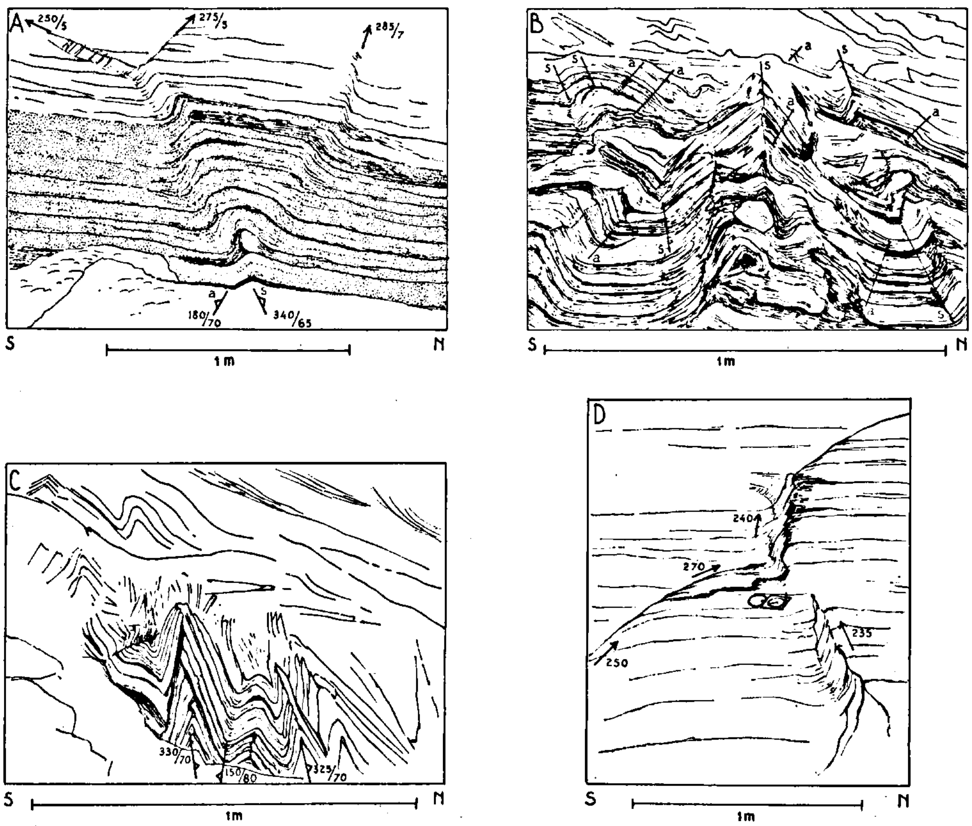


Fig. 35 Different fold types found in the griotte exposure.
A. conjugate fold system
B. irregular conjugate sets of folds
C. accordion folds
D. small contemporaneous cross folds causing an irregularity in the knick line. (s indicates axial planes dipping N, a indicates axial planes dipping S).

In addition to these conjugate fold systems other different fold types and fold shapes are visible in this exposure (fig. 35).

In the southern part of the griotte outcrop, single knick folds occur with knick planes dipping 65° S (fig. 36). The horizontal foliation is rotated to a dip of 40° N between the two knick planes. The wave length of these knick folds is about 30 cm. The knick lines are remarkable straight.

These single knick folds and the conjugate knick fold systems, described above, are developed as regular wide spaced minor structures. In the same outcrop, a very intensive and irregular minor folding pattern can also be observed (fig. 35 C). Fold orientations occur, which have axial planes parallel

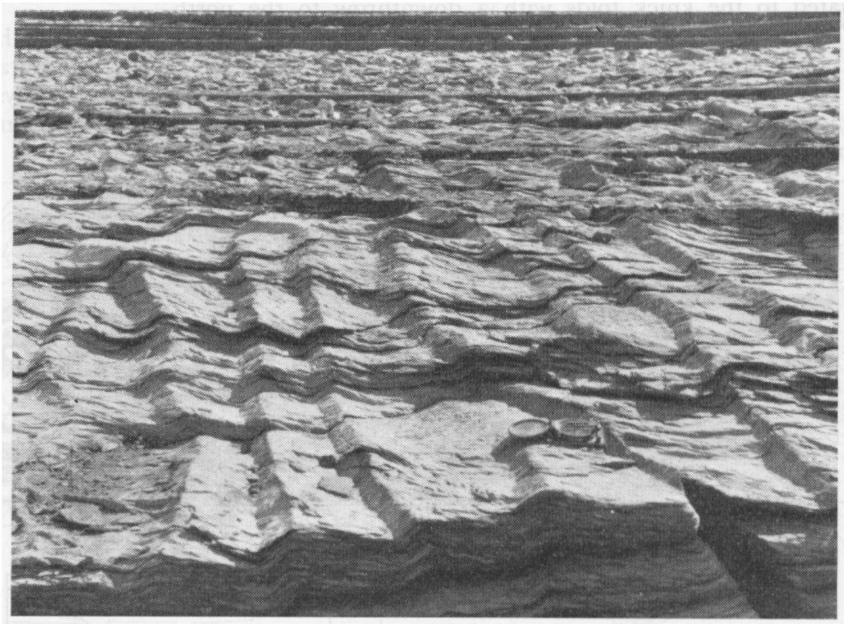


Fig. 36 Regular system of knick folds, griotte exposure.

to those of the conjugate fold systems. Axial planes are often curved. The disharmonic character of the folds means that knick folds, shear folds, accordion folds and concentric folds are found next to one another.

Axial plane cleavages are not developed. The straight limbs are sometimes sheared and thinned with respect to the thickness of the layers in the hinges.

In this brief description of the minor structures of the griotte exposure, we have seen that:

1. Different fold shapes, as disharmonic folds, conjugate fold systems, knick folds etc. occur in a small outcrop, in the same rocks in which no obvious lithological differences occur.
2. Fold directions of the minor folds diverge over an angle of 60°.
3. Fold directions making an angle of 30° with one another, occur in one and the same conjugate fold system.

4. Axial planes with a uniform strike often show a different dip direction, to the south as well as to the north.
5. Small irregular folds occur, crossing the knick folds, which plunge 30° E. Both folds are simultaneous developed.

It is thought that this fold pattern developed in one folding phase. The two orientations of the knick planes are to be regarded as a conjugate set of oblique shear folds, also the dispersion of the fold axes is due to this conjugate set.

The irregular folds plunging 20—30° E are related to the knick zones and are never found on the flat limbs.

All these variations clearly shown as minor structures in this griotte exposure, also occur as crenulation cleavage and related lineations in the shales and siltstones. Two lineation directions are observed making an angle of 30° with each other in the siltstones. In thin section they appear to be connected with each other, forming a conjugate set. In diagram R (fig. 20) these two directions can be observed. As those micro crenulations are superimposed on N—S directed minor folds, their poles are orientated on two girdles. The thick dots representing the poles of the cleavage planes, dipping to the south as well as to the north, show a same picture as on diagram D of fig. 34.

So we must be very careful with our conclusions, when we encounter two lineation directions and different attitudes of the cleavage planes, whether we have two sets of surfaces, inclined to one another and to the local fold axial plane, developed in one folding phase, or two separated generations of micro folds.

The Arroyo de Miranda exposure

At low water level where the Arroyo de Miranda flows out into the Pantano de Camporredondo, an exposure of strongly folded sandy siltstones and shales is visible.

The age of these rocks is not certain because of the lack of fossils; only some indeterminable plant debris could be found in the siltstones. The rock has been regarded as Yuso group here, because of its position with respect to the Devonian.

The exposure is a vertical wall of 4 to 5 m height and some 100 m long. The plane of exposure runs N—S perpendicular to the fold axes. A detailed section has been made of the most northern part of the outcrop (fig. 37) on a scale 1 : 25, composed of field measurements, sketches and photo interpretations. Not all the layers are worked out in detail in this section. The inset figures are drawn from thin sections of orientated samples. The sandy siltstones are concentrically folded with a tendency to accordion folding whereas the interbedded shale layers have reacted in an incompetent way and show an intensive micro folding, shearing and a development of crenulation and fracture cleavage, with the result that the folding is disharmonic. The average thickness of layers of the sandy siltstones is about 2—2.5 cm. The shales have a fine fissility and the amplitude of the folds is very variable.

The macro folding is not at all clear, as we can only follow this formation over a very short distance towards the west. Further north it is cut off by a fault which runs in easterly direction into the reservoir.

The minor folds on the contrary are beautifully exposed, their wave length

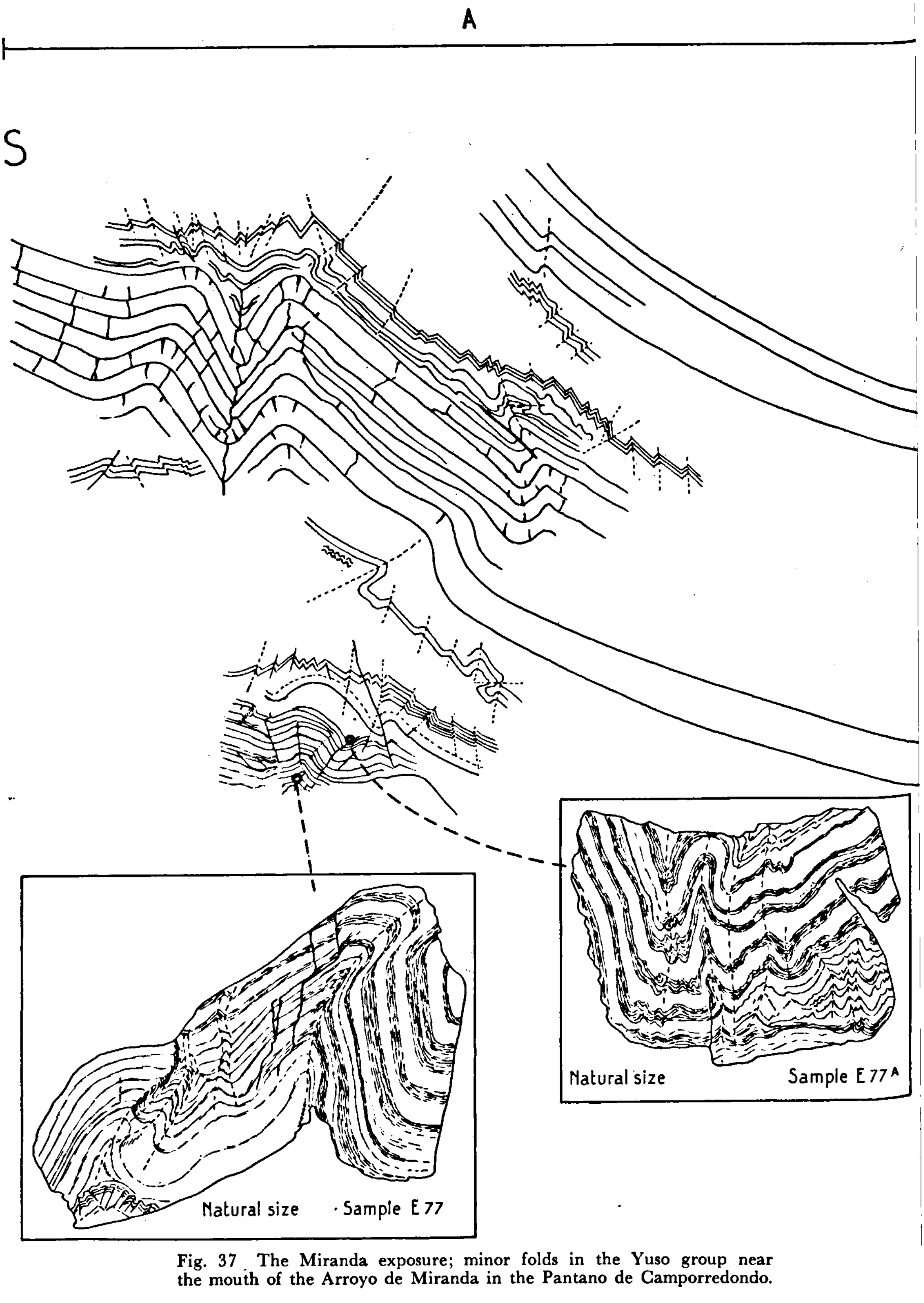
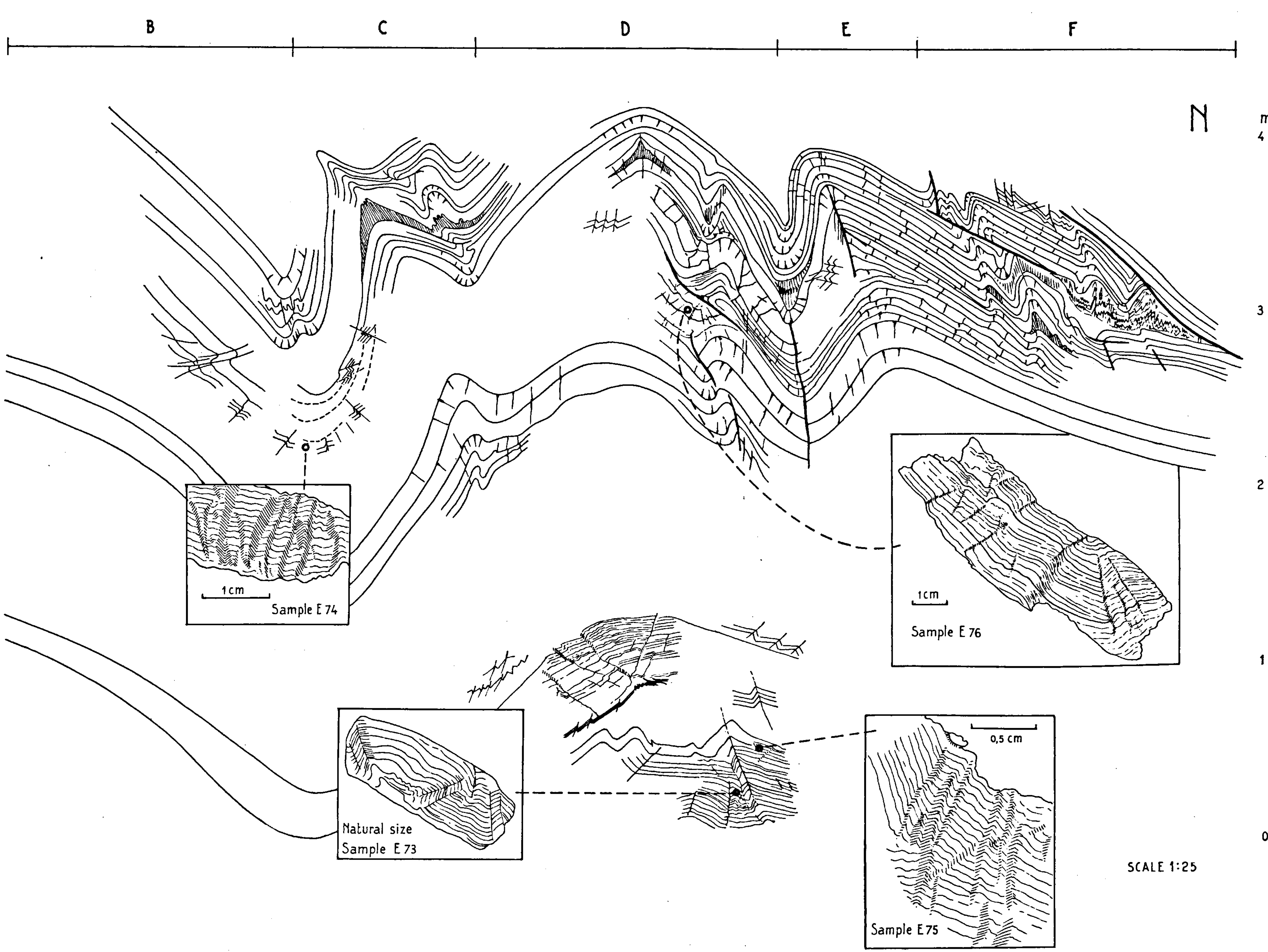


Fig. 37 The Miranda exposure; minor folds in the Yuso group near the mouth of the Arroyo de Miranda in the Pantano de Camporredondo.



B
C
D
E
F
N
m
4
3
2
1
0
1cm
Sample E 74
1cm
Sample E 76
Natural size
Sample E 73
0,5 cm
Sample E 75
SCALE 1:25

varies from 10 to 150 cm. Next to these folds we can recognize micro folds with an amplitude from 0.05 to 2 cm.

The fold axes plunge at 10° to N 65° E and N 110° E. The axial planes of the minor folds are vertical or dip steeply N. The orientation of the axial planes of the micro folds shows a relation to the orientation of the bedding on which they are developed. The relation between minor and micro folds will be referred to later. The minor folds have been named A, B, C, D, E and F from south to north respectively (see top fig. 37). Generally the minor folds are asymmetric with a long northern limb and a steep shorter southern limb (E and F). The anticline D is asymmetrical with a vertical axial plane. Between the synclinal trough of B and the anticlinal crest of A there is a long undisturbed limb. Most of the axial planes are slightly curved. The asymmetric folds have often closely associated thrusts.

Fold D shows different fault types merging into one another in a vertical sense. In the syncline between D and E and in the small anticline on the north flank of D at the relative altitude of 2 m, small crest faults are developed parallel to the axial plane, dying out upwards. A steep thrust fault is formed in the upper prolongation of the latter also dying out quickly after a 20 cm. Higher up in the sequence a low-angle overthrust is visible, parallel with the northern limb of the anticline, and dying out to both sides in the bedding plane. When we follow the axial plane again somewhat higher, no faulting is observed, but here we find some incompetent shale layers between the competent sandy siltstones. The deformation of these layers which shows a similar micro folding with cleavage development, takes over the function of the faults, to compensate the lack of space which occurs in the core of a concentrical folded anticline.

Tension joints are obviously related geometrically to individual folds. Stretching of the outer side of the layer in the flexure caused radially arranged joints, often opening upwards in the anticlinal crests, and downwards in the synclinal troughs, forming open tension cracks (D and E). In the core of the fold the compression develops crumpling, faulting and possible diapiric movement.

A slaty cleavage is developed parallel to the bedding. Examination of thin sections has shown that the cleavage can be traced around the hinges of the minor folds. A polishing and slickensiding of the bedding surfaces has been found, the direction of the slickensides being perpendicular to the B axis of the concentric folds. The deduction seems warranted that during the folding the bedding plane has acted as plane of deformation and recrystallization occurred.

When we consider fig. 37 it becomes evident that the micro folding shows a clear relation to the concentric minor folding and that both are contemporaneous. In most cases the micro folds occur in the incompetent layers as parasitic folds of the minor folds.

When we consider the anticline D at an altitude of about one metre in the section, we can distinguish two sets of parasitic folds.

1. One system showing a strong convergence of the axial planes upwards to the hinge of the anticline.
2. The second system with a divergence of the axial planes upwards to the hinge of the same fold.

The first system occurs in relatively incompetent thinly foliated beds. On the southern limb of the anticline asymmetric micro folds are developed with axial planes dipping 70° S and thrusted from south to north. On the hinge of the anticline symmetrical Z shaped micro folds with vertical axial planes exist (Crestal folds, de Sitter 1958). On the northern limb, dipping 15° N, the micro folds have an axial plane dipping 65° N. On the hinge only the lateral stress direction is responsible for the development of the crestal folds and local cleavages. On the limbs a frictional drag is superimposed which results in a convergence of the axial planes of the micro folds upwards to the top of the anticline.

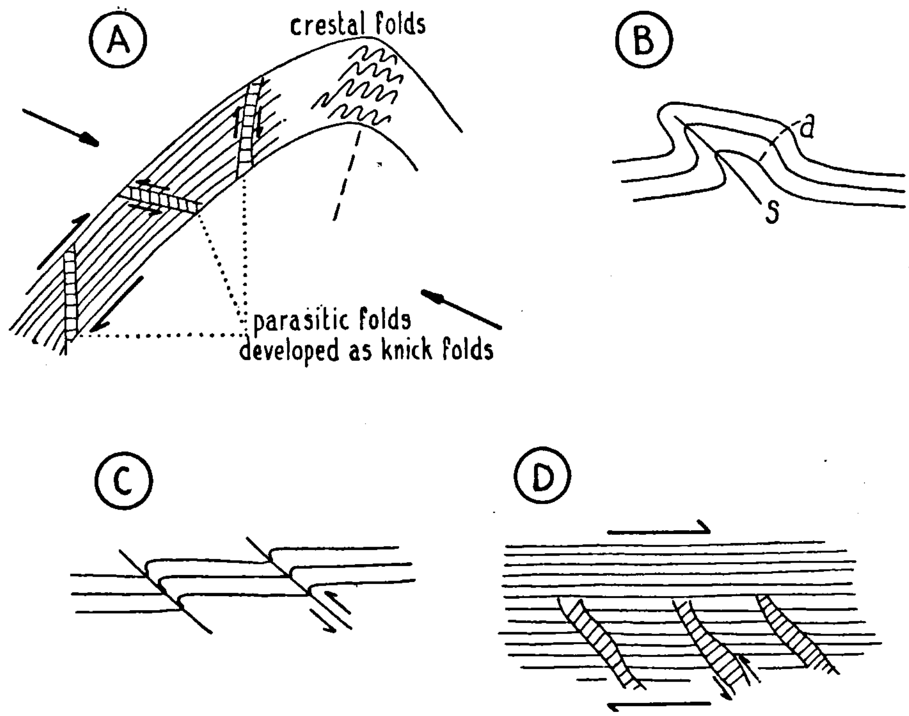


Fig. 38 A. Schematical sketch of parasitic micro folds found in a minor fold
B. Boxfold developed as a conjugate set of folds
C. Down dip folds with a shear direction along the bedding planes as indicated in D
D. Down dip knick zones
The direction of shear of the down dip structures C and D is due to an opposite couple, "antithetische Flexuren", (C and D after Hoeppener 1956 Abb. 21).

The second system is developed in more sandy siltstone layers immediately above the shale layers in which the first system occurs and generated by an opposite couple (fig. 38 A). On the northern limb of the anticline the axial planes of the micro folds dip about 60° S, meanwhile on the southern limb the axial planes dip 65° N. The sense of movement is complementary.

The micro folds often show a great resemblance to knick zones, a pair of joint planes enclosing a narrow zone in which the foliation is knicked or flexured. The two joints might run into each other, and thus forming one joint. They can die out and the deformation continues as a monoclinal flexure or S shaped micro fold. In the second system the relative movement is reversed to the direction of slip, up the dip of the beds of the minor fold. These folds are known in the English literature as "down dip folds" (amongst others Nevin 1942, Wilson 1950).

In the syncline B, at a relative height of about 2.50 m, down dip folds are also visible. On the southern limb even with horizontal thrusting and relative to the bedding in a down dip direction. The axial planes or knick planes on the northern limb vary from 25° to 70°. Their attitude is related to the attitude of the bedding plane on which they are developed.

Both sets of micro folds can occur together in one and the same layer, cross cutting each other as a conjugate set of two related directions. (Drawings of thin sections samples E 74, 75, 76). The two directions are also developed in the axial planes of the minor folds. In anticline A several folds are developed with a southern dipping axial plane as well as folds with a northern dipping axial plane. The same picture has already been described in the development of the conjugate fold systems in the griotte.

Hoeppener (1956 fig. 21) shows a correspondance of "Knitterung id. Knickzonen" with "Flexurfalten".

He compares the synthetic system with drag folds (Schleppfalten) in the direction of the bedding plane slip, up the dip towards the hinge. The antithetic system is on the contrary down dip; "Die Verschiebungsrichtung auf den S-Flächen verlauft entgegengesetzt zu der, die für Schlepp- und Gleitfaltung gefordert wird" (fig. 38 C and D) This picture given by Hoeppener appears to bear a strong resemblance to the two systems developed in anticline D, just described.

In a box fold the synthetic curvature is often more strongly developed than the contemporaneous formed antithetic curvature, resulting in a special type of fold (fig. 38 B). A recent publication by M. Lindström (1960) on tectonic structures of Scania (Sweden), calls attention to such folds. He concludes that this special type of fold is an effect to the combination of deformations: p. 324 "In these folds the short SW limb is sharply overturned towards SW, and is due to the first deformation. The long NE limb has, however, by a later deformation been thrown into a hump shaped rather smooth subsidiary fold, the axial plane of which is subvertical with a tendency to be overturned towards the NE".

In the Miranda exposure the minor folds as well as the micro folds form an interfering set. The close relationship between the attitude of the axial or knick planes of the micro folds and the attitude of the bedding in the minor folds on which they are developed, indicates their simultaneous formation. Closely spaced micro folds in incompetent beds often pass into an axial plane crenulation cleavage in the hinges.

SUMMARY AND CONCLUSIONS

As shown in the preceding pages, three structural units can be distinguished in our area. Several folding phases have effected different types of deformation in these structural units, according to the lithological characteristics of the outcropping rocks.

In the present area the following sequence of events has been worked out:

A. Upheaval of the zone of San Martin-Camporredondo.

The sedimentary history during the Devonian was strongly influenced by the vertical upheaval of this E—W trending instable zone, probably along a set of fault lines or sharp flexures.

Uplift in both the Middle and Upper Devonian have caused two hiatuses in the zone of San Martin-Camporredondo, which correspond with 900—1000 m of sediments in the Valsurvio dome (see chapter III).

B. Subsidence of the zone of San Martin-Camporredondo.

During sedimentation of the Ruesga group the instable zone subsided, resulting in great lithological differences. South of this zone, in the Valsurvio dome, there are thick massive limestones, whereas in the E—W zone itself clastic and incompetent pelitic sediments occur. The exact thickness of these sequences could not be determined. The changes are very rapid, suggesting movements along faults.

C. First generation structures.

The lithological differences caused by the subsidence and upheaval of the zone of San Martin-Camporredondo have determined the mode of deformation in the various areas.

The Middle and Lower Devonian slate-limestone-sandstone alternation is intensively folded in E—W recumbent isoclinal folds, with a nearly horizontal slaty cleavage. This cleavage is almost parallel to the bedding on the limbs and only cross cutting in the hinges. The thick competent Upper Devonian quartz sandstones together with the massive limestones of the Ruesga group are folded in large E—W trending isoclinal folds with axial planes dipping from 60° S, down to nearly horizontal and without a cleavage development.

When passing from the Valsurvio dome into the zone of San Martin-Camporredondo mode of deformation changes with the lithological change from large isoclinal folds into low-angle overthrusts, which have ESE—WNW trends and are thrusted towards the north over the incompetent Lower Carboniferous griotte horizon. The maximum measured lateral displacement is about 2.5 km.

D. Second generation structures.

In the Middle and Lower Devonian isoclinal folds have developed with an axial plane, fracture or crenulation cleavage, which deforms the first generation slaty cleavage. The attitude of the axial planes is rather flat lying. First and second generation folds are often difficult to separate in the field. Near Besande NE—SW cross folding has been found deforming the first generation folds.

In the zone of San Martin-Camporredondo the second deformation has caused large asymmetrical folds with E—W trends and axial planes dipping 30°—40° N. These folds deform the low-angle thrust planes and other first generation structures. Contemporaneous with this E—W refolding, N—S directed minor folds occur; both have a same attitude of their axial planes.

The N—S fold direction has also been found in the Yuso group, developed as minor folds on the limbs of large contemporaneous formed isoclinal E—W folds, which dip about 35° N. Sometimes, the slaty cleavage can be seen cutting through the bedding on the hinges of the tight recumbent minor folds. This first deformation of the Yuso group corresponds to the second deformation in the Devonian and Ruesga group.

In the core of the Valsurvio dome, between the Otero and the Hornalejo anticlines, a N—S cross fold has also been found, the Valsurvio syncline, which appears older than the third generation structures. A similar age as the second generation folds is implied, but could not be proved.

E. Third generation structures.

The present day geometry of the Valsurvio dome has been formed by the third deformation. The Otero and Hornalejo anticlines both broad open E—W structures, with nearly vertical axial planes and respectively plunging to W and E, form together the Valsurvio dome. First and second generation structures have been deformed by this late E—W refolding. These broad open structures can be traced into the Cea basin (Tejerina syncline, Besande anticline, Cueto syncline).

In the zone of San Martin-Camporredondo only small scale third generation structures have been found, described in detail from the griotte and the "Miranda" exposures. A concentric minor folding with a conjugate set of parasitic micro folds often occurs. Crenulation cleavage or widely spaced micro knick zones also frequently form conjugate sets, resulting in two sets of contemporaneous lineations at an angle of 0—30° to each other.

F. Tertiary movements.

Some small epirogenetic movements during Eocene were followed in the Oligo-Miocene by the large upheaval of the Cantabric-Asturian mountains. Together with this uplift a set of border faults have been formed and a narrow E—W zone has been up-ended and overturned. Large WNW—ESE trending wrench-faults were formed. The Cotolorno wrench-fault show a clear connection with a flexure-like fold, developed in the Cretaceous.

From previous papers (Kanis 1955, Wagner 1955 and de Sitter 1959, 1961) we know that at least three folding periods could be distinguished in the Cantabric-Asturian mountains. De Sitter (1961) gives the following tectonic succession:

4. Saalic phase	Post-Stephanian—Pre-Triassic	E—W folding
3. Asturic phase	b. Stephanian A—B a. Westphalian D—Stephanian A	NNW and NNE cross folding
2. Curavacas phase	b. Westphalian A—B a. (Post-Visean—Westphalian A)	E—W isoclinal folds and nappe structures
1. Bretonic phase	Famennian	tilting of the Leonides

The upheaval of the zone of San Martin-Camporredondo during Upper Devonian can be correlated with the tilting of a block in the Esla-Bernesga region during Famennian (cf. chapter III).

The first generation structures are unconformably overlain by the Yuso group near the Arroyo de Valderrianes and thus belong to the Curavacas folding phase. Similar low-angle overthrusts, but of a much larger scale, are found in the Esla-Bernesga region (de Sitter 1959). The thrust movement in both areas is to the N or NE.

The second generation structures are thought to belong to the Asturian folding phase because:

1. Cleavage developed throughout the Yuso group has never been found in the Cea group.
2. The slate fragments probably derived from the Devonian found in the graywackes of the Cea group show pre-erosion, slaty and crenulation cleavages, which indicate that both must have been developed before sedimentation of the Cea group, the first in the Curavacas folding and the second during the Asturian folding phase.
3. In the Cea group only E—W folds with vertical axial planes have been found, whereas in the Yuso group E—W folds with vertical axial planes are superimposed on isoclinal folds with rather flat N dipping axial planes, the latter must be formed during Asturian folding.
4. Small chloritoid crystals with random orientation, crystallized after the development of the slaty cleavage of the Yuso group, have been found in the Devonian, the Ruesga and Yuso groups, but never in the Cea group. This might indicate that the slaty cleavage as well as the chloritoid phenocrysts were formed before the sedimentation of the Cea group. However, not enough is yet known of the distribution of these chloritoid crystals in adjacent areas.
5. Similar approximately N—S cross folds are also known from the Asturian phase in the adjacent areas, as pointed out by de Sitter (1961).

A clear angular unconformity between Yuso and Cea group has not been found, but is known from the Tejerina syncline further west. The fossil flora of the Cea group indicates a pre-Stephanian age for this folding. However, dating is not well established yet. De Sitter (p. 58, 1961) writes: "En el estado actual de conocimientos de las discordancias en la tectónica astúrica no sabemos si hay una fase principal con subfaces, o bien si una determinada fase predomina en una cierta área y otra en algún otro sitio".

The third generation structures deformed the Guardo-Cervera coal basin and can be dated as Post-Stephanian A. If this folding phase belongs to the Saalic phase (Post-Stephanian—Pre-Triassic) could not be proved in our area.

SUMARIO

Geomorfología

La zona de que se trata aquí está situada en las laderas meridionales de las montañas cantábrico-asturianas, cuyo mapa ha sido trazado a escala de 1 : 50.000.

· A lo largo del Río Carrión cuatro niveles de acumulazión han sido encontradas, en las alturas relativas de 100 m, 55 m, 30 m y 1—15 m sobre el nivel del río, que, a imitación de Mabesoone (1959), han sido llamados respectivamente las terrazas T_1, T_3, T_4 y T_5. En el Monte Orvillo se encontraron algunos restos glaciales en una altura de 1730—1800 m. Los depósitos aluviales de pie de monte, situados al pie de la Sierra del Brezo, han sido formados como "sheet flood" depósitos, el establecimiento de cuya edad queda discutible.

Estratigrafía

El espesor de las capas devónicas en el domo de Valsurvio es de 1000—1200 m. Se distinguen cinco formaciones, llamadas por orden de edad, empezándose por la más vieja: las formaciones de Compuerto, Otero, Hornalejo, Valcovero y Camporredondo.

Las formaciones de Compuerto y Hornalejo constan ambas principalmente de pizarras oscuras, en la que difícilmente puede distinguirse una estratificación. Están separadas por las calizas de Otero, de estratificación espesa con muchos corales y estromatoporoideos. Algunas veces se han desarrollado en esta formación pequeños arrecifes, como por ejemplo en el Arroyo Abianos. En el lindero entre las formaciones de Compuerto y Otero, se ha encontrado una tierre roja de suelo fósil al oeste del Río Carrión, lo que indica una interrupción de la sedimentación. Una segunda capa de caliza se encuentra encima de la formación de Hornalejo. Cerca del pueblo de Valcovero, la formación de Valcovero consta de calizas macizas de un espesor de 145 m. Hacia el nordeste, la facies cambia y los 80 m superiores se han desarrollado como lutitas negras, con una fauna abundante de fósiles de tamaño uniforme, depositados en cayos, en un ambiente probablemente litoral. Las areniscas cuarzosas de Camporredondo tienen un espesor de 300—500 m.

A base de determinaciones de los braquiópodos y las características lithológicas, se estableció una correlación con el Devónico descrito por Comte (1959), de la región del Esla en la provincia de León.

El Carbonífero se ha subdividido en tres grupos, los de Ruesga, Yuso y Cea, separados unos de otros por discordancias angulares. El grupo de Ruesgo se ha desarrollado en parte en la facies "Caliza de Montaña" como calizas macizas, y en parte en la facies "Culm", como una sucesión espesa de pizarras, areniscas, sub-grauvacas y conglomerados polimixtos.

El grupo de Yuso, que se encuentra encima del grupo de Ruesga y sedimentos de más edad, con discordancia angular, es, por sus características lithológicas, muy semejante a la facies de Culm. El grupo de Cea, que consta de sedimentos

de menos edad que la fase del plegamiento asturiano, aflora en una zona estrecha situada al pie sur de la Sierra del Brezo. Es una sucesión de capaz de un espesor mínimo de 1100 m en que se encuentran yacimientos de carbón. La parte superior de la sucesión no se conoce, porque hacia el sur está cubierta de forma discordante por sedimentos cretáceos y terciarios.

Las facies y la infuencia de sucesos tectónicos sobre la historia sedimentaria

Una correlación del Devónico con las zonas adyacentes enseña un rápido cambio lateral de la facies en dirección norte, quedando, en cambio, la facies bastante constante en dirección este-oeste. Entre las facies norte (Rio Arruz, Kullmann 1960) y sur (domo de Valsurvio) debe de haber estado un levantamiento estrecho y alargado, que separaba las dos zonas de características tanto lithológicas como paleo-ecológicas distintas. Cambios de las facies, conjutamente con un repentino adelgazamiento de las capas y la presencia de dos importantes interrupciones estratigráficas en el Devónico medio y superior respectivamente, de esta zona de San Martin - Camporredondo demostraron la existencia de este levantamiento de dirección este-oeste. El espesor del Devónico del domo de Valsurvio de unos 1100 m dismuye cerca de Triollo hasta unos 50 m. Aquí areniscas cuarzosas del Devónico superior cubren directamente calizas de edad Siegenense.

Se observa el fenómeno interresante de que esta faja inestable se ha desarrollado durante el Carbonífero como una "cubeta". Las calizas macizas de la facies Caliza de Montaña, situadas a los dos lados de esta cubeta, han sido depositadas en una plataforma bastante estable; las rocas pelíticas y clásticas de la depresión, en cambio, dejan ver muchas pruebas de una re-deposición del sedimento y un deslizamiento de material calcáreo desde los dos lados hacia la depresión. El papel desempeñado por esta zona durante la sedimentación del grupo de Yuso es menos evidente.

Estructura

Se distinguen tres unidades estructurales:

1. El domo de Valsurvio
2. La zona de San Martín - Camporredondo
3. La zona de fallas al límite sur,

las que han sido subdivididas, según sus características tanto litológicas como estructurales, en ocho sub-zonas. Se han encontrado a lo menos cuatro ciclos de deformaciones.

En el Devónico del domo de Valsurvio (sub-zonas 3 y 4), las deformaciones primera y segunda han causado pliegues isoclinales casi horizontales de menor tamaño, con una equistosidad de plano axial y un clivaje de corrugamiento de plano axial respectivamente. Los dos han sido deformados por el repliegue tardío de dirección este-oeste, que tiene planos axiales casi verticales y que forma dos anticlinales anchos; el anticlinal de Otero y el de Hornalejo, que están separados uno de otro por un pliegue cruzado anterior de dirección norte-sur, el sinclinal de Valsurvio. Proyección estereográfica deja ver los buzamientos de los clivajes primero y segundo en los flancos y la "nariz" del anticlinal de Otero, que se curvan sobre la "charnela" de este pliegue. Las estructuras volcadas del segundo ciclo, halladas cerca del Arroyo Abianos, y que están completamente invertidas, también indican un replegamiento posterior.

Cerca de Besande, encontramos el mismo replegamiento (sub-zone 2). El anticlinal de Besande de dirección este-oeste, con plano axial casi vertical, deforma los pliegues isoclinales de buzamiento 60° S, producidos en las calizas carboníferas del flanco sur del domo de Valsurvio. El anticlinal de Besande está estrechamente relacionado con los sinclinales de Tejerina y Cueto, que se presentan al norte y al sur repectivamente en la zona que linda al oeste. Las dos estructuras sinclinales deforman los sedimentos del grupo de Cea y, por consiguiente, deben haberse producido en una fase de plegamiento, datada como Post-Estefaniense. En otras partes, este plegamiento ha sido datado como Post-Estefaniense—Pre-Triásico (de Sitter 1961).

Cuando se pasa del domo de Valsurvio a la zona de San Martin-Camporredondo (sub-zonas 5 y 6), la litología cambia a corta distancia y, conjuntamente, el tipo de deformación. El repentino acuñamiento de las areniscas cuarzosas de Camporredondo, conjuntamente con la conversión S—N de la facies "Caliza de Montaña" en la de "Culm", dió en el norte origen a corrimientos de ángulo bajo, contrastando con lo que se encuentra en el sur, a saber los grandes pliegues isoclinales con buzamiento hacia el mediodía. Estas cobijaduras casi horizontales, empujadas de aproximadamente sur a norte, con rumbo ONO—ESE, han sido replegadas en pliegues asimétricos E—O, cuyos planos axiales tienen un buzamiento norte. Simultáneamente con estos pliegues E—O, se han desarrollado pequeños pliegues menores con ejes en dirección N—S. Especialmente en las pizarras se ha encontrado localmente un replegamiento tardío en dirección E—O, con planos axiales verticales, y, frecuentemente, con el desarrollo de un corrugamiento de un clivaje pre-existente. Estos últimos dos tipos de deformación se han hallado también en los sedimentos del grupo de Yuso (sub-zone 7) que cubren de forma discordante las cobijaduras de ángulo bajo, desarrolladas en el grupo de Ruesga.

La zona de fallas del límite sur (sub-zone 8) consta de sedimentos del grupo carbonífero de Cea, el Cretáceo y el Terciario, que todos están en posición invertida. Con el levantamiento de las montañas cantábrico-asturianas en el Terciario, se produjeron fallas con rumbos ONO—ESE y OSO—ENE, a lo largo del cual se ha levantado la parte montañosa. Los cambios del rumbo de esas fallas están relacionados con el cambio de dirección de los planos axiales de los pliegues isoclinales del grupo de Ruesga, y también con semejante cambio de la inclinación de los estratos del grupo de Cea y los conglomerados cretáceos y terciarios. La falla de Cotolorno, de tipo "dextral wrench-fold", está relacionada con un pliegue o flexión del grupo de Cea y el cretáceo cerca de Villanueva de la Peña, y pudo, por consiguiente, ser datada como Post-Cretáceo. La influencia de la deformación terciaria, sin embargo, queda limitada.

En resumen, encontramos las siguientes fases de deformación:

1. La fase de plegamiento de Curavacas:
   Cobijaduras de ángulo bajo, con rumbo ONO—ESE, desarrolladas en la zona de San Martín-Camporredondo.
   Plegamiento isoclinal E—O con planos axiales de buzamiento hacia el S, desarrollado especialmente en el domo de Valsurvio.

2. La fase de plegamiento asturiano:
   Replegamiento en dirección E—O, con planos axiales de leve buzamiento hacia el norte, simultáneamente con un plegamiento cruzado en dirección N—S, desarrollado localmente.

3. La fase post-Estefaniense—pre-Triásica:
Replegamiento tardío en dirección E—O con planos axiales verticales. Formación del domo de Valsurvio.
4. La fase de deformación terciaria:
Desarrollo de las fallas del límite sur y el empinamiento de una estrecha zona transitoria entre la meseta y la cordillera. Formación de grandes fallas de tipo "wrench-faults" con rumbo E—O o ESE—ONO.

Una descripción detallada de dos afloramientos cerca del Pantano de Camporredondo de algunas particularidades de "conjugate fold systems", producidos tanto en escala menor como en micro-escala.

Coordinate references of samples and fossil localities described in the text:

E 2	1°06'16" — 42°53'40"	K 119	1°01'10" — 42°52'46"
E 6	1°06'02" — 42°52'29"	K 127	1°03'09" — 42°50'50"
E 7	1°13'47" — 42°53'24"	K 130	1°01'08" — 42°52'44"
E 19	1°07'37" — 42°52'23"	K 139	1°02'45" — 42°48'56"
E 20	1°07'34" — 42°52'24"	K 149	1°02'38" — 42°48'46"
E 23	1°07'33" — 42°52'25"	K 171	1°02'57" — 42°48'27"
E 35	1°06'58" — 42°52'46"	K 181	1°02'59" — 42°47'46"
E 41	1°06'47" — 42°53'02"	K 182	1°02'55" — 42°47'43"
E 59	1°04'12" — 42°53'32"	K 183	1°02'54" — 42°47'42"
E 78	1°00'48" — 42°54'08"	K 186	1°07'36" — 42°48'10"
E 79	1°00'48" — 42°54'08"	K 192	0°59'20" — 42°49'29"
E 81	1°03'26" — 42°54'16"	K 205	1°05'48" — 42°52'32"
E 85	1°00'03" — 42°54'54"	K 216	1°01'10" — 42°52'41"
E 94	1°01'09" — 42°52'46"	K 217	1°12'38" — 42°52'09"
E 100	1°04'15" — 42°53'25"	QC	0°58'49" — 42°50'27"
E 101	0°56'55" — 42°53'57"	QD	0°58'36" — 42°50'10"
E 105	1°03'55" — 42°54'11"	QU	0°55'52" — 42°49'40"
E 106	1°03'58" — 42°54'10"	Q 37	0°58'49" — 42°50'27"
E 108	1°00'05" — 42°54'23"	Q 72	1°02'25" — 42°52'09"
E 111	1°01'10" — 42°54'14"	Q 75	1°02'14" — 42°52'02"
E 114	1°12'49" — 42°54'06"	Q 78	1°02'07" — 42°52'27"
K 9	1°05'24" — 42°49'25"	Q 84	1°02'10" — 42°52'51"
K 96	1°02'36" — 42°48'47"	Q 115	1°00'07" — 42°54'48"
K 97	1°02'36" — 42°48'48"	BB	1°02'14" — 42°52'03"
K 103	1°04'45" — 42°48'20"	XCVII	0°55'40" — 42°50'32"
K 116	1°02'03" — 42°52'07"	XCV	0°54'25" — 42°53'52"

The plant fossils cited in the text are partly in the collections of the "Geologisch Bureau voor het Nederlandse Mijngebied", Heerlen, Netherlands and partly in the collections of "Institut Royal des Sciences Naturelles de Belgique", Bruxelles, Belgium. All other cited fossils can be found in the collections of "Rijksmuseum voor Geologie en Mineralogie", Leiden, Netherlands. The holotype of the cited Psychopyge n. sp. is in the collection of the "Senckenberg Museum", Frankfurt a. M., Germany.

Op verzoek van de Faculteit der Wis- en Natuurkunde volgt hieronder een kort overzicht van de academische studie van de schrijver.

In 1951 werd hij ingeschreven als student in de Faculteit der Wis- en Natuurkunde aan de Rijksuniversiteit te Leiden.

Op 8 juni 1954 werd het candidaatsexamen (letter I) afgelegd, het doctoraalexamen geologie op 25 maart 1958.

Hij studeerde onder leiding van de hoogleraren Dr. B. G. Escher, Dr. Mr. F. Florschütz, Dr. E. Niggli, Dr. A. J. Pannekoek, Dr. L. U. de Sitter en Dr. I. M. van der Vlerk.

Voor het eerste bijvak structurele geologie verrichtte hij veldwerk in het zuidelijk gedeelte van het Cantabrisch gebergte gedurende de zomers van 1954 tot en met 1958 onder leiding van Prof. Dr. L. U. de Sitter, bovendien deed hij een gedetailleerd sedimentologisch onderzoek in een koolbekken gelegen aan de zuidrand van dit gebergte.

Na een onderbreking van 2 jaar voor het vervullen van de militaire dienstplicht, zette hij zijn veldonderzoek in 1960 voort en werden de verzamelde gegevens uitgewerkt en vastgelegd in dit proefschrift.

Sinds februari 1956 is hij als assistent aan de afdeling Structurele Geologie van het Geologisch-Mineralogisch Instituut te Leiden verbonden.

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