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THE GEOLOGY OF THE SAN ISIDRO-PORMA AREA (CANTABRIAN MOUNTAINS, SPAIN)

BY

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ABSTRACT

The age of the non-metamorphic sedimentary series in the Cantabrian mountains ranges from the Cambrian through the Tertiary. The Lower Palaeozoic deposits mainly show a sandy or quartzitic facies, and were deposited in an extensive Cambro-Ordovician geosyncline of which only the miogeosynclinal part is accessible for investigation, the orthogeosynclinal part now being covered by the Mesozoic-Tertiary deposits of the Spanish Meseta.

The Caledonian orogeny is expressed in phenomena resulting from igneous activity. The Hercynian orogeny started with epeirogenetic movements during the Upper Devonian (Bretonic phase of Stille). The uplift caused a varying hiatus on which Upper Famennian transgressive sediments are found unconformably overlying anything between the Lower Famennian and the Middle Cambrian.

The E-W-trending León line played an important role during these epeirogenetic movements. In the Asturides north of it, a Central Asturian dome shows an almost constant amount of uplift. Over a very wide area the Upper Famennian rests unconformably on the Lower Ordovician quartzite. Two main stratigraphic and tectonic units are distinguished:

1. The Isidro-Tarna-Ponton subarea.
2. The Mampodre-Fontasguera-Ten subarea.

The Upper Famennian transgression was followed by the sedimentation of a uniform limestone during the Visean and Namurian. In the Central Asturian dome region a period of non-deposition or slow sedimentation set in after the sedimentation of the Namurian limestone. Locally, even this limestone was deposited under such conditions. South of the León line, the sedimentation continued with a thick series of mainly clastic deposits (San Emiliano Formation).

During the Upper Namurian-Lower Westphalian, tectonic forces (Sudetic phase of Stille) caused nappe and overthrust structures south of the León line, whereas the areas north of this line, the Central Asturian dome region included, acted as a marginal trough to this orogene.

The Asturian folding phase started in the Upper Westphalian, during which the Carboniferous basins north of the León line became folded. In the Central Asturian dome region, for example, SW-NE-moving thrust sheets developed, and were refolded in a predominantly E-W direction.

Stephanian intramontanous coal basins are found in the present area along the León line. These basins were folded in an E-W direction at some time between the Triassic and Upper Stephanian (Saalic phase of Stille).

Features indicative of a Würm glaciation are frequently found in the area under consideration.

CONTENTS

INTRODUCTION

The study reported here formed part of a systematic program of mapping being carried out in the Cantabrian mountains by the Department of Structural Geology of the Geological Institute of Leiden University. The present paper concerns the rocks and structures occurring north of the León line between the upper reaches of the Curueño and Porma rivers (province of León) and the Isidro and Sella rivers (province of Oviedo).

The investigation concerned the sedimentary series ranging in age from the middle of the Cambrian through the Carboniferous. Middle Cambrian trilobites, Upper Devonian brachiopods and conodonts, Carboniferous conodonts, fusulinids, and algae were used for age determination and correlation. Columnar sections were taken through various lithostratigraphic units for the construction of isopach maps. Some remarks on facies and palaeogeography are collected in a separate chapter. The development of the Upper Devonian epeirogenetic movements and associated erosion are discussed, and the Hercynian low-angle overthrusts and refolding structures are treated in detail. A geological map and structural sections will be found at the back of the volume.

The field work on which this paper is based was done during seven summer seasons between 1959 and 1966. The maps used were enlargements to 1 : 25,000 and 1 : 10,000 scale of the Mapa de España (*Instituto Geográfico y Catastral, Madrid*), scale 1:50,000, sheets 79 (Puebla de Lillo), 80 (Burón), 104 (Boñar), 105 (Riaño). Aerial photographs were put at our disposal by the courtesy of the *Instituto Geológico y Minero de España*. The runs covering the area of the present study are: 53006—52989, 53407—53427, 53598—53618, 43728—43744.

The plant fossils mentioned in the text are to be found in the collections of the *Geologisch Bureau voor het Nederlandse Mijngebied*, Heerlen, The Netherlands. All the other fossils and thin sections can be found in the collections of the *Rijksmuseum voor Geologie en Mineralogie*, Leiden, The Netherlands.

I wish to express my sincere thanks to Professor L. U. de Sitter, whose critical approach to the subject and intimate knowledge of the region contributed so much to the successful completion of this study.

I am also greatly indebted to Dr. D. Boschma for his assistance in the field.

The fusulinids, conodonts, and plant fossils were determined by Dr. A. C. van Ginkel, Mr. H. A. van Adrichem Boogaert, and Mr. H. W. J. van Amerom; the brachiopods by Dr. Th. F. Krans and Mr. C. F. Winkler Prins; and the algal faunas by Dr. L. Rácz and Mr. G. J. B. Germs.

The thin sections were prepared by Mr. J. Schipper, Mr. M. Deyn, and Mr. J. Verhoeven. Most of the photographs and their technical adaptation were made by Mr. J. Hogendoorn. The excellent panorama photo was made by my friend Mr. M. Zindler, who did not fail to live up to his name.

Mrs. I. Seeger corrected the English text. The maps and illustrations, which form such a substantial part of the work, were drawn by Miss C. P. J. Roest and Mr. B. Lieffering. My thanks are also extended to Miss T. W. Terpstra for the typing of the manuscript.

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STRATIGRAPHY

INTRODUCTION

In the area described in this paper Cambrian, Ordovician, Devonian, and Carboniferous rocks crop out. Historical reviews show clearly that rocks of this age have been known in the Cantabrian mountains for a long time, although there has been considerable conflict with regard to the age determination, correlation, and denomination of the various formations.

Concerning the known Cambrian and Ordovician rock units in Asturias, we may refer to the numerous publications of F. Lotze and his collaborators, who introduced a number of different rock units for Asturias and the province of León, and to the work of Comte, who proposed the names Lancara Formation and Oville Formation for Cambrian deposits and Barrios Formation for the sediments of Ordovician age. In the present work we will follow mainly the subdivisions established by Comte for the Lower Paleozoics.

Silurian and Devonian rocks do not occur in our region, with the exception of Upper Famennian (the Strunian included) transgressive deposits. These unconformably-deposited sediments vary strongly, depending on the area of deposition.

We distinguish the following subareas:

I. A central Mampodre-Fontasguera-Ten subarea, with Upper Devonian deposits in unconformable contact with either the Oville Formation or the Lancara Formation.

II. The surrounding Isidro-Tarna-Ponton subarea, with Upper Devonian sediments unconformably deposited on the Barrios Formation.

The situation thus sketched is the result of a Bretonic upheaval, followed by denudation cutting down to different levels.

In both subareas the rock-stratigraphic unit Ermita Formation is used here; in the central Mampodre-Fontasguera subarea a further subdivision distinguishes a Valverde Member and a Mampodre Member. The former includes decalcified ferruginous sandstones, the latter bituminous shaly limestones and coarse-grained greyish limestones.

The lowermost part of the Carboniferous, as found in the Cantabrian mountains, is composed of a sequence of black phosphatic shales alternating with thin cherty layers, sometimes including levels with phosphatic nodules. The age of this Vegamián Formation (Comte, 1959) is still under discussion, but there seem to be strong arguments in favour of a Tournaisian age (Higgins *et al.*, 1964). These black shales do not occur everywhere. The Vegamián Formation is followed by an alternation of red and grey nodular limestone with red shales and radiolarites which directly overlies the Ermita Formation when the Vegamián Formation is missing. This Alba Formation, so designated by Comte (1959), is of Visean age and is followed by the so-called Caliza de Montaña Formation consisting of fine-grained bituminous and intraclastic limestones. Its age, as indicated by goniatites and fusulinids assembled from the top of the formation, is considered to be Lower Namurian B (Kullmann, 1962) or Bashkirian (van Ginkel, 1965). The Vegamián and Alba Formations are excellent examples of a condensed sequence, the whole Tournaisian and Visean being present here in an unusually thin sequence.

The Caliza de Montaña Formation is followed by about 25 m of brownish-red to greenish mudstones with chert nodules, for which the name Ricacabiello Formation is introduced here.

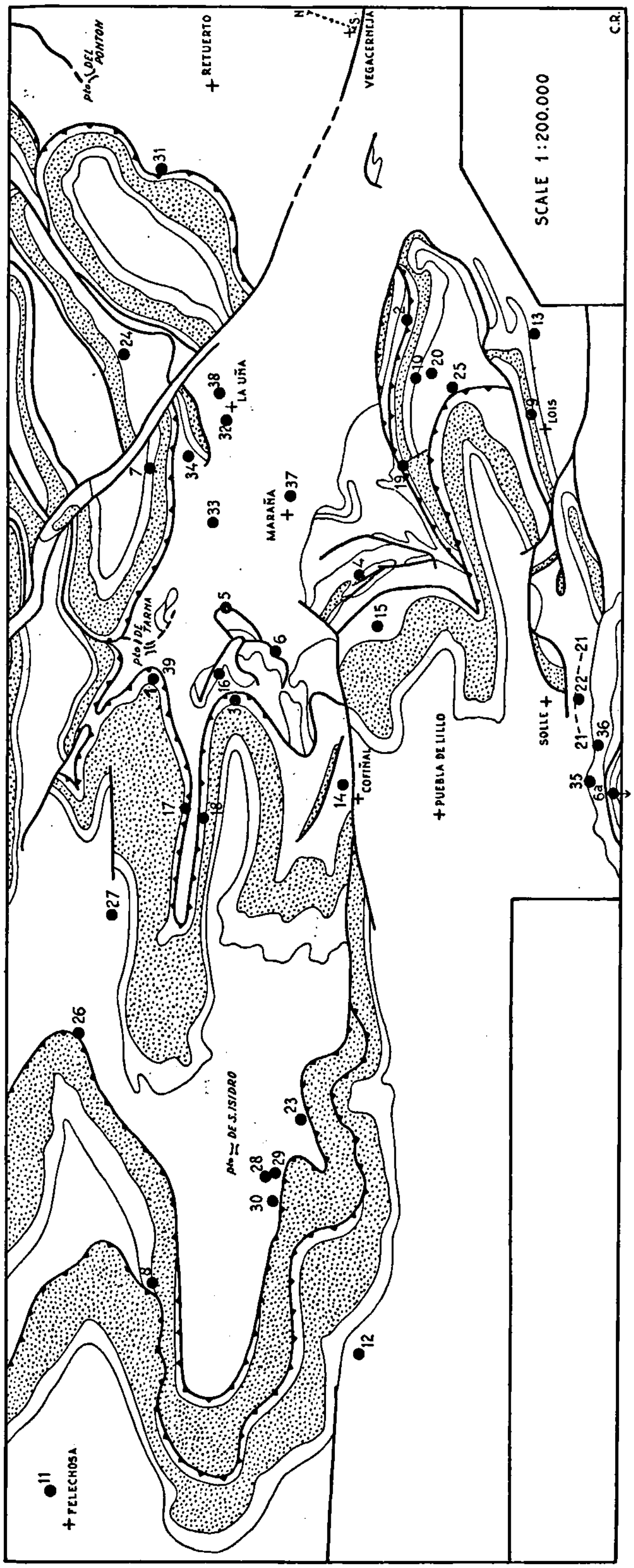
In the province of León (San Emiliano area), the oldest sediments overlying the Caliza de Montaña constitute the San Emiliano Formation (Brouwer and van Ginkel, 1964), which consists of an alternation of greywacke, shale, and limestone lithosomes. Fusulinids, indicating an Upper Bashkirian age, have been found in some of these limestones. The thickness of this formation is about 1750 m. In the area treated in this paper, only deposits corresponding in age to the uppermost part of this formation are found; these deposits are composed of unfossiliferous clastic sediment with few fusulinid-bearing limestone layers. The missing part (about 1500 m) is represented by the Ricacabiello Formation. This suggests a period of non-deposition or slow sedimentation after the deposition of the Caliza de Montaña, a suggestion sustained by the fact that the top of the latter limestone has a rather high manganese content. The locally slow sedimentation had already started during the deposition of the Caliza de Montaña.

Koopmans (1962) subdivided the Carboniferous into three rock-stratigraphic units separated by the two major Hercynian folding phases:

Cea Group	
	Asturian folding phase
Yuso Group	
	Curavacas folding phase
Ruesga Group	

The Curavacas folding phase, however, is not present in the area treated in this paper, and the boundary line between the Ruesga and Yuso Group has therefore become a matter of controversy. We propose to draw this boundary at the top of the Ricacabiello Formation, where the deposition of the mudstones was followed by the sedimentation of the well-known greywackes, shales, and limestones. In this way the Vegamián, Alba, Caliza de Montaña, and Ricacabiello Formations are considered to comprise the Ruesga Group.

In the Yuso Group a Lena and Sama Formation have been distinguished, the terms Beleño Formation and Lois-Ciguera Formation being applied locally. These



1 Remelende section
2 Ricacabiello section (lower part)
3 Pico del Lago section
4 Valverde section (lower-and upper part)
5 Riosol section
6 Cueto Joracao section
6a Armada section
7 La Uña section
8 Felechosa section (lower part)
9 Lois section
10 Ricacabiello section (upper part)
11 Felechosa section (upper part)
12 Vegarada section
13 Lois-Ciguera section

14 Loc. C.L.-1
15 Loc. Tr-1
16 Loc. Tr-2
17 Loc. Tr-3
18 Loc. Sc.64
19 Loc. Sc.33
20 Loc. L.405
21 Loc. 763-777
22 Loc. 766
23 Loc. 13-64
24 Loc. 18-64
25 Loc. B.20
26 Loc. A.2
27 Loc. A.1
28 Loc. 8-64
29 Loc. 9-64
30 Loc. 10-64
31 Loc. L.406
32 Loc. L.34
33 Loc. 888
34 Loc. L.302
35 Loc. L.300
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37 Loc. 1-64 B
38 Loc. 2-64 B
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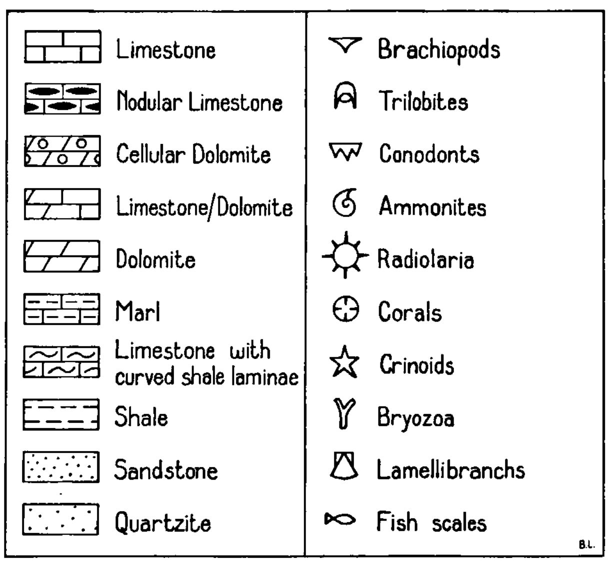

Fig. 1. Location of stratigraphic sections and localities.

Fig. 2. Symbols pertaining to stratigraphic sections.

formations vary in age between the uppermost Bashkirian and Upper Moscovian (van Ginkel, 1965). A quite different alternation of Carboniferous rocks is formed by a thick sequence of greywackes, sandstones, shales, limestones, and conglomerates of Moscovian age. This sequence is the direct continuation to the west of the Curavacas Conglomerate Formation and the Lechada Formation of the Cardaño area (van Veen, 1965).
In the area under consideration, the age of two limestones was analysed in detail: 1. The Lázaro Limestone Lens and 2. The Parme Limestone Lens.
The Asturian folding phase produced an extremely sharp line between the Yuso and the Cea Group. The latter is here represented by a sequence of conglomerates, shales, and coal beds deposited with strong angular unconformity on older rocks. For these deposits the name Rucayo Formation is introduced here. From plant determinations, an age of Stephanian B has been established (van Amerom, 1965).
The location of the stratigraphic sections and localities and the relevant symbols are shown in Figs. 1 and 2.

CAMBRIAN

Historical review

The first description of the Cambrian in the Cantabrian mountains was given more than a century ago by Casiano de Prado (1860), who discovered the so-called "Primordial Fauna" near Sabero (province of León) and in the neighbourhood of Belmonte (Asturias). Determinations were made in this material by de Verneuil and Barrande (Casiano de Prado, 1860).
In 1877 Barrois made his first tour in Spain. He reported the discovery of two new locations of Cambrian trilobites near Vegadeo and Tineo in Asturias, but questioned the occurrence of Cambrian deposits north of Belmonte mentioned by Casiano de Prado. To clear up this problem, Mallada and Buitrago (1878) again visited the controversial area. They confirmed the occurrence of Cambrian rocks, and in fact it was their stratigraphic division which was given by Barrois (1882, pp. 431—432).
Lotze also studied this region (Lotze and Sdzuy, 1961, Teil I, p. 100). His detailed stratigraphic section of the Cambrian near Gorias de Arriba (Asturias) is very similar to the stratigraphy given by him for the Cambrian in the province of León (Lotze and Sdzuy, 1961, Table 7, p. 88).
Comte (1959, pp. 70—74, 80—81, 145) also described the Cambrian of the province of León, and introduced three lithostratigraphic units. The subdivisions of both these authors are shown in Table I, page 59.
A well-known exposure of Cambrian deposits is situated along the Asturian-Galician border, between the villages of Vegadeo (La Vega de Rivadeo) and Ribadeo (Rivadeo). Already in 1858 Schulz had described the limestones of La Vega de Rivadeo, and in 1882 Barrois, who had found the first Cambrian trilobites, gave the following division (Barrois, 1882, pp. 416, 433):

Grès de Cabo Busto (base du Silurien)	Grès blancs et schistes Grès versicolores, poudingues et schistes
Calcaire et schistes à *Paradoxides* de la Vega	Schistes grossiers, fossilifères, et bancs épais de quarzites verts, 50 à 100 m. Calcaires (20 à 60 m), schistes, et lit de minerai de fer (1 m 50 à 2 m).

The same locality was described by Sampelayo (1935), who gave a useful division which was later considered in more detail by Comte (1959, pp. 111—113).

TABLE I. Subdivision of the Cambrian in the province of León, after Lotze & Sdzuy (1961) and Comte (1959).

Lotze & Sdzuy, 1961	
Hangendes: "Armorikanischer Quarzit"	
Oberkambrium	Boñar-Schichten, ca 300 m Obere Cerecedo-Schichten, 60 m (Obere Luna-Schichten, Obere Babia-Schichten)
Mittelkambrium	Untere Cerecedo-Schichten, 10 m (Untere Luna-Schichten, Untere Babia-Schichten) Rote León-Mergel, ca 4 m Rote León-Kalke, 14 m Helle León-Kalke, 6 m
Unterkambrium	León-Dolomite, 60 m Barrios-Schichten, 45 m Cándana-Quarzite, 250 m

Comte, 1959			
Tremadoc et Potsdamien	Grès et Schistes d'Oville	Petits bancs de grès quartziteux ou de quartzites séparés par des feuillets micacés. Des schistes verts. Des grès massifs, des grès ferrugineux, de couleur jaunâtre ou ocre vif, des schistes verts.	120 à 240 m
Acadien	Calcaires de Lancara	Un calcaire marneux, souvent noduleux, d'un rouge intense tirant sur le violet. Des calcaires cristallins associés à des calcaires très fins.	60 à 100 m
Géorgien Précambrien	Grès de la Herreria	Des calcaires dolomitiques compacts gris ou jaunâtres à patine claire. Grès feldspathiques roses, grossiers, disposés en bancs épais.	> 1400 m

Lotze, too, treated the above-mentioned Cambrian in great detail in his discussion of a section north of Vegadeo, near El Fondon (Lotze and Sdzuy, 1961, pp. 125—126; Barrois, 1882, p. 420).

The Cambrian in the mapped area

In 1963 it was conclusively shown that Cambrian sediments can be found in various localities in the region between the San Isidro Pass and the Tarna Pass, including the source areas of the Porma and Esla rivers (province of León) (van Adrichem Boogaert *et al.*, 1963). These Cambrian deposits consist of limestones, dolomites, shales, and sandstones, concordantly overlain by a thick series of hundreds of metres of massive quartzose sandstones and quartzites of Ordovician age.

The lithological properties of these sediments show a strong resemblance to those of the Cambrian deposits of the province of León. This justifies the application to the present area of the lithostratigraphic units used by Comte for the province of León. Thus, the Cambrian in the present area is composed of:

Top a. The Oville Formation with sediments having a mainly clastic character.
Bottom b. The Lancara Formation, in which the calcareous element dominates.

The geological study made by Julivert (1960) of the Cuenca de Beleño and surrounding areas, includes a part of the present area. He distinguished sediments below the massive quartzite without, however, being able to establish their Cambrian age. His subdivision is as follows (Julivert, 1960, pp. 31—36):

Top Nivel de cuarcitas compactas (150—400 m.)
Nivel de pizarras metamórficas y cuarcitas (100—190 m.)
Bottom Nivel inferior con calizas marmóreas (10—150 m.)

Martínez, working in the more westerly areas, did not mention deposits below the quartzites and overlooked the Cambrian zone between the Pico Torres and the Peña del Alba, just north of the Puerto de San Isidro (Martínez, 1962, pp. 30—36).

Lancara Formation

In this formation, three rock units of lower rank can be distinguished:

Top c. Nodular limestone member
b. Coarse-grained limestone member
Bottom a. Limestone-dolomite member

a. Limestone-dolomite member. — The lowermost member is made up of a thick series of limestones, dolomitic limestones, and dolomites, in layers from 10 to 30 cm thick, sometimes intercalated with black sandy and marly shales, the latter varying in thickness from 20 cm to about 5 m near the base. The limestone may display a very fine-grained texture, in which case it shows a dark-blue colour. In many exposures this limestone is very platy and often somewhat bituminous. Pyrite crystals are found, sometimes abundantly, in both the limestones and dolomites. Large parts of this member consist of an allochemical limestone, in which trilobite fragments sometimes occur abundantly. Dolomitization, causing a marked yellow weathering, is not restricted to the bedding plane; small tongues of yellow-weathered dolomite penetrate into the limestone layers perpendicular to the bedding.

b. Coarse-grained limestone member. — In the upward direction, these limestones become more massive, coarser-grained, and show a bluish-grey colour turning to violet. The Remelende section (Fig. 3, p. 61) shows that the coarse-grained limestones contain a certain layer which is highly fossiliferous, the main component being parts of brachiopods and trilobites. A striking pecularity is the fact that this fossiliferous layer always contains a high percentage of iron. Possibly it represents the equivalent of the "lit de minerai de fer" in the Calcaire de la Vega described by Barrois (see p. 59).

Hernandez Sampelayo (1935) denies the occurrence of sedimentary iron ore in the "Caliza de Vegadeo" and regards it as metasomatic iron ore deposited along a fault zone. Although the base of the Lancara Formation in the mapped area is frequently a thrust-fault plane, the presence of both iron and fossil fragments in one and the same layer suggests a sedimentary origin.

c. Nodular limestone member. — The first limestone of the nodular limestone member is very coarse-grained, and can be distinguished from the identical limestones only by many small curved seams of hematitic shaly material. Depending on their shale content, the rocks display all transitions between a limestone with only thin sutures, via a true nodular limestone, to a marly shale with only a few scattered limestone nodules. The colour of the limestone varies from grey via greyish-green to red. There is a striking resemblance to the "Graue, z. T. schwach rötliche, flaserige bis plattige Kalke" from the Gorias de Arriba section, Asturias (Lotze and Sdzuy, 1961; see also p. 59) and the "Calcaire marneux, souvent noduleux, d'un rouge intense tirant sur le violet", (Comte 1959; see also Table I, p. 59).

The tendency of the shale content to increase upward is typical for the uppermost part of the Lancara Formation. No sharp boundary can be established with the overlying shale-sandstone sequence of the Oville Formation which contains occasional limestone nodules at the base. Oele (1964) proposed that the line be taken at the first sandstone bed, a suggestion followed by the present author.

In many sections, for example that of Cueto Joracao (p. 73), part of the nodular limestone is highly silicified, presenting a striking resemblance to the uppermost part of the "Vegadeo Limestone" of which Lotze says: "Die Folge, schliesst mit einer stärkeren

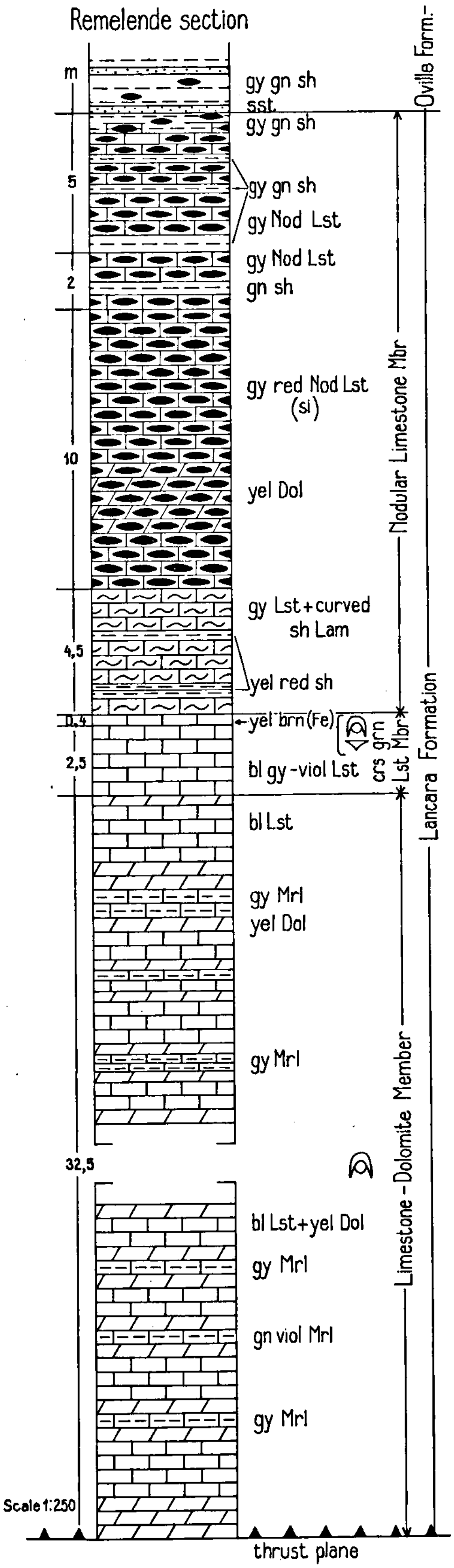

Fig. 3. Columnar section through the Lancara Formation west of Puerto de Tarna (Pico Remelende).

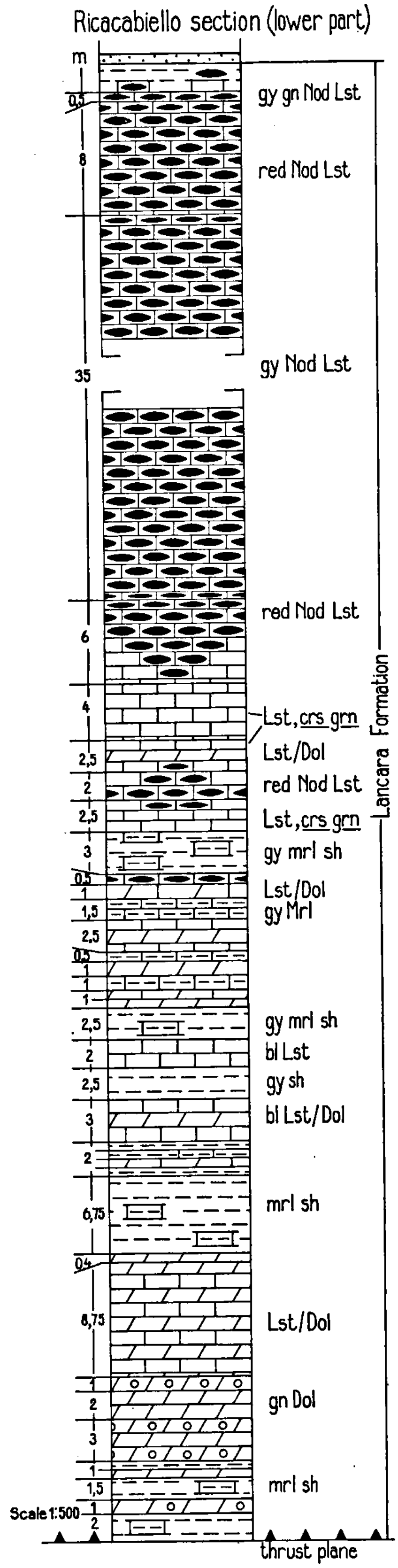

Fig. 4. Columnar section through the Lancara Formation north of Pico Ricacabiello.

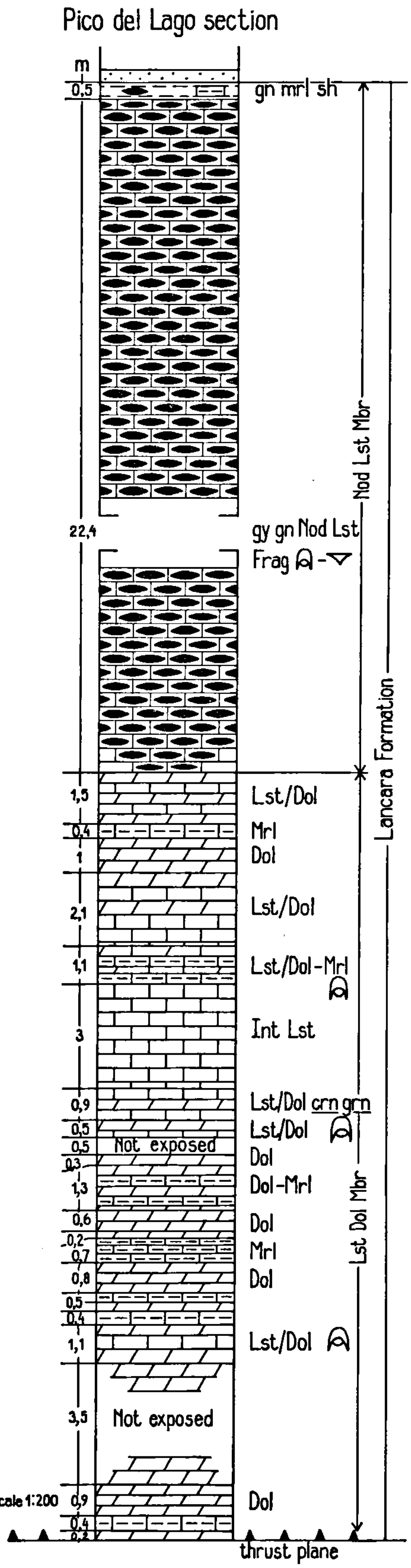

Fig. 5. Columnar section through the Lancara Formation east of Pico del Lago.

Bank verkieselten Kalkes" (Lotze and Sdzuy, 1961, p. 125).

In rare cases the nodular limestone also shows some dolomitization, which causes a colour change to a yellowish-brown. Solution of the remaining limestone gives the rock a very typical cellular dolomitic appearance.

Some detailed columnar sections of the Lancara Formation are provided. Since the base of the Lancara Formation in all exposures is a thrust-fault plane, the measured sections are incomplete.

It was often difficult to establish the exact line between the members of the formation, since thin nodular limestone beds may occur below the coarse-grained limestones (Ricacabiello section, Fig. 4, p. 61). In some places the coarse-grained limestone member is nearly absent in the gradual transition of the limestone-dolomite member to the nodular limestone member (Pico del Lago section, Fig. 5, p. 62).

That the succession can vary rapidly is demonstrated by the fact that along the thrust fault at the basis of the Lancara Formation east of the Pico del Lago, about 150 m to the south, the development is already different. Besides the greyish-green nodular limestones there are also reddish nodular limestones, and occasionally these rocks are partly dolomitized. The top of the formation here consists of a strongly silicified greyish nodular limestone, about 5 m in thickness.

Lithology (*thin sections*). — As we have already seen, the lower part of the formation (i.e. the limestone-dolomite member) consists mainly of carbonate rocks showing different grades of recrystallization.

We can distinguish:

1. carbonate rock, partly recrystallized, with parts of the sediment as dark turbid fine-grained patches in the surrounding recrystallized coarser material, giving the rock a clotted or grumous texture (Pettijohn, 1957, p. 412; Bonet, 1952, p. 172).
2. carbonate rock, wholly recrystallized, with small dolomite rhombs among larger calcite crystals, the whole considerably lighter in colour and coarser in texture than the original sediment.

Some thin sections show an abundance of organic remains. The dolomite of the trilobite and brachiopod debris shows a coarser grain-size and a clearer colour than the surrounding dolomite. This is in full accordance with the data given by Oele (1964) for the organic remains in the Lancara dolomites of the province of León, from locations about 25 km to the south. Consequently, we completely agree with his conclusion that dolomitization took place after sedimentation (see also p. 60).

In one locality, a few hundred metres northeast of the village of Cofiñal (Loc. C.L.-1), a limestone has been found which contains abundant, randomly orientated trilobite fragments, clearly concentrated in the bedding-plane of layers with few clastic quartz grains (Thin sections R.M.G.M. 137501 and R.M.G.M. 137502). Using Folk's terminology (Folk

1959, p. 32), this type of rock is a trilobite biomicrite or, when partly recrystallized, a biomicrosparite (Fig. 6, p. 63). Small clastic quartz grains occur throughout the whole sediment; the rounded to subrounded grains show good sorting, the grain-sizes varying between 40 and 150 μ. Where there is an alternation of carbonate rocks with thin curved limonite-bearing layers of muscovite, the quartz grains show a concentration in these layers (Thin section R.M.G.M. 137503).

Besides these detrital grains, excellent euhedral bipyramidal quartz crystals are much in evidence, ranging in size between 30 and 800 μ (Fig. 7, p. 63). These authigenic crystals, unlike the detrital grains, show only a slight undular extinction. Secondary growth around a nucleus was observed frequently. This phenomenon is found in several forms:

1. Quartz around a detrital quartz grain, the authigenic material replacing both matrix and trilobite fragments (Fig. 8, p. 63). A ring of separate carbonate inclusions indicates the boundary of the original quartz grain (Thin sections R.M.G.M. 137504 and R.M.G.M. 137505).
2. Calcite around a detrital calcite grain, the overgrowth being somewhat coarser and more limpid than the calcite of the detrital grain (Thin section R.M.G.M. 137506).

Only a few of the thin sections show very small felspar crystals, 60—120 μ in diameter, disseminated through the enclosing rock. These crystals frequently show euhedral forms, are limpid, and mostly lack any recrystallization. Some of them display excellent polysynthetic twins. Carbonaceous inclusions are randomly orientated (Thin section R.M.G.M. 137507; Fig. 9, p. 64). The negative extinction angle of 15.13° in a plane normal to 010, indicates an anorthite content of about 3 %. Since it is known that the authigenic felspars in sediments are nearly always pure alkali-felspars (Berg, 1952; Honess and Jeffries, 1940), the unmixed character, coupled with the euhedral form of the crystals, supports a diagenetic origin of the crystals. Many investigators have gone deeply into the problem of the exact moment of origin

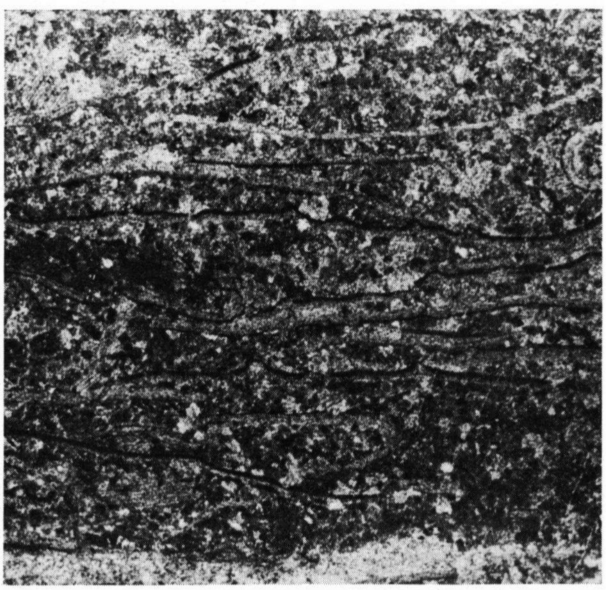

Fig. 6. Lancara Formation, limestone-dolomite member; trilobite biomicrite with trilobite fragments concentrated in the bedding plane (section taken perpendicular to the bedding plane). (15 ×)

Fig. 7. Lancara Formation, limestone-dolomite member; authigenic bipyramidal quartz crystal. (100 ×)

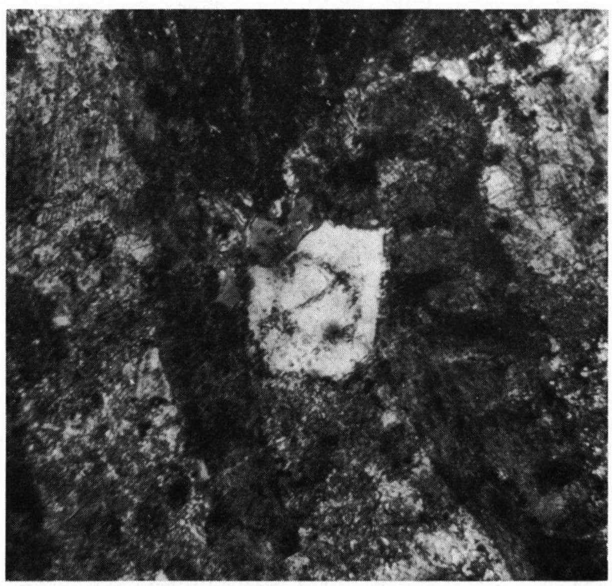

Fig. 8. Lancara Formation, limestone-dolomite member; authigenic quartz crystal with ring of carbonate inclusions indicating the boundary of the original quartz grain. Authigenic material has replaced part of the trilobite fragment. (100 ×)

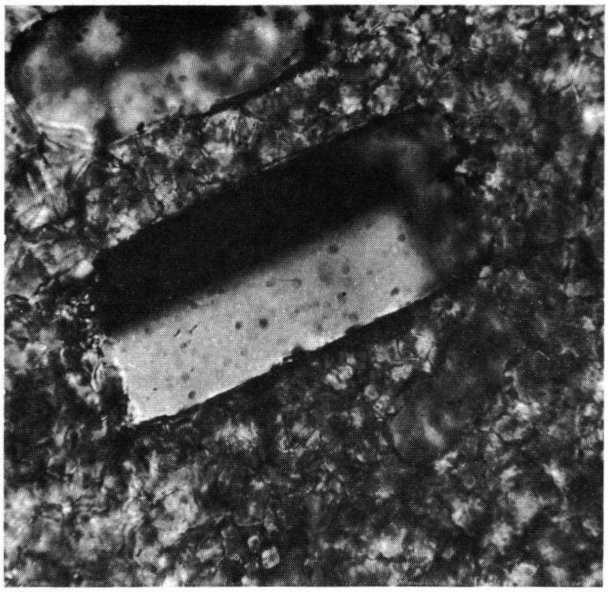

Fig. 9. Lancara Formation, limestone-dolomite member; authigenic albite crystal. (600 ×)

Fig. 10. Lancara Formation, limestone-dolomite member; authigenic quartz crystal with carbonate inclusions showing successive stages in the development of the authigenic material. (420 ×)

of the authigenic mineral. Pettijohn (1957, pp. 666—667) gives a general view of the various theories on this subject. He comes to the conclusion that, "it seems likely that the feldspar is a postconsolidation product, as no growing feldspar crystals have been found in recent carbonate muds, and fossils and oolites are replaced by authigenic feldspar".

Thin section R.M.G.M. 137508 shows clearly that there were successive stages in the development of the authigenic minerals, each consecutive stage being faintly discernable by small carbonate inclusions (Fig. 10, p. 64).

The present author holds the view that the situation described on page 62, in which dark, fine-grained patches of calcite and/or dolomite are found in an environment of coarser and more limpid material, need not necessarily be the result of recrystallization of the original sediment. Some of the thin sections, especially those from the Pico del Lago section, reveal that these dark patches are often rounded, with sharp boundaries between them and the surrounding matrix. We should prefer the term intraclasts for these patches, that is to say, ". . . . fragments of weakly consolidated carbonate sediments that have been eroded from adjoining parts of the sea bottom and reincorporated into newly formed calcareous deposits" (Krumbein and Sloss, 1963, p. 117; see also Folk, 1959, p. 4).

In fact, in some thin sections all four allochem grains (Folk, 1959) are present: intraclasts, pellets, oölites, and fossil fragments. For these particular layers, with the matrix built up of rather coarsely crystalline-textured (sparry) calcareous material, the term "sparry allochem limestone", or, when the intraclasts comprise more than 25 % of the allochem grains, "intraclastic limestone" (or fossiliferous intrasparite) may be applied (Thin sections R.M.G.M. 137509, R.M.G.M. 137510 and R.M.G.M. 137511, Fig. 11, p. 65). Occasional the intraclasts are much bigger than 1 mm; in that case the rock is called an intrasparudite. Both intrasparite and intrasparudite locally show strong recrystallization.

Oölite-bearing limestones scarcely occur, in contradistinction to the occurrence of an important group of this kind of rocks in the lower part of the Lancara Formation in the province of León (Oele, 1964, p. 37). Only superficial oölites, i.e. an envelope consisting of only one layer (Carozzi, 1960), have been found. Different kinds of cores have been observed, mainly trilobite fragments but also quartz grains or grains built up of a mozaic of small calcite crystals.

In one locality, i.e. the "Valle de Carcedo" just north of the village of La Uña, a most interesting kind of rock (forming a 50 cm thick layer) is developed in the lower part of the limestone-dolomite sequence. This layer is made up of quartz and calcite fragments set in a microcrystalline matrix of calcareous material. The calcite fragments are rounded to well rounded, sometimes reach a size of 6 mm, and are built up of micrite, trilobite biomicrite, and intramicrite. Three types of quartz grains can be distinguished (Thin sections R.M.G.M. 137504 and R.M.G.M. 137506):

Fig. 11. Lancara Formation, limestone-dolomite member; intraclastic limestone. (40 ×)

1. Composite quartz grains (rock fragments), built up of an aggregate of small quartz crystals and showing a strong undular extinction. They are subrounded to well rounded.
2. Quartz grains consisting of only one quartz crystal showing only a slight undular extinction. They are subangular to subrounded and measure from 40 to 850 μ.
3. Authigenic quartz around a nucleus, the latter being visible as a ring of small carbonate inclusions within the newly-formed quartz crystal.

Oele (1964, p. 43) compares this kind of rock with shoal breccias as described by Dunbar & Rodgers (1956, p. 179); the present author prefers the term sandy intramicrudite.

Thin sections of the nodular limestone frequently show that the rocks can have an organo-clastic appearance, being built up mainly of trilobite fragments in a matrix of calcite alternating with thin layers of clayey material. In Folk's classification (1959) the rock varies between a trilobite biomicrite and a fossiliferous intrasparite. Locally, this nodular limestone also displays strong recrystallization.
Especially in the top, the glauconite percentage is rather high, and not only small clastic quartz grains (40—90 μ) but also authigenic quartz crystals and felspar crystals are readily observed. The glauconite grains are mainly light green, although grains of a darker shade are not exceptional. Both kinds often display a rim of ferric material. Grain sizes vary between 10 and 550 μ. Replacement by calcite is frequently found.
Oele (1964, pp. 43—44, 54—55) discerned different kinds of glauconite in the Cambrian sediments of the province of León. In the Lancara Formation (nodular limestone member), the present author observed crystalline forms of glauconite which had originated from the conversion biotite-glauconite, the resulting shape closely resembling the elongated mica flakes (Thin section R.M.G.M. 137512). This is in contradiction to the data given by Oele (1964). For more detailed information on the lithology and origin of the Lancara griotte, the reader is referred to Oele (1964, pp. 45—50).
In closing the description of the lithology of the Lancara Formation, it may be said that ferric material is much in evidence throughout the entire sediment in the form of grains and rims around grains, as well as very fine microstylolitic seams. Pyrite crystals show excellent cubic forms; zircon is found in many thin sections. Calcite veins cross through the whole rock, cutting oölites, fossils, and various sedimentary structures.

Oville Formation

In the area under consideration, beds of different lithologic composition can be distinguished: sandstones, shales, mudstones, and occasionally limestones. With respect to age and lithology we may compare these deposits with the Oville Formation of Comte (1959).
No consistent order is known for these beds, although an ochreous ferruginous mudstone seems to be restricted to the base of the formation. All the various lithosomes are not even necessarily present in a single section. No key beds can be distinguished.
When we find Barrios quartzite on top of the Oville formation, the thickness of the latter may vary strongly, i.e. between 60 and 140 m. Where the quartzite is entirely missing due to the Pre-Upper Famennian erosion which sometimes cut away all or parts of the Barrios and Oville Formations (see p. 70), the present thickness in the mapped area can even come to much less than 60 m.

a. Ochreous ferruginous mudstone. — At many localities, a thin-bedded ochreous mudstone (0—25 m) occurs at the base of the Oville Formation, concordantly overlying the Lancara Formation.
Very typical are the open bore holes lying perpendicular to the bedding.
The best exposure is situated just northwest of Pico Ten, in a small cleft perpendicular to the A° de la Majada de la Castellana. Thin sections show an opaque groundmass of limonite with a small amount of authigenic quartz. Floating in the groundmass are

small clastic quartz grains (max. 60 μ), felspars, and muscovite and biotite flakes (Thin section R.M.G.M. 137513).

b. Ferruginous sandy shales and sandstones. — The main part of the formation under consideration consists of ferruginous sandy shales and sandstones. In the mineralogical composition the most frequently encountered mineral is quartz, very irregular in outline due to diagenetic solution, and limpid. Microcline, muscovite, biotite (in all transitional stages of glauconitization), and glauconite occur in varying amounts.
The following accessoria are also much in evidence: zircon, apatite, green and blue tourmaline, staurolite, and pyrite (Thin sections R.M.G.M. 137514, R.M.G.M. 137515, R.M.G.M. 137516). The over-all picture is in full accordance with the facts given by Oele (1964, pp. 52—55) and Comte (1959, pp. 72—73) for the Oville Formation in the province of León. Strong quartz cementing exists, as well as cementing by iron-rich material and calcite.
In some layers the iron percentage is rather high, giving the rock a typical yellowish-brown to brownish-red colour, an appearance also described by Lotze in, for example, the El Fondon section (Lotze and Sdzuy, 1961, p. 126). A striking pecularity lies in the fact that it is precisely in these iron-rich layers that trilobite fragments occur abundantly.
Frequently, an important part of the rock consists of glauconite grains, as for example in the sandstones of the Valle de Valverde. The majority of these grains are well rounded; crystals which clearly betray their autochthonous character as an alteration product of biotite flakes are not exceptional, however.
The well-rounded glauconite grains sometimes present a change of colour from green to yellow or yellowish-brown due to oxidation (Thin section R.M.G.M. 137517; Pettijohn, 1957, p. 149). Especially when the section consists of an alternation of sandstone and shales, such glauconite-rich layers are frequent.
Stratigraphically higher in the section the sandstones have a tendency to become more quartzitic, a development leading to the quartzites of the Barrios Formation above them. These upper parts, too, frequently contain glauconite-rich inlayers (Thin section R.M.G.M. 137518). The occurrence of glauconite in such great quantities almost throughout the entire sedimentary sequence is in contradiction to the facts given by Oele (oral communication) and Comte (1959, p. 73), the latter having stated: "La glauconie n'est jamais abondante et semble manquer dans les assises supérieures".
In various localities, such as those east of the village of Cofiñal (Loc. Tr. 1), west of the Tarna mercury mine (Loc. Tr. 2), and along the road Puerto de Tarna-Cofiñal (Loc. Tr. 3), a determinable trilobite fauna is found, consisting predominantly of *Conocoryphe* species and what are probably a few *Solenopleuropsis* specimens, and in all localities *Carpoids* are abundant. The following trilobite assemblage has been determined (van Adrichem Boogaert *et. al.*, 1963):

Conocoryphe heberti
Conocoryphe cf. *ovata*
Solenopleuropsis sp.

This assemblage is closely comparable to the fauna described by Lotze and Sdzuy (1961, pp. 77—78), from the base of the Oville Fomation ("Untere Luna Schichten" in their terminology, see page 59) near Los Barrios de Luna (León), and some other localities in the Luna, Porma, and Esla valleys, as well as to the fauna of the "Obere Artedo Schichten" in Asturias (Lotze and Sdzuy, 1961, pp. 113—114). These faunas indicate a middle Cambrian age.
Comte, too, places the lower part of the formation in the Acadian, whereas ". . .. la partie supérieure des Grès et Schistes d'Oville représente très probablement le Potsdamien et le Tremadoc" (Comte, 1959, pp. 129—132).

c. Greyish Scolithus sandstone. — The best exposures of this type of rock are found along the road between Puerto de Tarna and Cofiñal (Loc. Sc. 64) and south of the Cuesta Rasa (Loc. Sc. 33).
The mineralogical composition, as demonstrated by thin sections, is the same as that of the above-mentioned sandstones, with only the following differences of minor importance: glauconite occurs only in small quantities, calcite-cementing dominates over cementing by quartz or iron material, and the iron content of the rock is very low.
In this type of rock there are many bore holes, animal traces about which Shrock stated: "The ubiquitous *Scolithus*, however it may have been formed, appears to represent a burrow of some sort" (Shrock, 1948, p. 183). These bore holes (Fig. 12, p. 67) are always perpendicular to the bedding-plane and filled with sediment, unlike the bore holes described above (see p. 65).
The diameter of the *Scolithus* varies between 3 and 10 mm. There is no difference in the mineralogical composition of the material in and outside the bore holes, the latter being visible by a ring-like arrangement of mica flakes and small iron particles (Thin section R.M.G.M. 137519). Similar animal traces in Cambrian deposits have been described by Barrois (1882, pp. 431—432), Lotze and Sdzuy (1961), Comte (1959), and Rupke (1965).

d. Moss-green silky shales. — Like the above-mentioned beds, greenish shales occur throughout the whole section, either in rather thick beds or in alternation with ferruginous sandstones. Under the microscope the shales prove to be built up of small quartz crystals, mica flakes, and glauconite grains; zircon, apatite, tourmaline, staurolite, and pyrite occur only in small quantities, the whole being cemented by quartz and iron-rich material.

e. Limestones. — As we have seen above, cementing by calcite occurs frequently. Occasionally, however, the carbonate forms the major constituent of the rock,

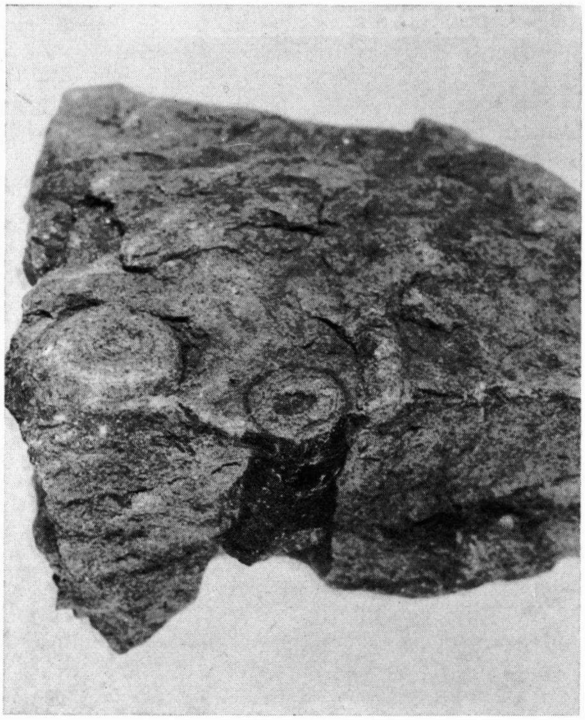

Fig. 12. Oville Formation; greyish *Scolithus* sandstone. (natural size)

producing a limestone which does not differ much from those of the Lancara Formation. This rock consists mainly of microcrystalline calcite, strongly recrystallized, together with parts having a more clayey and iron-rich appearance. Due to the iron content of the rock, the calcite may have a very turbid appearance.

Within the groundmass, sub-rounded to rounded quartz grains are scattered throughout the whole rock. Felspars, glauconite, brown tourmaline, zircon, and pyrite are also present. Trilobite fragments are not exceptional.

In one case, a locality just east of the village of Cofiñal, the limestone is strongly oölitic. According to Folk's classification (1959), the limestone is a fossiliferous oömicrite (Fig. 13, p. 67).

The cores of the oölites are built up of one of the following: an agglomerate of small calcite grains accompanied by some small quartz crystals, a single calcite crystal, iron-rich material, or zircon crystals.

The oölites, circular in form, are usually normal ones, the envelope consisting of more than one layer, the outermost rim always being somewhat more iron-rich than the others, which are built up of an alternation of dark and light calcite (Thin section R.M.G.M. 137520). A radial-concentric pattern, as described by Oele (1964, pp. 38—39) for the oölites in the limestones of the Lancara Formation, is only faintly discernible.

As compared with other beds of the formation, there are very few quartz grains, accessoria, and mica flakes.

In Asturias, as we have seen on page 59, comparable deposits can be found between the "Calcaire de Vegadeo" and the "Quartzites de Cabo Busto". The same picture holds for the province of Galicia; there the thickness of the Cambrian deposits seems to vary considerably between the various localities. The highest values are given by Sampelayo for the surroundings of the Ribadeo river (700 to 800 m), but other places show a rapid decrease in thickness (see Comte, 1959, p. 112).

Comte uses the name "Schistes et grès de Puente Radical", a denomination sometimes also applied by Barrois. According to the Table given by Comte (1959, p. 114), the above-mentioned "Schistes et grès de Puente Radical" together with the "Grès versicolore, poudingues et schistes de Cabo Busto" correspond to the "Schistes et grès d'Oville" in the province of León.

ORDOVICIAN

Historical review

In many places in northern Spain it can be observed that Cambrian sediments are sometimes unconformably overlain by a thick series of sandy quartzites of presumably Ordovician age (Lotze and Sdzuy, 1961, p. 190). From a historical point of view, it will be useful to distinguish between quartzites cropping out in the western areas, south from the line Ribadeo-Gijon, and quartzites found in eastern Asturias, between the Central Basin of Asturias and the Picos de Europa.

Among the early investigators who studied the quartzites in the *western* part of Asturias, Schulz (1837), Paillette (1845), and Barrois (1882) are prominent;

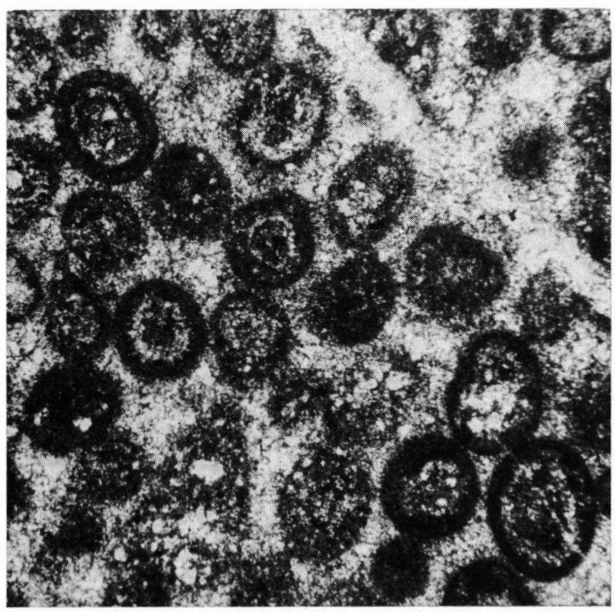

Fig. 13. Oville Formation; fossiliferous oömicrite (partly recrystallized). (40 ×)

during the years 1852, 1855, and 1860, the co-operation between de Verneuil, Collomb, Casiano de Prado, and Barrande, proved to be fruitful and resulted in various publications in the Bulletin de la Société Géologique de France.

Barrois (1882, p. 442), who introduced the name "Grès de Cabo Busto", stated: "La base de ce système est formée d'après nous, par des *grès et quarzites versicolores*, violacés, verts ou rouges, passant à leur partie supérieure aux grès blancs (No 3), bien connus en Espagne sous le nom de *grès à Scolithes* ou à *Bilobites*".

Later, the following terms were proposed: Mallada (1896): "Cuarcitas de Cabo Busto", Adaro (1916): "Cuarcita de los Cabos, Cuarcita armoricana", Hernandez Sampelayo (1942): „Cuarcitas de *Cruciana*". These terms were often in simultaneous use, but in more modern publications the name "Cuarcita Armoricana" has been adopted (Carcia-Fuente, 1952, 1953; Llopis Llado; Julivert, 1957, 1960; Lotze and Sdzuy, 1961; Martínez, 1962).

Barrois (1882, p. 464) gives for Asturias the following lithostratigraphic division of the "Silurien" ("Silurien" being the French equivalent of Ordovician + Silurian in English terminology).

Silurien supérieure

Schistes et quartzites de Corral, *ampélites*.
Sch. calcar. de el Horno à *Endoceras duplex*.

Silurien moyen

Sch. ardois de Luarca à *Calymene Tristani*.
Lit de minerai de fer.
Grès de Cabo Busto à Scolithes.

Silurien inférieure

Grès versicolore, poudingues et schistes.

In the above-mentioned "Schistes de Luarca", which belong to the upper part of the Ordovician (Comte, 1959, pp. 118—119), C. de Prado found the first "Silurian" fossils known for Asturias. The age of the underlying quartzites could be anything from Arenegian to Llanvirnian.

Comte introduced the name Barrios quartzite for the Ordovician deposits in the province of León. This author, comparing the Ordovician and Silurian rocks in this province with those cropping out in Asturias (Comte, 1959, pp. 123—125, 132—141), found strong arguments in favour of a direct correlation of the Barrios Formation with the "Quartzites de Cabo Busto." In that case, the "Schistes de Luarca" of Asturias are missing in the province of León. Another possibility is that the Barrios quartzites correspond to the combined" Quartzites de Cabo Busto" and "Schistes de Luarca."

As a result of a similarity in the lithology of the quartzites cropping out in the *eastern* part of Asturias, coupled with a lack of fossils, it was originally supposed that in this region only one quartzite could be distinguished. It will be useful here to give a short review of this controversial point, which arose as soon as early investigators attempted to date these quartzites.

The geological map of Schulz (1858) shows only one quartzite of Carboniferous age; and after him, various geologists came to the same conclusion, including E. and F. Hernandez Pacheco (1935—1936), who after the discovery of *Lepidodendron* decided to assign a Lower Carboniferous age to all quartzites in direct contact with the so-called Caliza de Montaña.

More recently, Schindewolf and Kullmann (1958) concluded that the quartzite of the Sella river, concordantly underlying the nodular limestone (griotte) and Caliza de Montaña, must be of Upper Devonian or Lower Carboniferous (Tournaisian) age.

The view that quartzites of Devonian age occur in the eastern and central parts of Asturias has also been generally accepted. Gascue (1875) and Barrois (1882) were the first to study these Devonian quartzites, the latter introducing the term "Grès de Cué", for quartzite deposits occurring just below the Lower Carboniferous griotte. Much later, Saenz Garcia (1943, 1944) found fossils indicating a Devonian age for the quartzites of Camporredondo (Palencia), and argued that all the quartzites of that region therefore belong to the Devonian.

The occurrence of Devonian nodular limestones, stratigraphically below and above a quartzite east of the Puerto de Ponton (north of the village of Riaño, province of León), led de Sitter to express the opinion that the quartzites west of that pass have the same age. Nevertheless, throughout the years many investigators have taken the position that the quartzites are much older than Carboniferous or Devonian. Quirago (1887) found *Scolithus* and supposed a Cambrian age on that account, but Mallada (1896, 1898), who had first favoured a Devonian age, was convinced by Quirago's publication that a "Silurian" age was more probable. In 1916 Adaro was the first to see the connection between the above-mentioned "Grès de Cué" and the "Cuarcita Armoricana" in the west.

Hernandez Sampelayo (1928, 1942) opposed the views of Saenz Garcia and Hernandez Pacheco; after the discovery of a *Cruziana*, this author assigned a "Silurian" age to the "Grès de Cué". According to de Sitter (1949), the "Grès de Cué" is Ordovician in age.

Delépine (1932) and Comte (1938*a*, *b*) also came to the conclusion that the quartzites below the Lower Carboniferous griotte are "Silurian" in age; these investigators are of the opinion, however, that both Devonian and Carboniferous quartzites can occur locally.

Julivert (1957, 1960) distinguished more than one quartzite in his investigation of the Cuenca de Beleño and surrounding areas:

1. The Devonian quartzites of Casasuertes, east of the Puerto de Ponton.
2. A quartzite in which *Scolithus* and *Cruziana* frequently occur; this would, according to its facies and fossil content, be the equivalent of the "Cuarcita Armoricana".
3. Carboniferous quartzites and quartzitic sandstones overlying the Caliza de Montaña; these are of minor importance.

Llopis Llado was of the same opinion; in his reply to the publication of Schindewolf and Kullmann, he came to the conclusion that there was not sufficient proof, on paleontological grounds, to assume the existence of only one quartzite for the whole of Asturias.

Recent work has shown that the geologists who supposed an Ordovician age for the quartzites in the upper reaches of the Esla, Porma, and Curueño rivers, were right. The discovery of trilobites in the Oville Formation below the quartzite (van Adrichem Boogaert *et al.*, 1963) demonstrated that this quartzite, in which *Cruziana* and *Scolithus* are frequently found, nearly everywhere overlies sediments of Cambrian age. This discovery, which threw new light upon the problem, coupled with the fact that there is a gradual transition between the Cambrian and the quartzite, justifies the assumption of an Ordovician age for the quartzites in the area under consideration.

In this area, younger Ordovician or Silurian sediments overlying the quartzite are not known because the Bretonic erosion always cut very deep, as far as or into the quartzite or even deeper. These Ordovician quartzites seem to correspond to the "Cuarcita Armoricana" in the west and the Barrios Formation in the province of León (see p. 68).

Devonian quartzites, such as those of Casasuertes, do not crop out in the mapped area, and only small quartzitic sandstone beds occur in the various Carboniferous formations.

Barrios Formation

As we have already stated, the exact boundary between the Oville Formation and the overlying Barrios Formation is difficult to draw.

For practical purposes, the first rather thick massive quartzite layer has been taken here as the boundary between the two formations. It must be kept in mind, however, that the more sandy and shaly layers following this quartzitic bed are lithologically exactly the same as those occurring in the upper part of the Oville Formation. Therefore, the thickness of the Oville Formation as well as that of the Barrios Formation is dependent on the varying lithology of the transitional beds between the two formations, in which, furthermore, rapid wedging out of the quartzite layers is frequently observed.

Stratigraphically higher in the Barrios Formation, the rock becomes a rather massive quartzite, but in some places it alternates with thin iron-rich shaly layers, especially where the quartzite, too, is rich in iron.

The uppermost part of the formation is a massive white quartzite.

In thin sections the sandy basal layers of the formation contain the following minerals: quartz, microcline, mica flakes, glauconite, tourmaline, zircon, apatite, and rutile, i.e. the same picture as we have already seen in the Oville Formation (Thin sections R.M.G.M. 137521 and R.M.G.M. 137522).

The white quartzite has the following mineralogical composition (Thin section R.M.G.M. 137523):

Quartz, strongly predominant, often showing secondary overgrowth.
Felspars, scarce, mainly represented by microcline.
Muscovite, biotite (slightly pleochroic), and phengite (Oele, personal communication), the presence of the latter being a diagnostic characteristic of the formation.
Tourmaline (green, brown, and blue), zircon, rutile, and apatite.

The matrix between the detrital grains is formed by small quartz and sericite grains; cementing is by quartz and opaque material, the latter sometimes forming typical concentrations, giving the white quartzite a spotted appearance which seems to be another diagnostic characteristic for this part of the formation.

Cementing by quartz, which forms strong secondary overgrowths with dust rings only faintly discernable, probably led Comte (1959, p. 75) to the observation that: "Au microscope, ces quartzites se présentent comme une mosaïque de grains de quarts engrainés les uns aux autres sans interposition de ciment".

Thus, there is a striking similarity in lithology between the Barrios Formation as described by Oele (1964, Chap. V) for the province of León and the Ordovician sediments in the mapped area. In this area we found no conglomerates such as are described for the environment of Boñar, but only scattered pebbles (well rounded and varying in size between 0.5 and 4 cm) locally in the white quartzite of the upper part of the formation.

DEVONIAN

The only Devonian sediments occurring in the area under consideration are of Upper Famennian age (the Strunian included). These sediments have a transgressive character, and are in unconformable contact with mainly the Ordovician Barrios Formation, although contacts with older (Cambrian) sediments are not exceptional.

In the province of León, Comte (1938) discerned the so-called "Grès de l'Ermitage", transgressive Upper Famennian sediments overlying, with only a very small angular unconformity, rocks of different ages.

This is a common picture in the Leonide thrust zone (de Sitter, 1962 b, c) where, going from south to north, a differentiated Bretonic uplift caused an erosion that cut down to ever deeper levels. In the Bregon unit, south of Santa Lucia (Bernesga river), the Upper Famennian overlies part of the Nocedo Formation (Frasnian in age), while in the Gayo thrust sheet, north of Nocedo de Curueño, we can discern the unconformable contact between the Ermitage and La Vid Formations, Lower Devonian in age (Evers, in press). South of Tolibia de Abajo in the Bodon thrust sheet we find the Ermitage Formation lying on the Ordovician Barrios Formation (Comte, 1959, p. 201; Evers, in press).

That the Upper Devonian hiatus can even be much greater, is demonstrated by observations of the Ermitage Formation overlying part of the Cambrian Oville Formation south of Campillo (Porma river) in the Forcada thrust sheet (Evers, in press) or in the Armada unit, east of Utrero, Porma river (Comte, 1959, pp. 203—204).

The present author has studied an area north of the above-mentioned Leonide thrust zone, separated from it by the León line (de Sitter, 1962 b, c; Rupke, 1965; Evers, in press) and some Carboniferous basins. There too, the Ermitage Formation is found overlying the Barrios, Oville, or Lancara Formations (Fig. 14). There was certainly a difference in the development of the Upper Devonian hiatus between the two areas, south and north of the León line. The former shows a gradually increasing hiatus going from south to north, the latter presents an Upper Devonian erosion nearly always cutting down to the Barrios Formation, a situation which arose from the fact that the area under consideration was uplifted in its entirety, the erosion cutting down to a more or less constant stratigraphic level (the Barrios Formation). In only a few places, such as the Mampodre-Fontasguera-Ten regions, was

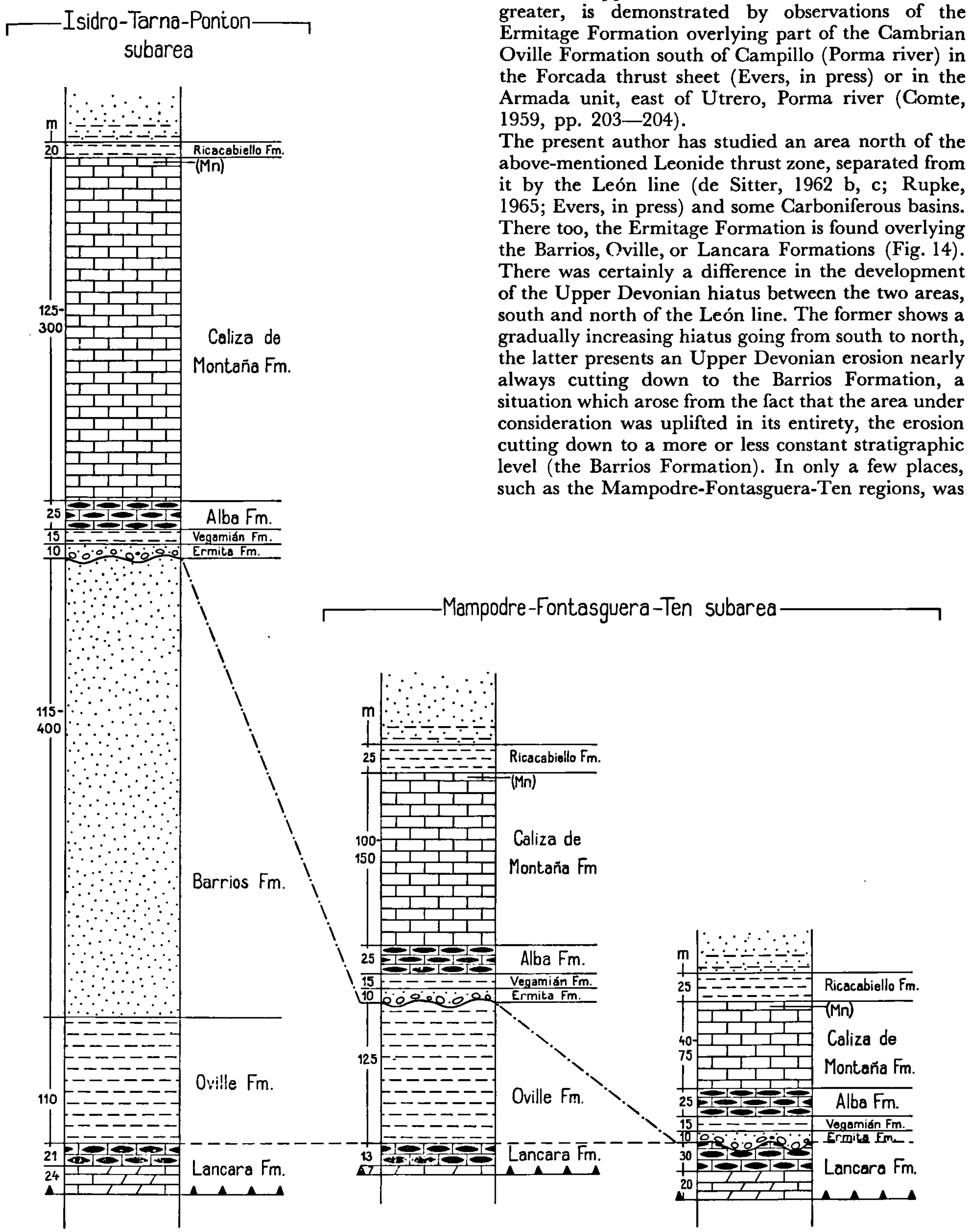

Fig. 14. The Upper Devonian Ermita Formation in unconformable contact with various older Paleozoic rock-stratigraphic units.

the uplift more pronounced; there we find the Upper Devonian sediments overlying the Oville Formation or the Lancara Formation (Fig. 14, p. 70).

Transitional zones do exist, with Upper Famennian transgressive sediments resting on only about twenty metres of the Barrios Formation, but we believe the majority of these zones to be covered by the Hercynian thrust planes.

Apparently the upheaval was dome-shaped; the name *Central Asturian dome region* has been introduced for this particular area (Fig. 41, p. 103).

As we shall see below, there are many differences in lithology between the Upper Devonian, as developed in those parts where the erosion cut down to the Barrios Formation (the Isidro-Tarna-Ponton subarea) and the Devonian in the central Mampodre-Fontasguera subarea.

Instead of the term "Grès de l'Ermitage", we prefer to use the lithostratigraphic denomination Ermita Formation. To apply this unit in the area treated in this paper, it will be necessary to extend the lithologic conception of the formation.

Rupke (1965, pp. 25—27), in his geological study of the Esla region, added his own observations to the lithology given by Comte (1959, pp. 233—234). He distinguished in the Peña Corada unit a reddish-white, often calcareous quartzitic sandstone (40—50 m thick), in the Agua Salio syncline thick-bedded white quartzites (80 m), and along the Arroyo de Remolina a decalcified ferruginous sandstone. In addition, a cross-bedded limestone occurs in various localities in the Las Salas zone and in the autochthone of Valdoré.

In the region treated in this paper, all the above-mentioned lithosomes occur with exception of the Upper Devonian thick-bedded white quartzite. A sequence of limestones, shales, and rare nodular limestones is found in the central Mampodre-Fontasguera-Ten subarea.

The best way to show the wide variation within our Ermita Formation is to give some detailed lithostratigraphic sections through the Upper Devonian as found in the investigated area.

Mampodre-Fontasguera subarea

1. *Valverde section* (*lower part*). — (Fig. 15, page 71). In this section the Ermita Formation starts with a thin conglomerate of small quartz pebbles (0.5 cm), lying on 10 m of the greenish-grey quartzite composing the upper part of the Oville Formation or the lower part of the Barrios Formation. This conglomerate forms the base of a 35 cm thick decalcified ferruginous sandstone layer. Thin section R.M.G.M. 137524 contains small, rounded quartz grains and a few accessoria such as tourmaline, zircon, apatite, and rutile, together with well-rounded calcite fragments. The matrix is calcite, sometimes with a high iron content. Strong limonite concentrations at the base of the layer are not exceptional.

For this part of the Ermita Formation, together with the overlying 10 cm ferruginous sandstone, the writer

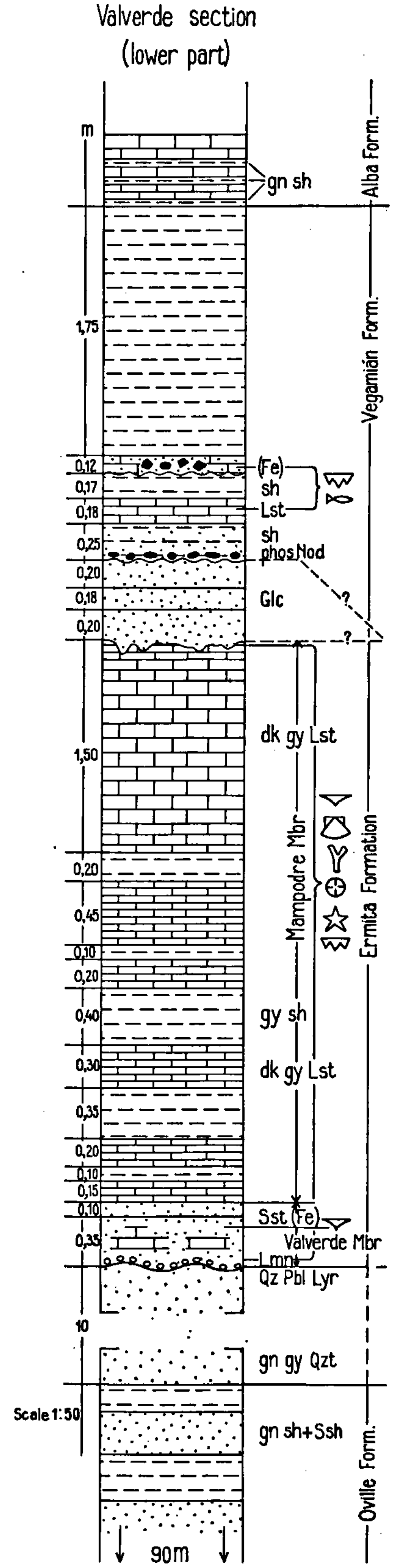

Fig. 15. Columnar section through Upper Devonian and Tournaisian deposits west of Pico de Mampodre (Valle de Valverde).

proposes the name Valverde Ferruginous Sandstone Member.
The lithology of this Valverde Ferruginous Sandstone Member is in conformity with the description given by Comte for the "Grès de l'Ermitage" south of Montuerto (Prov. of León): "Les Grès de l'Ermitage sont représentés par quelques bancs descontinus de grès très ferrugineux rouge foncé (0 à 2 mètres d'épaiseur totale). Ils offrent le même aspect au N.O. de Valdorria où un échantillon a donné 30 % de fer métallique" (Comte, 1959, p. 203).
Dr. Th. F. Krans has carefully investigated the brachiopods collected in this Valverde Member. The following spiriferid fauna could be distinguished:

Tylothyris mesacostalis (Hall, 1843).
Cyrtospirifer cf. *aquisgranensis* Paeckelmann, 1942.
Cyrtospirifer almadenensis Paeckelmann, 1942.

This fauna is of Famennian age.
Besides the above-mentioned brachiopods, indeterminable fragments of lamellibranchs, corals, bryozoans, and crinoids occur frequently.
The next 2.45 m of the formation is formed by an alternation of thin limestone and shale layers, called the Mampodre Limestone and Shale Member. Thin sections of the limestone show a very fine-grained texture and contain a few small clastic quartz grains as well as small amounts of tourmaline, rutile, zircon, and iron material. Small authigenic quartz concentrations occur frequently.
Fossil fragments are abundant (crinoids, bryozoans, brachiopods, lamellibranchs, conodonts, corals). This rock is a partly recrystallized, rather coarse-grained, biomicrite. Sometimes very thin limestone layers are slightly oölitic (Thin section R.M.G.M. 137525).
The top of this member is formed by a 1.50 m thick coarse-grained limestone containing a large quantity of detrital material. In the upper part of this limestone the first glauconite grains can be found, as well as quartz grains, micas, and accessoria (Thin sections R.M.G.M. 137526 and R.M.G.M. 137527). Again there are many fossil fragments, i.e. bryozoans, crinoids, shell fragments, etc. The limestone ranges between a biomicrite and a biosparite; part of the rock is recrystallized.
The boundary between this limestone and the overlying sandstones with a high glauconite content is a very pronounced disconformity. This strongly-marked line is taken here as the boundary between the Upper Devonian limestone and the Tournaisian Vegamián Formation.

2. *Riosol section.* — Fig. 16, page 72). The Ermita Formation in this section starts with a 4 m thick limestone, unconformably overlying an alternation of thin-bedded limestones, dolomites showing yellowish-brown weathering, and light-grey marly shales. The writer assumes this sequence to be part of the Lancara dolomites, the Upper Devonian erosion having here cut away even the Cambrian nodular limestone, although there is no evidence other than a strong

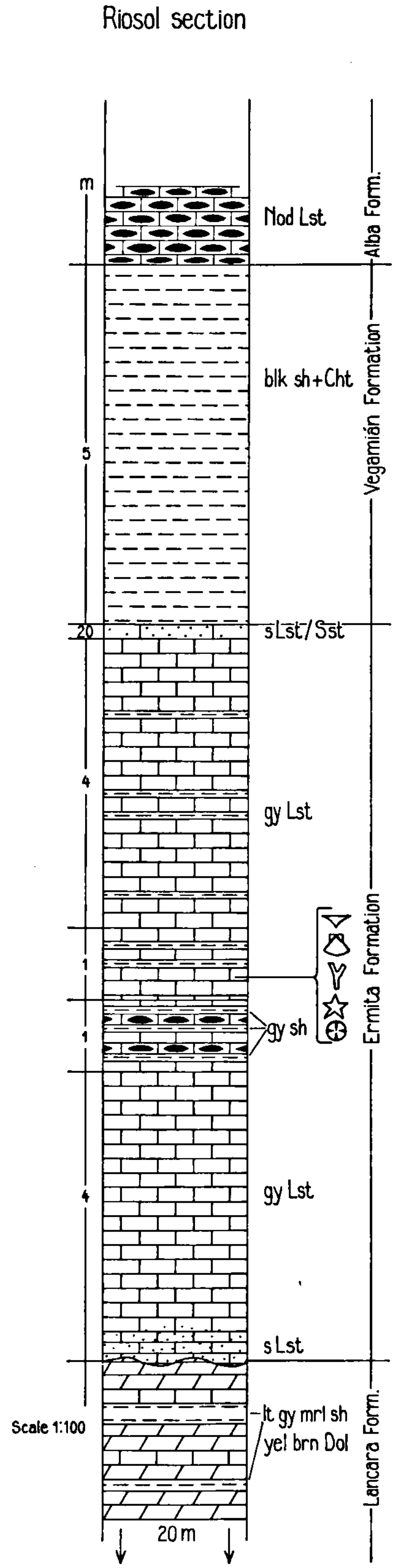

Fig. 16. Columnar section through Upper Devonian and Tournaisian deposits south of the Chapel of Riosol.

resemblance in lithology and the presence of numerous trilobite fragments.

The limestone just above the unconformity is thick-bedded, greyish, and shows a fine-grained texture. Some thin silicified inlayers can be observed. The lower part is coarse-grained, and contains much detrital material; the lowermost 5 cm proved to be a sandstone with a strong calcite cementing. In this section the well-known thin quartz pebble layer is not developed.

In the limestone on top of a thin-bedded, 1 m thick, rather nodular limestone alternating with greyish shales, the writer collected the following spiriferid fauna, as determined by Dr. Th. F. Krans:

Spinocyrtia struniana (Gosselet, 1879).
Cyrtospirifer verneuili (Murchison, 1840).

This fauna indicates an Upper Famennian (Strunian) age. As a matter of fact, the whole Upper Devonian section shows an abundance of brachiopods, lamellibranchs, corals, bryozoans, and crinoids. A sandy limestone layer or sandstone with strong calcite cementing can be found between the above-mentioned Devonian sediments and the overlying black shales.

In this section, it is difficult to draw an exact boundary between the Ermita Formation and the Vegamián Formation.

3. *Cueto Joracao section.* — (Fig. 17, page 73). This section, located only a few hundred metres south of the Riosol section, shows a quite different alternation. The transgression starts with 1.75 m of sandy limestone, originally a biosparite but due to recrystallization now presenting a very coarse-grained texture (Thin section R.M.G.M. 137528).

In this limestone, clastic quartz grains and accessoria such as tourmaline (green, brown, and blue), zircon, apatite, and rutile occur frequently. Authigenic quartz grains, the crystals showing distinct euhedral forms, are not exceptional.

Fossil fragments including remains of brachiopods, bryozoans, crinoids, and corals are abundant.

A gradual transition can be observed between this limestone and the overlying quartzitic sandstones (4.35 m), the first 3.25 m of which presents many holes due to solution of pre-existing fossils.

The sandstone contains many tourmaline, zircon, and rutile crystals (Thin sections R.M.G.M. 137529 and R.M.G.M. 137530). Higher in the section, we again find a gradual transition to a sandy biomicrite, partly recrystallized and representing the top of the Ermita Formation (Thin section R.M.G.M. 137531).

4. *Armada section.* — (Fig. 18, page 74). A corresponding section through the Upper Devonian sediments can be taken in the Armada unit, south of the León line (outside the mapped area). This section (Pröpper, internal report) shows great similarity to the Valverde section. Again we find, now resting on the Cambrian Oville Formation, a quartz pebble layer at the base of a sandstone. Then come 2.5 m of limestone, in which

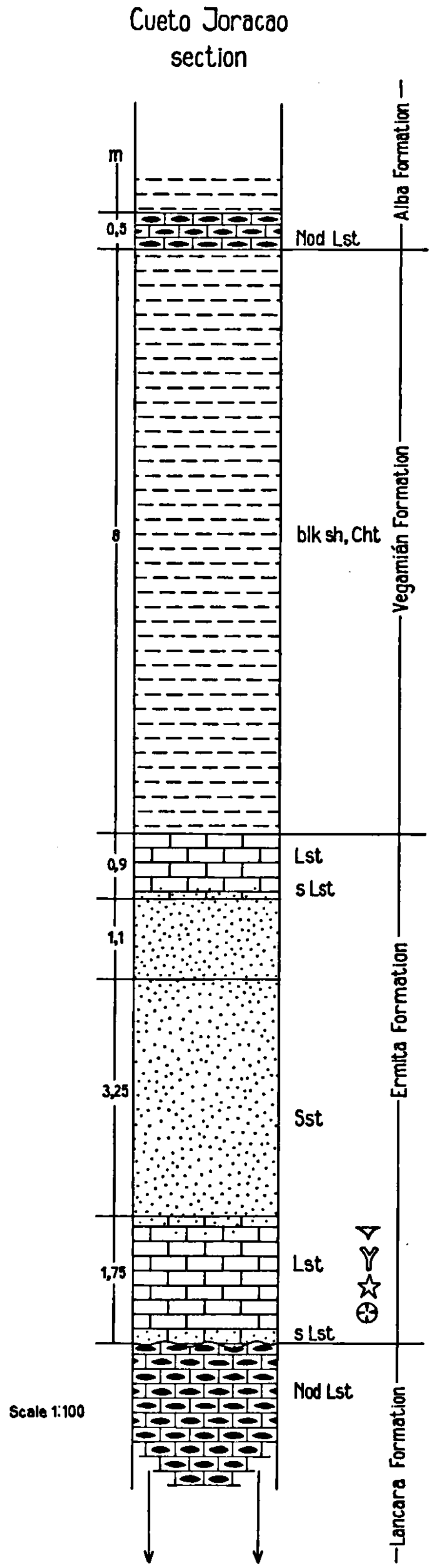

Fig. 17. Columnar section through Upper Devonian and Tournaisian deposits west of the village of Maraña.

Dr. Th. F. Krans could establish the following spiriferid fauna:

> *Fusella tornacensis* (De Koninck, 1883)
> *Cyrtospirifer verneuili* (Murchison, 1840),

which again indicates an Upper Famennian (Strunian) age.

Isidro-Tarna-Ponton subarea

As we have already seen, in the Isidro-Tarna-Ponton subarea, which covers the major part of the mapped area, the Upper Devonian sediments are always deposited on top of the Ordovician Barrios Formation. It will be useful to describe two sections representative of the Upper Devonian as developed in this subarea.

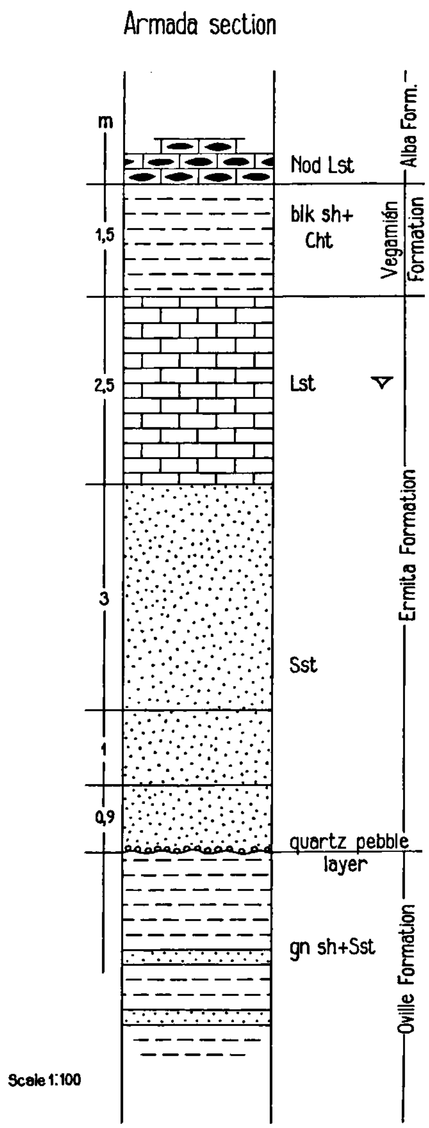

Fig. 18. Columnar section through Upper Devonian and Tournaisian deposits north of the village of Vegamián (Armada unit).

1. *La Uña section.* — (Fig. 19, page 74). The Ermita Formation in this section starts with the well-known quartz pebble layer at the base of a light-grey, brownish-weathered, massive, 5 m thick cross-bedded sandstone which gradually changes into a sandy dolomite.

On top of this dolomite there is 0.5 m of sandstone, followed by a coarse-grained, dark-grey limestone

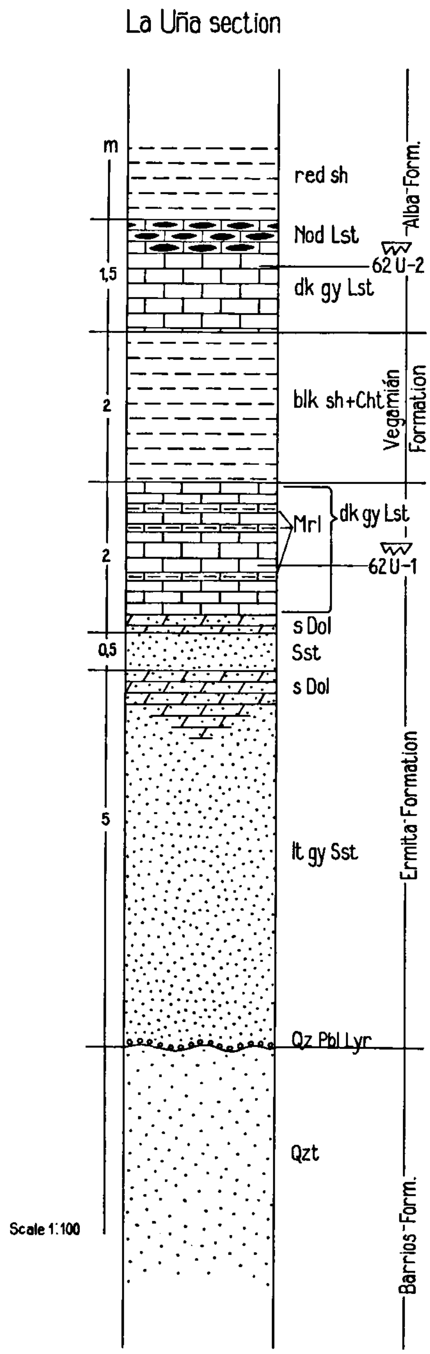

Fig. 19. Columnar section through Upper Devonian and Tournaisian deposits north of the village of La Uña.

alternating with some thin marly layers. The limestone has at its base some dolomite layers with a high content of clastic material.
Mr. H. A. van Adrichem Boogaert has studied the conodont fauna in this limestone; the following species were distinguished (van Adrichem Boogaert *et al.*, 1963):

Loc. 62 U-1:
Polygnathus communis Branson & Mehl, 1934.
Polygnathus cf. *inornata* E. R. Branson, 1934 s.l.
Spathognathodus aculeatus (Branson & Mehl, 1934).
Spathognathodus costatus (E. R. Branson, 1934).

This fauna belongs to the *Costatus* zone of Ziegler (1962), a zone comprising the upper part of the Lower *Gonioclymenia* zone and the Upper *Gonioclymenia* zone (or *Wocklumeria* subzone) of goniatite biozonation. This zone is of Upper Famennian age (the Strunian included).

2. *Felechosa section* (*lower part*). — (Fig. 20, page 75). In this section, a brown dolomite overlying the Barrios Formation is assumed to form the first Devonian sediment, in spite of the fact that evidence for this dating could only be found in the middle of the overlying 7 m thick, coarse-grained, highly detrital limestone. At that level, Mr. H. A. van Adrichem Boogaert could establish an Upper Famennian age (the Strunian included) because of the presence of the following conodont species (van Adrichem Boogaert *et al.*, 1963):

Loc. 62 Fe-1:
Polygnathus communis Branson & Mehl, 1934.
Spathognathodus aculeatus (Branson & Mehl, 1934).
Spathognathodus costatus costatus (E. R. Branson, 1934).

The next layer, a 5 cm thick greyish shale, is followed by a fine-grained, in the upper part often nodular, limestone of Lower Visean age (see p. 79). In this section the Vegamián black shales are not developed.
We correlate the Strunian limestone of the Mampodre-Fontasguera subarea, as dated from a spiriferid fauna, with the Upper Famennian limestone in the Isidro-Tarna-Ponton subarea, as dated by conodonts.

CARBONIFEROUS

Historical review

Carboniferous rocks in the Cantabrian mountain chain have been the subject of many publications since early in the nineteenth century. The report "Minas de Carbón de piedra de Asturias", published in 1831 by Ezguerra del Bayo, Banza, A. de la Torre, and Garcia, was the first really scientific work dealing with the geology of Asturias. The later works of Schulz (1837, 1858), Buvignier (1839), Paillette (1845, 1855), Casiano de Prado (1860), and Verneuil (1882) were of primary importance, and showed that the Carboniferous could be subdivided into three units. Thus, the geological map of Asturias accompanying the important work "Descripción geológica de la provincia de Oviedo" (Schulz, 1858), shows three different colours, reserved for the following subdivision:

Top	Carbonífero rico
	Carbonífero pobre
Bottom	Caliza Carbonera (encrinera)

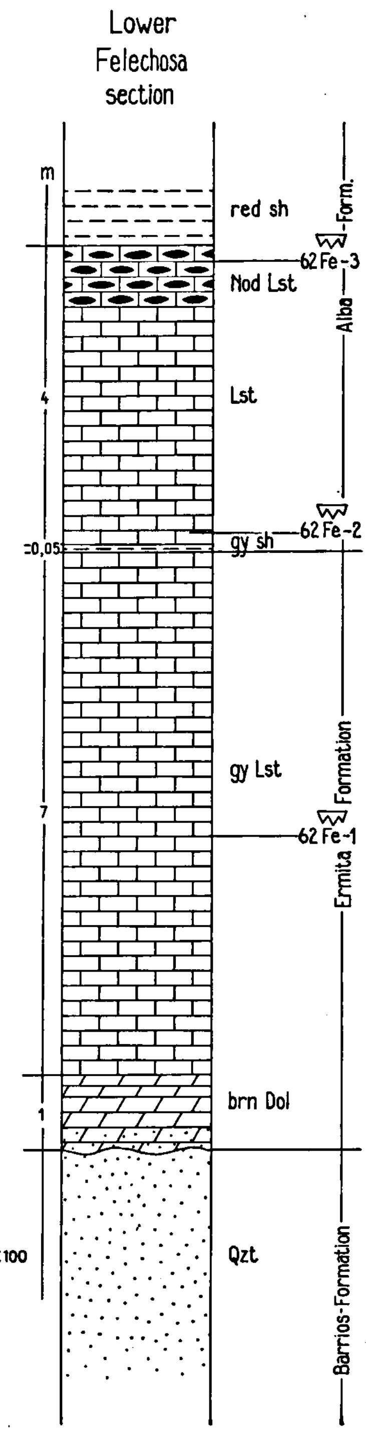

Fig. 20. Columnar section through Upper Devonian and Tournaisian deposits northeast of Puerto de San Isidro.

A few years later, another important work was published by Mallada (1896, 1898), entitled "Explicación del Mapa Geológico de España; Sistema Devónico y Carbonífero", and in 1926 the "Atlas del estudio estratigráfico de la cuenca hullera asturiana" of L. de Adaro was posthumously published by Junguera. After this publications by Patac (1920, 1927, 1943), Hernandez Sampelayo (1946), Kukuk (1927), Quiring (1939), Delépine (1922, 1928, 1932, 1935, 1943), Llopis Llado (1952, 1954), Almela and Rios (1953) and Alvarado (1952) followed.
For more recent data, the reader is referred to Comte (1959), Jongmans (1951, 1952), Julivert (1960), Martinez (1962), Wagner (1959, 1962, 1964), Wagner-Gentis (1962, 1963), van Amerom (1965), Higgins (1962), Higgins *et al.* (1964), Kullmann (1961, 1962, 1963, 1964), Schindewolf & Kullmann (1958), de Sitter (1949, 1955, 1962 b, 1965), van Ginkel (1959, 1965), Brouwer & van Ginkel (1964), Koopmans (1962), Kanis (1956), Helmig (1965), Rácz (1965, 1966), and de Sitter & Boschma (1966).
In the area under consideration the Carboniferous is divided into three groups: The Cea Group, the Yuso Group, and the Ruesga Group.

Ruesga Group

In the Ruesga Group the following formations have been distinguished:

Ricacabiello Formation
Caliza de Montaña Formation
Alba Formation
Vegamián Formation

Vegamián Formation. — In many places a sequence of black phosphatic cherty shales occurs at the base of the Carboniferous. These deposits are assembled in the Vegamián Formation, a term first used by Comte (1959, p. 330) who described the "couches de Vegamian" from a locality south of the village of Vegamián (province of León). The best exposures found in the area treated in this paper are situated in the Valle de Valverde (Valverde section, lower part, p. 71), to the east of the village of Lois (Lois section, p. 78), and west and northwest of the village of Maraña (Riosol section and Cueto Joracao section, pp. 72 and 73).
Because fossils are lacking, the boundary between the Upper Devonian sediments and the first Carboniferous deposits cannot be accurately drawn in any of these sections. There is certainly a very pronounced disconformity between the Strunian argillaceous limestone and an overlying strongly glauconite-bearing sandstone (0.55 cm) (Valverde section, lower part, p. 71), but whether these sediments belong to the Carboniferous is not certain.
Thin sections show that in these sandstones the mineral glauconite occurs frequently, as well as many accessoria such as rutile, tourmaline, zircon, and apatite.
Strong cementing by calcite is present. Some of this calcite has a certain iron content, and this phenomenon, being restricted to certain layers, gives the rock a thin-bedded appearance. Often, small cross-bedded laminae can be observed (Thin section R.M.G.M. 137532). The quartz grains are subangular to subrounded and range in size from 30 to 100 μ, though occasionally well-rounded grains of 400 μ are found. There are no signs of secondary overgrowth. Sericitization and replacement by calcite, on the other hand, are not exceptional. The chances are that at least part of the glauconite, i.e. the grains displaying a well-rounded appearance, derived from the Oville Formation, uncovered by the Upper Devonian erosion.
The main part of the formation, however, consists of a few metres (0—15 m) of black, paper-thin bedded, phosphatic shales, sometimes showing various shades of bluish-violet. Frequently, an alternation with thin, predominantly black chert layers is observed.
The phosphate locally causes a strong yellowish-white weathering of the rock, most clearly demonstrated in an exposure just east of the village of Lois (Lois section, p. 78). At other localities the phosphate occurs as phosphatic nodules a few centimetres in diameter. Van Veen (1965, p. 63) carried out a chemical analysis of these nodules and found about 30 % P_2O_5. The nodules are restricted to certain layers, especially where intraformational disconformities can be demonstrated.
Small markasite nodules occur occasionally (Riosol section), but small pyrite crystals are scattered throughout the whole sequence.
In rare instances thin limestone layers can be noticed, as for example in the Valverde section. In thin sections this limestone shows a very fine-grained texture, with only few clastic quartz grains. Small iron-rich lenses of calcite, showing large amounts of clastic material such as quartz grains, glauconite, and accessoria, are much in evidence.
In the Valverde section another thin sandy limestone with well-rounded iron-rich pebbles and chert fragments, can be found. The iron-rich pebbles, often more than 1 cm in diameter, contain much clastic material and sometimes show iron-oölites around a quartz nucleus (Thin section R.M.G.M. 137533).
The abundant occurrence of calcium-phosphatic fossil fragments consisting of conodonts, fish scales, etc. in both limestone layers, proved to be an important characteristic of the formation. In view of the sedimentary history, it is not astonishing that the conodonts assembled from both limestone layers show a strong admixture of older (Upper Devonian) conodonts (van Adrichem Boogaert, in press).

The age of the Vegamián Formation. — The age of the black shale deposits in the Cantabrian mountains has been the subject of much speculation. Barrois, for example, mentioned the black shales below the "marbre griotte" in the neighbourhood of Vallota (Asturias), and was of the opinion, in agreement with Schulz, that the shales must be of Devonian age (Barrois, 1882, p. 548). Adaro, too, supposed a Devonian age, from a comparison of the formation with the "schistes de la Llama". Llopis Llado (1951) put forward the view that these beds overlying the "Cuarcita Armoricana" are comparable with the "Schistes de Luarca" and are consequently of Silurian age. Comte (1959, pp. 330—331) stated that: ". . . . entre les schistes noirs et les grès sous jacents la discontinuité est évidente . . .", and supposed an age differing little from that of the overlying Visean griotte. Ziegler (1959) and de Sitter (1962) saw an analogy with some Pyrenean sections, and for that reason the latter said, ". . . . there is a black shale with some thin chert layers on the top of the Ermitage Formation which in analogy with some Pyrenean sections might represent the Tournaisian" (de Sitter, 1962-b, p. 258).

On the basis of the finding of some goniatite fragments, Budinger & Kullmann (1964, p. 421), on the other hand, assumed an Upper Visean age for black sandy shales in the Gildar-Montó area, southwest of Posada de Valdeón. Higgins, Wagner-Gentis and Wagner (1964) have demonstrated the presence of Tournaisian in the province of León. Because of the resemblance between the Santiago de las Villas section (province of León), treated below, and the Valverde section (upper part), as worked out by the present author (p. 78), it will be useful to discuss the former in more detail. Higgins has studied the conodont fauna from this section (Higgins *et al.* 1964, pp. 211—213).

In a 0.25 m thick fine-grained limestone at the base of the Vegamián Formation, Higgins observed *Gnathodus kockeli*, together with other species that can be found in both the Upper Devonian and Tournaisian. The occurrence of the former, known in Europe only from the *Gattendorfia*-Stufe (Tournaisian), together with an abundance of *Polygnathus communis*, is interpreted as indicating a Tournaisian age for the sediment under consideration. The next 0.02—0.05 metres of sandy shales containing phosphate nodules has been demonstrated, on the basis of its conodonts, to be of Middle to Upper Tournaisian age.

The next determinable conodont fauna comes from a grey nodular limestone (base of the Alba Formation) located about 3 m higher in the section, and is of Lower Visean age. Therefore, the intermediate sediments, which include 2.40 m of black and grey shales, must belong either to the Upper Tournaisian or the lowermost Visean.

The same publication has a description of another very similar section, the Genicera section, to which the interested reader is referred (Higgins *et al.* 1964, p. 218—219). In this Genicera section a spiriferid and productoid fauna found in the top part of the black shales, points to a Lower Carboniferous age (Upper Tournaisian or Lower Visean) (Winkler Prins, in preparation).

Alba Formation. — A red, green, or grey nodular limestone in alternation with red or green shales and radiolarites can be found overlying the Vegamián Formation, or, when the latter is absent, the Upper Devonian. A limestone-shale alternation of this character is present over a very wide area, and for that reason has been known for a long time.

The accurate description given by Barrois for the Cantabrian mountains contains, according to structure and colour, the following terminology and definitions:

"Calcaire amygdalin" for a sequence built up by limestone nodules surrounded by shaly material, the nodules bearing a resemblance to amygdaloides.

"Calcaire entrelacé" for a variety of limestone layers in alternation with thin shaly layers.

In addition, the local stone masons (the rock being very useful as building material) had their own denominations, using the terms "marbre griotte" and "marbre campan", depending upon whether the shales covering the nodules were red or green, respectively.

In recent publications the rock-stratigraphic unit Alba Formation is used (de Sitter, 1962-b; Koopmans, 1962; Helmig, 1965; Rupke, 1965; van Veen, 1965 etc.) in analogy with Comte (1959, p. 330) who distinguished the "Griotte de Puente de Alba" already described by Verneuil (1882) and Barrois (1882, p. 577).

In the map and sections this formation has been drawn on an exaggerated scale; in reality, the excellent marker horizon varies in thickness between only 3 and 25 metres. An accurate, detailed description of the lithology of the formation has been given by Koopmans (1962, pp. 151—153).

To give some idea of the Alba Formation in the present area the examples given below will be useful, although none of them is really representative because these condensed sequences of nodular limestone-shale deposits are quite variable in their lithological characteristics. Two basic types of the griotte-sequence are nevertheless recognized here:

1. A gradual transition of the black shales (Vegamián Formation) to greenish mudstones, followed by an alternation of nodular limestone, shales, and radiolarites (Lois section, Fig. 21, p. 78). The exact boundary between the Vegamián and Alba Formations in such sections is difficult to draw.
2. A pronounced transition between black shales, overlain by very fine-grained violet-brownish limestones, the top part being a true nodular limestone, followed by red or green shales and radiolarites, in their turn overlain by nodular limestone (La Uña section, p. 74; Valverde section, upper part, Fig. 22, p. 78).

In the area considered in this paper, the writer could not demonstrate any disconformity between the Vegamián black shales and the overlying Alba Formation. However, an unconformable relationship is known from some sections in the Leonide thrust zone, i.e. the Olleros and Santiago de las Villas sections (Higgins *et al.*, 1964, p. 221).

In alternation with the nodular limestone, some shales, shaly cherts, and cherts in varying thicknesses are frequently found. Generally, the chert and associated shales are red, brownish-red, purple, or green; less commonly they are black or yellowish-grey.

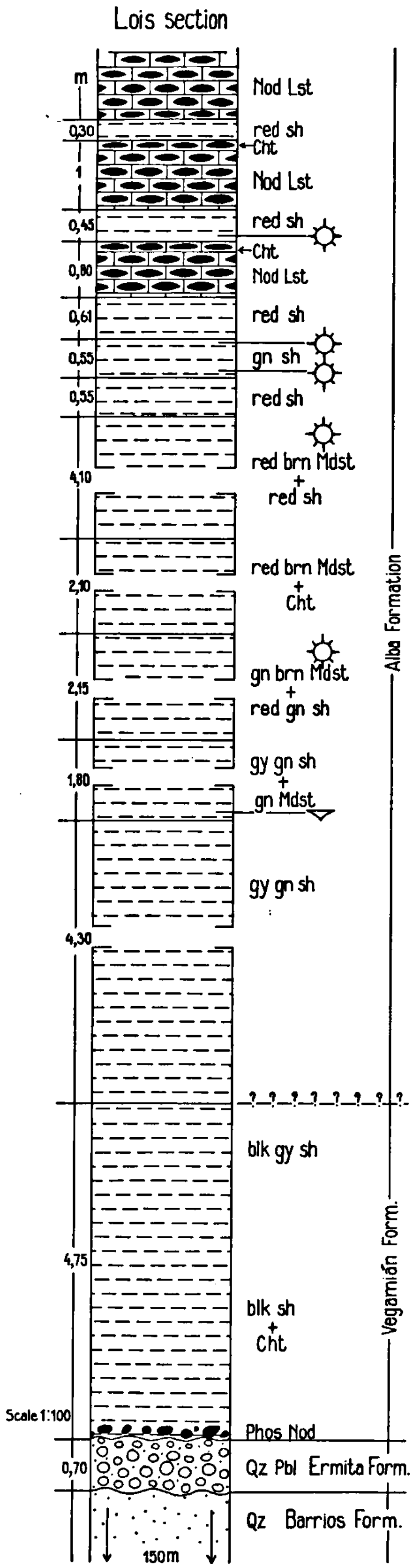

Fig. 21. Columnar section through the Vegamián and Alba Formations east of the village of Lois.

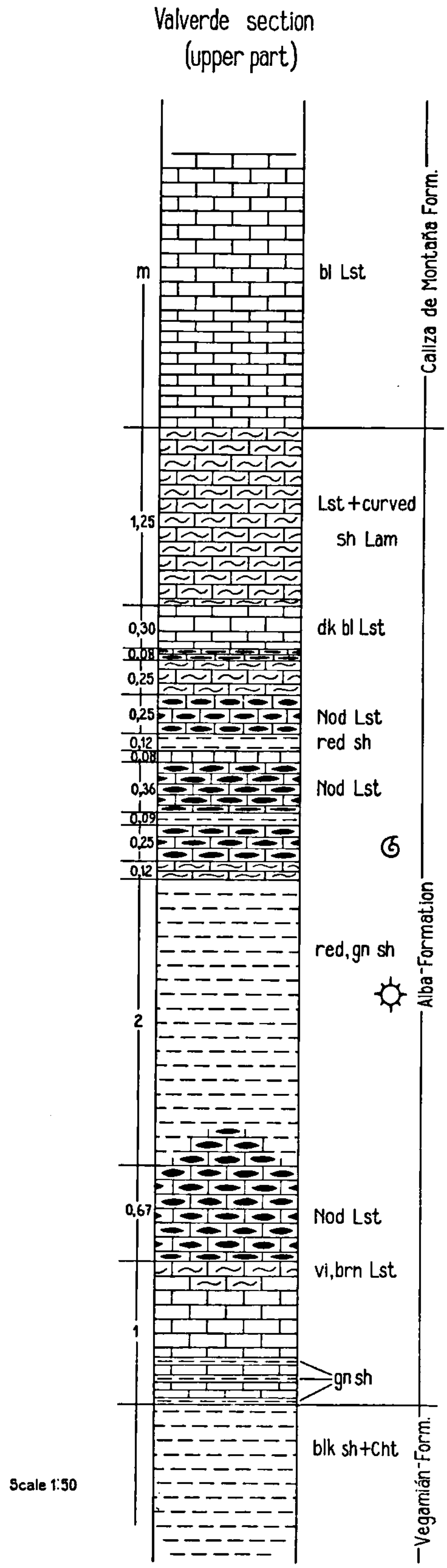

Fig. 22. Columnar section through the Alba Formation west of Pico de Mampodre (valle de Valverde).

The average thickness of the chert layers is about 4 cm, whereas the shales are paper-thin bedded. The separation between the chert and shale is very sharply defined.

Almost invariably the colour of the shale is the same as that of the associated cherts. In the radiolarites, or better shaly radiolarites or radiolarian cherts, the radiolaria are scattered abundantly throughout the sediment. Under the microscope the radiolaria show a circular or elliptical outline and lie in a fine-grained siliceous matrix, coloured red by iron oxide (Thin section R.M.G.M. 137534), Fig. 23, p. 79.

For detailed information dealing with the lithological characteristics of the cherts and shales, the cause of their rhythmic alternation, and the origin of the abundance of silica in the sediment, the reader is referred to the description of the radiolarian cherts of the Franciscan Group given by Davis (1918), the report on the radiolaria-bearing deposits investigated by Grunau (1947, 1959), and lastly, to Pettijohn's "Sedimentary rocks" (1957, pp. 432—445).

The origin of the nodular limestone is still under discussion. Oele (1964, pp. 49—50) has discussed the various theories and relevant publications. A concretionary, detrital, or tectonic origin has been presumed. Oele himself considers the Cambrian Lancara nodular limestone and limestone nodules to be the remainders of ordinary beds in which part of the limestone was removed by solution processes in a submarine environment.

Recently, Nagtegaal (pers. comm.) studied the problem of the origin of nodular limestone, and came to the conclusion that, in a lime-mud deposited in an oxidizing environment and coming to lie, because of continued sedimentation, in a zone in which the pH ranges between 7 and 7.8, the calcium carbonate in the lime-mud will dissolve into the circulating water. This solution, when it penetrates into the overlying zone (pH $>$ 7.8), will give local precipitation of calcium carbonate, the initial stage of the later nodule.

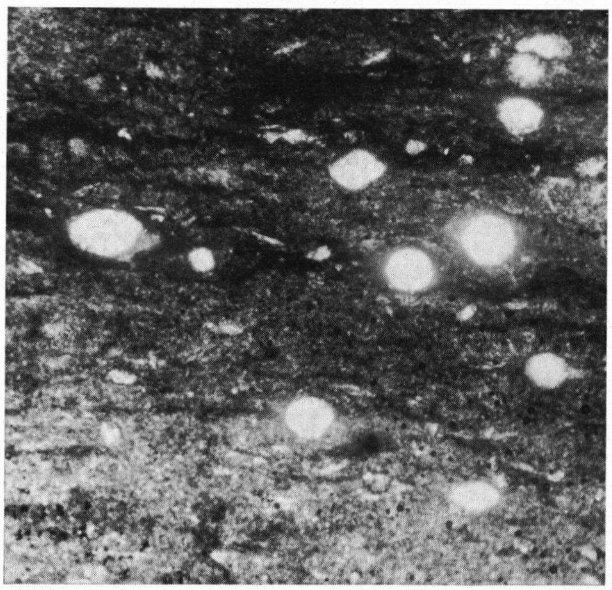

Fig. 23. Alba Formation; radiolarian chert with circular and elliptical radiolaria. (40 ×)

The age of the Alba Formation. — Already in 1882 Barrois demonstrated an Upper Visean age for the Lower Carboniferous nodular limestone in the Cantabrian mountains, a conclusion which was confirmed by Delépine (1928, 1932, 1935, 1943), who described several goniatite faunas of this age. In spite of this, Hern. Sampelayo & Kindelan (1950) and Garcia Fuente (1952) supposed a Tournaisian age. More recently, the formation yielded another comprehensive goniatite fauna, which was studied by Kullmann (1961, 1962, 1963, 1964) and Wagner-Gentis (1960, 1962, 1963). Furthermore, a wealth of data has been acquired by means of conodont-stratigraphy (Higgins, 1962, Higgins *et al.*, 1964; Budinger, 1965; van Adrichem Boogaert, 1965, van Adrichem Boogaert *et al.*, 1963). Thanks to this new approach, we now know that in the area treated in this paper, the base of the Alba Formation was deposited during the Lower Visean (van Adrichem Boogaert *et al.*, 1963). The top of the formation is considered to be of Upper Visean age, although elsewhere, outside the mapped area, it may locally continue into the Lower and Middle Namurian A (Higgins, 1962; Wagner-Gentis, 1963).

According to van Adrichem Boogaert (van Adrichem Boogaert *et al.*, 1963), the La Uña section (p. 74) shows the following Lower Visean conodont fauna:

Loc. 62 U-2:
- *Gnathodus semiglaber* (Bischoff, 1957).
- *Polygnathus inornata* E. R. Branson, 1934 s.l.

The Felechosa section, lower part (p. 75) has yielded a Carboniferous conodont fauna from two levels (van Adrichem Boogaert *et al.*, 1963):

Loc. 62 Fe-2:
- *Gnathodus semiglaber* (Bischoff, 1957).
- *Gnathodus texanus* Roundy, 1926.
- *Hindeodella segaformis* Bischoff, 1957.
- *Pseudopolygnathus triangula pinnata* Voges, 1959.

Loc. 62 Fe-3:
- *Gnathodus delicatus* Branson & Mehl, 1938
- *Gnathodus semiglaber* (Bischoff, 1957).
- *Gnathodus texanus* Roundy, 1926.

Both faunas indicate a Lower Visean age, corresponding to the middle and upper part of the *Pericyclus* zone of goniatite biozonation, respectively. From similar sections located just outside the mapped area, we know that nodular limestone, overlying the red shale-radiolarite sequence, which in turn overlies the Lower Visean part of the griotte-section just discussed, has yielded conodont faunas of Upper Visean age.

To conclude the discussion of the Alba Formation, we may mention a spiriferid fauna found together with an abundance of *Lingula* c.f. *mytilloides* Sowerby in the greenish mudstones of the Lois section (see p. 78), a fauna which also indicates a Lower Carboniferous age (Winkler Prins, pers. comm.).

Caliza de Montaña Formation. — The Alba Formation gradually passes upwards into dark-coloured well-bedded limestone, which in its turn changes into a light-grey massive limestone. In our area the uppermost part of this limestone (0.5—2 m) always consists of a manganese-bearing, dark-grey limestone weathered to a violet-brown. The whole formation varies strongly in thickness, i.e. from 50 to 400 m.

Historical review. — In 1858 Schulz called the thick Lower Carboniferous limestone "Caliza Carbonera". A few years later, Barrois (1882) introduced the name "Calcaires des Cañons", a denomination occasionally still used in modern times (Delépine, 1943; Comte, 1959).

Mallada (1898) and Adaro (1916), taking into account the locally very rich ore deposits in this limestone, used the now rejected term "Caliza Metalifera". The latter (1916) also used the term "Caliza de Montaña", a denomination frequently found in recent publications (de Sitter 1961, 1962, 1963, 1965; Julivert, 1960; Martinez, 1962; etc.).

In 1964 Brouwer and van Ginkel proposed that the rock sequence be given a rock-stratigraphic name, and introduced the term Escapa Formation without, however, giving a well-described type section of the limestones occurring on the Sierra de Escapa (Asturias).

The age of the Caliza de Montaña Formation. — As we have seen (p. 79), the uppermost part of the Alba Formation, where it gradually passes into the overlying limestone, has locally yielded a goniatite and conodont fauna of Lower and Middle Namurian A age. Generally,. however, only Upper Visean faunas have been found in the upper part of the Alba Formation. Furthermore, goniatite fragments found in the limestone overlying the griotte near Santibañez de Resoba (Palencia) even indicate an Upper Visean age (Kullmann, 1961, p. 228). From these data, it is clear that the dividing line between the Alba and Caliza de Montaña Formations is not a synchronic one.

Some fusulinids have been found locally (San Emiliano area) in the upper part of the limestone. The very scanty faunas indicate a Bashkirian age (van Ginkel, 1965, p. 186). In the area treated in this paper, however, we were unable to confirm this because the limestone did not yield fusulinid faunas.

The manganese-bearing top layer of the formation has yielded an excellent goniatite fauna in a locality along the Porma river, just south of the mapped area (Armada unit), a fauna which belongs to the *Reticuloceras* zone (Lower Namurian B) (Kullmann, 1962). In addition to goniatites, a brachiopod fauna, consisting mainly of *Martinia* sp., can be found in many places in this top layer, although without indicating anything more than a Carboniferous age (determination by Krans, pers. comm.). By far the greatest part of the Caliza de Montaña Formation, and especially its dark-coloured, well-bedded, rather bituminous base, is very poor in fossils, however; only undeterminable crinoid debris, coral fragments, and small shell particles can be distinguished.

As already mentioned, the thickness of the Caliza de Montaña Formation varies considerably. The formation was found to be thinnest where the Barrios Formation (+ Oville Formation) is lacking, and consequently at places where, during the Bretonic phase, elevation was maximal (Mampodre-Fontasguera-Ten subarea).

It seems likely that during the deposition of the Caliza de Montaña Formation, the subsidence of the basin was relative small in this subarea. (For details, see p. 106). It seems justifiable to assume that during the deposition of the Caliza de Montaña Formation and the brownish-red to greenish mudstones overlying it, there where periods of non-deposition or at least slow sedimentation in this subarea. The following evidence supports this supposition:

1. The thinning out of the Caliza de Montaña Formation, going from the west and north toward the Central Mampodre-Fontasguera-Ten subarea (see the isopach map in Fig. 44, p. 107).
2. The occurrence of a manganese-bearing level on top of the formation (see p. 106) the manganese ore being the richest where the formation has minimal thickness. (See also Krumbein and Sloss, 1963, p. 308).
3. The absence of thick Bashkirian deposits, as known from the San Emiliano area (San Emiliano Formation, van Ginkel, 1965, p. 186), normally overlying the Caliza de Montaña Formation in the Leonide thrust zone (see p. 83).
4. The occurrence of reddish-brown mudstones containing chert nodules, overlying the above-mentioned manganese-bearing level (see p. 82).

No angular unconformity between the Caliza de Montaña Formation and the overlying sequence could be demonstrated anywhere.

Lithology of the Caliza de Montaña Formation. — Thin sections of the Caliza de Montaña indicate that the rock is built up of:

a) Allochemical grains, such as intraclasts, pellets, oölites (scarce), and fossil fragments (scarce).
b) Terrigenous constituents, such as quartz grains and clayey material.
c) Orthochemical constituents, such as microcrystalline calcite and sparry calcite.

According to Folk's (1959) classification, various types of rocks can be distinguished within the Caliza de Montaña Formation.

1. Micrite. — This rock (Thin section R.M.G.M. 137535), which is well developed in the lower parts of the formation, consists almost entirely of microcrystalline calcite, the original bedding being faintly visible due to thin, sometimes stylolitic, laminae of clayey and iron-rich material.

The rock is platy, well layered, and in places bituminous, with a dense and homogenous texture. Only very small quantities of allochem grains (fossil fragments) can be demonstrated.

In some places an alternation of limestone and thin chert layers can be observed; locally, the chert has grown out into thin lenses varying in thickness (Fig. 24, p. 81). In thin sections, part of the micrite shows all the peculiarities of the rock called by Folk "dismicrite", i.e. rounded to elliptical forms of sparry calcite, with sharp boundaries, are enclosed by the micrite (Folk 1959, p. 28) (Thin section R.M.G.M. 137536).

Often, due to recrystallization which gave the rock a coarser texture, the original micrite has become a microsparite (Folk 1959, p. 32) (Thin section R.M.G.M. 137537). Even much coarser textures occur. Thin section R.M.G.M. 137538, for example, shows a biomicrosparite which was originally a biomicrite whose former ooze matrix was completely recrystallized to 30—90 μ microsparite, leaving the shell fragments unaffected. In a small zone bordering the shell fragments, the original microcrystalline calcite is still visible (Fig. 25, p. 81).

2. Intrasparite. — In this type of rock (Thin section R.M.G.M. 137539), Fig. 26, p. 81, by far the most common allochem grains are intraclasts, sparry calcite filling the pores between them. These intraclasts consist mainly of micrite or pelmicrite. Oölites, pellets, and fossil fragments (bryozoans, algae, corals, shell fragments, and crinoid stems) occur only in minor quantities. Occasional intraclasts exceed 1 mm, in which case use of the term intrasparudite is justifiable.

In one locality (Mampodre region) the light-grey massive limestone contains a rather large amount of oölites; there this rock is an oölitic intrasparite.

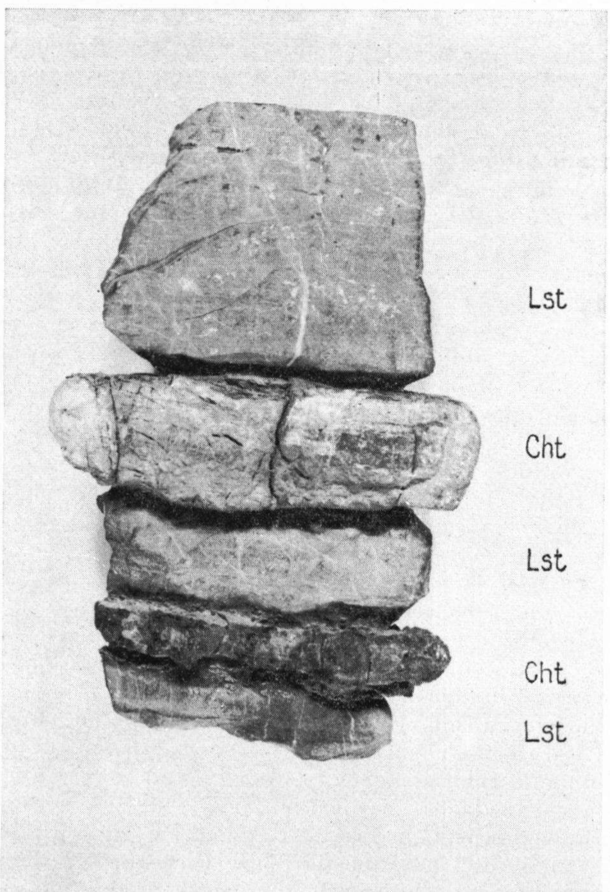

Fig. 24. Caliza de Montaña Formation; alternation of thin chert and limestone layers. (natural size)

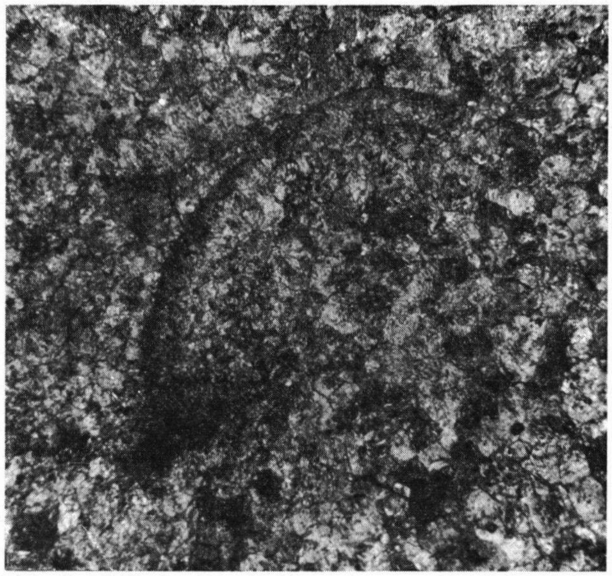

Fig. 25. Caliza de Montaña Formation; biomicrosparite, the shell and its surroundings unaffected by recrystallization. (100 ×)

Fig. 26. Caliza de Montaña Formation; intrasparite, partly recrystallized. (10 ×)

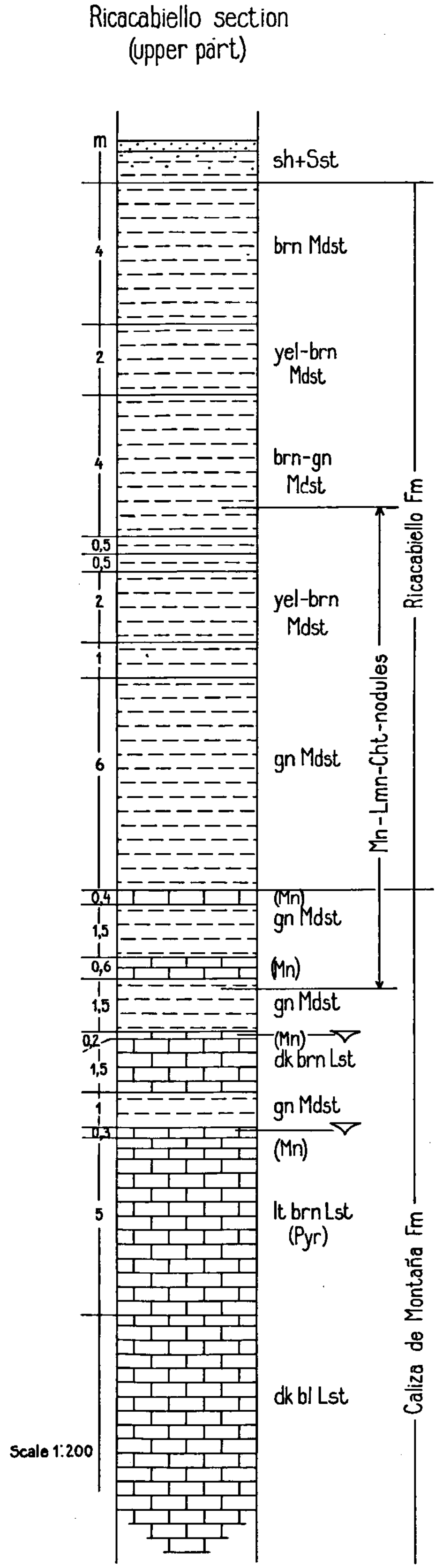

Fig. 27. Type section of the Ricacabiello Formation south of Pico Ricacabiello.

As in the limestone-dolomite sequence of the Lancara Formation, authigenic bipyramidal quartz crystals can be found in the Caliza de Montaña Formation. These authigenic quartz crystals have been known for a long time. De Maestre (1864, p. 47) and Barrois (1882, p. 40) already regarded them as useful stratigraphic marker objects for this Lower Carboniferous limestone.

The crystals, which vary in size from 80 μ to 2 cm, do not show any evidence of a detritic core; only zones of small carbonate inclusions parallel with the crystal faces of the bipyramidal quartz could be established. For more detailed information on this subject, the reader is referred to Koopmans (1962, pp. 154—155). The present writer believes the authigenic mineral to be restricted to the light-grey, rather massive intrasparite overlying the well-bedded dark micrite (microsparite).

At many places the limestone is strongly converted into a white or yellowish-brown dolomite. This dolomitization is in no way related to the bedding; Julivert (1960, p. 108) and Martínez (1962, p. 44) speak of "bolsadas de cotorno irregular".

Where dolomitization took place, fossil fragments in the upper part of the Caliza de Montaña Formation were also converted into dolomite. The best exposure of dolomitization is visible just north of the village of Solle and west of the village of Cofiñal. It is a striking peculiarity of the latter locality that dolomitization appears to be restricted to the rather complicated fault zone of the "Minas de Talco", just northwest of the village of Puebla de Lillo, a fault relationship that could not be assumed for the dolomites north of Solle. The dolomitized fossil fragments, the irregular outline of the dolomite mass, and the observed relationship with a fault zone, all suggest a secondary replacement of calcite by dolomite.

In some places there was a strong mineralization of the limestone, the mercury deposits south of the Tarna pass and those east of the village of Lois being the most important. In addition, antimony, fluorite, azurite, malachite, and talc can be found in varying quantities (see p. 124).

Calcite veins and occasional quartz veins cut through the whole rock. The calcite veins occasionally contain few quartz crystals.

Ricacabiello Formation. — The sediments directly overlying the Caliza de Montaña Formation consist, as we have seen, of brownish-red to greenish mudstones, varying in thickness from 5 to 35 metres. In these mudstones, for which the rock-stratigraphic name Ricacabiello Formation is introduced here, chert-, limonite-, and manganese-bearing nodules (0.5—2 cm in size) are found frequently.

In the type section (Fig. 27, p 82 and Fig. 28 B, p. 83), located just south of the Pico Ricacabiello, these mudstones are overlain by a sequence of almost fossil-free clastic sediments in which there is a 9 m thick oölitic limestone (the Lázaro Limestone Lense) lying

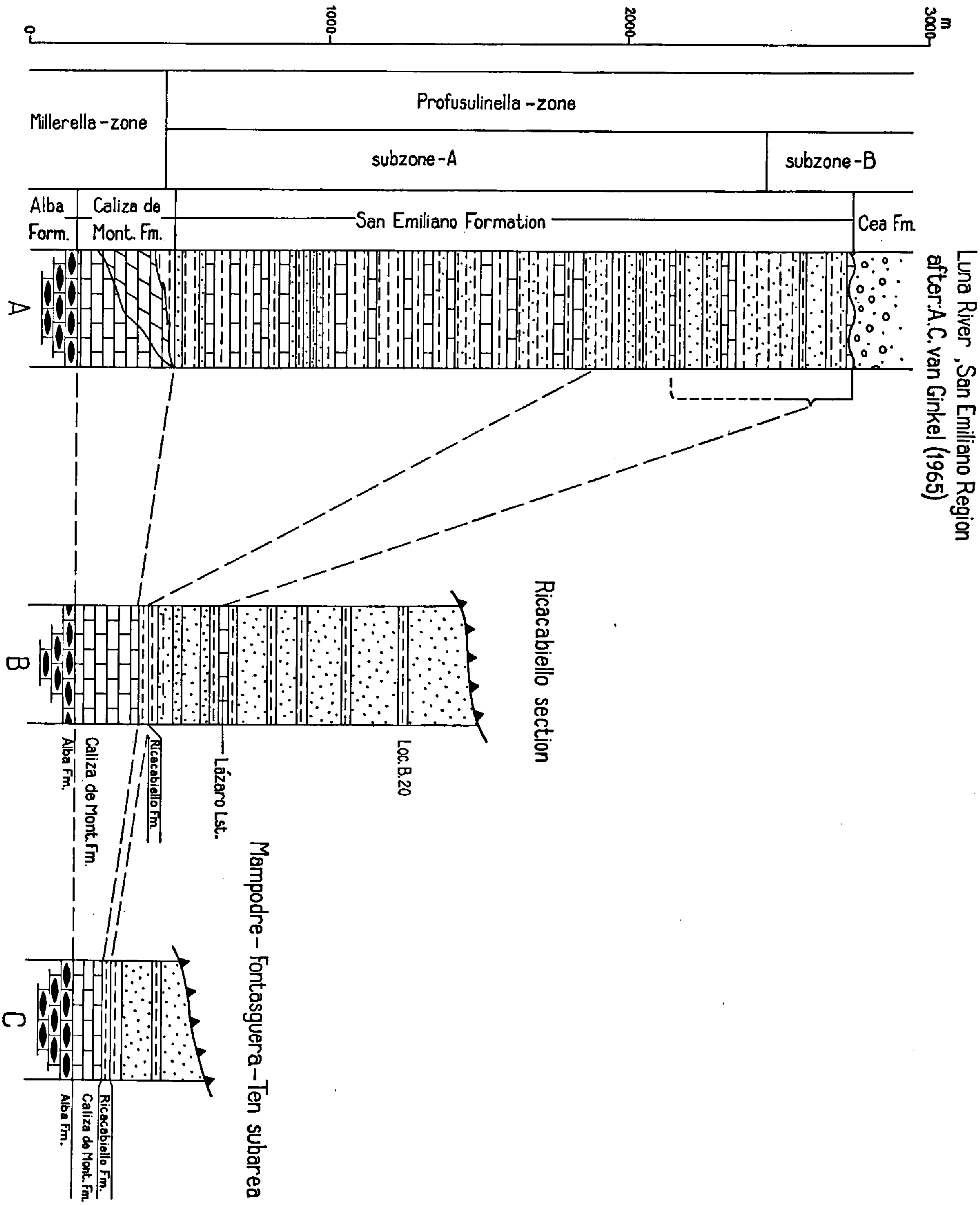

Fig. 28. Comparison between the Ricacabiello Formation and the San Emiliano Formation (Luna river).

about 250 m above the Ricacabiello Formation. This limestone is of Vereyan age (see p. 92).
In the stratigraphic column given by van Ginkel for the San Emiliano region (Fig. 28 A, p. 83) a horizon of this same age could be expected at the top of the San Emiliano Formation (van Ginkel, 1965), which in this case would mean that its position would be at least 1500 m stratigraphically above the top of the Caliza de Montaña Formation.
If it is assumed that the top of the Caliza de Montaña Formation is more or less of the same age in both areas, the greatest part of the San Emiliano Formation in the Leonide thrust zone seems to be represented by the mudstones of the Ricacabiello Formation in the mapped area. This suggests a period of slow sedimentation or non-deposition after the deposition of the Caliza de Montaña Formation, which, in its turn, was also deposited locally under conditions of slow sedimentation.
In the Mampodre-Fontasguera-Ten subarea, excellent examples of the Ricacabiello Formation can also be found. Due to structural events, however, any limestone outcrops within the clastic sediments overlying the Ricacabiello Formation are concealed below the thrust sheets (Fig. 28 C, p. 83), the whole subarea being surrounded by low-angle overthrusts cutting away a much greater part of the stratigraphic column than in the Ricacabiello section.
In the absence of the Curavacas folding phase, the boundary line between the Ruesga and Yuso Groups has been drawn at the top of the Ricacabiello Formation.

Yuso Group

The following list shows the formations of the Yuso Group according to the basins in which they can be found (Fig. 46, p. 109).

1. Piedrafita-Lillo basin
   a. Lena Formation.
2. Lois-Ciguera basin
   a. Lois-Ciguera Formation.
3. Beleño basin
   c. Fito Formation
   b. Caliza Masiva Formation
   a. Beleño Formation.
4. Maraña-Retuerto basin
   b. Lechada Formation
   a. Curavacas conglomerate beds and lenses of the Lechada Formation.

The deposits of the Yuso Group consist principally of an alternation of shales, greywackes, sandstones, limestones, and occasionally some conglomerate beds and coal seams.
Stratigraphic and paleontological investigations were carried out mainly by van Ginkel (1959, 1965) and Rácz (1965, 1966), who studied respectively the fusulinid fauna and algal flora in the various limestone members. More recently, algal associations were studied by Mr. G. J. B. Germs. From these studies, a biozonation could be established (Fig. 29, p. 85).

Piedrafita-Lille basin

Lena Formation. — The name "Grupo de Lena" was introduced by Barrois (1882) for the Carboniferous rocks overlying the Caliza de Montaña Formation in the Central Basin of Asturias near the village of Pola de Lena.
The rock-stratigraphic term Lena Formation, given by van Ginkel (1965), is here used for deposits occurring in the Carboniferous basin between the Upper Torio and Porma rivers. This basin, for which the name Piedrafita-Lillo basin was introduced (Rácz, 1965, p. 13), is the direct continuation to the southeast of the Central Basin of Asturias.
An important section, the Vegarada section, which is situated north of the village of Redipuertas (Curueño river), extends from the Caliza de Montaña Formation, which crops out at the Peña de la Carva, to the limestones crossing the valley betweeen Pico Huevo and Pico Morala in the southwest. This Vegarada section (Fig. 30, p. 86), has been presented by Rácz (1965), although he called the sediment of this section the Lois-Ciguera Formation, following Brouwer and van Ginkel (1964). For a detailed description of the shales, (quartzitic) sandstones, limestones, and coal seams composing the Lena Formation, the reader is referred to Rácz, 1965, pp. 19—22.

TABLE II. Algal flora and fusulinid fauna in limestones of the Lena Formation (Piedrafita-Lillo basin).

Locality number	Calcareous algae determined by Rácz (1965)	Fusulinids determined by van Ginkel (internal report)
Loc. C. 11	*Dvinella comata* Chvorova, 1949 *Girvanella* sp. *Cuneiphycus aliquantulus* Johnson, 1960	
Loc. B. 42	*Dvinella comata* Chvorova, 1949 *Amorfia jalinki* Rácz, 1964	*Eofusulina* sp. *Pseudostaffella* sp. *Ozawainella angulata* (Colani, 1924) *Paræofusulina* sp.
Loc. B. 29	*Dvinella comata* Chvorova, 1949 *Amorfia jalinki* Rácz, 1964 *Pseudokomia cansecoensis* Rácz, 1964 *Komia abundans* Korde, 1951	*Profusulinella* ex. gr. *prisca* (Deprat,1912) *Paraeofusulina* sp. *Ozawainella* sp. *Pseudostaffella* sp. *Profusulinella* sp.
Loc. C. 21	*Pseudokomia cansecoensis* Rácz, 1964 *Komia abundans* Korde, 1951	*Fusulina* sp.

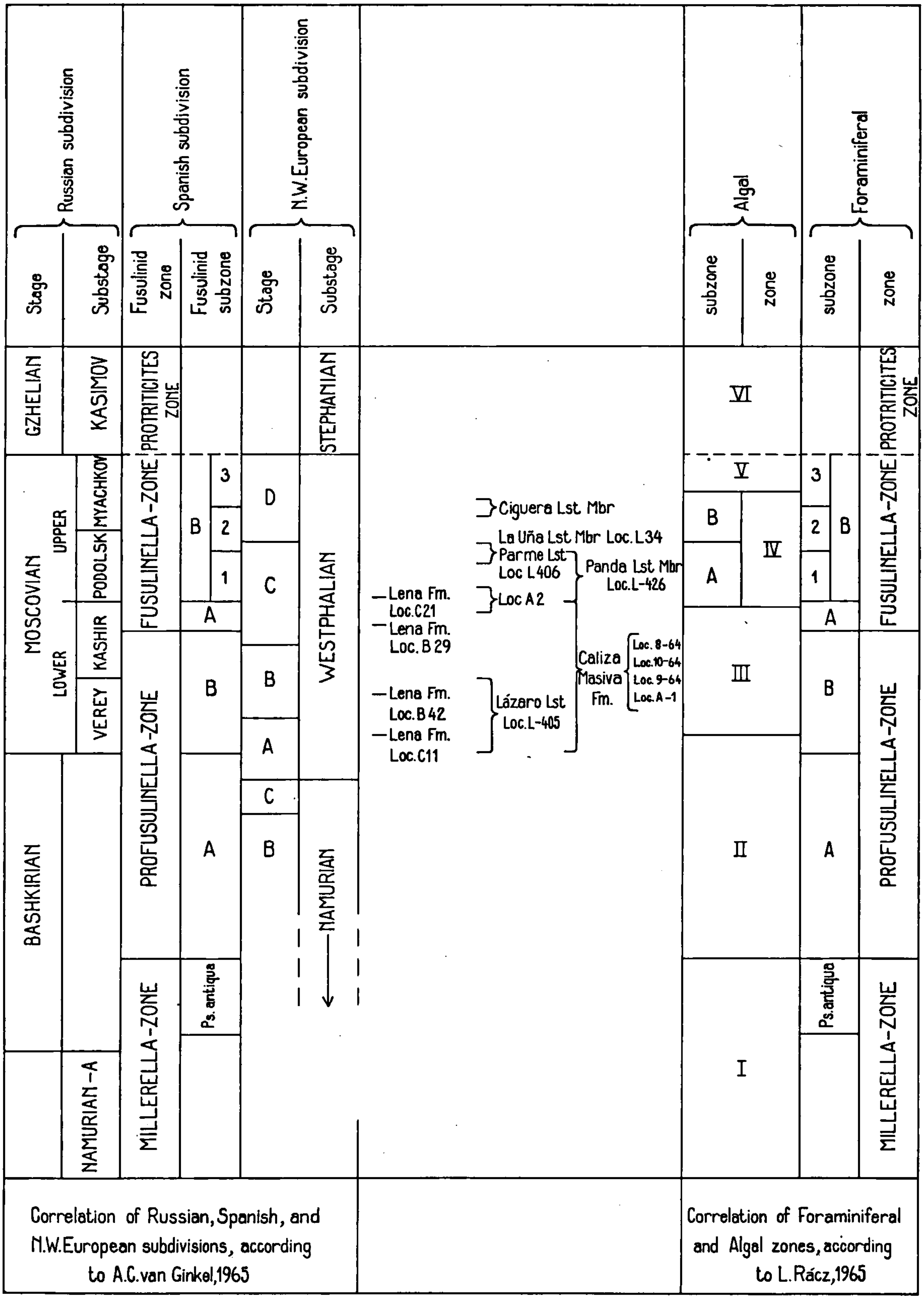

Fig. 29. Correlation charts after van Ginkel (1965) and Rácz (1965).

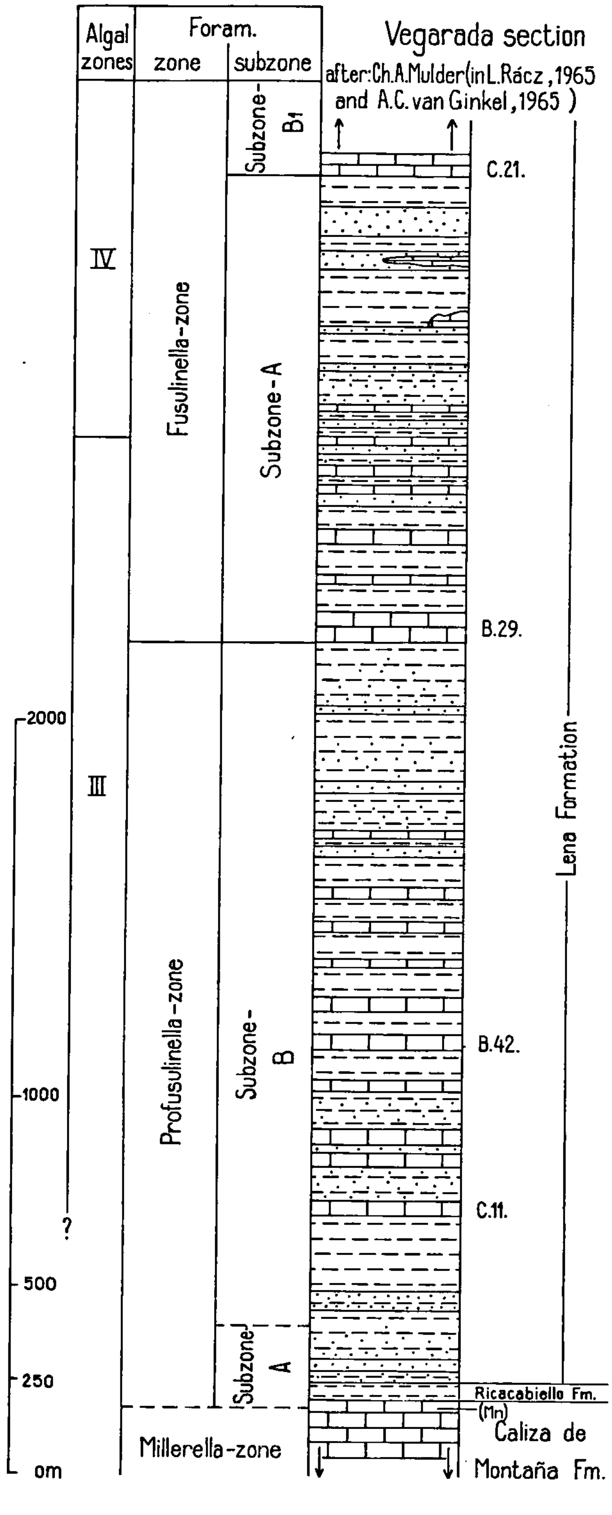

Fig. 30. Columnar section through the Westphalian Lena Formation northeast of the village of Redipuertas.

Biostratigraphy, — In addition to fusulinids and calcareous algae, the limestones of the Lena Formation contain an abundance of brachiopods, gastropods, corals, echinoderms, trilobites, ostracods, and bryozoans.

In the Vegarada section, the algal flora of four limestone layers have been examined (Table II, p. 84). The locality numbers are shown to the right of the stratigraphic column given in Fig. 30, p. 86.

The fusulinid assemblages of the limestones cropping out in the Piedrafita-Lillo basin have not yet been studied in detail throughout the Vegarada section. We know for certain, however, that the first limestone above the Caliza de Montaña Formation in a corresponding section northwest of the village of Puebla de Lillo (Porma river), is of Lower Moscovian age (Profusulinella zone, base subzone B) (van Ginkel, pers. comm.). This author has stated: "The Lena Formation except for the uppermost and possibly the lowermost part of this unit as present in León, has been deposited during the Lower Moscovian as based on fusulinid evidence". (van Ginkel, 1965, p. 187).

The biostratigraphic interpretation of the algal assemblages listed in Table II is in agreement with the data given by van Ginkel. The limestones occurring in the lower part of the formation belong to algal zone III, and the overlying sediments clearly belong to algal zone IV. Again we are faced with the striking fact that comparison of the Vegarada section with the thick column of strata belonging to the San Emiliano Formation, which occurs in the Leonide thrust zone, shows that the same period of deposition is represented in the Piedrafita-Lillo basin by either a much thinner sequence or a hiatus. On this basis, van Ginkel concluded in 1965 that "In León the Lena Formation is found in the Asturides north of the León-line resting paraconformably upon the Escapa Formation. The lowest part of the Lena Formation here is probably a condensed sequence with local nonsequences and even some local angular unconformities" (p. 187). Furthermore, the top of the Caliza de Montaña Formation has a high manganese content, and along the whole northern and eastern border of the Piedrafita-Lillo basin, especially northeast of the village of Puebla de Lillo, excellent examples of the Ricacabiello Formation can be found.

The limestone sequences cropping out in the neighbourhood of the village of Solle, east of the Porma river, are the direct eastern continuation of the Piedrafita-Lillo basin. Still more to the east, these limestones pass laterally into the lower part of the Lois-Ciguera Formation (p. 27).

It will be useful to give some information about the limestones in this easternmost part of the Piedrafita-Lillo basin.

Thin sections examined by Dr. A. C. van Ginkel for fusulinids and by Mr. G. J. B. Germs for algal associations yielded the information collected in Table III.

TABLE III. Fusulinid and algal assemblages of the Lena Formation (sampled near the village of Solle).

Locality number	Fusulinid assemblages	Algal assemblages
763	*Schubertella* sp. *Profusulinella* sp.	*Uraloporella* sp.
764	*Schubertella* sp. *Staffella* sp. *Millerella* sp. *Profusulinella* ex. gr. *librovitchi* (Dutkevitch, 1934a) *Profusulinella* ex. gr. *prisca* (Deprat, 1912) *Profusulinella* sp. *Eofusulina* sp.	*Dvinella comata* Chvorova, 1949 *Uraloporella* sp.
765	*Millerella* sp. *Schubertella* sp. *Pseudostaffella* sp. *Eofusulina* sp. *Profusulinella* sp.	
766	*Fusulina* sp.	
767	*Schubertella* sp. *Pseudostaffella* sp. *Eofusulina* sp. *Profusulinella* sp. *Profusulinella* ex. gr. *parva* (Lee et Chen, 1930)	*Dvinella comata* Chvorova, 1949 *Epimastopora rolloensis* Rácz, 1964
768	*Profusulinella* sp.	*Dvinella comata* Chvorova, 1949 *Epimastopora rolloensis* Rácz, 1964
769	*Profusulinella* ex. gr. *librovitchi* (Dutkevitch, 1934a) *Profusulinella* sp.	*Dvinella comata* Chvorova, 1949
770	*Schubertella* sp. *Millerella* sp. *Eofusulina* sp. *Profusulinella* ex. gr. *prisca* (Deprat, 1912) *Profusulinella* ex. gr. *librovitchi* (Dutkevitch, 1934a)	*Dvinella comata* Chvorova, 1949 *Komia abundans* Korde, 1951
775	*Profusulinella* sp.	*Dvinella comata* Chvorova, 1949
776		*Epimastopora* sp. *Anthracoporella spectabilis* Pia, 1920
777		*Donezella lunaensis* Rácz, 1964 *Dvinella comata* Chvorova, 1949

The fusulinid associations of most of these limestones point to a Lower Moscovian age (Profusulinella zone, subzone B). No Bashkirian deposits could be demonstrated here. These facts justify the supposition of a period of slow sedimentation or non-deposition succeeding the Caliza de Montaña Formation. The youngest limestone found in this part of the basin is of Upper Moscovian age (Fusulinella zone, subzone B; (Loc. 766).

A sharp angular unconformity with limnic Stephanian B deposits (see p. 97), overlying the Lena Formation, can be easily observed.

Lois-Ciguera basin

Lois-Ciguera Formation. — In the southeastern part of the mapped area, west of the Esla river near the villages of Lois, Ciguera, and Anciles, a Lois-Ciguera Formation has been distinguished. This formation, as described by Brouwer and van Ginkel (1964, pp. 310—311) and van Ginkel (1965, p. 188), is composed of a thick sequence of about 1250 m of greywacke, shale, and limestone lithosomes. A striking peculiarity is the high proportion of limestone in this formation, of which Brouwer and van Ginkel wrote, "Ce développement culmine dans les synclinaux de Lois et de Ciguera, où plus d'un tiers de l'épaisseur totale est constituée par des calcaires".

The type section (Fig. 31, p. 88; after de Meyer, in van Ginkel, 1965) runs from Pico Montote, east of the village of Lois, as far as the centre of the Lois syncline. The top of the Caliza de Montaña Formation again shows a high manganese content; the oldest sediments overlying the Ricacabiello Formation belong to the Lower Moscovian, a picture which is common throughout the whole Piedrafita-Lillo basin and confirms our assumption of a period of slow sedimentation or non-deposition. Sedimentation in the Lois-Ciguera basin continued until close to the end of the Moscovian, since fusulinids belonging to the Fusulinella zone, subzone B_2 (or base subzone B_3), have been found in the Ciguera Limestone Member at the top of the formation (van Ginkel, 1965).

To the south and southwest, the formation is unconformably overlain by the above-mentioned Stephanian B deposits.

Beleño basin

Beleño Formation. — In the Beleño basin (Asturias) west of the Picos de Europa, another interesting section through Carboniferous rocks has been described by Julivert (1960). The Caliza de Montaña varies here in thickness between 100 and 250 m, its top showing a considerable manganese content. The Ricacabiello Formation occurs throughout the whole region above this manganese layer. Julivert (1960, p. 50) therefore distinguished in the Alto Corina section a "serie roja inferior", 25 m in thickness and consisting predominantly of red and green mudstones. In the Collado Baxeñu section (Julivert, 1960, p. 61) this author gave the following subdivision for the

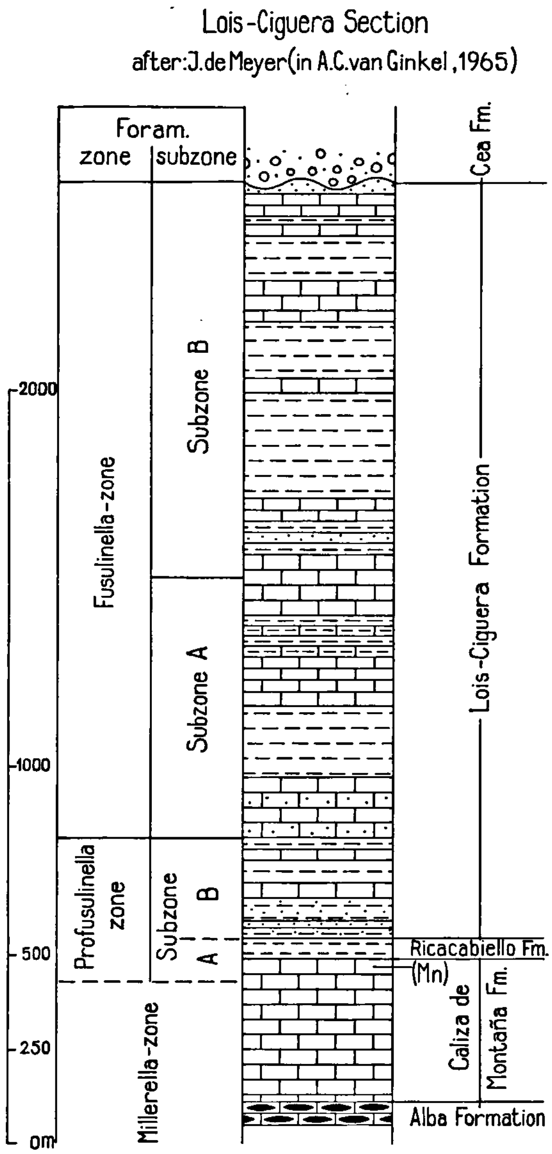

Fig. 31. Columnar section through the Westphalian Lois-Ciguera Formation east of the village of Lois.

sediments directly overlying the Caliza de Montaña Formation, i.e. our Ricacabiello Formation.

	Top
Serie	4. Pizarras verdes y marrones (6 m)
	3. Pizarras con capas manganesíferas (2.5 m)
abigarrada	2. Pizarras con nódulos limoníticos y de manganeso (4 m)
inferior	1. Pizarra roja y verdosa con nódulos de manganeso y limonita (15 m)
	Bottom

The next 250—350 m of sandstones, shales, and thin limestone layers are called "Conjunto pizarroso" by Julivert; the rock-stratigraphic name Beleño Formation, introduced by van Ginkel (1965, p. 190), covers the above-mentioned red and green mudstones plus the "Conjunto pizarroso".

For our area, the present writer proposes to reinterpret the lower part of the Beleño Formation by distinguishing a Ricacabiello Formation and a Beleño Formation overlying it (see Fig. 35, p. 94). A striking aspect is the varying thickness of the Beleño Formation in this area. In the neighbourhood of Puerto Forno a few metres of clastic sediment can be found between the Caliza de Montaña and the Caliza Masiva, whereas to the west this quantity rapidly increases with the splitting up of the Caliza Masiva Formation. A somewhat comparable situation has been mentioned by van Ginkel (1965), who noted the possibility of a thickening of the Escalada Formation (see p. 92) at the expense of the Beleño Formation in the Picos de Europa.

Thin limestone layers in the Beleño Formation have yielded excellent fusulinid faunas and algal floras from various localities. Dr. A. van Ginkel and Mr. G. J. B. Germs have kindly provided full details about these associations (Table IV).

TABLE IV. Fusulinid and algal associations in limestones of the Beleño Formation.

Locality number	Fusulinid assemblages	Algal assemblages
13—64	*Millerella* spp. *Parastaffella* sp. *Profusulinella* sp.	*Epimastopora rolloensis* Rácz, 1964
18—64	*Millerella* spp. *Schubertella* spp. *Parastaffella* sp.	*Epimastopora bodoniensis* Rácz, 1964 *Anthracoporella spectabilis* Pia, 1920

The fusulinid faunas are not useful, but the algal floras point to at least a Lower Moscovian age, because so far, *Epimastopora rolloensis* Rácz, 1964 has never been found in older sediments in the Cantabrian mountains.

The Lázaro Limestone Lense. — (Loc. L. 405, Ricacabiello section, page 83), is also one of the limestones of the Beleño Formation. The present writer has carried out a detailed investigation into the age of the Lázaro Limestone Lense. According to Folk's classification (1959), this limestone is an oösparite with abundant fusulinids and algae (Fig. 32, p. 89).

The age of the Lázaro Limestone Lense has been established by comparing its fusulinid content with closely similar Russian species. From 44 specimens considered to belong to a single species, orientated thin sections were prepared. These specimens were identified as typical examples of the genus *Aljutovella* Rauser-Chernoussova, 1951, showing strong resemblance to the species *Aljutovella elongata* Raus. et

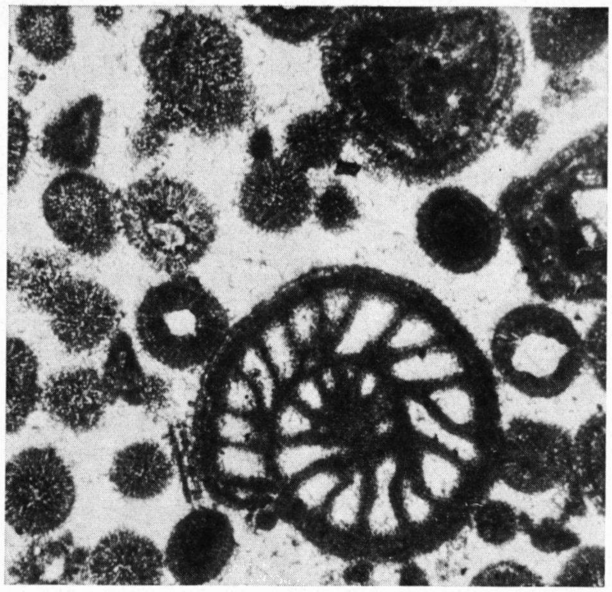

Fig. 32. Beleño Formation, Lázaro Limestone Lense; oösparite with abundant fusulinids. (40 ×)

Saf., 1951+) (= *Profusulinella aljutovica* Rauser-Chernoussova var. *elongata* Rauser-Chernoussova, 1938). We regard our present material to be sufficiently different from the above-mentioned *Aljutovella elongata* to justify the introduction of the subspecies *Aljutovella elongata* Raus. et Saf., 1951 subsp. *lazarensis* subsp. nov. The description of the new subspecies follows below.

Systematic description:

Genus ALJUTOVELLA Rauser-Chernoussova, 1951

+) Rauser-Chernoussova, D. M. *et al.*, Middle Carboniferous fusulinids of the Russian Platform and adjacent regions. Moscow, Akad. Nauk. S.S.S.R., Inst. Geol. Nauk. Minist. Neftianoi Prom. S.S.S.R., 1951, p. 182.

Aljutovella elongata Raus. et Saf., 1951 subsp. *lazarensis* subsp. nov.
Pl. I, Figs. 1—12.

a) Type specimen: Specimen 9 (Pl. I, Fig. 4) is designated as the holotype.
b) Locality: Lázaro Limestone Lense (Loc. L 405, see Fig. 1).
c) Description: Radius vector: 246—505 (in μ)
Form ratio: 1.88—2.76
Number of whorls: 4—5½

From 1st to 5th whorl, test changes from nautiliform, spherical or oval (1st wh.); oval, short fusiform or fusiform (2nd wh.); short fusiform, fusiform or elongated fusiform (3rd wh.); to fusiform, elongated fusiform or subcylindrical (4—5th wh.). The lateral sides are convex to nearly straight, straight, or occasionally concavo-convex; the poles bluntly pointed to rounded.

Dimensions:

Specimen 10: L = 2.39 mm D = 0.97 mm $\frac{L}{D} = 2.47$

Specimen 5: L = 2.31 mm D = 0.90 mm $\frac{L}{D} = 2.58$

Septa straight to weakly folded in inner whorls, in outer whorls weakly to strongly folded in the axial regions. In the last two whorls septal loops are occasionally observed in axial as well as in the lateral portions of the test. In inner whorls these septal loops are indistinct or absent. In sagittal sections, septa slender triangular to rod-shaped, in outer whorls often club-shaped.
The wall shows in some whorls a distinct three-layered structure; occasionally, however, in the outer whorls a weakly differentiated diaphanotheca has been observed. Chomata symmetrical or asymmetrical, distinct up to the ultimate or penultimate whorl, rather massive, rounded to subquadrate, sometimes—especially in inner whorls—asymmetric and relatively wide; steep or gentle slopes at the tunnel side.
Tunnel path varies from symmetric to asymmetric, generally almost symmetric. Average and range of maximum deviation of symmetry respectively 15° and 1—30° (N = 13).
Average and range of tunnel angle from 1st to 4th whorl are respectively 26° and 16—34°, 34° and 22—45°, 44° and 33—60°, 55° and 43—72°.
Axis generally maintains original position throughout growth; occasionally first whorl coiled at an angle to subsequent whorls.

d) Measurements: See Tables VI and VII, pp 90 and 91.
e) Comparisons and remarks:

Our new subspecies resembles *Aljutovella elongata* Raus. et Saf., 1951. *Aljutovella dagmarae* Safonova, 1951, *Aljutovella intermixta* Safonova, 1951 and *Aljutovella devexa* Safonova, 1951.
It is most similar to *Aljutovella elongata* Raus. et Saf., 1951, which differs in the more microspheric state (i.e. smaller proloculum, more volutions, and smaller

TABLE V. Variations in the diameters of the tests, by whorls (in mm).

Specimen number	Initial chamber	1st wh.	2nd wh.	3rd wh.	4th wh.	5th wh.
10	0.056	0.12	0.22	0.39	0.64	0.97
5	0.060	0.13	0.25	0.44	0.69	0.88

TABLE VI

MEASUREMENTS OF *ALJUTOVELLA ELONGATA* RAUS. *ET* SAF., 1951 SUBSP. *LAZARENSIS* SUBSP. NOV.

AXIAL SECTIONS

(in μ)

SPECIMEN		2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	RANGE	AVERAGE
RADIUS VECTOR	0	39	40	34	30	32	36	40	39	28	44	31	32	26	30	30	30	40	29	26–44	34
	1		84	92	62	64	74	77	86	68	92	59	72	61	64	69	70	78	71	59–92	71
	2	131	157	175	117	119	140	143	152	127	161	118	131	109	122	122	132	132	143	109–175	135
	3	209	265	310	201	187	239	241	248	225	257	194	215	207	219	220	225	234	240	187–310	230
	4		414	462	328	304	382	422	394	363	367	299	350	347	363	333	353		372	299–462	366
	5				486					525											
PERCENTAGE INCREASE OF RADIUS VECTOR	1–2		88	91	90	87	90	87	77	88	76	100	82	81	92	79	90	69	103	69–103	86
	2–3	59	68	77	73	57	70	68	63	77	59	65	64	89	78	80	70	77	68	57–89	70
	3–4		56	49	63	63	60	75	59	61	43	54	63	68	66	51	57		55	43–75	59
	4–5				48					45											
FORM RATIO	1		1,48	1,13	1,03	1,01	1,19	1,17	1,20	1,13		0,97	0,91	1,04	0,94	1,27	0,89	0,99	0,97	0,89–1,48	1,08
	2	1,36	1,77	1,70	1,36	1,13	1,77	1,69	1,68	1,46	1,50	1,50	1,61	1,38	1,38	1,96	1,78	1,52	1,47	1,13–1,96	1,55
	3	1,56	2,06	1,46	2,10	1,77	2.09	1,88	1,84	2,00	1,95	1,82	2,22	2,16	2,78	2,31	2,67	2,12	2,06	1,46–2,78	2,04
	4		2,18	2,36	2,35	2,39	2,10	1,88	2,23	2,11	2,20	2,37	2,07	2,26	2,61	2,76	2,44		2,41	1,88–2,76	2,29
	5				2,38					2,28											
WALL THICKNESS	1	10	12	14	8		9		12	10	16			13		12	11			8–16	12
	2	16	16	16	16				18	12	20	13		16		20	20		17	12–20	17
	3		16		24		16		21	16	24				20	22	25		24	16–25	21
	4				40					28											
TUNNEL ANGLE	1						16		34			25			24	30			25	16–34	26
	2	33	45	32	22		28		37	39	35	28		27	43	33	28		45	22–45	34
	3	41	43	53	41		33		46	42	44	39		35	52	43	44		60	33–60	44
	4			57	63		43		54	48	72	45			61	56				43–72	55
REL. HEIGHT OF CHOMATA	½			0,27	0,19										0,18				0,21	0,18–0,27	0,21
	1			0,26	0,17		0,29		0,23		0,30	0,33			0,21	0,22			0,21	0,17–0,33	0,25
	1½	0,25		0,27	0,22		0,31		0,40			0,31			0,36	0,28	0,23		0,39	0,22–0,40	0,30
	2	0,36	0,43	0,36	0,29	0,31			0,43	0,35	0,35	0,31			0,23	0,29			0,33	0,23–0,43	0,34
	2½	0,39	0,42		0,39	0,42	0,41		0,35		0,29	0,38			0,37	0,40	0,35		0,30	0,29–0,42	0,37
	3	0,39	0,48	0,30	0,38		0,42		0,40		0,37	0,42			0,30	0,42			0,36	0,30–0,48	0,39
	3½	0,28		0,50	0,36	0,35	0,33		0,37	0,38	0,29	0,54			0,35	0,40	0,30			0,28–0,54	0,37
	4			0,40			0,24		0,28		0,20									0,20–0,40	0,28
	4½				0,29	0,41			0,29			0,37								0,29–0,41	0,34

TABLE . VII

MEASUREMENTS OF *ALJUTOVELLA ELONGATA* RAUS. *ET* SAF., 1951 SUBSP. *LAZARENSIS* SUBSP. NOV.

SAGITTAL SECTIONS

(in μ)

SPECIMEN		51	52	54	55	56	57	58	59	60	61	62	63	64	65	66	67	68	69	70	71	72	73	74	75	76	77	RANGE	AVERAGE
RADIUS VECTOR	0	28	39	24	33	30	25	38	32	30	24	34	20	32	36	29	36	34	33	22	31	34	26	42	29	33	30	20 - 42	31
	1	74	92	68	86	76	60	92	76	92	64	88	43	93	84	64	80	79	72	64	76	66	70	93	96	76	74	43 - 96	77
	2	143	183	143	167	141	109	184	129	175	111	158	82	185	166	131	156	144	133	111	135	141	153	164	167	153	156	82 - 185	146
	3	261	310	269	263	244	209	243	217	287	211	281	148	295	289		289	247	227	183	219	271	249	273		263	247	148 - 310	250
	4	433					323	352			328	418	246		474		462	370		316	334			410		426	394	246 - 474	378
	5							505			462		366								484							366 - 505	454
PERCENTAGE INCREASE OF RADIUS VECTOR	1	95	100	112	95	87	83	69	70	91	75	81	88	98	98	106	96	82	86	75	79	115	118	75	74	103	112	69 - 118	89
	2	82	70	90	57	72	91	57	69	63	89	77	81	59	75		85	71	75	64	62	92	63	66		71	58	57 - 92	72
	3	66					55	45			56	57	66		64		60	50		73	53			50		62	60	45 - 73	58
	4							44			41		49								45							41 - 49	45
	5																												
WALL THICKNESS	1	11	8	10	9	12	8	12		16		13	12	14	14	11	10	11	12	12	14		12	8		9		8 - 16	11
	2	16	12	16	14	16	12	16		21		20	16	16	18		20	19	16	15	16		16	16		13		12 - 21	16
	3	19	19		18	16	20	19				28	20		20		28	20		17	20			22		19		16 - 28	20
	4	36						31					31		28			28			19			32				19 - 36	29
	5							24																					
SEPTAL COUNT	1	5	7	7		7		6		6	5	6		5	6		5	7	6	7	6			6	6	7		5 - 7	6
	2	8	10	9	9	8	8	9		10	9	9	9	9	8	9	9	8	9	10	9	9	10	8	8	10		8 - 10	9
	3	11	11	13	10	11	9	12		11	9	12	9	12	11		11	11	11	12	11	12	12	11		12		9 - 13	11
	4	16					11	14			14	15	12		16		17	14		13	14			14		15		11 - 17	14
	5							14			17		16								15							14 - 17	16

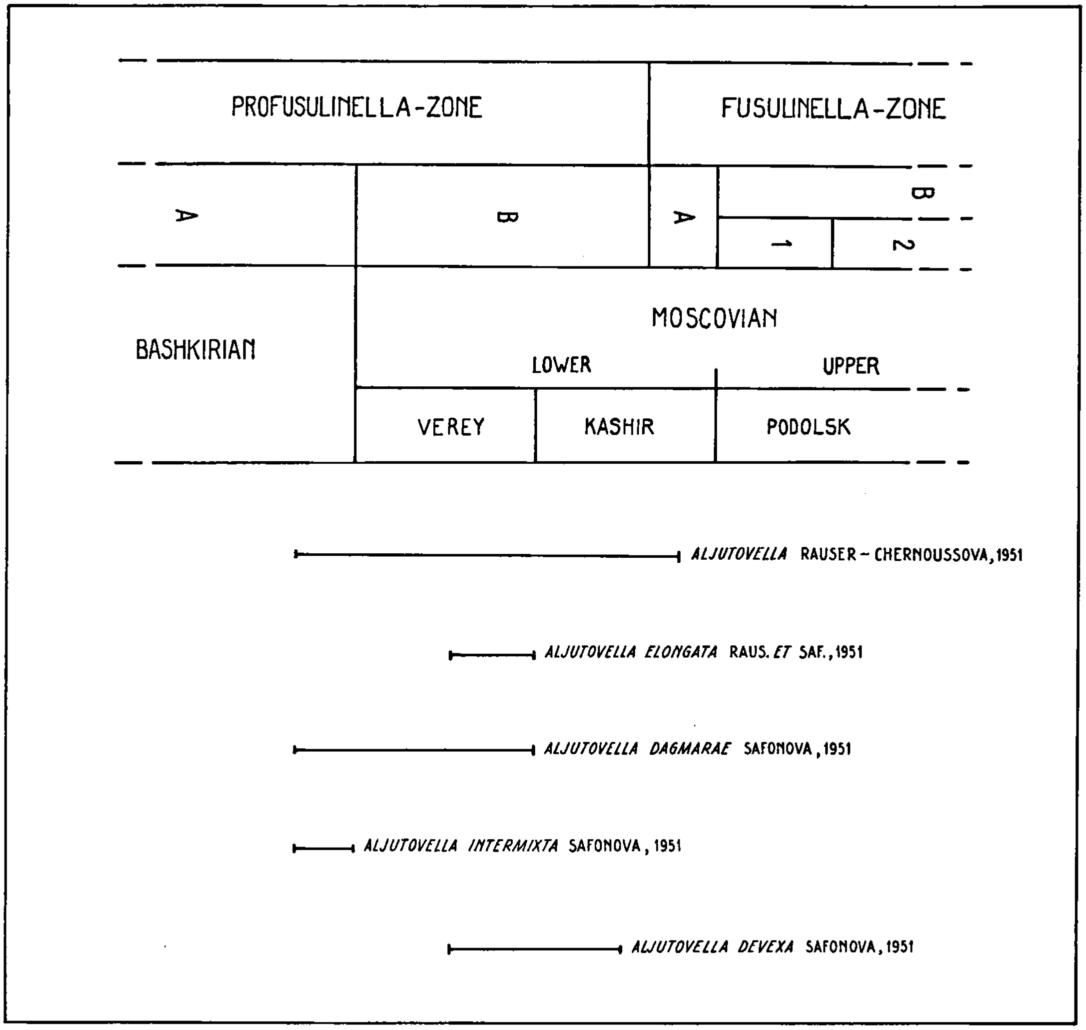

Fig. 33. The range of four species of the genus *Altjutovella* Rauser-Chernoussova, 1951.

diameter for corresponding whorls) and a somewhat greater $\frac{L}{D}$ ratio.

The range of the Genus *Aljutovella* Rauser-Chernoussova, 1951 in the U.S.S.R. is from the uppermost part of the Bashkirian up to and including most of the Kashirian of the Lower Moscovian.

The range of the four species mentioned above, as given in the original descriptions of these species, is shown in Fig. 33, p. 92. It is therefore concluded that the Lázaro Limestone Lense was most probably deposited in Vereyan times, although a late Upper Bashkirian or Lower Kashirian are also to be considered.

Calcareous algae occur in the Lázaro Limestone Lense. Mr. G. J. B. Germs distinguished the following species:

Anthracoporella spectabilis Pia, 1920.
Epimastopora rolloensis Rácz, 1964.
Girvanella sp.
Donezella lutugini Maslov, 1929.

A thick sequence of greywackes, shales, and sandstones overlies the Lázaro Limestone Lense. In this sequence and about 625 m above the top of the limestone lense (Loc. B 20, Fig. 28, p. 83), a brachiopod fauna has been found, consisting of:

Karavankina aff. *praepermica* Ramovs, 1966
Plicatifera sp.
? *Eomarginifera* sp.

This fauna indicates a Moscovian age, probably Upper Kashirian or Lower Podolskian (determinations carried out by Mr. C. F. Winkler Prins, personal communication).

Caliza Masiva Formation. — The Beleño Formation is conformably overlain by a thick (80—250 m) massive limestone called "Caliza Masiva Superior" by Spanish geologists. Van Ginkel (1965, p. 191), however, later introduced the name Escalada Formation for this unit. The limestone can be found throughout the whole San Isidro-Tarna-Ponton subarea, in the Beleño basin, in the Felechosa-Tarna syncline north of Puerto

Forno, and along the whole eastern border of the Central Basin of Asturias.

The Caliza Masiva Formation gradually passes into the limestones belonging to the Lena Formation. This is demonstrated clearly west of Puerto Forno, where the thick massive limestone of the Felechosa-Tarna syncline gradually splits up into the relatively thin limestone layers occurring in the north flank of the San Isidro anticline.

By far the greatest part of the rock composing the Caliza Masiva consists of intraclasts, 60—1200 μ in size; among the other allochem grains, however, the algae (mainly *Epimastopora rolloensis*) predominate. These constituents, coupled with a sparry calcite cementing, give the rock the appearance of a fossiliferous intrasparite or intrasparudite (Fig. 34, p. 93).

In the Beleño basin and in the area mapped by Martínez (Central Basin of Asturias, 1962) the age of the Caliza Masiva Formation has been given as Lower Podolskian and Upper Kashirian, respectively (i.e. Fusulinella zone, subzone B_1—A).

In our area a thin limestone layer just above the Caliza Masiva Formation, at Puerto Forno (Loc. A—2, Fig. 1), yielded a fusulinid fauna indicating an Upper Kashirian or Lower Podolskian age (van Ginkel, personal communication).

Loc. A—2 *Schubertella* cf. *subkingi* Putrya
Fusulinella sp.
Beedeina sp.
Millerella sp.
Ozawainella sp.
Pseudostaffella sp.
Schubertella sp.

At Loc. A—1, Fig. 1, situated about 4 km east of Loc. A—2, the base of the Caliza Masiva has yielded the following fusulinid fauna (van Ginkel, pers. comm.):

Millerella sp.
Parastaffella sp.
Pseudostaffella sp.
Ozawainella sp.

This fauna, in which characteristic genera are lacking, does not permit age determination. The same locality, however, has also yielded an algal flora (Germs, personal communication):

Epimastopora rolloensis Rácz, 1964
Donezella lutugini Maslov, 1929
Anthracoporella spectabilis Pia, 1920
Ungdarella sp.
Parachaetetes sp.

This flora incontestably points to at least a Lower Moscovian age (see page 88).

South of Puerto de San Isidro, a few more limestone localities have yielded the following fusulinid faunas and algal floras (Table VIII), determined by Dr. A. C. van Ginkel and Mr. G. J. B. Germs.

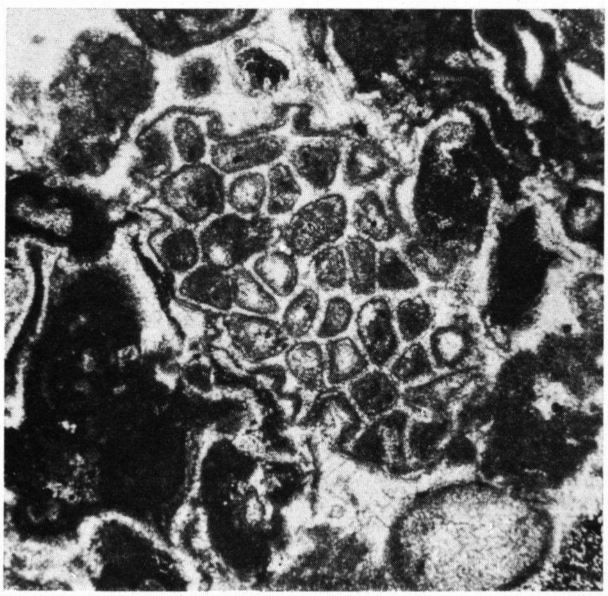

Fig. 34. Caliza Masiva Formation; fossiliferous intrasparudite with *Epimastopora rolloensis* Rácz, 1964. (40 ×)

TABLE VIII. Fusulinid and algal associations in limestones south of Puerto de San Isidro.

Locality number	Fusulinid assemblages	Algal assemblages
8—64	*Ozawainella* sp. *Millerella* sp. *Pseudostaffella* sp. *Eofusulina* sp.	*Mellporella* sp. *Donezella lunaensis* Rácz, 1964
9—64	*Parastaffella* ex. gr. *mathildae* (Dutkevitch, 1934a) *Profusulinella* aff. *cavis* Dalm., 1961 subsp. *arbejalensis*, v. Ginkel 1965 *Profusulinella* sp.	*Epimastopora rolloensis* Rácz, 1964 *Epimastopora bodoniensis* Rácz, 1964 *Donezella lutugini* Maslov, 1929
10—64	*Profusulinella* spp. *Profusulinella prisca* (Deprat, 1912) subsp. *rauserae*, van Ginkel 1965 *Aljutovella* sp. *Pseudostaffella* spp. *Pseudostaffella* ex. gr. *parasphaeroidea* (Lee et Chen, 1930) *Pseudostaffella* ex. gr. *gorskyi* (Dutkevitch) *Ozawainella* sp. *Millerella* sp. *Staffella* sp. *Schubertella* sp.	*Epimastopora rolloensis* Rácz, 1964 *Epimastopora bodoniensis* Rácz, 1964 *Parachaetetes* sp.

The fusulinid fauna of Loc. 8—64 indicates a Lower Moscovian age (Profusulinella zone, subzone B, or Fusulinella zone, subzone A). The fusulinid fauna and algal flora of both Loc. 9—64 and Loc. 10—64 also point to a Lower Moscovian age (Profusulinella zone, subzone B).

These age determinations, the thinness of the deposits between the Caliza Masiva and the Caliza de Montaña Formations, the presence of excellent examples of the Ricacabiello Formation, and the manganese content of the uppermost part of the Caliza de Montaña Formation all argue that a period of slow sedimentation or non-deposition succeeded the Caliza de Montaña Formation.

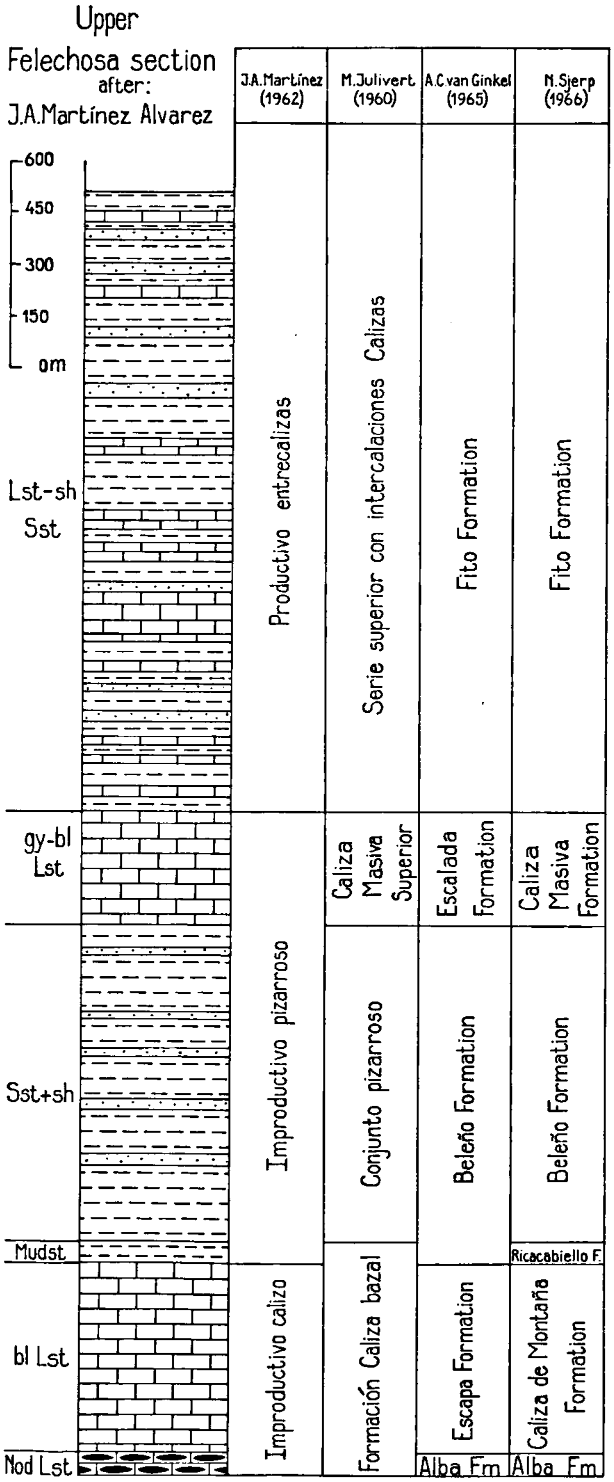

Fig. 35. a. Columnar section through Westphalian deposits northeast of the village of Felechosa.
b. Nomenclature used in the literature for the Asturian Carboniferous basins.

Fito Formation. — In its turn, the Caliza Masiva in the Beleño basin is conformably overlain by a thick series of about 700 m of sandstones, shales, and limestones, called the Fito Formation by van Ginkel (1965, p. 191) and grouped by Julivert in the "Serie superior con intercalaciones calizas" (Fig. 35, p. 94).

The whole sequence has yielded Upper Moscovian fusulinid assemblages (van Ginkel, 1965), and consequently this author considers that "The Fito Formation was probably contemporaneous with the Sama Formation and may well have graded into this unit".

In the western part of the mapped area, along the southeastern border of the "Cuenca Central de Asturias" north of the village of Felechosa, Martínez (1962, pp. 59—62) studied a very interesting section (see Fig. 35, p. 94).

Maraña-Retuerto basin

General remarks (the majority of the localities mentioned below are situated outside the mapped area and are shown in Fig. 36, p. 95).

In the province of Palencia, the Yuso Group developed as a thick sequence of conglomerates, sandstones, shales, and, rarely, limestones, unconformably overlying older formations.

East of the Yuso river, near Peña de Curavacas (Oriol, 1876) and Pico Los Cintos, the lower part of this group, consisting of a 500 m thick quartzite conglomerate with several interfingering sandstone and shale layers, has been called Curavacas Formation by Kanis (1956, p. 405).

The formation is conformably overlain and, laterally to the west, replaced by an interfingering sequence of sandstones and shales, the upper part of which is of Westphalian D age. This sequence was called the Lechada Formation by Savage (1961, internal report) after the Lechada river east of the village of Portilla de la Reina.

Starting at Pico Los Cintos and going westward, there is the following development in the quantities of conglomerate and the sandstone-shale in the Yuso Basin.

1. Pico Los Cintos:

   A 500 m thick quartzite conglomerate, alternating with shale and sandstone layers of minor importance.

   From plant fragments occurring in a large lens of sandstone and shale, a Westphalian B—C age

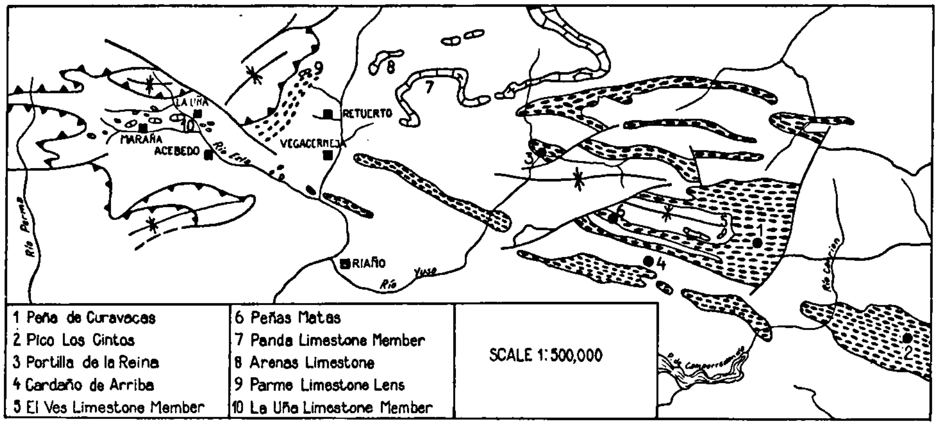

Fig. 36. Survey map of the Yuso Basin with locations mentioned in text.

could be ascertained (Kanis, 1956, pp. 416—417; Wagner, 1960).

2. West of Peña Curavacas:
   A Lechada Formation, consisting of shale and sandstone layers, interfingers and overlies the Curavacas Formation (van Veen, 1965, enclosure 2, section II).
   At a locality north of Cardaño de Arriba, a flora collected by van Veen (1965, p. 70) from the base of the formation indicates a Westphalian A age (Stockmans, in press).
   The El Ves Limestone Member has been distinguished in the lower part of the Lechada Formation overlying the conglomerates. Recently, a limestone from a corresponding unit (e.g. the Peña Prieta limestone) yielded a fusulinid association belonging to the Lower Moscovian (Profusulinella Zone, top subzone B) (van Ginkel, personal communication).
3. West of Peñas Matas:
   Rapid increase of the Lechada Formation at the expense of the Curavacas Conglomerate.
   The Panda Limestone Member is developed in the Lechada Formation; fusulinid assemblages indicate an Upper Moscovian (Podolskian) age. (Fusulinella Zone, subzone B, van Ginkel, 1965).
4. Maraña-Retuerto basin:
   In this basin, which is the direct continuation to west of the Yuso Basin, still younger sediments crop out.
   a. The Arenas Limestone Bed (south of Pico Las Arenas), yielded both fusulinid and algal material, indicating a somewhat younger age than the Panda Limestone for this unit (van Ginkel and Rácz, internal reports).
   b. West of the main road running through the vilage of Vegacerneja to the Puerto de Ponton, the sequence has a somewhat different appearance. At least six conglomerate beds, varying in thickness between 5 and 75 m and separated by sandstone-shale alternations, can be distinguished.
      A few limestone lenses could be mapped, of which the Parme Limestone Lense, just east of Pico Parme, yielded a fusulinid association indicating an Upper Moscovian age (see p. 97).
   c. The westernmost part of this Carboniferous basin is developed around the villages of Acebedo, La Uña and Maraña.
      Again we find conglomerates, showing rapid wedging out, which lie in a thick sequence of shales and sandstones. A striking fact is the rather high percentage of limestone in this sequence. One of these limestones, the La Uña Limestone Member (see p. 97), belongs to the Upper Moscovian.

Savage (personal communication) and van Veen (1965, p. 68) suggested emphasizing the lithofacies of the Curavacas Conglomerate by using this name for all the conglomerates occurring in the Yuso Basin. The same idea has been applied to the northern and southeastern parts of the Pisuerga basin, where Curavacas Conglomerate Beds and Lenses unconformably overly the Piedrasluengas Limestone Member, and are incorporated into the Molino Formation (Frets, 1965, p. 135; de Sitter and Boschma 1966, pp. 209—210).
The present author has adopted the same lithostratigraphic unit in an analogous way for conglomerates occurring in the Lechada Formation of the Maraña-Retuerto basin. These conglomerates do not differ in any way from those described by Kanis (1956), Koopmans (1962), and van Veen (1965). They show the following characteristic features:

1. a) The conglomerate consists mainly of orthoquartzite pebbles, cobbles, and boulders, varying in size between 5 and 45 cm.
   b) These constituents are well rounded and generally show an elongated shape.
   c) The conglomerate beds show very poor sorting.
   d) The matrix is a greywacke type of sandstone.
   e) The constituents do not touch each other.
2. Besides the orthoquartzite constituents, well-rounded limestone pebbles and cobbles consisting mainly of Caliza de Montaña fragments, occur locally.

3. The third type of constituent in the conglomerate beds are angular limestone fragments.
At locality 888, northwest of the village of La Uña, a fusulinid fauna and an algal flora were found in the limestone fragments (Table IX).

TABLE IX. Fusulinid and algal associations in limestone fragments of the Curavacas Conglomerate Beds northwest of the village of La Uña.

Fusulinid fauna (determinations by van Ginkel)	Algal flora (determinations by Germs)
Millerella spp.	*Anthracoporella spectabilis* Pia, 1920
Parastaffella sp.	*Komia abundans* Korde, 1951
Pseudostaffella sp.	
Fusulina sp.	

The fusulinid fauna indicates an Upper Moscovian age (Fusulinella zone, subzone B_1 or B_2).

A large block about 2.5 m long was found northeast of the village of La Uña.
The conglomerate is in places made up completely of subangular to rounded limestone fragments, 1—15 cm in size, with little greywacke matrix between them (intraformational conglomerates). In this type of rock the present writer could not distinguish an imbricated arrangement of the limestone fragments (Koopmans, 1962, p. 161).
4. In the conglomerate beds, thin shale and sandstone intercalations occur frequently.
5. Well-rounded quartzite pebbles and cobbles are often found dispersed in a mudstone matrix (see Fig. 37, p. 96) due to the action of mud-flows.

Fig. 37. Maraña Retuerto basin, Curavacas Conglomerate Beds; well-rounded quartzite pebbels and cobbles dispersed in a mudstone matrix.

The thickness of the poorly exposed sandstone-shale sequence found in the Maraña-Retuerto basin is very difficult to establish, because strong isoclinal folding occurred throughout the whole region and the hinges of the folds are not readily observable (see p. 122).
Many of the sandstones occurring in the sequence have a detrital matrix, whereas others display strong cementing by quartz or rare calcite. The felspar content is very low. The rock thus varies between a lithic greywacke and a subgreywacke, while rather thick beds of protoquartzites can be found locally (Pettijohn 1957, pp. 291—292). The greywackes show graded bedding, and bottom structures such as load casts, groove casts, and flute casts are much in evidence (Fig. 38, p. 96).
Generally, the sandstone-shale sequence does not yield many fossils, although indeterminable plant fragments occur locally in large quantities. However, at Loc. L. 302 situated in the bank of the Riosol river northwest of the village of La Uña, the following flora were collected:

Calamites suckowi Brongniart
Paripteris veeni Stockmans & Willière
Neuropteris tenuifolia Von Schlotheim
? *Linopteris* sp.
cf. *Diplothmema* sp.

This flora, found only about 200 m above the Caliza de Montaña Formation, definitely points to a Lower Westphalian age (van Amerom, personal communication).
The percentage of limestone in this westernmost part of the Yuso Basin is rather high. The Parme Limestone Lense (Loc. L. 406) is a fossiliferous intramicrite, but strong recrystallization has changed part of the

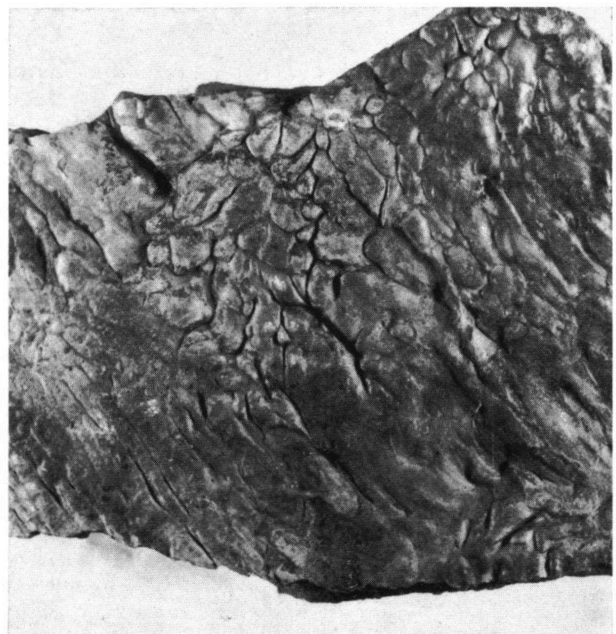

Fig. 38. Maraña Retuerto basin; graded greywackes with flute casts deformed by loading. (× ½)

rock into a microsparite. The La Uña Limestone Member (Loc. L. 34), varies between an intramicrite and a foraminiferal biomicrite, and shows only slight recrystallization.
The Parme Limestone Lense, lying only a few metres from the thrust plane below Pico Parme, is of the utmost importance for the dating of the movement along this thrust plane. The present writer therefore decided to carry out a detailed investigation to establish the age of this limestone. Fortunately, this limestone contains fusulinid faunas which are quite appropriate for this purpose. From the fusulinid genus *Fusulinella*, 25 sagittal and 14 axial sections were prepared. The data obtained from measurements of the various specimens are given in Tables X and XI, pp. 98 and 99.
There is no doubt that the fusulinids found in the Parme Limestone Lense belong to the Fusulinella zone, top subzone B_1 — base subzone B_2.
The La Uña Limestone Member occurring in the neighbourhood of the villages of La Uña and Maraña was examined in the same way by Mr. H. J. W. G. Schalke (internal report). The fusulinid association belonging to this limestone also points to the transition between subzones B_1 and B_2 of the Fusulinella zone.

The Parme Limestone Lense. — (loc. L. 406, see Fig. 1, p. 58) is a strongly recrystallized intramicrite with abundant fusulinids.
The age of the Parme Limestone Lense was established by comparing its fusulinid content with closely similar Spanish species. From 39 specimens considered to belong to a single species, orientated thin sections were prepared. The specimens were identified as belonging to *Fusulinella pandae* van Ginkel, 1965; see Plate II. The data given in Tables X and XI were plotted on the overlays in van Ginkel, 1965, pp. 180—181. The present assemblage of *Fusulinella* specimens was found to be comparable to specimens of *Fusulinella* from either the top of subzone B_1 or the lower part of subzone B_2. This indicates an Upper Podolskian age for the limestone member.

Cea Group

Rucayo Formation. — The name Cea Group, which is derived from the Rio Cea area, was assigned by Koopmans (1962, pp. 162—164) to Upper Carboniferous deposits succeeding the Asturian folding phase and lying with an angular unconformity on older rocks.
The sediments in this group are shales, sandstones, conglomerates, and coal seams. The Cea basins found in the southern part of the Cantabrian mountains, between the Porma and Carrión rivers, have been reported in detail by Helmig (1965), who considered, however, that, ". . . . the Cea unit is too monotonous to justify the term 'group' for it, and in this paper the Cea sequence will be classified as a 'formation'." He therefore distinguished a Carrión Member of Westphalian D age, overlain by a Prado Member of Stephanian A age whose base is formed by two limestone conglomerate horizons, the Villacorta Beds. Locally (Sabero basin), the Carrión Member contains a flora assemblage indicating a Stephanian A age, but the uppermost part of the sequence in this basin could have a Stephanian B age (Helmig, 1965, pp. 115, 118). Still further to the west, in the Matellana basin, only flora of Stephanian B age were observed.
Recently, de Sitter & Boschma (1966, pp. 213—216), in their paper on the Pisuerga basin, confirmed the opinion of Koopmans and considered the Cea unit to be a group.
In the mapped area, no subdivision of the Stephanian deposits is possible. Therefore, these rocks can only be considered as a formation. The present writer proposes the name Rucayo Formation, after the village of Rucayo—situated to the southwest just outside the mapped area—which is surrounded by Stephanian deposits. These deposits occur as relatively small, narrow outcrops along the León line (de Sitter, 1962b), and lie with a sharp angular unconformity on the older folded rocks of the Lois-Ciguera basin and adjoining areas.
The sediments in the Rucayo Formation are conglomerates, shales, sandstones, and occasionally "bolsas de Carbon". The conglomerates are made up of well-rounded quartz pebbles and cobbles in addition to well-rounded limestone pebbles in minor quantities. The matrix consists mainly of small quartz grains, though well-rounded glauconite grains and small rock fragments occur abundantly, as well as mica flakes, tourmaline, rutile, and zircon. Calcite veins cut through the whole rock.
A Stephanian B age has been established by Wagner (1963) and van Amerom (1965) from plant determinations in samples from various localities. Recently, the latter visited the area treated in this paper and made an excellent collection of flora from two localities.

Loc. L. 300: Mina Abandonada (Empresa: Hulleras del Norte, S.A.), situated about 1.2 km southwest of the village of Camposolillo, along the main road to Boñar.

This mine had already been visited by Wagner (1963, loc. 1172), who found the following plant association:

Neuropteris auriculata Brongniart
Callipteridium zeilleri Wagner
Pecopteris unita Brongniart
Polymorphopteris polymorpha (Brongniart)

To this flora the following species have now been added (determinations by H. W. J. van Ameron):

Pecopteris feminaeformis (Von Schlotheim) Sterzel
Pecopteris sp.
Annularia sphenophylloides (Zenker) Von Gutbier
Linopteris neuropteroides (Von Gutbier) H. Potonié (abundant)
Mixoneura ovata (Hoffmann) Zalessky
Alethopteris zeilleri Wagner
Aphlebia sp.
Cordaites sp.

TABLE X
MEASUREMENTS OF *FUSULINELLA PANDAE* VAN GINKEL, 1965
AXIAL SECTIONS
(in µ)

SPECIMEN		1	2	3	4	5	6	7	8	9	10	11	12	13	14	RANGE	AVERAGE
RADIUS VECTOR	0	51	53	44	55	47	39	56	54	41	41	39	35	49	47	35–56	47
	1	99	108	77	101	88	64	97	103	74	88	79	56	100	88	56–108	87
	2	163	183	136	178	176	103	165	179	113	140	143	102	196	157	102–196	152
	3	233	263	225	278	275	172	289	356	165	243	232	181	314	244	165–356	248
	4	320	404	350	422	453	238	470	599	251	278	361	304	461	334	238–599	382
	5	442	645	564			390		853	368	564		511	684		368–853	558
	6	584	975	862			578			506						506–975	701
	7	783															
PERCENTAGE INCREASE OF RADIUS VECTOR	1																
	2	65	69	77	76	100	61	70	74	53	59	81	82	96	78	53–100	74
	3	43	44	65	56	56	67	75	99	46	74	62	77	60	55	43–99	63
	4	37	54	56	52	65	38	63	68	52	56	56	68	47	37	37–68	54
	5	38	60	61			64		42	47	49		68	48		38–68	53
	6	32	51	53			48			38						32–53	44
	7	34															
FORM RATIO	1	1,25	1,30	0,99	1,28		1,18	1,21	1,38	1,00		0,81	0,93	1,00	1,32	0,81–1,38	1,14
	2	1,55	1,52	1,55	1,58	1,28	1,43	1,34	1,67	1,34	1,15	1,09	1,25	1,50	1,57	1,15–1,67	1,42
	3	1,66	1,66	1,85	1,67	1,59	1,53	1,42	1,56	1,71	1,28	1,40	1,29	1,50	1,73	1,28–1,85	1,56
	4	1,61	2,12	1,99	1,64	1,84	1,82	1,51	1,60	1,62	1,48	1,60	1,66	1,57	1,88	1,48–2,12	1,71
	5	1,77	2,17	1,90			1,75		1,80	1,71	1,78		1,60	1,81		1,60–2,17	1,81
	6	2,10	2,05	2,27			1,94			1,90						1,90–2,27	2,05
	7	2,05															
WALL THICKNESS	1	12	20	10	18	20	12	16	16	8	16	16	10	24	16	8–24	15
	2	16	22	22	26	26	16	32	36	16	26	16	14	30	25	14–36	23
	3	25	40	33	51	36	28	40	40	22	44	32	23	45	30	22–51	35
	4	28	57	47	60	20	39		64	26	50	44	32	40	20	20–64	41
	5	37	59	56			42			34	14			52		14–59	42
	6	41	44	35												35–44	40
	7	48															
TUNNEL ANGLE	1	17			30			20		18		13		26		13–30	21
	2	21	15	30	25	26	28	25	27	26	14		20	18	28	14–30	23
	3	20	14	30	20	35	35	23	38	21	28	34	26	21	22	14–38	26
	4	17	27	29	33	43	31	21	42	18	25	38	27	25	33	17–43	29
	5	19	43	35			35			24			43	35		19–43	33
	6	29	42	50			53									29–53	44
	7	35															
REL. HEIGHT OF CHOMATA	½	0,17	0,25	-	0,29	-	-	-	0,18	-	-	-	-	0,20	-	0,17–0,29	0,22
	1	0,32	0,32	0,20	0,42	-	0,21	0,25	0,25	0,25	-	0,18	-	0,30	0,25	0,18–0,42	0,27
	1½	0,34	0,33	0,33	0,39	0,40	0,20	0,29	0,33	0,26	-	0,27	0,26	0,30	0,37	0,20–0,40	0,31
	2	0,39	0,47	0,45	0,34	0,37	0,29	0,21	0,45	0,34	0,27	0,22	0,38	0,50	0,42	0,21–0,50	0,36
	2½	0,48	0,48	0,49	0,47	-	0,25	0,43	0,37	0,30	0,29	0,29	-	-	0,42	0,25–0,49	0,39
	3	0,29	-	0,48	0,39	-	0,33	-	0,37	0,36	0,31	-	0,33	0,50	-	0,29–0,50	0,37
	3½	0,39	0,48	0,50	0,44	0,38	0,30	-	0,46	0,41	0,41	0,26	0,40	-	-	0,26–0,50	0,40
	4	0,31	-	-	0,30	0,24	-	-	-	-	-	0,43	0,47	0,46	0,50	0,24–0,50	0,39
	4½	0,26	-	0,43	-		0,47	-	-	0,38	0,23		0,46	0,43		0,23–0,47	0,38
	5	0,33	0,56	0,37	-		0,37		-	0,26			0,42	0,37		0,26–0,56	0,38
	5½	0,33	0,28	0,36			0,48			0,33						0,28–0,48	0,36
	6	0,39	0,31	0,30			-			0,43						0,30–0,43	0,36
	6½	0,28															
	7	0,35															

TABLE XI

MEASUREMENTS OF *FUSULINELLA PANDAE* VAN GINKEL,1965

SAGITTAL SECTIONS

(in μ)

SPECIMEN		51	53	54	55	56	57	58	59	60	61	62	63	64	65	66	67	68	69	70	71	72	73	74	75	76	RANGE	AVERAGE
RADIUS VECTOR	0	42	52	61	50	47	50	37	51	38	51	47	49	51	43	69	37	44	40	47	46	51	60	37	38	49	37 - 69	47
	1	94	100	142	93	88	110	85	95	78	93	93	108	100	88	145	89	106	78	98	98	110	106	80	88	116	78 - 145	114
	2	152	173	247	172	158	192	147	153	136	157	172	176	155	165	233	146	166	129	162	176	189	163	147	156	219	129 - 247	170
	3	240	285	400	306	263	335	230	247	221	264	267	291	216	278	352	246	250	203	274	274	354		224	266	342	203 - 400	276
	4	335	426	588	446	438	555	374	372	331	386			332		527	383	407		490		581		356	441	547	331 - 588	438
	5	485	593		642	691	784			524	588					784		627						532	664		485 - 784	629
	6	706	843																									
	7		1176																									
PERCENTAGE INCREASE OF RADIUS VECTOR	1	61	74	74	84	79	75	72	61	74	68	84	64	55	87	61	64	57	65	65	80	72	54	83	77	90	54 - 90	71
	2	58	64	62	67	66	74	57	62	62	68	55	65	39	69	51	68	51	57	70	56	87		53	70	56	39 - 87	62
	3	40	49	47	56	67	65	63	51	50	46			54		50	56	63		79		64		59	66	60	40 - 79	57
	4	45	39		44	56	41			58	52					49		54						50	51		39 - 58	49
	5	45	42																									
	6		40																									
	7																											
WALL THICKNESS	1	16	16	22	16	14	16	16	14	15	9	14	16	13	16	21	14	10	12	17	17	14	18	12	13	24	9 - 24	16
	2	20	39	38	28	27	22	21	24	21	16	22	24	22	24	28	18	16	14	28	28	25		20	16	36	14 - 39	25
	3	22	47	47	30	40	36	28	40	30	29	20	22	28	32	42	26	24	16	40	30	34		28	24	40	16 - 47	31
	4	24	56		36	32	48			36	38			28		48		47						34	40		24 - 56	39
	5	31	65		30		28			24	49							33						40	28		24 - 65	36
	6	56	70																									
	7		36																									
SEPTAL COUNT	1	6	7	6	8	7	7	7	7	7	8	8	8	8	8	7	7	7	7	8	7	8	7	6	6	8	6 - 8	7
	2	11	12	12	12	12	10	10	12	11	11	10	11	12	10	13	11	11	11	11	10	11	13	12	11	12	10 - 13	11
	3	13	16	15	16	14	11	12	14	14	12	11	15	17	12	15	11	12	12	12	14	12		11	11	15	11 - 17	13
	4	19	20	16	19	15	14	15	16	16	15			15		17	12	14		13		12		15	15	17	12 - 20	16
	5	20	22		22		17	14		18						19		14							15		14 - 22	18
	6	19	21																									
	7		19																									

Loc. L. 301: Mina Nieves (Razón Social: Felix Población, Boñar), situated about 1 km east of Loc. L. 300.

The species collected from the tip of this mine are listed below (determination H. W. J. van Amerom).

Pecopteris polymorpha Brongniart
Pecopteris subelegans H. Potonié
Pecopteris acuta-dentata Brongniart
Pecopteris waltoni Corsin
Pecopteris candolleana Brongniart
Pecopteris unita Brongniart
Pecopteris feminaeformis (Von Schlotheim) Sterzel
Annularia stellata (Von Schlotheim) Wood
Annularia sphenophylloides (Zenker) Von Gutbier
Pinnularia capillacea Lindley & Hutton
Linopteris neuropteroides (Von Gutbier) H. Potonié
Mixoneura ovata (Hoffmann) Zalessky
Cyclopteris fimbriata Lesquereux
Pseudomariopteris dimorpha (Lesquereux) Stockmans & Willière
Callipteridium gigas (Von Gutbier) Weiss
Alethopteris zeilleri Wagner
Sigillariostrobus sp.
Sigillariophyllum sp.
Dicksonites sterzeli (Zeiller) Danzé-Corsin
Sphenopteris sp.
Calamites sp.
Cordaites sp.

These assemblages clearly indicate a Stephanian B age. Besides this flora, a few specimens of the fresh-water lamellibranch *Anthraconaia* sp. have been found.
Along the Rucayo Formation a few miles to the east, the tip of a mine northeast of the village of Viego has yielded an important flora identified by van Amerom (1965). Renewed sampling provided the following additions (van Amerom, pers. comm.):

Pecopteris polymorpha Brongniart
(=*Polymorphopteris polymorpha* (Brongniart) Wagner)
Taeniopteris jejunata Grand'Eury
Odontopteris minor cf. *zeilleri* H. Potonié
Calamostachys calathifera Weiss
Sphenophyllum costae Sterzel
Sphenophyllum cf. *incisum* Wagner

This flora confirms the Stephanian B age of the Rucayo Formation.

CHAPTER II

SOME REMARKS ON THE FACIES AND PALAEOGEOGRAPHY OF THE CANTABRIAN PALAEOZOICUM

The palaeogeography of the Cambrian has been reported recently in a paper by Lotze and Sdzuy (1961, pp. 190—204), to which the present writer refers. A short review of the picture submitted by these investigators follows below, accompanied by their palaeogeographic map (Fig. 39, p. 101).
The Cambrian, as known from the Leonide thrust zone (de Sitter, 1962b, c), the Esla thrust zone, and the area treated in this paper, belongs to the Cantabrian facies type, of which Lotze and Sdzuy state: ". . . . der Kantabrische Typus repräsentiert die Entwicklung eines stabilen Schelfes, und sein Verbreitungsgebiet ist als 'Vorlandsporn' auffassbar, um den sich eine stark sinkende, sehr aktive Randsenke orthogeosynklinalen Charakters, verdeutlicht durch den Westasturischen und Iberischen Typus, schlingt" (Lotze und Sdzuy, 1961, p. 202).
The Herreria Formation (de Sitter, 1962a, b, c, 1965; Oele, 1964) known from the Leonide thrust zone, the Montes Pardominos (Rupke, 1965), and the western Asturian coastal areas, lies unconformably on the folded Precambrian Mora Schists (de Sitter, 1962-a; Pastor Gomez, 1962), and is made up of coarse-grained sandstones and conglomerates, in alternation with shales and medium-grained quartzites, with some dolomites near the top. The formation is not found in either the Esla nappe or in the area treated in this paper because in both the base of the Lancara Formation, which overlies the Herreria Formation, acted as a detachment plane for the overthrusts.
The depositional environment of the Herreria Formation must have been marine; the sedimentary structures point to agitated shallow-water conditions (Oele, 1964, p. 29).
Observations on sedimentary structures in the Herreria Formation of the Montes Pardominos led Rupke to conclude a generally N-S transport direction (Rupke, 1965, p. 14).

No hiatus in deposition seems to exist between the upper part of the Herreria Formation (Barrios Schichten, Lotze, 1965) and the overlying dolomite-limestone sequence of the Lancara Formation. This Lancara Formation consists of micrites and biomicrites, sandy shales, and marls, in addition to all kinds of such allochemical limestones as intrasparites (intrasparudites) and oösparites, of which Krumbein and Sloss say: "Cambrian and Ordovician limestone terrains in many parts of the world are characterized by intraclastic limestone layers, commonly with fine interstitial material forming intramicrite" (Krumbein and Sloss, 1963, p. 181).
The intrasparite specimens clearly imply a two-phase genesis in environments of prolonged calm-water conditions interrupted by sudden pulses of greatly strengthened wave or current energy, that is to

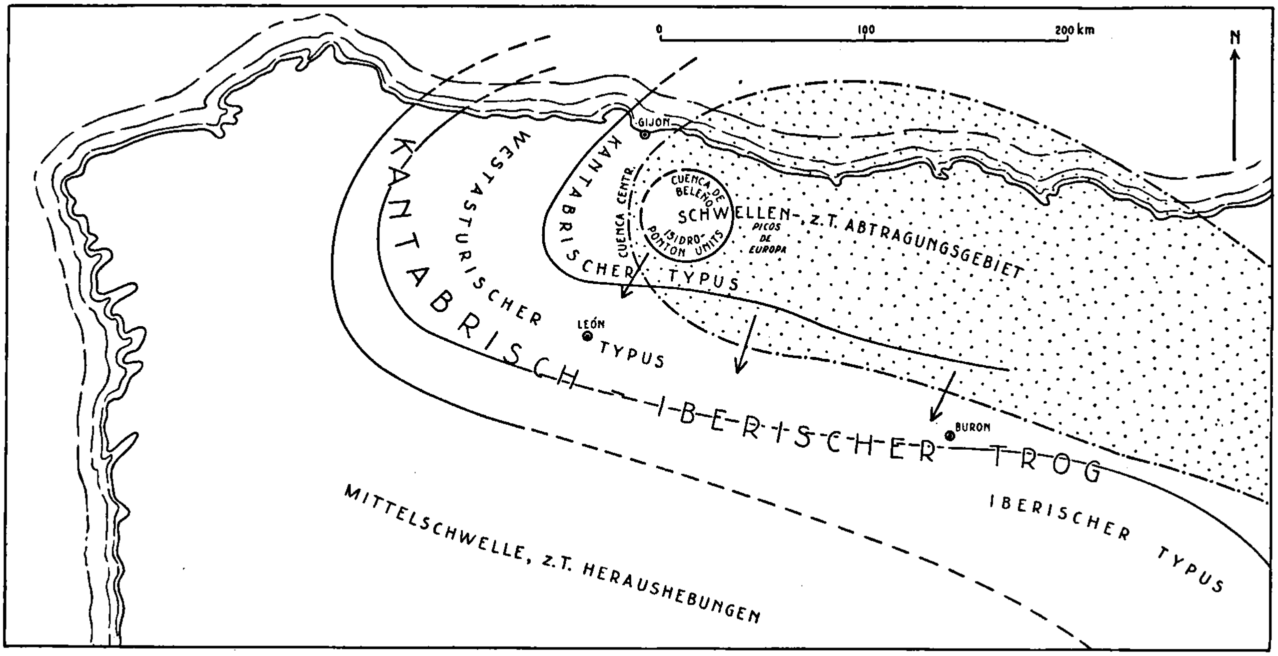

Fig. 39. The palaeogeography of the Cambrian in northern Spain after Lotze and Sdzuy.

say shallow-water conditions (Folk, 1959, p. 22). The oösparite fits very well with such an environment, and also indicates fairly vigorous current action. Accordingly, Illing (1954) stated that: "Oösparite forms in high-energy environments", and Krumbein and Sloss (1963, p. 133) that: "Oölites are formed in saline waters under agitated conditions along shores or in shallow places where waves break".

On the other hand, micritic rocks point to a depositional area in which currents were calm. Such conditions could prevail either in deep water or in shallow protected areas. The arrangement of the trilobite fragments, parallel to the bedding plane, indicates, however, that even the micritic rocks must have been deposited, at intervals, in rather agitated waters.

Clastic quartz can be found in all the thin sections; it occurs abundantly in the sandy intramicrudite, mainly as subangular grains, indicating land to have been adjacent.

The present writer therefore believes that the allochemical limestones were deposited in a shallow-neretic to littoral environment, whereas at least some of the micritic rock was formed in calm, somewhat deeper water.

This limestone-dolomite member of the Lancara Formation, which is found in large areas of the Cantabrian mountains, fits very well with the picture of a "Thin sheet-like microcrystalline limestones, closely associated with shales, and passing laterally into coarser grained carbonates, found with other sediments of the cratonic shelf" (Krumbein and Sloss, 1963, p. 568). The nodular limestone, with its organoclastic appearance (trilobite biomicrite—fossiliferous intrasparite associations), must have originated in a shallow-neritic depositional environment.

Nagtegaal (in press, see p. 79) holds the view that the lime-mud must have been deposited in an oxidizing environment; the original red colour, caused by hematite, points to the same circumstances (van Houten, 1961). Furthermore, Oele (see p. 79) supposed the existence of submarine solution processes to explain the origin of the uninterrupted nodular limestone sequence.

A similarity in lithology exists between this Cambrian nodular limestone and the Visean griotte, for which a condensed character has been assumed on the basis of conodont and goniatite stratigraphy (see p. 79).

Between the Lancara Formation and the overlying Oville Formation there is a gradual transition. The sediments of the Oville Formation are mainly clastic deposits, among which glauconite-bearing sandstones and shales occur frequently. Glauconite, once formed, was sometimes transported over a wide area. This has been assumed for the Oville Formation, in which the majority of the glauconite grains display a well-rounded shape.

The percentage of glauconite in the sediment seems to increase to the north, suggesting its source area to have lain in that direction. However, glauconite crystals were also formed *in situ* from biotite flakes occurring in the shales and sandstones.

There are various theories concerning the environment in which glauconite can be formed (Pettijohn, 1957, p. 468), but its presence is generally considered a reliable criterion for a marine origin of the enclosing sediment. This agrees with the occurrence of such sedimentary structures as load casts, slumps, and ripple marks, the load casts occurring especially in the upper part of the formation.

These features point to a rapid accumulation in shallow water, a depositional environment fitting very well with the occurrence of rare oölitic limestones in the Oville Formation. Trilobite fragments are concentrated in the bedding plane of layers with a relatively high iron content. Animal tracks and bore holes (*Scolithus*) can be found frequently in the more sandy layers.

No interruption of sedimentation can be demonstrated in the gradual transition from the Oville to the Barrios Formation. Consequently, there is no evidence for the Iberian folding phase.

During the development of the Barrios Formation, clastic material was deposited hundreds of metres thick over a very wide area. Thus, the tendency toward an increasing rate of sedimentation, already started with the deposition of the Oville Formation, continued into the Ordovician.

Everything points to the deposition of the Barrios quartzites in a shallow water-environment. Here, again, we find various sedimentary structures such as cross-bedding (Fig. 40, p. 102), ripple marks, and load casts. Scattered, well-rounded pebbles occur frequently throughout the whole formation. Oele (1964, p. 85) has already described the rapid wedging-out of the quartzite beds. This author carried out some measurements on the long axes of grains occurring in the Barrios Formation as well as in the Oville Formation. Although the number of samples investigated was not sufficient to permit the drawing of conclusions for a wider area, the majority of the measurements suggest a roughly north-south current direction.

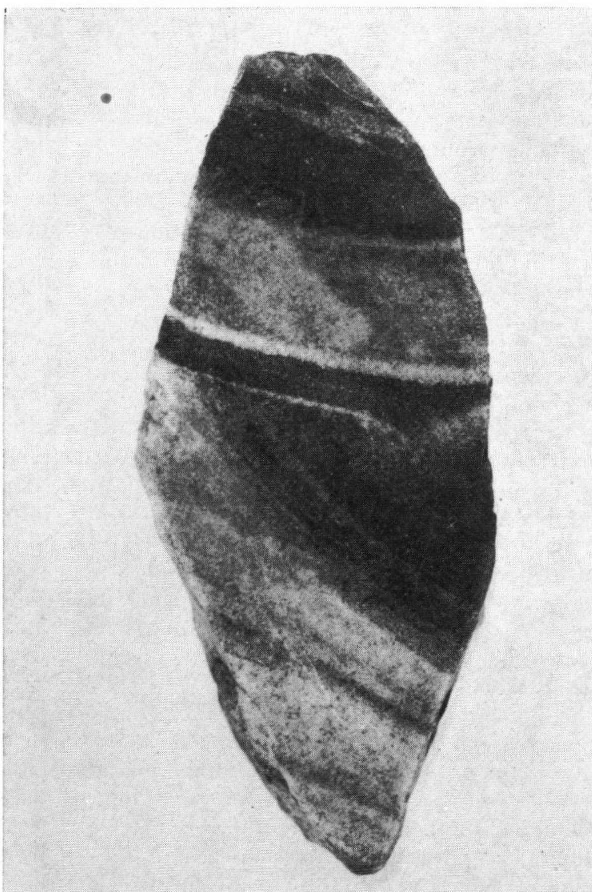

Fig. 40. Cross-bedding in the Barrios Formation. (natural size)

The Barrios Formation is very poor in fossils. Besides burrows, only animal tracks (*Cruziana furcifera*) occur in any abundance.

A basic vulcanism, indicated by the occurrence of doleritic sills, subaguous tuffs, and tuffaceous sandstones (see p. 123), can be found at many places in the area treated in this paper. As everywhere in the Cantabrian mountains, the occurrence of this kind of rock is restricted to the Oville and Barrios Formations (the Silurian Formigoso Formation does not occur in the mapped area).

A doleritic sill was mapped at the base of the Oville Formation, south of the village of Liegos. Tuffs and tuffaceous sandstones occur in the Oville Formation as well as in the lower parts of the Barrions Formation near the villages of Lois and Redipuertas. This volcanic activity is one of the few indications of the Caledonian orogeny known from regions outside the Cantabrian mountains (Lotze and Sdzuy, 1961).

To summarize, it is evident that the Cambrian and Ordovician formations were all deposited under shallow-water conditions, the great lateral extension of the deposits pointing to a very extensive shallow sea. It is striking to find the Lancara Formation, with its high content of calcareous sediment, lying between the clastic sediments of the Herreria Formation and the Oville and Barrios Formations respectively. Apparently, the supply of clastic material decreased rapidly during the deposition of the Lancara Formation, although the presence of sandy intramicruditic rocks, with their subangular quartz grains, and the rather large amount of clastic material in the limestones as well as in the marls, constitute evidence of some supply of clastic material, and occasionally even of a source area at relatively short distance.

There seem to be arguments to support the assumption that the transport direction was roughly north-south during the deposition of the sediments under consideration. This fits very well with the palaeogeographic picture of Lotze and Sdzuy. North of the Cantabrian-Iberian trough, these investigators assumed a "Schwelle" that occasionally served as the source area of the clastic material. However, the newly-discovered Cambrian sediments, which are overlain by the Ordovician quartzites, cover great parts of the geanticline of Lotze and Sdzuy. The same holds for their continuation below the Carboniferous rocks of

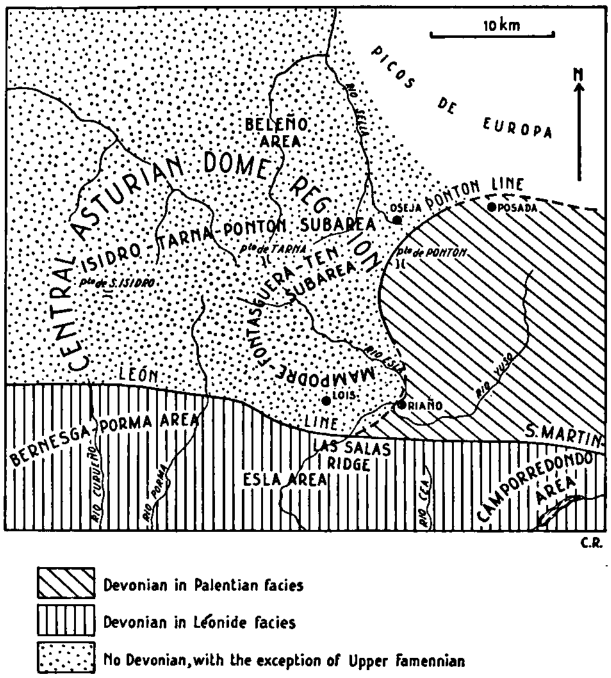

Fig. 41. Devonian facies zones in the Cantabrian mountains.

the Cuenca de Beleño (Julivert, 1960) and would hold for their assumed existence below the Barrios quartzites north of the Puerto de San Isidro (Martínez, 1962). These Cambrian deposits, like the Cambrian of the province of León, belong to the Cantabrian facies type.

Since no interruptions in the sedimentation could be demonstrated in either the Cambrian or the Ordovician, a great part of the "Schwelle" must have been continuously below the water surface, at least from the middle of the Lower Cambrian through the Ordovician. In no way could the area considered above have acted as source area during Cambrian or Ordovician times (Fig. 39, p. 101).

In the western part of Asturias the Ordovician quartzites are overlain conformably by the Luarca shales. In the province of León, however, there are strong arguments for assuming the existence of a hiatus, comprising the Upper Llanvirnian through the Lower Llandoverian, between the Barrios quartzite and the overlying Silurian and Devonian rocks (see p. 68).

For our area no such conclusion can be drawn because the Ordovician Barrios Formation is everywhere disconformably overlain by Upper Famennian transgressive sediments. Whether Silurian and Devonian rocks were ever deposited, cannot be ascertained.

According to Rupke (1965), a differentiation of the Devonian basin had already commenced in the Lower Devonian. This author found strong arguments for an early rising of the Las Salas ridge (León line), and stated: "These movements were epeirogenetic in character and probably continued through the Devonian to reach a climax during the Upper Famennian". For the region more to the east, Koopmans (1962) also mentioned the uplift of a narrow E-W-running San Martin-Camporredondo ridge (the eastward continuation of the León line), during two periods in the Middle and Upper Devonian. This ridge separates two different facies zones: in the south the Leonide facies of the Esla-Bernesga region and in the north the Palentian facies of the Upper Carrión region (de Sitter, 1962b; Brouwer, 1962; Koopmans, 1962; van Veen, 1965). This northern facies continues westwards as far as the Puerto de Ponton. There, a boundary line must be drawn between this area of Palentian facies and an area in which there is at present no Devonian (with the exception of Upper Famennian). For this boundary the name *Ponton line* is introduced here; its continuation to the south is difficult to establish (Fig. 41 p. 103).

During the Westphalian, the eastern border of the Lois basin acted as a facies boundary (see p. 108). This line may have already been active during the Devonian; in that case the Ponton line would join the León line north of the village of Las Salas. More to the west, the León line separates the Bernesga-Porma area, with Devonian developed in Leonide facies, from the Isidro-Tarna-Ponton region.

We know for certain that there was a pre-Upper Famennian erosion period which cut down to the Barrios, the Oville, or the Lancara Formations in a region west of the Ponton line, independent of the León line.

An isopach map (Fig. 42, p. 104) of the Barrios Formation in our region has been drawn, and if we assume that the variation in thickness was caused by the Bretonic erosion, there is evidence of the uplift along the Ponton line. The isopach map (which has not been corrected for tectonic movements) shows two striking features:

1. An absence of Barrios quartzite in the Mampodre-Fontasguera-Ten subarea.
2. A gradual decrease in the thickness of the Barrios Formation in the thrust units, in the direction of the central Mampodre-Fontasguera-Ten subarea.

This picture, however, was distorted by later tectonic movements. The present writer therefore considers the sudden jump in thickness from 150 m (thrust-units) to 0 m (central areas) to be due to a thrust movement extending over many kilometres.

Fig. 43, p. 105, shows the difference in erosion level for two sections, one lying north of the central areas and the other crossing them. Apparently the uplift was dome-shaped, and it is possible that its sharp eastern border, the Ponton line, was actually a fault. For this Bretonic structure the name Central Asturian dome is used.

An identical situation holds for the areas in the southwest along the León line. There, the Forcada

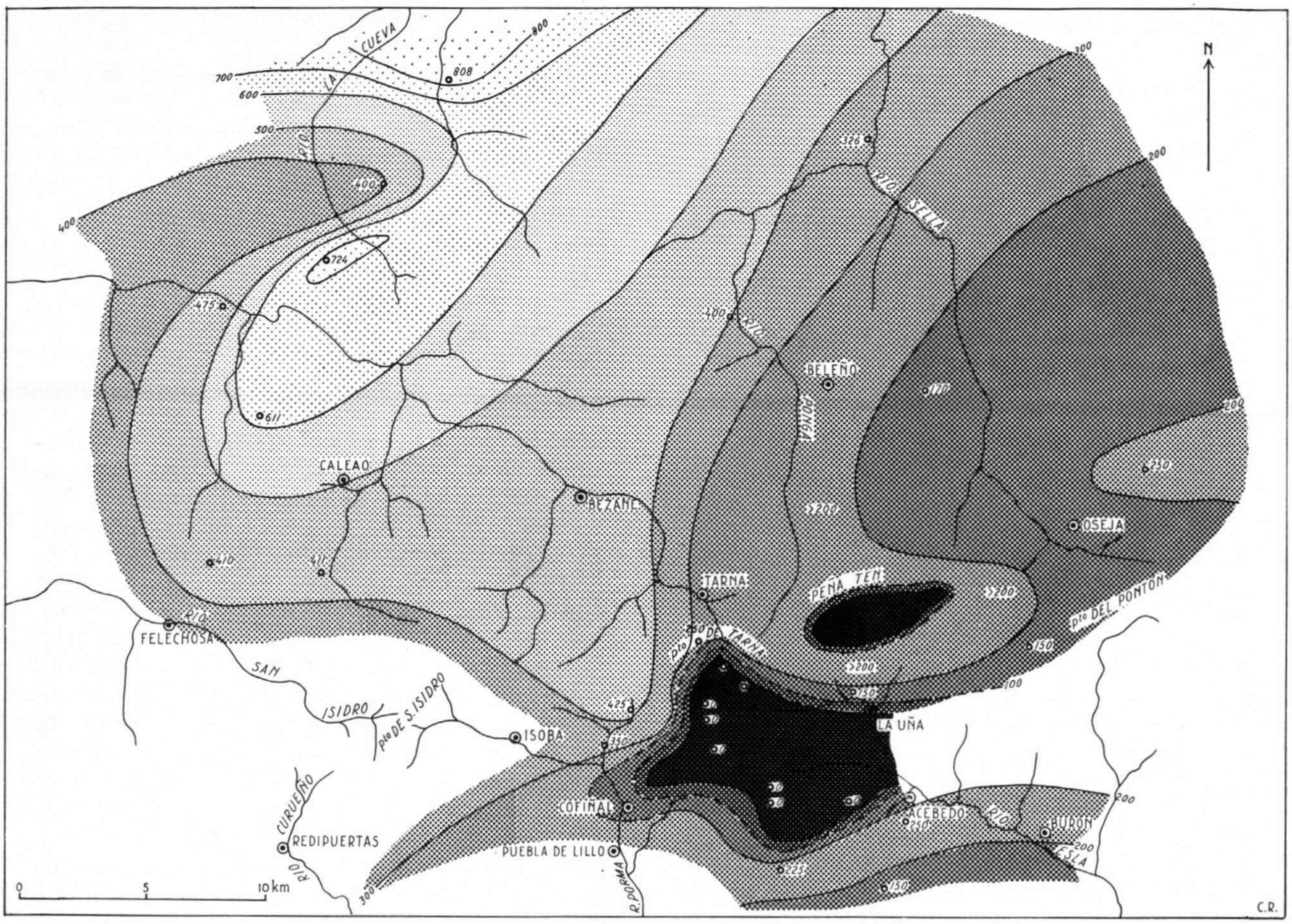

Fig. 42. Isopach map of the Barrios Formation, Central Asturian dome region.

thrust sheet and Armada unit show the same level of erosion as in our central areas. This is in accordance with the decrease in thickness of the Barrios quartzite south of the Mampodre subarea in the direction of the León line. We might therefore also expect a rapid southward thinning of the Barrios quartzite below the Piedrafita-Lillo basin in the direction of the Forcada thrust sheet. We shall see later that the absence of the competent Barrios quartzite below the Upper Devonian and Carboniferous sequence had a direct influence on the nature of later tectonic deformations.

After the above-mentioned periods of positive epeirogenetic movements and subsequent peneplanization, the sea encroached on the land in Upper Famennian times. This Upper Devonian transgression has already been reported by Comte (1938a, b; 1959, p. 326), who stated: "Au Famennien supérieur, le mouvement de subsidence a donc repris, et s'est accéléré momentanément au début du Strunien imprimant à la transgression les traits si curieux que nous a révélés l'analyse stratigraphique".
The Ermita Formation, representing a very short time-span, is, despite its very small thickness, extremely widespread, suggesting a rather rapid transgression over a relatively flat surface.
In the area treated in this paper, the sections through the Upper Famennian (Ermita Formation) point to a wide variation in lithology, but the post-hiatus sedimentation started with a thin quartz-pebble layer or a sandy dolomitic limestone layer.
In the eastern part of the Isidro-Tarna-Ponton subarea, the Upper Devonian bears a mainly clastic character, only its top showing a few limestone layers (La Uña section, page 74). In the western part of this subarea, the Ermita Formation is built up almost completely of sandy calcareous material (Felechosa section, page 75).
In the central Mampodre-Fontasguera subarea, the sequence shows much more differentiation. Biomicritic and biosparitic limestones, oölitic limestones, and some rare nodular limestones have been found. All show a rather high amount of clastic material. In addition, there are ferruginous sandstones, quartzitic sandstones, and greyish shales and marls. Well-preserved fossil faunas are much in evidence.
In our region the typical limestone-shale alternation of Strunian age is restricted to these central areas. An identical fine-grained limestone occurs in the Armada

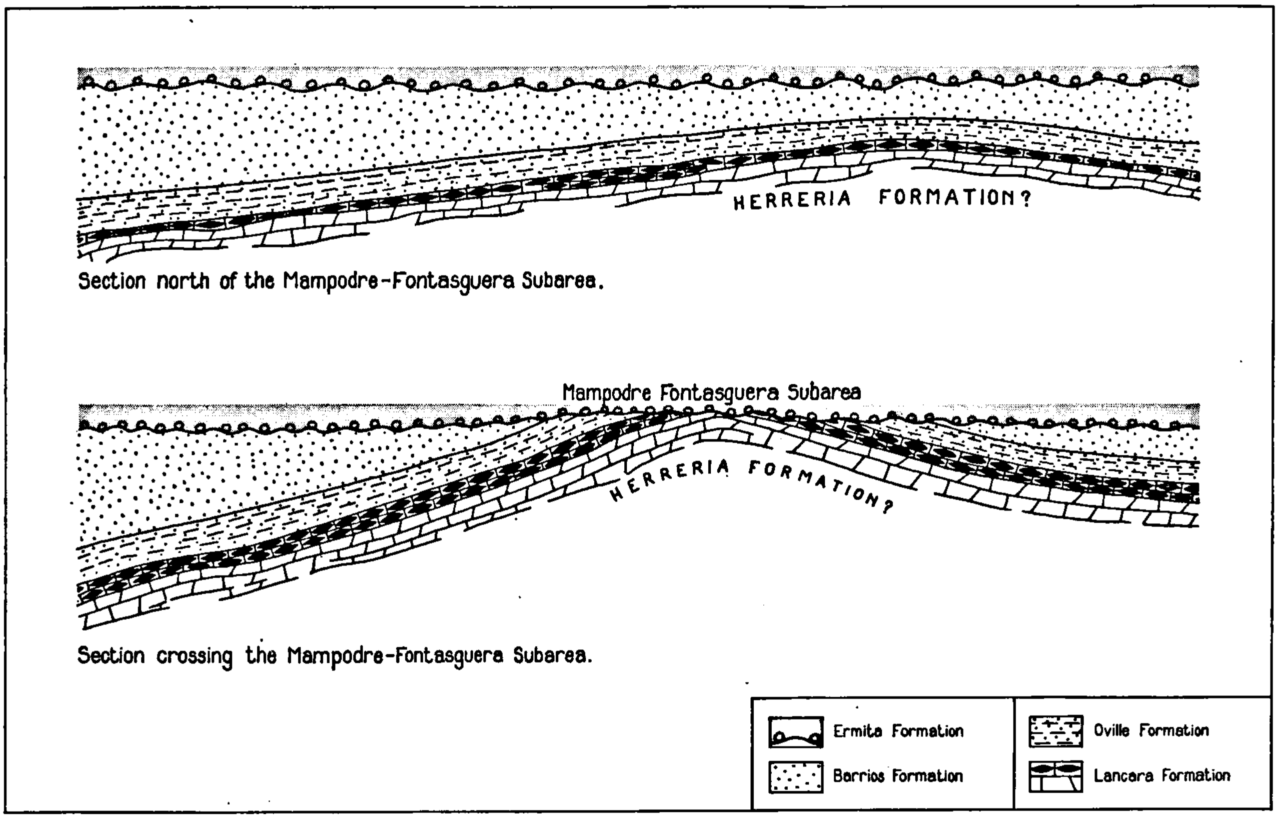

Fig. 43. The Bretonic erosion in the Central Asturian dome region.

unit, just south of the León line. A striking coincidence is the fact that this Armada unit, too, shows its erosion level below the Barrios Formation.

There is a great difference between the Upper Famennian transgressive sediments in our area and the Famennian deposits east of the Ponton line. According to van Veen (1965) and van Adrichem Boogaert (1965, pp. 168—173), the Upper Devonian in the Cardaño-Vidrieros region is complete, and is represented by two nodular limestone formations, Frasnian and Famennian in age, lying respectively below and above the Murcia quartzite. More to the west, north of the villages of Cuenabres and Casasuertes, the Famennian is represented by the "Montó Schichten", as designated by Kullmann (1960; see also van Adrichem Boogaert 1965, pp. 175—176).

There is therefore no doubt that the region east of the Ponton line remained below the water surface during the time that upheaval, peneplanization, and subsequent marine transgression took place in other parts of the Cantabrian mountains. This proves that even during the Upper Famennian, the Ponton line played an important role. It separated an area with transgressive Ermita Formation in the west from an area in the east with relatively calm though shallow water in which the limestones of the Vidrieros Formation were formed.

On the other hand, it is evident from a comparison of the area investigated with the Esla-Porma-Bernesga area that the León line loses its characteristic function as facies boundary line in the Upper Famennian, the Ermita Formation being developed identically north and south of it. This situation continued, as we shall see, during the deposition of the Tournaisian black shales, the Visean griotte and radiolarites, and part of the Namurian Caliza de Montaña.

After the Upper Famennian transgression the sea covered great parts of the Cantabrian mountains. In the beginning, sedimentation was certainly very slow, since both the Vegamián and Alba Formations show a considerably condensed sequence. Locally, some breaks in the sedimentation have been demonstrated (Budinger & Kullmann, 1964; Higgins *et al.*, 1964; Rupke, 1965). At some places, for example, the black shales of the Vegamián formation do not occur, but whether this is due to non-deposition or erosion cannot be definitely decided. The assumption of local positive epeirogenetic movements would explain both possibilities.

In attempting to explain local unconformities in the Bernesga-Porma area, Higgins and Wagner distinguished some more or less shallow ridges and periods of local uplift, with subsequent transgressions during Upper Tournaisian and Lower Visean times. The same holds for our Valverde section (page 71), where a basal transgressive glauconite bearing sandstone with phosphatic nodules can be found overlying the

Strunian limestone. A second disconformity occurs only a short distance above the first.
The occurrence of large quantities of well-rounded (clastic) glauconite grains, most probably derived from the underlying Oville Formation, suggests that locally some parts of the region were available as source area.
In contradistinction to the pre-Upper Famennian erosion, these Lower Carboniferous vertical movements were small, although their effect would have been quite considerable since ". . . . the removal of only a few metres of strata in this very condensed succession already results in considerable time gaps" (Higgins *et al.*, 1964, p. 224).
Kullmann (1963, p. 318) came to the conclusion that during the Visean the sea transgressed onto the land from the south and southeast to the north and northwest. However, according to this author, it would be better to consider such a transgression as ". . . . eine Belebung der Sedimentation, womit wirkliche Transgressionen über einige Inseln nicht geleugnet werden sollen" (p. 316).
On the other hand, Wagner found arguments to support a Visean transgression encroaching from the northwest onto a landmass somewhere in the south (Wagner, 1955, p. 155; Higgins *et al.*, 1964; Rupke, 1965, p. 29).
It is quite evident that, following the Upper Famennian transgression, a period of slow sedimentation took place under shallow but quiet water conditions. Locally, epeirogenetic movements were demonstrated to have been still active, having caused small-scale transgressions and regressions associated with non-deposition and erosion.
No boundary seems to have been in effect between the Bernesga-Porma area and the Central Asturian dome region. Even the Ponton line, which was still effective during the Upper Famennian, must have lost its function during the Tournaisian and Visean, because both west and east of it, black shales and nodular limestones can be found overlying Upper Devonian deposits.
There are no signs of interrupted sedimentation between the Visean griotte and the Namurian Caliza de Montaña Formation. The intraclastic and oölitic rocks point to shallow water conditions, whereas for the deposition of the micritic type of rocks a somewhat deeper and quieter water has to be assumed.
Such a microcrystalline calcite ooze is considered to have been formed by a rather rapid chemical or biochemical precipitation. The micrite is rather bituminous, platy, and mostly barren of organic material. Pyrite crystals are frequent, which could point to a lack of sufficient oxygen in some parts of the basin.
In the Central Asturian dome region, even during the Namurian, epeirogenetic movements were active, as is demonstrated by the isopach map of the Caliza de Montaña (Fig. 44, p. 107). Two striking points are evident:

1. The decrease in thickness of the sequence toward the central Mampodre-Fontasguera-Ten subarea.
2. The coincidence of the minima of this isopach map with those of the Barrios quartzite map (Fig. 42, p. 104).

Table XII shows the relation between the Bretonic uplift, the erosion, the thickness of the Barrios quartzite, and the thickness of the Caliza de Montaña Formation.

It is thus evident that the areas having the strongest epeirogenetic uplift during pre-Upper Famennian times later show a very thin Caliza de Montaña development.
The only acceptable explanation is that during the Namurian, the epeirogenetic uplift known from the Upper Devonian, the Tournaisian, and the Visean, was still active. Periods of non-deposition or slow sedimentation resulted, leading to the locally very thin Caliza de Montaña. Fig. 45, p. 107 explains the occurrence of only one manganese top-layer, the manganese content being highest at places where the Caliza de Montaña has its minimal thickness.

Tabel XII. Results of epeirogenetic movements in the Central Asturian dome region.

	Bretonic uplift	Erosion	Thickness of the Barrios quartzite	Thickness of the Caliza de Montaña	Mn – level
Western part of the isopach map	slight	slight	400—800 m	300—400 m	very thin
Eastern part of the isopach map	strong	strong	100—200 m	100—200 m	rather thick
Mampodre subarea	very strong	very strong, to the Oville Formation	0 m	100—150 m	very thick
Fontasguera subarea	very strong	very strong, to the Lancara Formation	0 m	< 100 m	very thick, local ore deposits

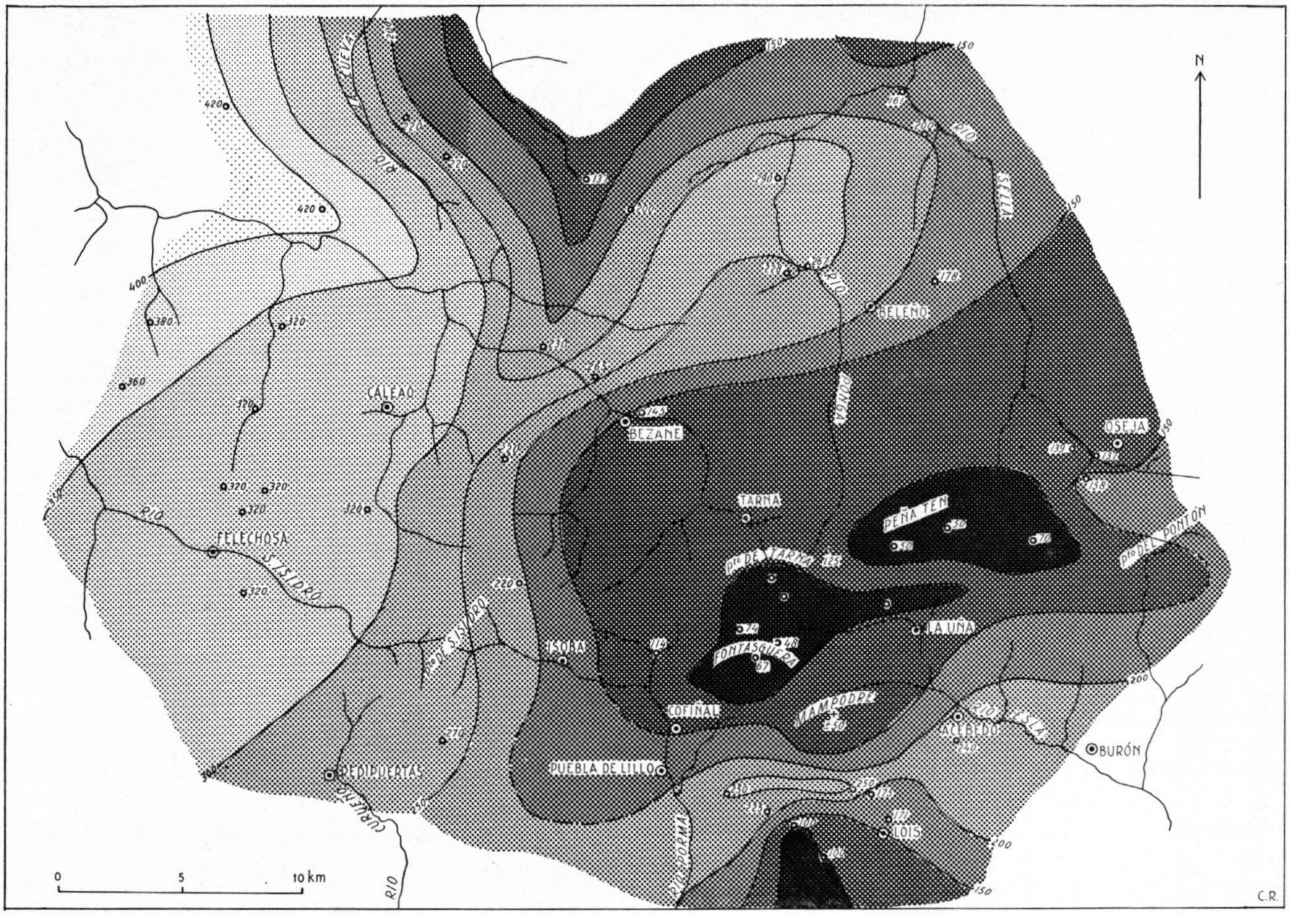

Fig. 44. Isopach map of the Caliza de Montaña Formation, Central Asturian dome region.

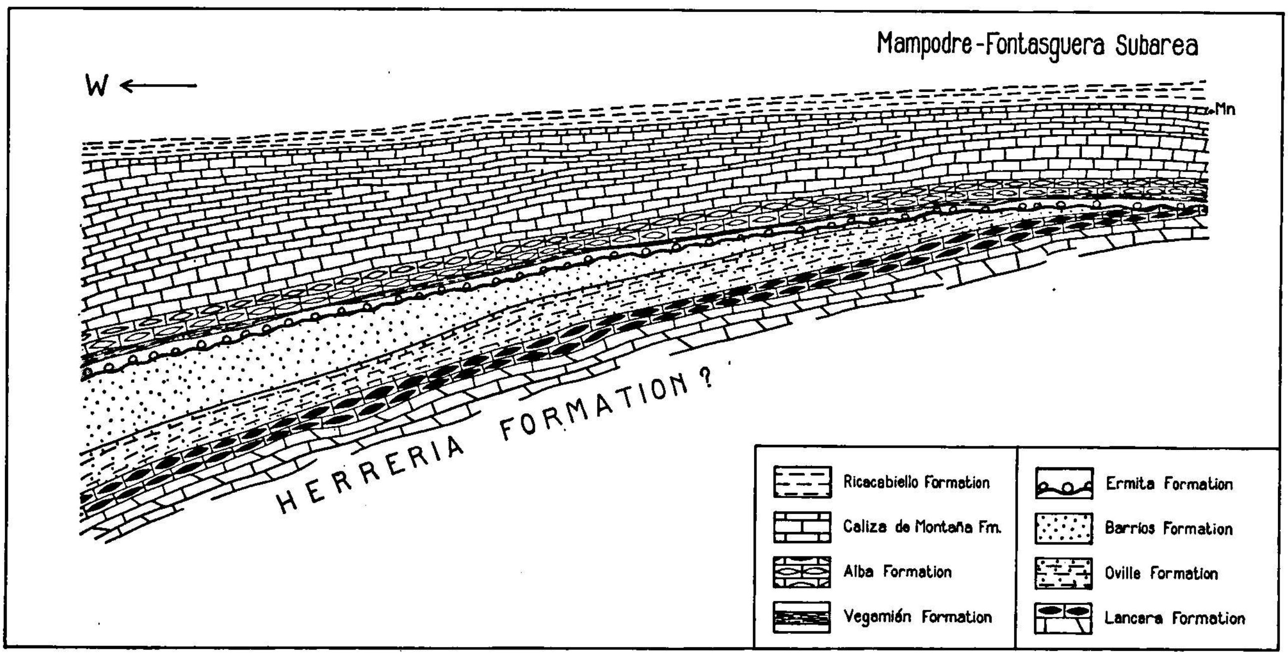

Fig. 45. The development of the manganese top-layer of the Caliza de Montaña in the Central Asturian dome region.

Apparently, the period of slow sedimentation and non-deposition first started at these places.
The same tendency of non-deposition alternating with slow sedimentation continued during the deposition of the Ricacabiello Formation, which probably corresponds in time and duration to the deposition of by far the greatest part of the relatively thick San Emiliano Formation in the Bernesga-Porma area.
It is striking that the manganese top-layer and subsequent Ricacabiello Formation are restricted to the Central Asturian dome region. Therefore, it was only after the sedimentation of the Caliza de Montaña that the León line resumed its function as a facies boundary, now separating the Central Asturian dome region with its slow sedimentation of the manganese top-layer and the similarly-deposited later Ricacabiello Formation from the Bernesga-Porma area with its thick limestone-shale-sandstone sequence of the San Emiliano Formation. Most probably, however, this separation was less clearly expressed in a small zone bordering the León line, where a similar manganese level and the Ricacabiello Formation cross the line over a short distance extending slightly to the south of it (Armada and Forcada units).
Again, the areas with the deepest Bretonic erosion level of the Bernesga-Porma area show in a later phase the closest resemblance to the development in our area, north of the León line.
This picture was gradually reversed by the end of the Namurian. Slow sedimentation changed to rather rapid sedimentation in the Central Asturian dome region, whereas shortly afterwards the sedimentation terminated in the areas south of the León line. There, sedimentary events were probably correlated with an important tectonic phase that produced many low-angle overthrusts (de Sitter, 1962 b,c; Rupke, 1965). With the exception of the Tejerina-Prioro area east of the Esla thrust zone and the Vegamián basin in the upper reaches of the Porma river, no younger Yuso deposits can be found south of the León line.
According to Rupke (1965, p. 67), "The thrusting formed the palaeogeographical limits during the Westphalian at the southern edge of the Asturian basin, the thrusted area forming relative highs due to stacking up of the thrusted series". This Leonide block was evidently being uplifted during the entire Westphalian, and thus formed the main source of sediment for the subsiding Asturian basin.
In the Cardaño area, van Veen (1965, pp. 73—74) supposed the Sudetic phase to be characterized mainly by epeirogenetic movements that caused strong upheavals in the source area of the Curavacas and Lechada Formations.
The main Asturian basin can be subdivided into various subbasins (Fig. 46, p. 109):

1. The Piedrafita-Lillo basin, in which the Lena Formation was deposited under marine to paralic conditions.
2. The Beleño basin, with a wholly marine depositional environment.
3. The Lois-Ciguera basin, characterized by its high amount of calcareous deposits.
4. The Huelde basin (Rupke, 1965, pp. 31—32; van Ginkel, 1965, pp. 188—189.)
5. The Maraña-Retuerto basin.

Without any doubt, the first two of these basins should be considered as one entity that once belonged to the Central Basin of Asturias, and perhaps even included the Lois-Ciguera basin.
Both the Huelde and Maraña-Retuerto basins, with their series of shales, sandstones, greywackes, conglomerates, limestones, and occasional coal deposits, are very distinct from the other basins mentioned above, with their regular series of limestones, shales, and graywackes. Since their graded bedding, slumping, and bottom structures all point to unstable conditions during sedimentation, the deposits of the Maraña-Retuerto basin are clearly the continuation of the thick molasse deposits in the east, e.g. the Curavacas and Lechada Formations described by van Veen (1965).
Van Ginkel (Brouwer & van Ginkel, 1963, p. 311; van Ginkel, 1965, p. 188), has already pointed to the difference in fusulinid associations occurring in the limestones of the Lois-Ciguera Formation and those of the Pando Formation. According to this author, two faunal provinces were in existence, possibly separated by tectonically active zones.
The present writer believes that the old highs, known from the Devonian up to and including the Namurian, were still active during the Westphalian. This implies the assumption of a ridge separating the Central Basin of Asturias to the west from the Yuso Basin to the east (Fig. 46, p. 109).
Yuso conglomerates are situated disconformably against this ridge along its whole eastern border, i.e. in the Huelde basin and west and northwest of the village of Riaño. Furthermore, the subsidence of a part of this high (Maraña-Retuerto basin) could have caused the assumed unconformable relationship between these upper Westphalian deposits and the Mampodre subarea. Locally, parts of this ridge protrude through the Upper Westphalian cover, as, for example, in the case of the Caliza de Montaña in the neighbourhood of Burón.

With the Asturian folding phase at the end of the Westphalian, sedimentation in its turn ceased in Asturias and northwestern León. Great parts of the Cantabrian mountains, including the area investigated in this paper, were folded, uplifted, and exposed to erosion. Shortly afterwards, however, Upper Westphalian D and Stephanian sediments were deposited in intramontane basins along the southern border of the Cantabrian mountains. These basins appear to be closely related to fundamental fault zones. According to Helmig (1965), onlap onto older Palaeozoic rocks coincides with a change in depositional environment from continental to paralic (Guardo-Cervera basin) into limnic (Sabero, Matallana, and Magdalena basins).

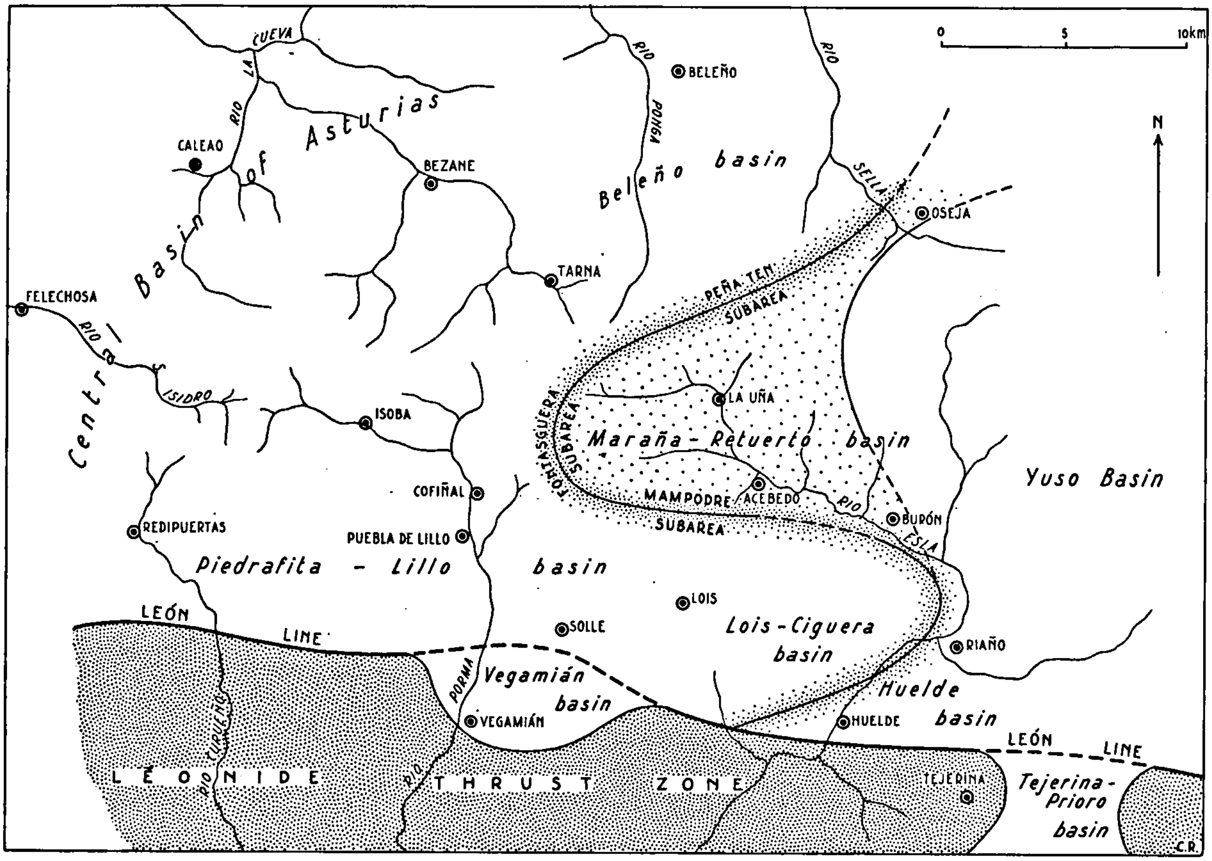

Fig. 46. The Asturian basin north of the Leonide thrust zone.

A second girdle of small restricted basins occurs along the León line. After Helmig (1965, Figs. 33 and 34), these Stephanian B deposits are the continuation to the northwest of the Prado Member; apparently sedimentation started somewhat later along this line.

In the area treated in this paper we find no record of younger sediments. However, Triassic, Jurassic, and Cretaceous rocks occur along the southern and southeastern border of the Cantabrian mountains, unconformably overlying older Palaeozoic deposits (Ciry, 1939; de Sitter & Boschma, 1966, pp. 216—219). Similar deposits occur in the northern part of the Cantabrian mountains (Schulz, 1858; Karrenberg, 1934; Llopis Llado, 1956). Martínez (1962) mapped the continuation to the west of the Mesozoic basin of Infiesto, north of the Isidro-Tarna-Ponton region. Julivert (1960) mentioned Triassic sediments unconformably overlying the Carboniferous between Soto de Sajambre and Amieva (Rio Sella, Asturias). Although upheaval, erosion, and blanketing throughout Mesozoic and Tertiary times can be demonstrated, the major morphogenetic uplift of the Cantabrian-Asturian mointain range is assumed to be of Oligo-Miocene date (Richter & Teichmüller, 1933; Solé Sabaris, 1952).

Consequently, along the southern border of the Cantabrian mountains, Eocene-Oligocene deposits overlie the Cretaceous with only a slight unconformity and there is a strong unconformable relationship with the overlying Miocene conglomerates. The intensive erosion after the upheaval gave the mountain chain its present relief. Features of a Würm glaciation have been found throughout the whole area investigated in this paper.

STRUCTURAL GEOLOGY

A sketch map of the main structural trends and areas is given in Fig. 47, p. 110. The main deformation of the investigated area took place during the Hercynian orogeny, which consisted here of thrusting, folding, and faulting.
Thrusting and refolding of the overthrusted units must have taken place during the Upper Westphalian, because:

1. Folded, low-angle overthrusted units are found covering Upper Westphalian C deposits (Parme limestone, page 97).
2. Middle Westphalian C deposits (Loc. A-2, page 93) are cut off by the thrust plane north of Puerto Forno.
3. Middle Westphalian D deposits were folded together with some thrust sheets (Lois-Ciguera basin, page 87, Felechosa-Tarna syncline, page 94).

In the southwestern part of the mapped area, Stephanian B deposits unconformably overlie the previously folded older rocks (pages 87 and 97).
All these facts place the thrusting and refolding in a post-Middle Westphalian D—pre-Stephanian B time, i.e. the Asturian folding phase of Stille (1924).
The deformation of the Stephanian B deposits in all likelihood reflects the Saalic folding phase of Stille.
The twofold subdivision of the mapped area already described (page 71), proved to be useful again, since each subarea shows differences in the character of its structural units. These subareas are the Isidro-Tarna-Ponton subarea and the Mampodre-Fontasguera-Ten subarea.

A. ISIDRO-TARNA-PONTON SUBAREA

In this subarea the Bretonic phase (page 69), with its epeirogenetic movements, caused an erosion that always cut down to the Ordovician Barrios Formation. Therefore, in this subarea a thick Barrios quartzite occurs below the Carboniferous and Upper Devonian sequence (Fig. 42, p. 104).
The structural features caused by the Asturian phase consist here of low-angle thrusting over great distances, followed by refolding and faulting. The low-angle overthrusts originated from NW-SE-striking anticlines that had broken through.
Due to the highly incompetent character of the Lancara Formation as compared to the Barrios quartzite, the Lancara Formation frequently acted as a detachment plane, bringing the Cambrian into contact with younger lithostratigraphic units.
Among the structural units distinguished, the Cabonero and Peña Cruz thrust units are typical examples. These structures are located at the southeastern corner of the map, between the Lois-Ciguera region and the Mampodre subarea.
Thrusting caused a threefold repetition of formations ranging in age from Cambrian through the middle of the Westphalian. Fig. 48, p. 111, shows this situation; the structural sections were taken perpendicular and parallel to the E-W-striking folds.
In both the Cabonero and the Peña Cruz units it was the Lancara Formation that acted as detachment plane; locally, however, the Oville Formation took over this function.
The map and sections show clearly that the northern

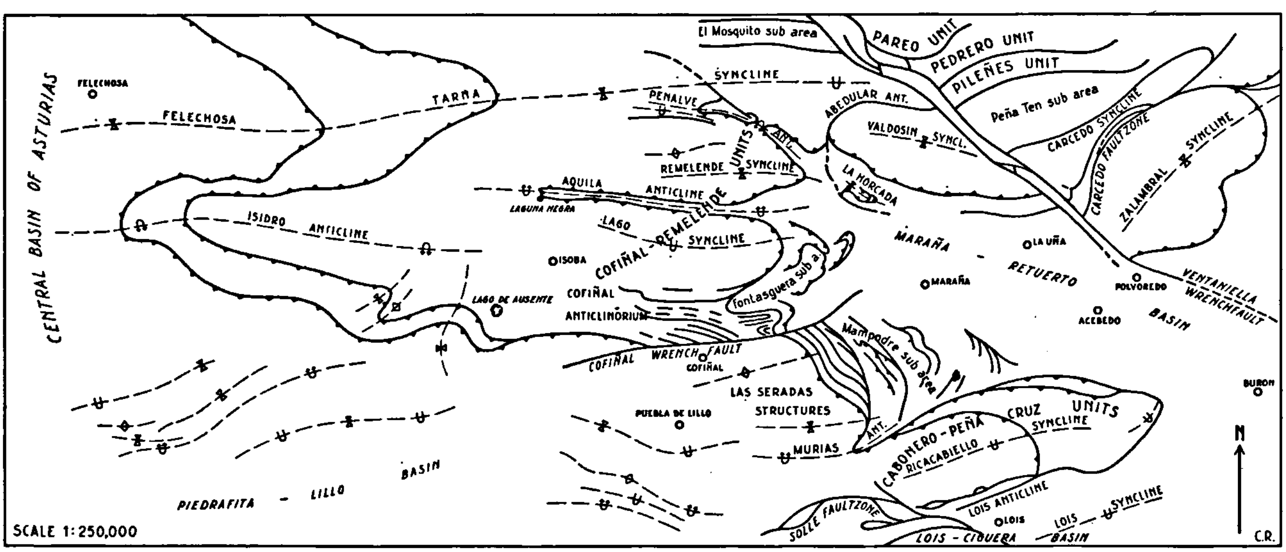

Fig. 47. Survey map of the main structural areas and trends in the San Isidro-Porma area.

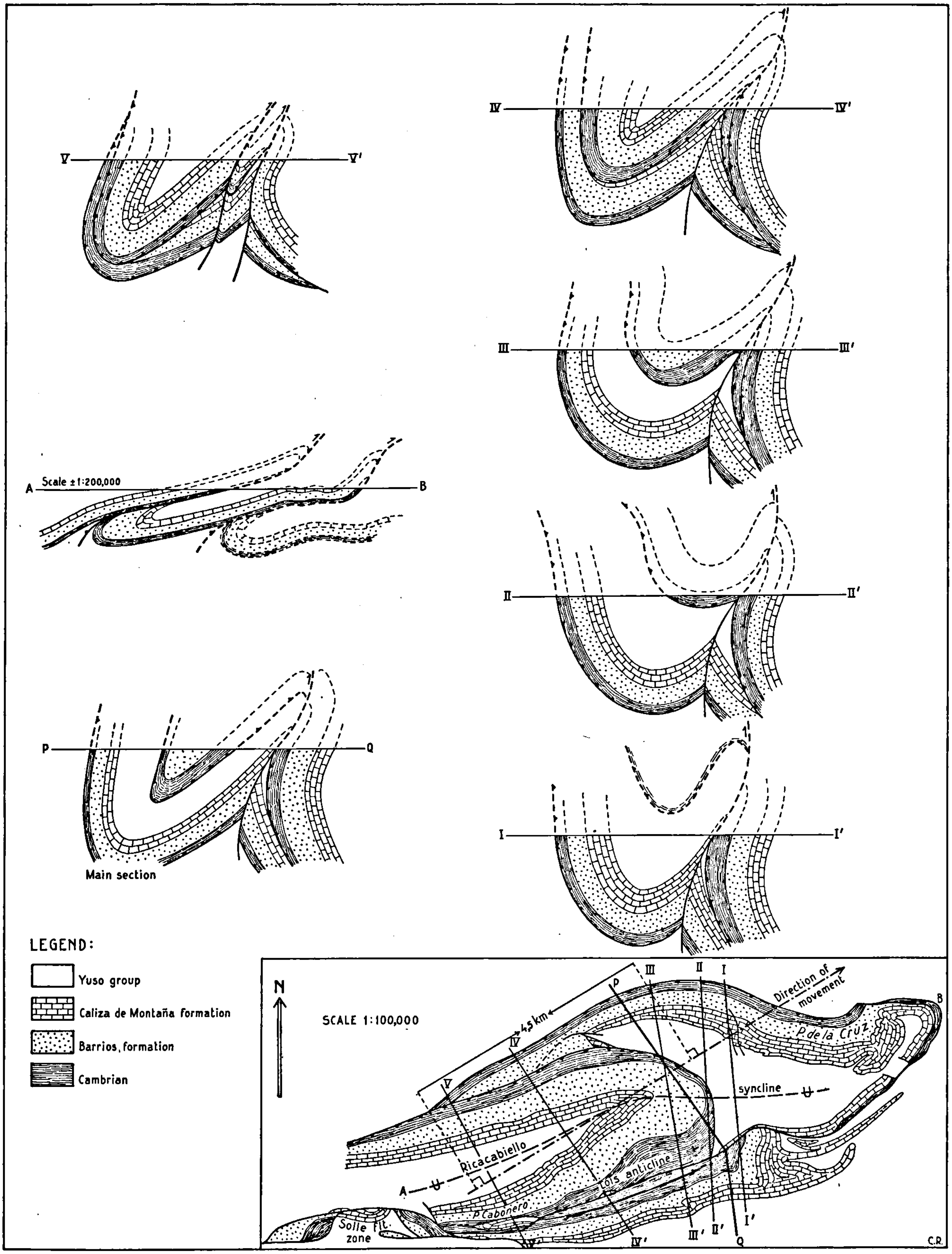

Fig. 48. The Cabonero and Peña Cruz thrust units, map and sections.

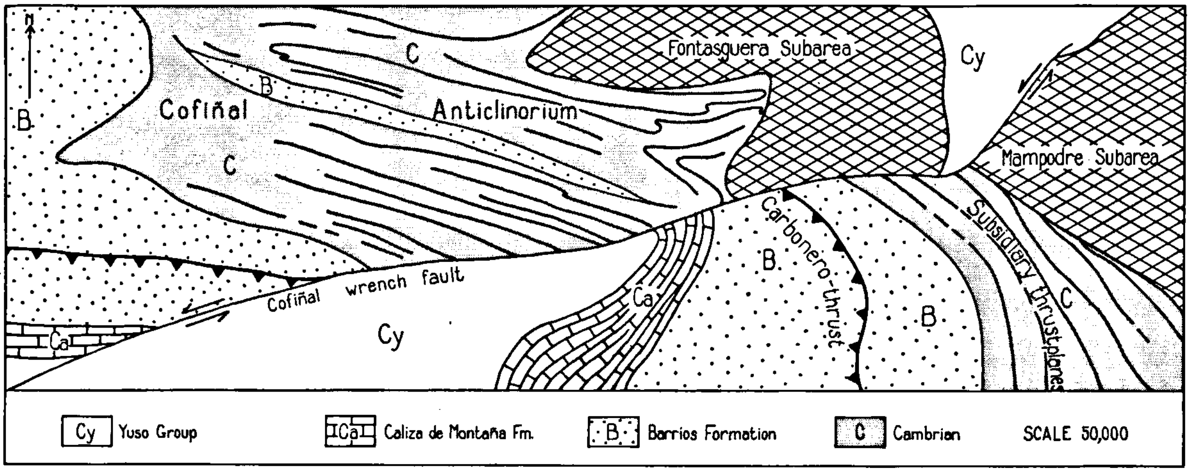

Fig. 49. The Cofiñal wrench fault.

part of the Cabonero thrust plane cuts upward sharply through the lithostratigraphy of the underlying Peña Cruz unit. Within a distance of about 5 kilometres, the thrust plane cuts through the Cambrian, the Ordovician, the Upper Devonian, the Lower Carboniferous and, lastly, the greywackes and shales of the Westphalian. In addition, the thrust plane cuts upward into the stratigraphy of the nappe, finally lying just below the lower limit of the Barrios Formation.

Successive to the thrusting, strong refolding in different directions has been established, the E-W-striking folds being the most important.

The main structures in the Cabonero and Peña Cruz thrust units are: the Ricacabiello syncline, the Lois anticline, and the Solle fault zone.

The Ricacabiello syncline is an overturned structure with a steep, mostly inverted, northern flank. To the east, its E-W trend changes rapidly to a NE-SW direction. The axis of the synclinal nose plunges steeply to the SW.

The Lois anticline is an overthrusted anticline with an axial plane dipping steeply to the north. South of Pico Cabonero, this anticline shows in its core a tectonic fenster which is the western continuation of the folded Peña Cruz thrust unit (Fig. 48, section V—V'. Further to the west, the anticline has been disturbed by the complicated fault zone east of the village of Solle. To the northeast, between Pico Montote and Pico de la Cruz, the amount of thrusting is maximal (section L—L').

The Carboniferous rocks of the Lois-Ciguera basin, bordering the thrust units, were folded in essentially the same E-W direction. In this basin a sequence from the Lower Westphalian through part of the Westphalian D shows uninterrupted sedimentation. There is a sharp angular unconformity with Stephanian B deposits. Consequently, thrusting and folding of both Lois-Ciguera basin and thrust units, subsequent to sedimentation, must be dated as post-Middle Westphalian D—pre-Stephanian B.

To the west, the Peña Cruz unit is cut off by the Cabonero thrust. The Cabonero thrust can be traced westward into an overturned anticline. From this Murias anticline the axial plane dips steeply to the north. In its core, the Lancara Formation has been thrust against the Oville Formation of the Mampodre subarea.

To the north, the thrust plane cuts upward in the stratigraphy of the thrust sheet until it reaches the base of the Barrios quartzite. The unit below the thrust consists of Barrios quartzite with in front at least six subsidiary thrust planes. The thrusted Barrios quartzite and its overlying Carboniferous deposits were refolded in a broad E-W syncline and anticline with subvertical axial planes.The axes plunge about 50° to the west. In combination with the Murias anticline, these structures have been called Las Seradas structures. To the north, these structures are cut off by the Cofiñal wrench fault. This fault brought the Westphalian shales and greywackes into contact with all the older lithostratigraphic units in this area.

Fig. 49, p. 112, and sections F-F', G-G', and H-H' show that:

a. The Cabonero thrust plane finds its continuation to the west in the Felechosa thrust plane. The associated Barrios quartzite and Caliza de Montaña can be correlated with identical deposits occurring west of the Porma river.
6. The continuation of the Barrios quartzite and the subsidiary thrust planes lying in front of the Las Seradas structures are found in similar structures occurring in the Cofiñal anticlinorium.
c. The Fontasguera subarea once formed an integral unit with the Mampodre subarea.

Therefore, everything points to a sinistral character of the wrench fault, from which a horizontal movement of 3 to 3.5 km can be established. The fault plane is vertical, and both its eastern and western extremities deviate from the roughly E-W general trend, now running in a NE and a SW direction, respectively. It

seems possible that the fault continues eastward, bordering the Mampodre subarea in the north.
The E-W trend points to a SW-NE-acting deformative stress as the cause of the movement along the fault plane. A stress having a similar direction caused the NW-SE folding responsible for the low-angle overthrusts.
Consequently, we assume the movement of the wrench fault to have been simultaneous with or slightly posterior to the thrusting and prior to the folding of the thrust plane, which has another character in the Mampodre and Fontasguera subareas.
South of the E-W-running quartzite ridge situated in the centre of the Cofiñal anticlinorium, the subsidiary thrust planes are not, or only slightly, refolded. This is in contrast to the strongly refolded thrust planes occurring north of the ridge, in a region bordering the Fontasguera subarea. The folding of the Fontasguera subarea and that of the thrust planes show the same type and trend. Consequently, we suppose that after the thrusting, the Cofiñal unit and the Fontasguera subarea were both refolded simultaneously.
The Lago syncline borders the Cofiñal anticlinorium to the north. The difference between this syncline and the structures lying in front of it to the east (see Fig. 55, p. 120) demonstrates clearly the influence of the Barrios quartzite, the former presenting a single overturned structure with a north-dipping axial plane, the latter displaying four strongly refolded small-scale overthrusts.
Further along the Lago thrust plane we reach the important overturned Aquila anticline north of the Lago syncline, with its axial plane dipping steeply to the north. The thrust sheet, with a constant stratigraphic level at its base (i.e. the Lancara Formation), can be followed over a distance of about 7.5 km, on top of Westphalian shales and greywackes. Finally, in the vicinity of the Laguna Negra this structure terminates as a small, strongly folded anticlinorium.
A comparable situation has been mapped north of the Remelende syncline. In the core of the Penalve anticline, Westphalian deposits crop out in a tectonic fenster. As an exception, the axial plane of this fold dips to the south.
The southern flank of the Remelende syncline has been disturbed by a normal fault. The fault plane is subvertical to steeply dipping to the north, the northern side of the fault being downthrown. The Penalve anticline shows a fault on its northern flank. This fault is primarily a wrench fault. Since the southern side of the fault is upthrown, movement in the vertical sense may be assumed.
The detachment plane of the Remelende syncline can be traced east of Puerto de Tarna, at the base of the Lancara Formation in the Valdosin syncline. Again, we find the complete stratigraphic sequence represented in this thrusted unit. North of this structure is the Abedular anticline. This anticline was strongly overthrust to the north, bringing the Lancara Formation, which formed its core, into contact with the Caliza de Montaña of its northern limb. At Puerto de Tarna the strike of this structure is NNE-SSW, which changes abruptly north of Pico Abedular into an E-W trend. Possibly the thrust fault continues to the south, joining the fault that cuts off the Horcada structures southeast of Puerto de Tarna.
North of the Penalve and Abedular anticlines we

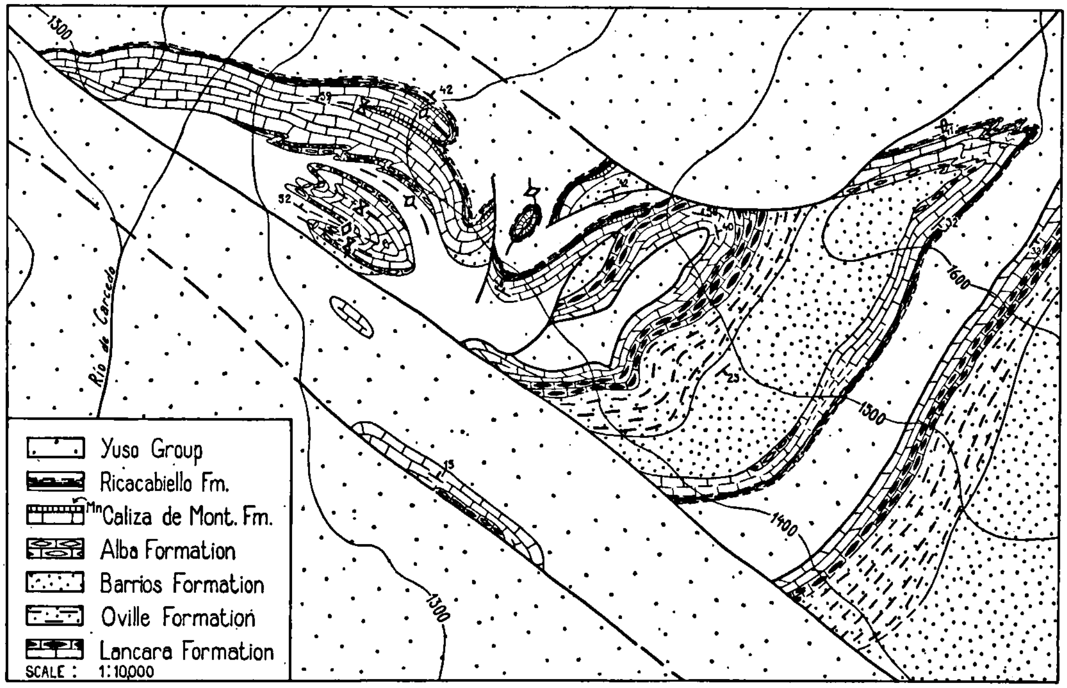

Fig. 50. The southern part of the Carcedo fault zone.

find the Tarna syncline, whose northern flank is steeply overturned, dipping to the north.
The Tarna and Valdosin synclines as well as the Abedular anticline are cut off abruptly by the important NW-SE-running Ventaniella fault zone. In a narrow zone between two parallel wrench faults, all formations from the Cambrian through the Westphalian were squeezed and faulted. Before describing this fault zone in more detail, it will be useful to review the structures occurring east of it.
The eastern extension of the Valdosin syncline is considered to be represented by the Zalambral syncline, its present position being due to the dextral displacement along this fault zone.
Perpendicular to the Ventaniella fault zone, another highly complicated fault zone occurs north of the Zalambral syncline. Part of this Carcedo fault zone is shown in Fig. 50, p. 113; it separates the Zalambral syncline from the strongly northward-overthrusted Carcedo syncline, only the north flank of which crops out.
Together with the Peña Ten subarea to the north, this situation resembles the southern border of the Mampodre subarea (sections L-L′ and K-K′). Here too, a wedging out of the Barrios quartzite has been assumed, now in the direction of the Peña Ten subarea. The detailed map and sections (Appendix I) of this particular subarea demonstrate the difference between the tectonic aspects of the Peña Ten subarea and its surroundings.
North of Pico Ten, the Pileñes, Pedrero, and Pareo overthrusts have caused a threefold repetition of stratigraphy, the thrust movement being here from north to south.
From the foregoing it is evident that the refolding, which produced large E-W-trending folds, occasionally resulted in overthrusting promoted by the incompetence of the Lancara Formation. South of the Peña Ten structure the direction of movement is then roughly from south to north, but north of this structure it goes from north to south.
The El Mosquito Mountain, situated north of the Tarna syncline just outside the mapped area, is the western continuation of the Peña Ten subarea. In fact, the sections drawn east and west of the Ventaniella fault zone are very similar (Fig. 51, p. 114).

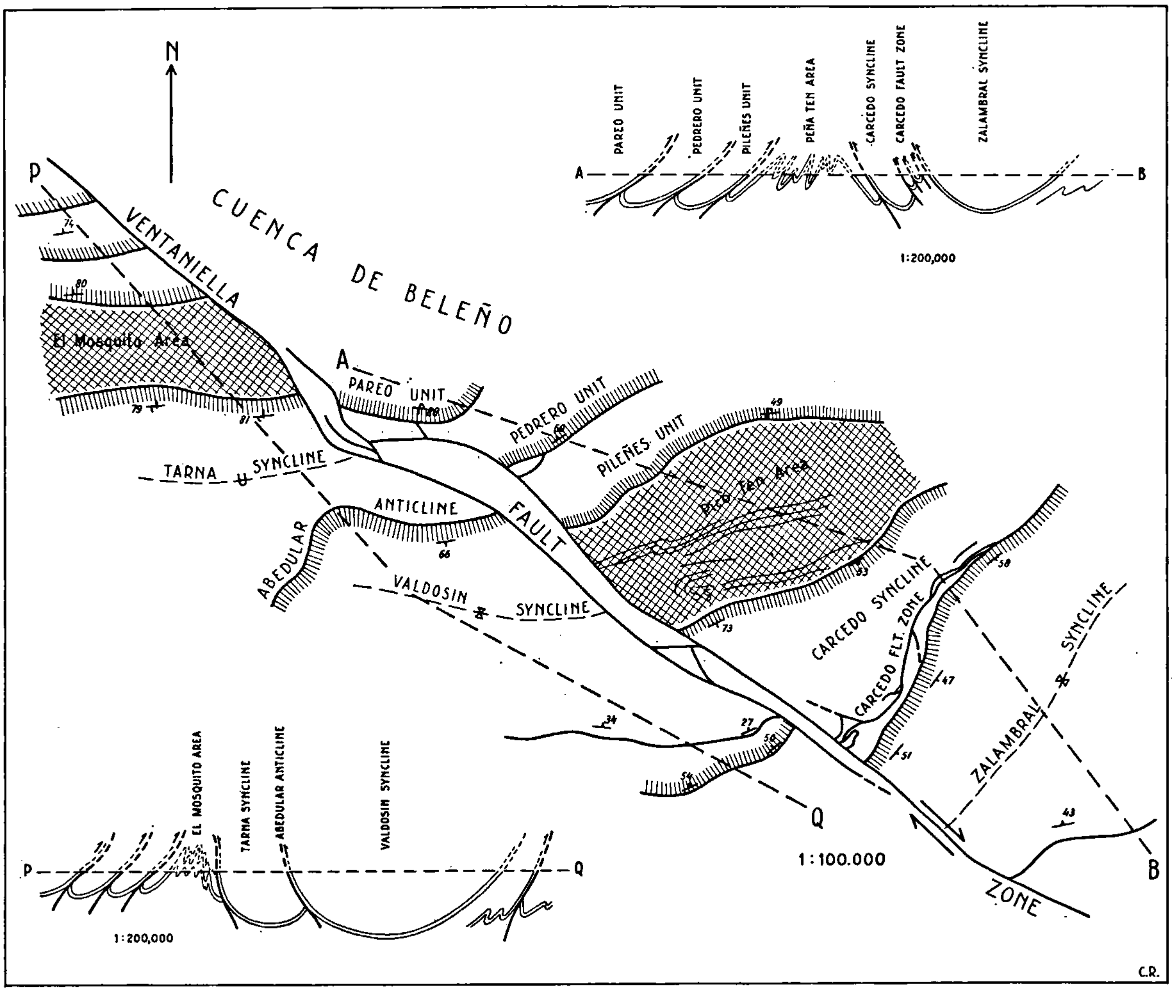

Fig. 51. The Ventaniella fault zone, map and sections.

South of the El Mosquito structure, the northern flank of the Tarna syncline was thrust over the El Mosquito region in a northerly direction, whereas north of this region at least three thrust units occur for which a north-south direction of movement has been established.
To summarize, two kinds of thrusting can be observed:

1. Low-angle overthrusts, which originated from NW-SE-trending folds.
2. Overthrusts, which originated from the E-W-trending refolding, the thrust movement having caused only a relatively small amount of upthrow.

The threefold repetition of stratigraphy displayed by the Cabonero and Peña Cruz units is due to low-angle overthrusting; later refolding produced the overthrusted Lois anticline. The threefold repetition of the units occurring north of the Peña Ten subarea is, to the contrary, the result of the refolding that produced the three overthrusted anticlines.

The great similarity between the structures west and east of the Ventaniella fault zone (Fig. 51, p. 114) characterizes the fault movement as an event successive to the refolding, but whether this took place during a late Asturian, a Saalic, or a Tertiary folding phase remains in doubt.
With regard to this problem, it strikes the observer that:

1. The SE-NW trend, together with the dextral character of the fault zone, points to a N-S-acting stress field, a direction which fits very well with the stress directions during both the Asturian E-W refolding phase and the Saalic phase (de Sitter, 1962, p. 263).
2. The presumed continuation of the fault in the northwestern direction, west of the Cuenca de Beleño, is cut off by the Alpine normal faults bordering the Mesozoic and Tertiary deposits east of the village of Pola de Siero (province of Oviedo).
3. To the southeast, the fault zone most probably continues into the Cardaño line (v. Veen, 1965), and this connection with the Cardaño line, too, makes a late Hercynian age very probable.

Consequently, a late Asturian or Saalic age seems to be the most likely.
In the area investigated, however, Saalic E-W folding seems to be restricted to the Stephanian B deposits. Hence, if late Asturian is the correct age of the fault movement, the present straight appearance of the fault line makes it unlikely that the Saalic folding phase affected the trend of the fault.
The fault zone separates an area with E-W-running structures, i.e. the Valdosin and Tarna synclines, from an area with roughly NE-SW directions, i.e. the Zalambral syncline. This suggests the existence of two blocks east and west of the fault zone. Apparently, the direction of the Asturian stress field during the refolding period was influenced by these blocks, causing different stress directions east and west of the fault zone.
From a comparison of two equivalent points on both sides of the fault, a horizontal shearing movement of about 4 to 5.5 km can be estimated. Furthermore, the present, nearly closed plunge of the southwestern part of the Zalambral syncline with respect to the broad Valdosin syncline, gives evidence of an important vertical component in the movement, the eastern side of the fault being upthrown.

West of the Cofiñal-Remelende unit, two more thrust units have been mapped, i.e. the Isidro and the Felechosa units. Like the previously-described units, both have been strongly refolded in a predominantly E-W direction.
In the northern part of the mapped area, all overthrusted units have been folded in a syncline called the Felechosa-Tarna syncline. The axial plane of this structure dips steeply to the north or south. At its southern extension we find the Isidro and Felechosa units folded in an anticline called the Isidro anticline. The axial plane of this structure dips steeply to the south.
The structures occurring west of Lago de Ausente show that mainly NE-SW and NNW-SSE folds disturbed these dominantly E-W directions, but whether these folds were formed simultaneously with or posterior to the E-W folds cannot be ascertained. Both directions can be distinguished in the diagram of the axial planes of folds found in the Lancara Formation of Pico Remelende, just west of Puerto de Tarna, Loc. 4—64B (Fig. 52, p. 116).
At a few locations (Fontasguera subarea, page 120; Fig. 55), even N-S-running folds seem to be deformed by E-W directed folds.
North of the village of Felechosa, the Felechosa thrust fault, which lies at the base of the Barrios quartzite, cuts sharply downward into the lithostratigraphy of the underlying Isidro thrust unit. Carboniferous deposits from the Lower Visean Alba Formation up to and including the Westphalian were cut off by the thrust, which brought the Barrios quartzites of both units into contact with each other. Somewhat more to the west, the thrust lies at the base of the Lancara Formation, but its continuation shows that by far the greatest part of the thrust action had the Oville Formation as detachment horizon. Over a great distance the thrust can be traced lying at the top of the Barrios quartzite of the Isidro unit. Only two places have been mapped, west of Pico de Fuentes, where Upper Devonian limestones and Caliza de Montaña can be found cut off by the Felechosa thrust plane. South of Lago de Ausente, the thrust has been traced cutting into the Barrios quartzite of the Isidro unit and finally cutting off the whole Isidro thrust unit. From there on, the Felechosa thrust is in contact with the Westphalian of the Cofiñal-Remelende unit, cutting rapidly downward through the Ricacabiello Formation, the Caliza de

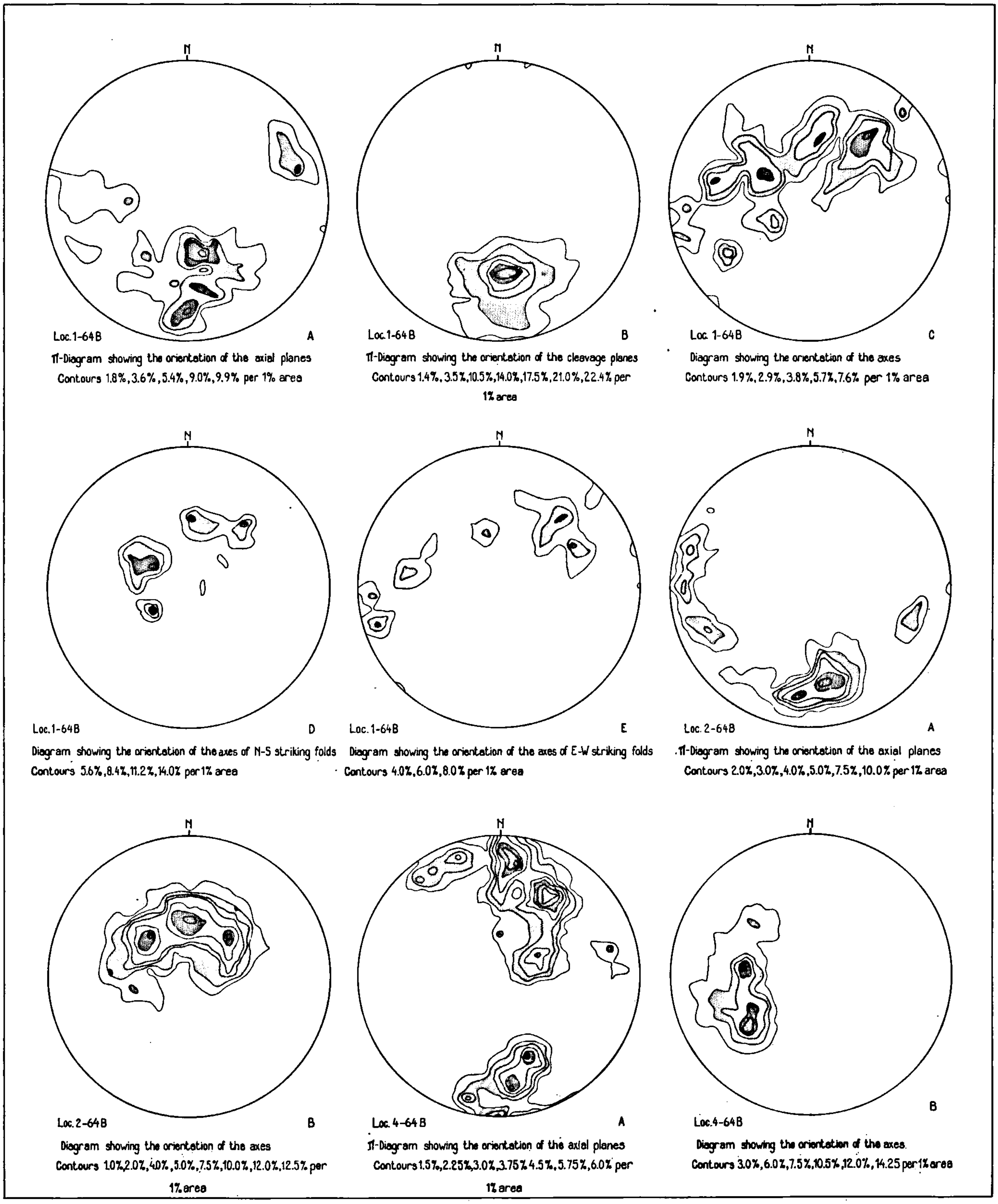

Fig. 52. Orientation of the axial planes, cleavage planes, and axes of Hercynian folds.

Montaña, the Alba Formation, and part of the Barrios quartzite, until it is itself cut off by the Cofiñal wrench fault. South of this wrench fault the same thrust can be traced in the Las Seradas structures, finally cutting off the fault plane of the Cofiñal-Remelende unit.

The Isidro thrust fault shows the same picture as the Felechosa thrust. It runs first along the base of the Barrios quartzite, bringing this unit into contact with the Westphalian of the Tarna syncline; then, more to the west, south of Pico Torres, the thrust passes through the Oville Formation; and finally it cuts down to the Lancara Formation. To the east, from Pico de Fuentes onward, it again lies at the base of the Oville or Barrios Formation. The thrust cuts downward through the shales and greywackes overlying the Caliza Masiva, until it cuts off a limestone of Westphalian C age (page 93) north of Puerto Forno. From there on, as far as Peña del Alba, it follows the same stratigraphic level just above the Caliza Masiva, but in the core of the Isidro anticline even this limestone is cut off sharply by the thrust fault.

In the vicinity of Lago de Ausente a sequence of Caliza de Montaña and Barrios quartzite, interpreted here as belonging to the Cofiñal anticlinorium, is exposed below the thrust (section D—D′). The thrust cuts downward into the Barrios quartzite of this sequence, then cuts upward again, and finally is cut off by the Felechosa thrust fault.

Thus, the Isidro anticline displays in its core a tectonic fenster exposing the Westphalian of the Cofiñal-Remelende unit.

As we have already seen, the Felechosa-Cabonero thrust plane can be followed till it disappears in the complicated fault zone north of the village of Solle, near the León line. The map prepared by Martínez (1962), can be interpreted as implying the continuation of the thrust plane outside our area far to the north, as far as Pico Trigueiro. The direction perpendicular to the line joining both extremities of the thrust plane is parallel to the SW-NE direction of movement established for the Cabonero and Peña Cruz units. A similar direction of movement was estimated for the thrusting north of the village of Beleño (province of Oviedo), where a whole set of at least five overthrusted anticlines occurs (Julivert, 1960). With this direction in mind, the amount of thrusting for both the Felechosa and the Isidro units can be estimated as 7.1 and 10.1 km, respectively. Similarly, the tectonic fenster of the Aquila anticline indicates a 7 km thrust movement for this part of the Cofiñal-Remelende unit.

As shown by the thrust movements mentioned above (4, 5, 7, 7.1, and 10.1 km long), the estimated amounts of thrusting vary widely. The thrusting of the Cabonero and Remelende units against the western border of the Mampodre-Fontasguera subarea was caused not by a single thrust fault but rather by a large number of subsidiary thrusts which resulted in an unknown sum total of thrusting.

In the core of the Murias anticline thrusting is not very apparent at first sight, but movement is demonstrated by the anomalous contact between the limestone-dolomite member of the Lancara Formation and the Oville Formation of the Mampodre subarea.

The amount of movement along the Valdosin-Zalambral thrust plane is difficult to estimate, but the occurrence of movement is confirmed by the abrupt cutting off of the Upper Westphalian deposits east of Pico Parma and Pico Pozua and by the presence of presumably Upper Westphalian deposits in the Carcedo fault zone (section L—L′).

According to the map prepared by Julivert (1960), the whole eastern border of the Cuenca de Beleño appears to be overthrusted, with our Lancara Formation again acting as detachment horizon. The thrusting took place against a limestone massif (forming part of the Picos de Europa) which must be considered the northeastern continuation of our Pico Ten subarea.

In the north, the thrust continues as one of the abovementioned NW-SE-running overthrusted anticlines located north of the village of Beleño; the three E-W-running units north of the Pico Ten subarea form the overthrusted southern border of this Cuenca de Beleño.

The reader's attention is directed to the fact that the thrust movement during this refolding period always seems to be in the direction of a presumed high. The movement south of the Pico Ten-El Mosquito structures was from south to north, whereas to the north of it, a movement in the opposite direction has been demonstrated. A similar situation has been established for the overthrusted Lois anticline, where movement occurred in the direction of the León line (Las Salas high).

To summarize, we find the following:

I. In the south (Cabonero and Peña Cruz units) and west (Felechosa, Isidro and Cofiñal-Remelende units), low-angle overthrusting can be established on the basis of four tectonic fensters. Thrusting was followed by a strong E-W folding, causing:

   1. Structures with vertical to north-dipping axial planes, e.g. the Cabonero and Peña Cruz units and the Cofiñal-Remelende units. There is only one overthrusted E-W anticline, i.e. the Lois anticline, and this one displays a thrust movement from north to south in the direction of the León line.
   2. Structures with vertical to south-dipping axial planes, e.g. the Felechosa-Isidro units. No thrusts associated with the refolding were found here.

II. East of Puerto de Tarna, no tectonic fenster is found except in the Carcedo fault zone, but movement along the Valdosin-Zalambral thrust plane is demonstrated by the cutting off of the Parme limestone sequence.

Overthrusted E-W anticlines with axial planes vertical to south-dipping occur frequently. These display a thrust movement from south to north towards the Peña Ten-El Mosquito high.

III. No low-angle overthrusting could be demonstrated north of the Peña Ten-El Mosquito high. There, E-W-running anticlines with vertical to north-dipping axial planes caused thrusting—associated with the refolding—from north to south in the direction of the presumed high.

Thus, the succession of low-angle SW-NE overthrusting, followed by E-W refolding, which can be clearly recognized in the west, becomes less identifiable in the east and northeast. Low-angle overthrusting and E-W folding are closely related, both belonging to the same Asturian folding phase. At some places these phenomena clearly occurred successively, but locally we found no evidence that thrusting had preceded the E-W folding.

B. MAMPODRE-FONTASGUERA-TEN SUBAREA

In this area Barrios quartzite does not occur below the Upper Devonian deposits. The absence of this competent layer influenced the tectonic style of the region, which displays a mainly isoclinal type of folding.

1. *Mampodre subarea*

The structure of this particular subarea is illustrated by Fig. 53, p. 118, and sections I—I', J—J' and K—K'. These sections show a strong isoclinal folding, often with overthrusted anticlines.

The NW-SE direction occurring in the neighbourhood of Rio de Valverde, in all probability corresponds to the NW-SE-running folds responsible for the low-angle overthrusts treated in the fore-going. Consequently, the intensively folded Cuesta Rasa must once have been such a NW-SE anticline later deformed by refolding.

An unconformable relationship between the Caliza de Montaña of the Mampodre subarea and the Upper Westphalian deposits of the Maraña-Retuerto basin is assumed (see page 108), but an exposure of the contact between these two units has not been found. The Cofiñal wrench fault may possibly continue to the east, bordering the Mampodre subarea in the north.

2. *Fontasguera subarea*

This particular region is shown in detail in Figs. 54 and 55, pp. 119 and 120. Isoclinal folding, overthrusting, and refolding occur frequently. The overthrusted units cut each other off, the extreme eastern ones also cutting the Upper Westphalian deposits of the Maraña-Retuerto basin (Fig. 56, p. 121). The Lancara Formation frequently acted as detachment horizon, as did the shales of the Oville Formation and locally even the Upper Devonian limestone. Refolding took place in a mainly E-W direction (e.g. section H—H', Fig. 55), producing folds with very steep plunging axes (e.g. the Buecicardiel syncline).

N-S-trending folds occur; at one place, south of the Chapel of Riosol, an overthrusted N-S-refolded unit was very probably deformed by an E-W anticline. This sequence of phenomena, however, appeared to be highly exceptional in the area investigated.

3. *Pico Ten subarea*

For detailed information the reader is referred to the map and structural sections of Appendix I. The whole region shows marked isoclinal folding in an E-W direction. The anticlines are overthrusted, with the Lower Visean nodular limestone acting as detachment horizon. The predominantly E-W direction corresponds to the E-W-running folds found north and south of the Pico Ten subarea. Other directions, although they occur frequently in both the Mampodre and Fontasguera subareas, are of only minor importance here.

To the west, the Ventaniella fault zone cuts off this subarea. Part of it has been squeezed to the north

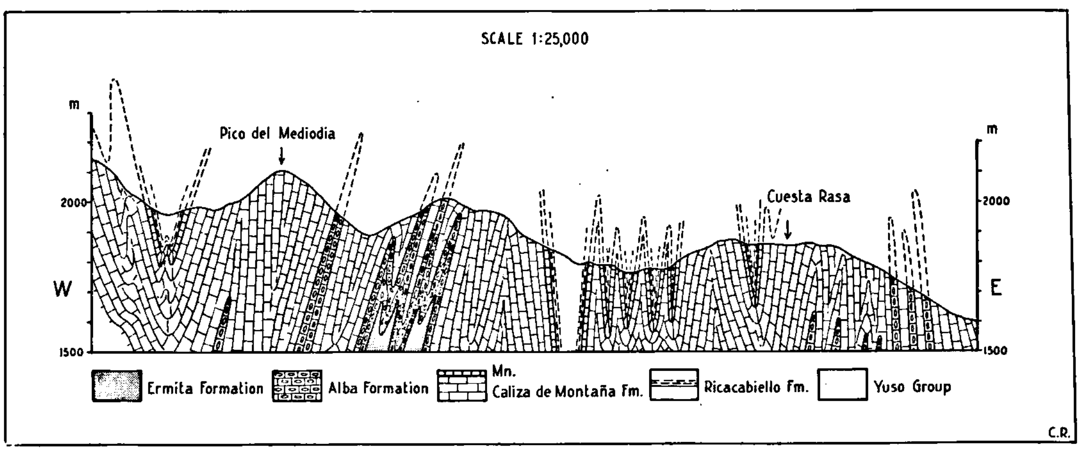

Fig. 53. Structural section through the Mampodre subarea.

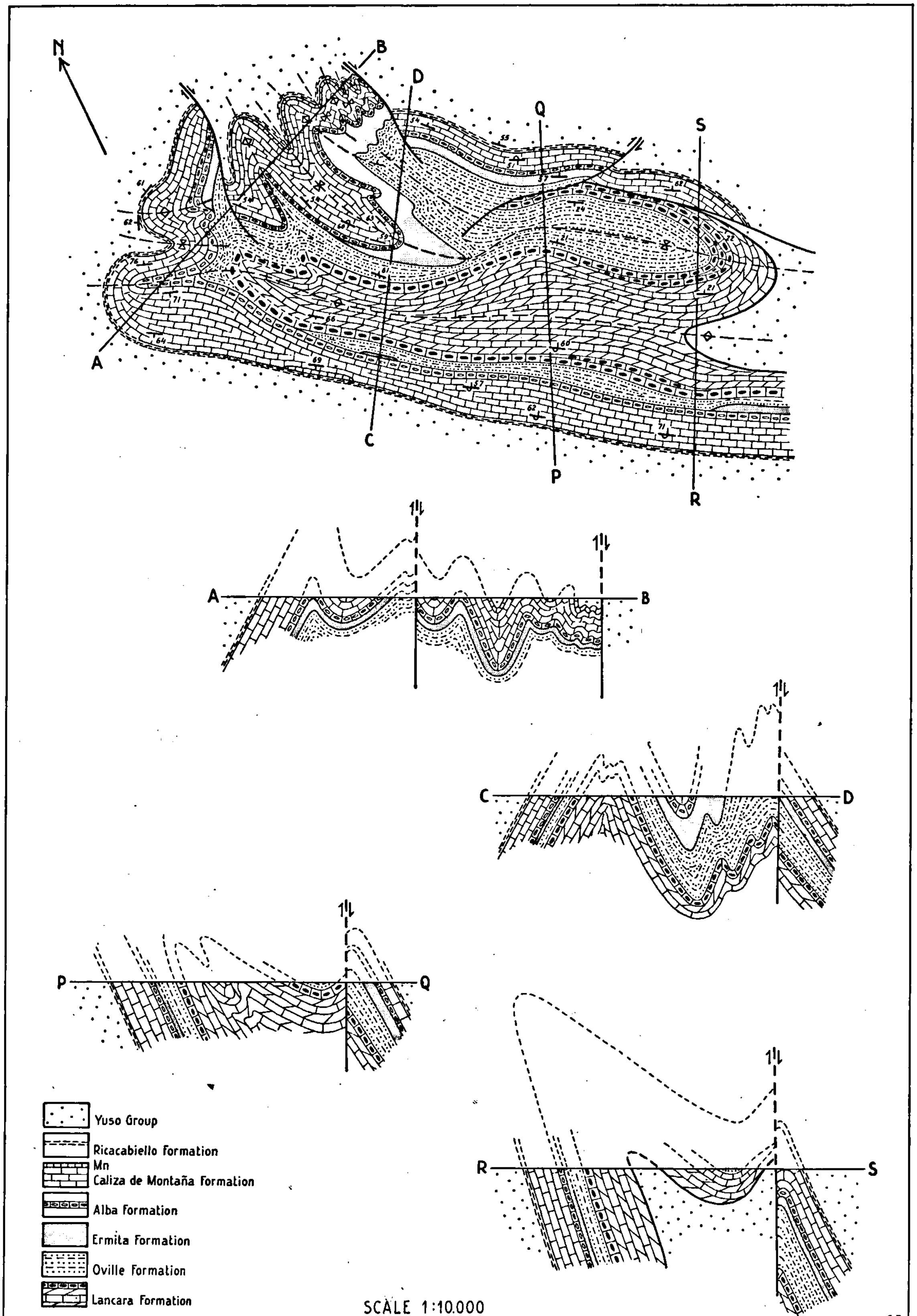

Fig. 54. The northernmost part of the Fontasguera subarea, map and sections.

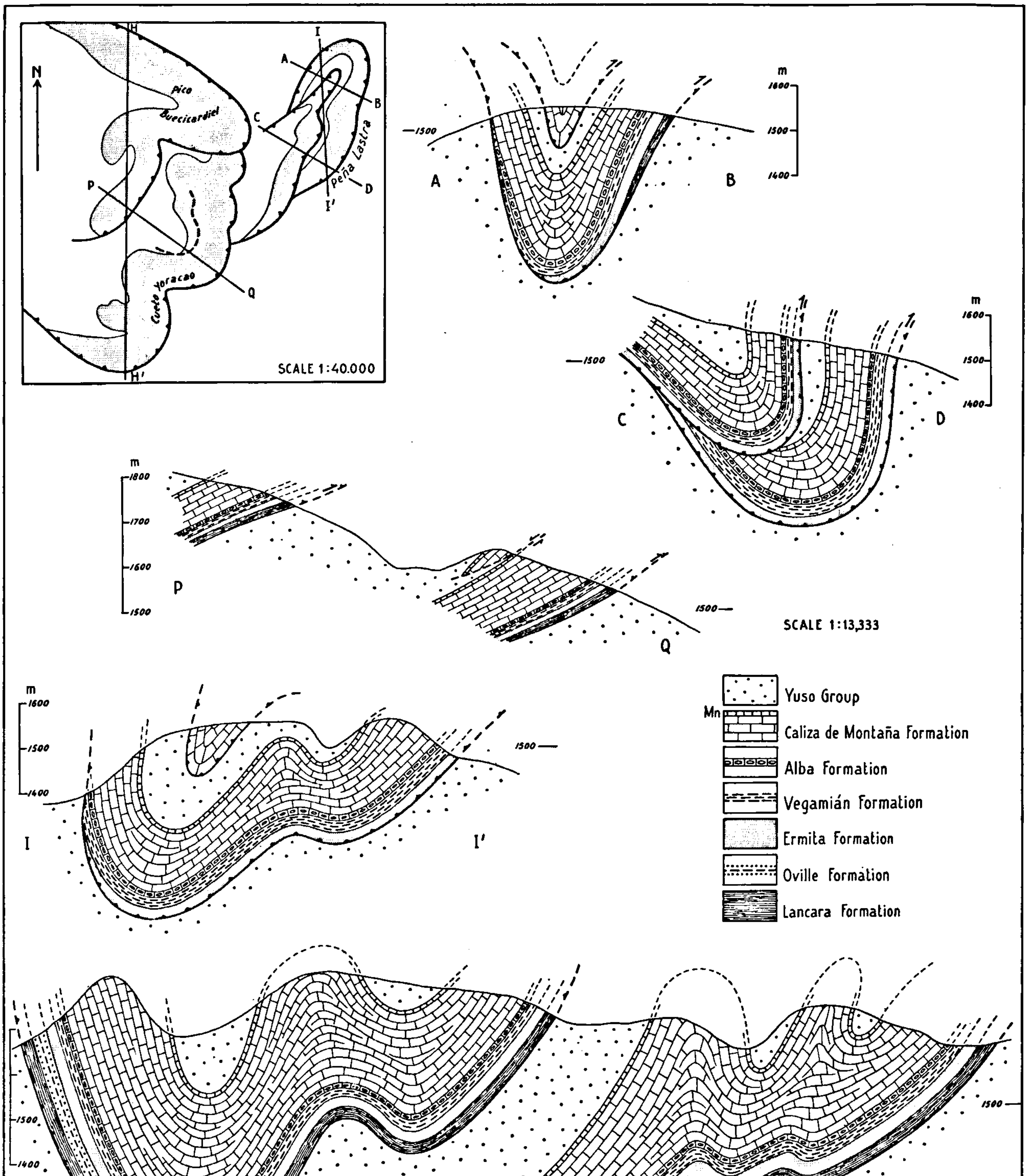

Fig. 55. Structural sections showing cross-folding in the Fontasguera subarea.

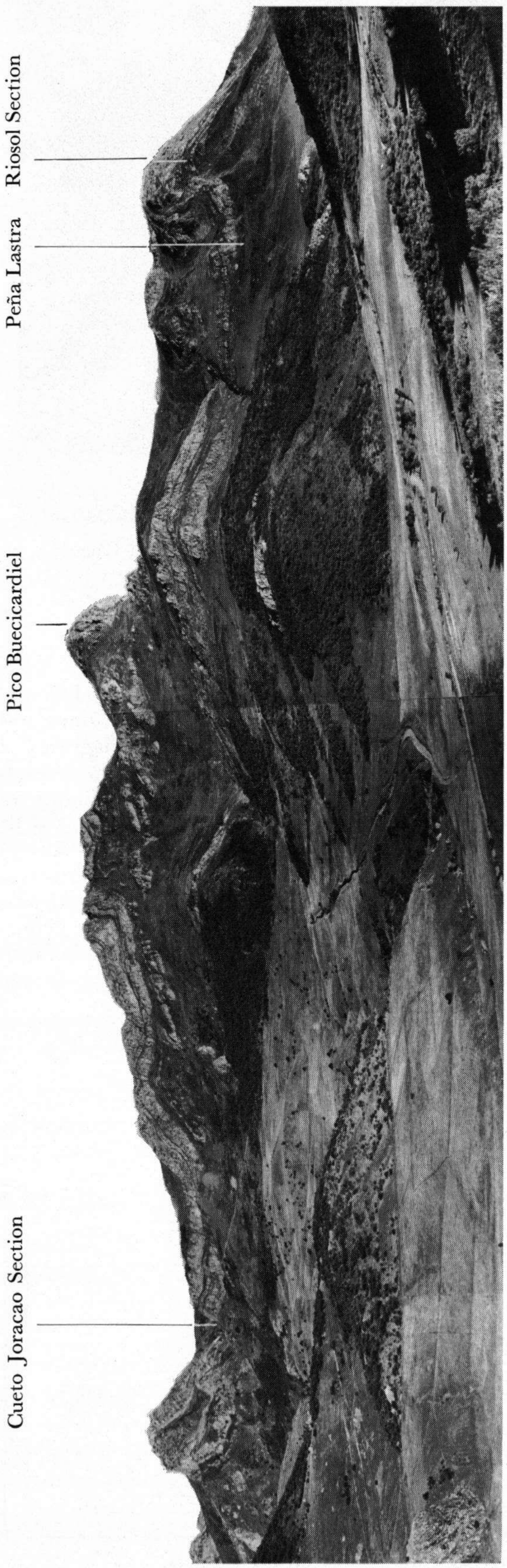

Fig. 56. The eastern side of the Fontasguera subarea, west of the village of Maraña.

Fig. 57. E-W isoclinal folding in Yuso deposits east of the village of Maraña (Loc. 1—64B).

Fig. 58. N-S minor folds in shales of the Yuso deposits east of the village of Maraña (Loc. 1—64B).

Fig. 59. N-S fold in greywackes exposed along the main road between the villages of Acebedo and Lario.

between the two parallel faults, the El Mosquito subarea being the western continuation of the Pico Ten subarea.

The structures present in the Yuso deposits of the Maraña-Retuerto basin resemble those of the subareas treated above. Strong isoclinal folding in a predominantly E-W direction is seen frequently (Fig. 57, p. 121). The hinges of these folds can rarely be found, but the isoclinal character of the folding can often be demonstrated by top and bottom determinations (graded bedding and bottom structures). The axial planes of the folds dip to the north, and an axial plane cleavage has been developed in the shales. Overthrusted anticlines occur frequently.

Folds trending N-S occur both as minor folds (amplitude 5 cm to 0.5 m) in the shales (Fig. 58, p. 121) and as much larger folds in sandstone and greywacke layers (Fig. 59, p. 122). The diagrams of the axial planes, axes, and cleavage planes of folds occurring at two localities east of the villages of Maraña (Loc. 1—64 B) and La Uña (Loc. 2—64 B) are given in Fig. 52, p. 116. North of the village of Vegacerneja (Fig. 1, p. 58), a broad overturned syncline with an axial plane dipping to the north was mapped (Fig. 60, p. 122). In front of this syncline, however, top and bottom structures seem to confirm the mainly isoclinal, overturned character of the folding in this part of the Yuso Basin.

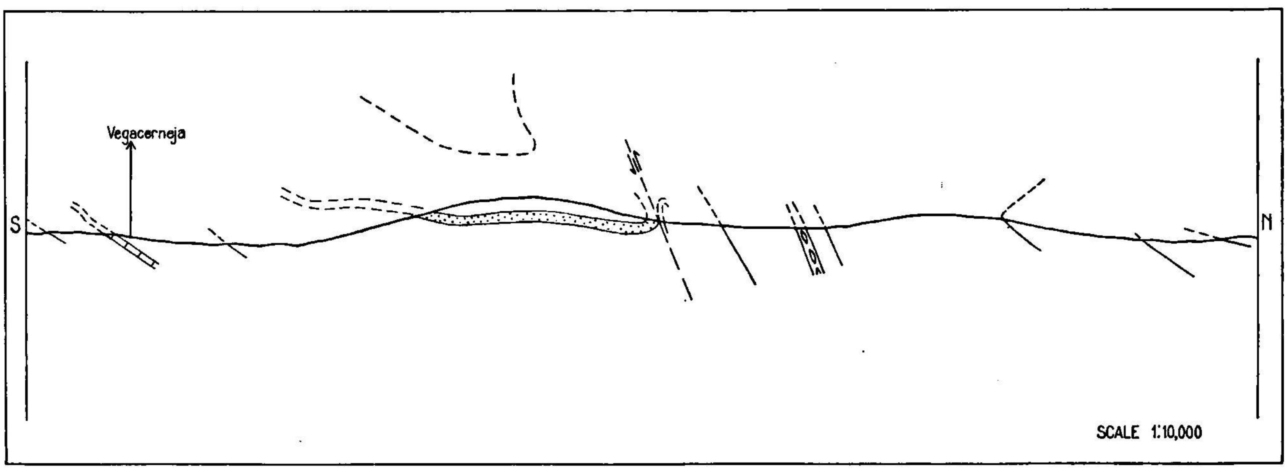

Fig. 60. Structural section through Yuso deposits, north of the village of Vegacerneja.

IGNEOUS ROCKS AND ORE DEPOSITS

In the mapped area, Cambro-Ordovician hypabyssal and volcanic rocks occur. A dolerite can be found as a sill in the shales and sandstones of the Oville Formation south of the village of Liegos and as a dyke in the quartzites of the Barrios Formation east of the village of Solle.
Tuffs and tuffaceous sandstones occur in both the Oville and the Barrios Formations east of the village of Lois and south of Pico de Fuentes. Similar rocks, which also occur in the Oville and Barrios Formations, have been studied by Mr. C. F. Winkler Prins (internal report) in a region south of the León line. A close resemblance between the Cambro-Ordovician igneous rocks north and south of the León line was established.

The dolerites are easy to discern in the field, because of their dark-green to brown colour, roundish forms, and resistance to weathering, which are in contrast to the shales and sandstones of the Oville Formation.
Thin sections show the following. The rock is fine-grained, holocrystalline, and typically non-porphyritic. It has two chief constituents, felspar and augite, both strongly decomposed. It frequently shows an intersertal texture, interstitial patches of sferulitic chlorite and leucoxene, both occurring in a plexus of lath-shaped plagioclase crystals. The ophitic texture commonly found in this type of rock, is seen only occasionally in the dolerites under consideration. The felspar of the dolerite varies between andesine (41 % An) and labradorite (62 % An).
Regeneration gave rise to fresh and glassy felspar (albite-oligoclase) with enclosed epidote (Fig. 61, p. 123). The basic type of felspar now shows strong decomposition into sericite and saussurite or calcite. The augite shows strong decomposition into pale-green chlorite, calcite, yellowish-brown pleochroic biotite, and yellowish-green pleochroic actinolitic hornblende. Chlorite is also seen, as very fine-grained sherulitic aggregates. Bowlingite occurs as alteration product of olivine. Secondary patches of clear quartz and calcite are found in small vesicles and veinlets. Rutile and apatite occur as minor constituents, the latter occasionally showing needle-like crystals. Ilmenite and magnetite are abundant in the form of crystals or skeleton crystals. The ilmenite gave rise to its characteristic decomposition product, a grey cloudy semi-opaque leucoxene.

The pyroclastic rocks occur as enormous, irregular tuff masses, as small bombs and lapilli, and as matrix in the Oville sandstone. In thin sections the lapilli consist mainly of a groundmass of dark-green chlorite, crowded with closely-packed small sferulitic growths. These sferulites vary in shape, being mostly roundish and spheroidal but sometimes irregular or fitted together. They are composed of fine fibres of quartz and chlorite. A concentric structure is often seen, showing rims of leucoxene and green chlorite around a sferulitic nucleus, the latter giving an excellent black cross between crossed nicols. Roundish and spheroidal forms of leucoxene, green chlorite, and, more rarely, calcite, can also be found. Ilmenite and hematite crystals are scattered throughout the whole lapillus, and leucoxene occurs as an outer rim around the whole volcanic fragment. Clear quartz is present rather abundantly, in scoriaceous forms.
Occasionally, rock fragments are enclosed in the lapilli. In such cases the surrounding volcanic material shows flow-structures, due to stretching, and a parallel arrangement of the spherulites and the leucoxene concentrations (Fig. 62, p. 124).
Besides the lapilli and bombs mentioned above, quite different, greyish lapilli occur. These consist of a cryptocrystalline groundmass of leucoxene, quartz, and felspar, with few small crystals of biotite, muscovite, quartz, felspar, chlorite, ilmenite, and magnetite.

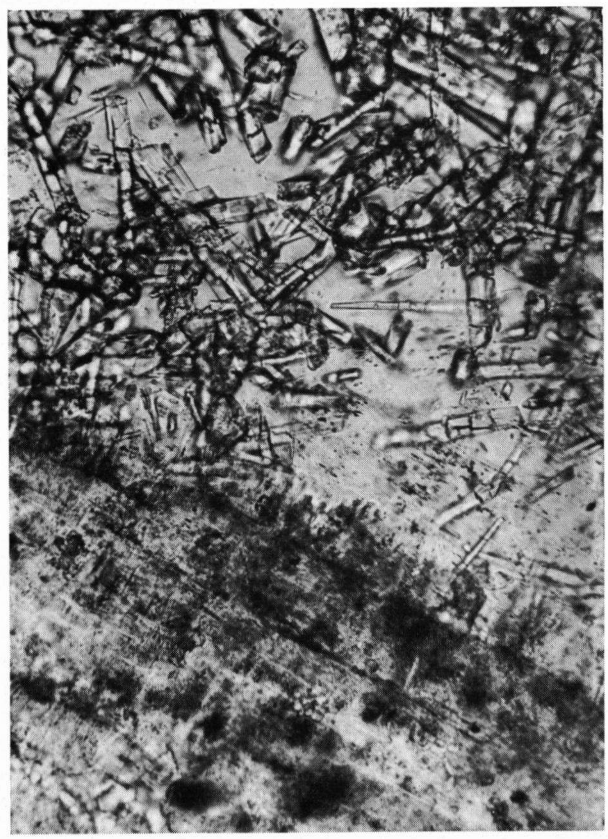

Fig. 61. Dolerite; regeneration of basic felspar giving rise to fresh and glassy felspar (albite-oligoclase) with enclosed epidote. (250 ×)

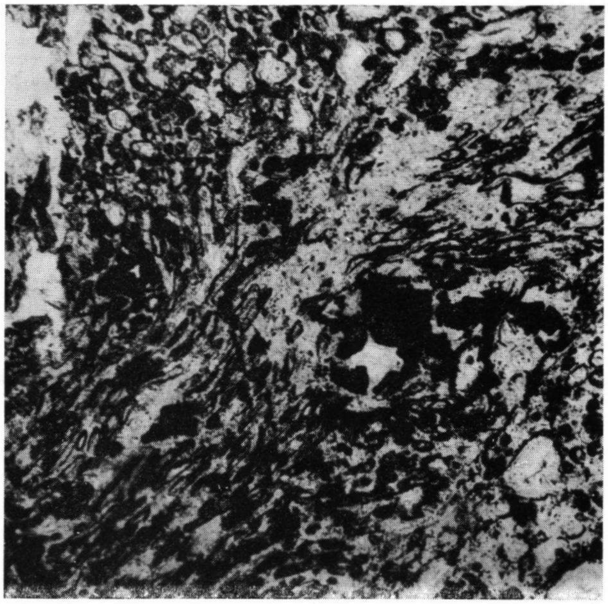

Fig. 62. Pyroclastic rock; flow-structures caused by stretching and a parallel arrangement of leucoxene concentrations and spherulites. (100 ×)

The colour of the tuffaceous sandstones varies strongly with the variations in the composition of the matrix material: they may be green, reddish-brown, or greyish, according to whether their matrix is chlorite, hematite, or leucoxene, respectively.

In the upper Westphalian shales and greywackes of the Maraña-Retuerto basin, intrusive rocks are found frequently near the village of Burón. Their composition varies from quartz-diorite to diorite, the principal mineral constituents being reddish-brown pleochroic hornblende, biotite, prehnite, plagioclase, quartz, and apatite. The hornblende is hypidiomorphic: twinning occurs frequently. The hornblende often altered to a pale-green chlorite (penninite), showing an abnormal blue interference colour. The chlorite occurs as an amorphous mass, as small curved lamellae, or as sherulitic aggregates. The biotite proved to be strongly subordinate to hornblende. The plagioclase is andesine (37 % An) and shows strong sericitization. In addition to Carlsbad twinning, twin lamellae of the albite type are much in evidence. Zoned plagioclase has been observed, especially in the more acid type of rocks.
The rock shows a medium to coarse texture, and closely resembles the late Hercynian intrusions described from other parts of the Cantabrian mountains (de Sitter & Boschma, 1966), where they are situated close to the León line on its northern side. Our intrusions are also situated not far from an important structural line, i.e. the Ventaniella fault zone-Cardaño line.

Epithermal ore deposits occur throughout the whole region in various sedimentary rocks of Carboniferous age. Stibnite is found in the Upper Westphalian shales, greywackes, conglomerates, and limestones of the Maraña-Retuerto basin. The antimony sulphide is beautifully crystallized, and frequently shows a yellowish or white colour due to strong oxidation.
Near the village of Burón, the stibnite occurs together with the intrusive rocks treated above, which suggests an affiliation of the stibnite with the intrusive phase. In most cases, however, a relationship with deeper, intrusive bodies cannot be established. The picture resembles that of the world's principal resources of antimony (south and southwestern China), where in general the ore is also associated with quartz-diorites and occurs in sedimentary rocks (Juan, 1946). In the "Valle de Pedrolla", east of the village of Polvoredo, a quartzite conglomerate is the host rock, a situation identical to the occurrence of antimony northwest of the village of Maraña. The ore occurs as veins, irregular veinlets, or lenticular bodies. The stibnite is usually curved around the quartz pebbles, but occasionally even the fissures in the quartzite pebbles are filled with stibnite.
A short distance west of the village of Maraña, stibnite is found in the Upper Westphalian limestones and occasionally in the surrounding shales. Arsenopyrite and pyrite occur in small quantities in the limestone.

Important cinnabar deposits are known for two localities in the area under consideration, i.e. south of the Puerto de Tarna and east of the village of Lois. In both cases the Caliza de Montaña acted as host rock. The cinnabar occurs as veins and irregular pockets. Subordinate amounts of stibnite and fluorite are frequently present, and pyrite is widely distributed throughout the limestone. Malachite and azurite occur, most probably as alteration products of chalcopyrite. Gangue minerals include quartz and calcite.
No Carboniferous intrusive rocks have been mapped in the vicinity of these ore deposits.

At one locality, just west of the village of Burón, fluorite is found in large amounts as veins and irregular pockets in the Caliza de Montaña. It occurs together with quartz and calcite veins and as a replacement deposit forming fine diffuse pockets in the limestone. The fluorite may be colourless, but is usually violet, blue, green, or white. Excellent idiomorphic cubic and octahedral crystals are easily found. The fluorite is in all probability related to the quartz-diorite that occurs not far from this locality. A fault borders the limestone in the north and west.

Talc is mined west of the village of Puebla de Lillo, Caliza de Montaña being the host rock. This hydrated magnesium silicate generally occurs as a white, grey, or pale-green, soft compact mass. Excellent idiomorphic pyrite crystals, cubes, and pentagonal dodecahedrons, are abundant.
The host rock at this locality is dolomitized to a

striking degree in a zone bordering the Felechosa thrust plane and the Cofiñal wrench fault. According to Lindgren (1933), the purest talc deposits are associated with metamorphosed limestones or dolomites, the talc being a ".... direct derivation from dolomite under influence of siliceous magmatic emanations ....". Intrusive rocks do not crop out in the vicinity of this important fault zone. At one locality, a small abandoned mine at Puerto de Tarna, talc is found together with stibnite, cinnabar, fluorite, malachite, and azurite.

Throughout the area, manganese occurs in varying amounts in the uppermost part of the Caliza de Montaña Formation. From the foregoing, we know that a period of slow sedimentation and non-deposition set in after the sedimentation of (part of) the Caliza de Montaña Formation. There is undoubtedly a relationship between the thickness of this Caliza de Montaña and the manganese content of its top layer (see page 106). In the Fontasguera subarea, where the limestone has its minimal thickness, the manganese enrichment of this level was so strong that there is sufficient manganese ore to warrant commercial exploitation. X-ray determinations have shown that the manganese ore consists mainly of pyrolusite with small quantities of todorokite (?). Wad, the typical amorphous oxidation product, is abundant.

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PLATES

PLATE I

Figs. 1—12 *Aljutovella elongata* Raus. et Saf., 1951 subsp. *lazarensis* subsp. nov.

Axial sections

Fig. 1 — specimen 7

Fig. 2 — specimen 5

Fig. 3 — specimen 4

Fig. 4 — specimen 9 — (holotype)

Fig. 5 — specimen 17

Fig. 6 — specimen 15

Fig. 7 — specimen 10

Fig. 8 — specimen 16

Sagittal sections

Fig. 9 — specimen 61

Fig. 10 — specimen 62

Fig. 11 — specimen 74

Fig. 12 — specimen 67

(30 ×)

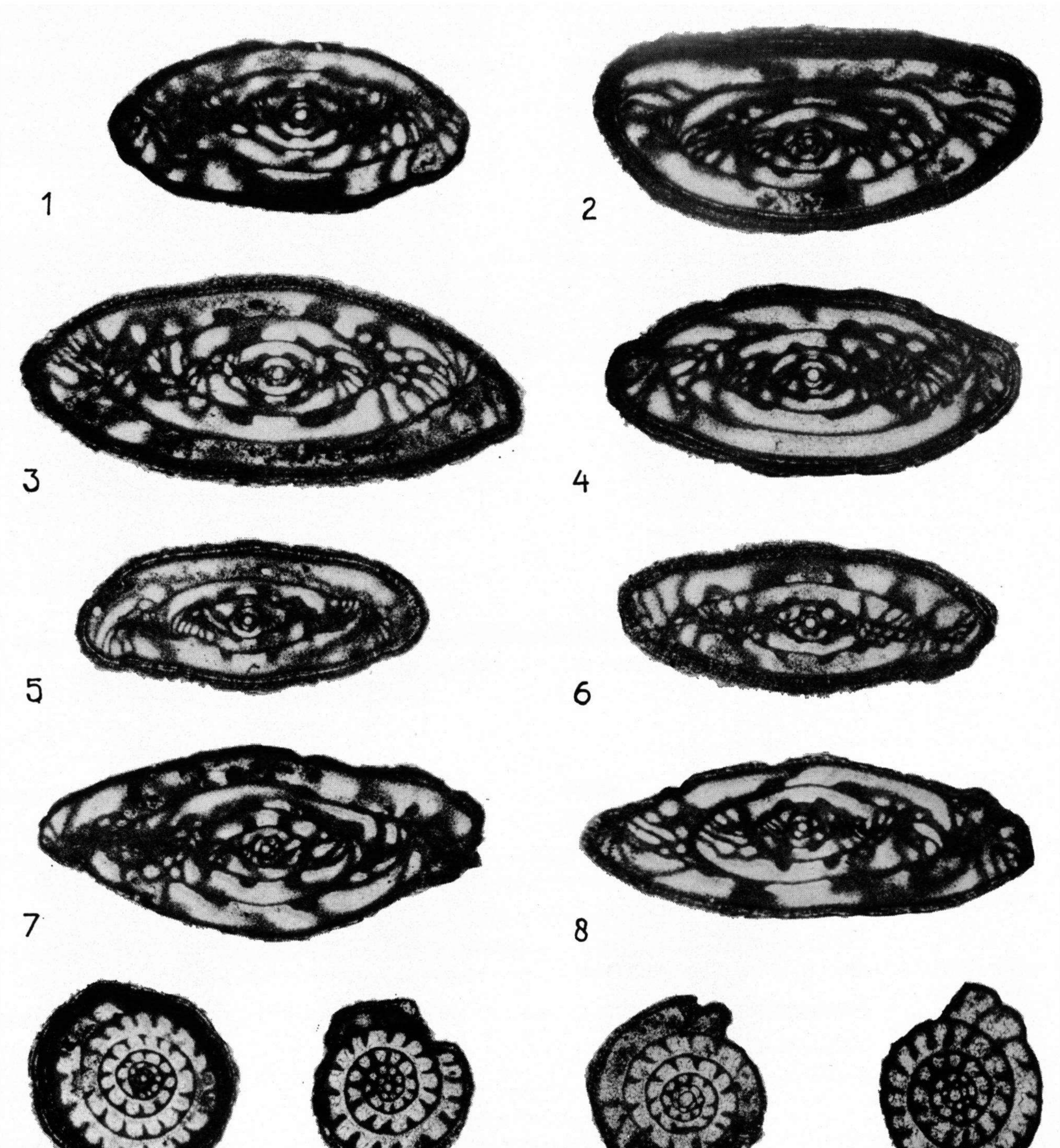
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PLATE II

Figs. 1—8 *Fusulinella pandae* van Ginkel, 1965

Axial sections

Fig. 1 — specimen 1

Fig. 2 — specimen 2

Fig. 3 — specimen 3

Fig. 4 — specimen 10

Fig. 5 — specimen 4

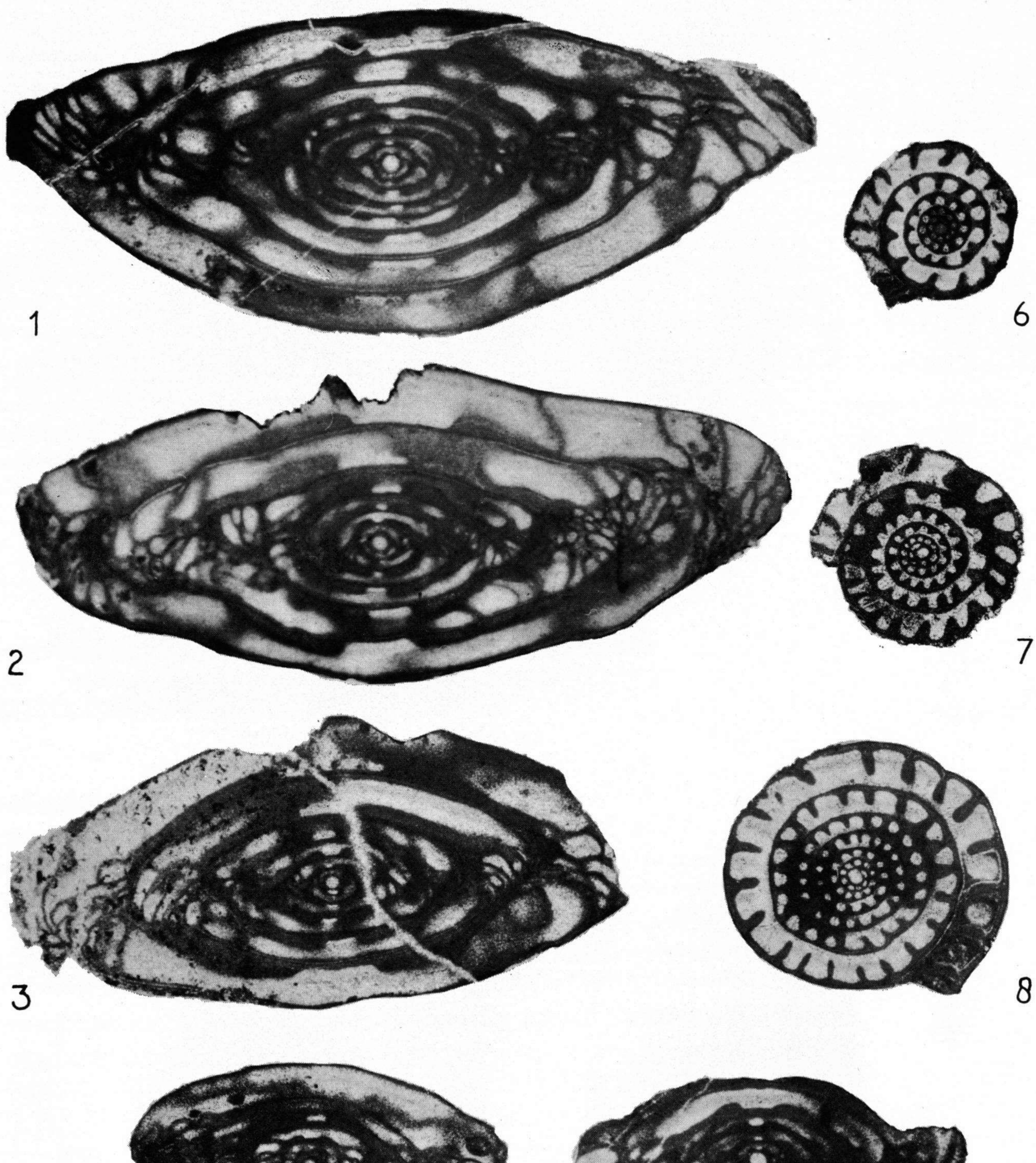
Sagittal sections

Fig. 6 — specimen 67

Fig. 7 — specimen 60

Fig. 8 — specimen 51

(30 ×)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