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THE TECTONIC AND STRATIGRAPHIC HISTORY OF THE CARDANO AREA, CANTABRIAN MOUNTAINS, NORTHWEST SPAIN

BY

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CONTENTS

ABSTRACT

The Devonian and Carboniferous rock-sequence in the Cantabrian Mountains is developed in two different facies which are separated by an E-W tectonic line, the León Line, and called the Leonide and Palentian facies, respectively to the S and N.

The Leonide facies is widely exposed throughout the Cantabrian Mountains and its tectonic history is now well known. The most complete sequence of the Palentian facies occurs in the present area, which lies across the boundaries of the provinces of Palencia, León and Santander. Towards the west and north the Devonian and Lower Carboniferous gradually plunge below the younger strata. The Palentian facies in this region is only found north of a second fundamental tectonic feature, the SE-NW Cardaño Line. This line joins the León Line in the SE of the area near the village of Santibañez but can be traced far to the NW into the province of Oviedo. Immediately south of the Cardaño Line there is the Siero basin filled with Upper Carboniferous clastics which even overlap in places the León Line further south. The pre-Westphalian rocks of the present area are limited to the north by the Peña Prieta Line, the fundamental nature of which is emphasised by a large granite intrusion. To the east the Polentinos fault separates the present area from the mainly Westphalian, Pisuerga basin. The fundamental lines are partly expressed at the surface as large fault systems.

It is remarkable that in the Palentian facies sedimentation was not interrupted by the major erosional periods known from the Leonide facies. The post-Silurian and pre-Westphalian sequence of the Palentian facies is on average about 850 m thick and consists mainly of shales with thin limestone intervals. The corresponding sequence of the Leonide facies is on average about 1750 m thick and shows important, thick Devonian reef limestone intervals. Only the lower and uppermost of the Devonian contain clearly higher-energy deposits (sandstone sequences). Both the lithology and fossil association confirm that the Palentian facies sequences developed in a more offshore (middle to outer neritic) environment of deposition than indicated for the Leonide facies (littoral to inner neritic),

Important epeirogenetic movements in the late Namurian and in the Westphalian disturbed the preceding long period of quiet sedimentation. Between the Cardaño Line to the south, the Peña Prieta Line to the north and the Polentinos fault to the east the Cardaño block was subjected to a regional tilting during the deposition of the rocks of the Yuso Group. The maximum uplift and erosion are indicated in the N and E whereas the deepest subsidence and maximum deposition have been detected in the S and W of the Cardaño block.

The pre-Westphalian rocks, where unconformably overlain by those of the Yuso Group, show local, gentle, pre- or synsedimentary folded structures. The present study has not revealed pre-Westphalian structures that would justify the use of the term orogenic phase (i.e. Sudetic) for their origin. They are rather interpreted as the results of local compression accompanying pre- and synsedimentary epeirogenetic movements.

The unconformable Yuso Group consists of a conglomerate facies — the Curavacas Formation — up to about 700 m thick and a sandstone-shale facies — the Lechada Formation — of at least 750 m N of Cardaño de Arriba (probably up to 2000 m to the W). The Westphalian rocks were deformed during the main compressive phase, which therefore is thought to correspond with the Asturian folding phase (pre-Stephanian).

The tectonic transport here was from north to south in contrast to the Leonides where it was from south to north. This correlates with the theory that the folding of the two areas took place at different times; Asturian in the Asturides (Palentian facies) and Sudetic in the Leonides.

The inhomogeneity of the Palentian facies rock sequence is reflected in the very complicated final tectonic picture. The Cardaño Area can be subdivided into 4 subareas (Northern, Central, Southern and Arauz), in each of which a different lithofacies is related with a corres-

ponding minor tectofacies. Simultaneous cross folding can be related to the rapid facies changes in the affected rocks.

The present area gives very instructive examples of the close interaction of tectonics and sedimentation. Epeirogenetic movements between fundamental tectonic lines controlled the deposition of the sedimentary sequence. These heterogeneous rocks were then acted upon by a relatively short compressive tectonic phase which created out of them the present architecture of the Cardaño Area.

INTRODUCTION

This paper describes the geology of an area of Upper Palaeozoic rocks on the southern flank of the Cantabrian Mountains of N. Spain. The work has formed part of a programme of geological mapping being carried out by students of the Geological Institute of Leiden under the direction of Prof. Dr. L. U. de Sitter. The author's fieldwork was carried out during the summers of the years 1959—1962, completed with a few weeks fieldwork during the summer of 1964.

The area lies about 150 km N of Valladolid, across the boundary between the provinces of León and Palencia. Only a very small part of the about 450 square kilometer area has been cultivated into fields and alps. The ground ranges between 1000 and 2500 m, the highest parts carrying almost the whole year remmants of snow cover.

The geology of adjacent areas has been described by Savage (1961) to the west and by Koopmans (1962), Kanis (1959) and Frets (1965) to the south, while a stratigraphic section to the NE has been described and dated by Kullmann (1960).

Investigations described in internal reports have been carried out by J. A. van Hoeflaken (1958) and P. A. C. de Ruiter (1963) in the E, by W. F. E. Michel (1961) to the SW, by P. Kamerling (1962) to the N. and H. Teer (1964) to the W. These internal reports are deposited at the Geological Institute of Leiden.

I am indebted to several specialists for fossil determinations; Mr. H. A. van Adrichem Boogaert (Leiden), Dr. J. Kullmann (Tübingen), Prof. F. Stockmans (Brussels), who are credited in the text. Their determinations gave important support because a comparable lithologic sequence to that found here had been described only partly before in the Cantabrian Mountains.

The progress and the results of this study have been greatly assisted by the discussions of all problems among the team of geologists in Leiden. I would especially like to thank Prof. L. U. de Sitter, Mr. J. F. Savage M.Sc., Dr. D. Boschma and Dr. A. Breimer. Also I want to thank here for their very willing and skillful technical help I received: Miss T. W. Terpstra (typing), Miss C. Roest and Mr. I. Sánta (for drawings), Mr. J. Hoogendoorn (for photographs), Mr. C. J. van Leeuwen (for thin sections) and Mr. J. Schipper (for laboratory work).

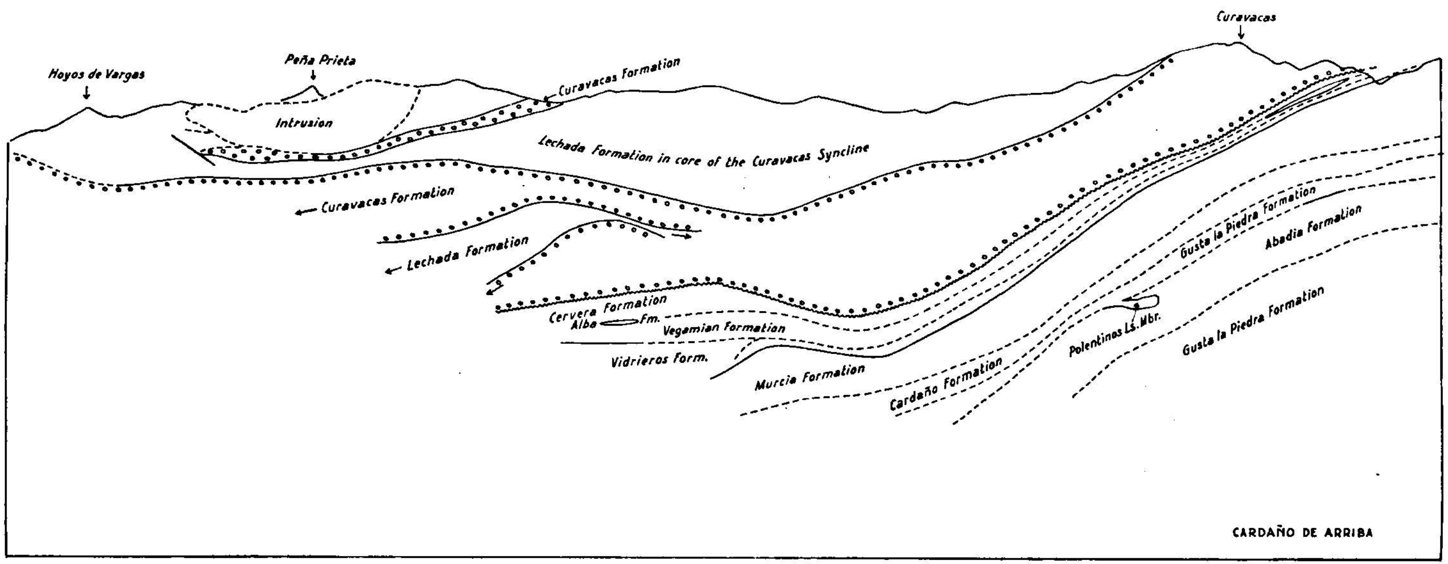

Fig. 1. View looking north from Cardaño de Arriba across the Murcia Anticlinorium and the Curavacas Syncline.

Part I

STRATIGRAPHY

INTRODUCTION

The area is underlain by a more or less complete sequence of Devonian and Carboniferous rocks which have been subjected to intense deformation (isoclinal folding, thrusting, cleaving, low-grade metamorphism).

De Sitter (1961, 1963) distinguishes the Leonide and the Asturide facies bordering one another with a straight E-W running contact which he called the León Line. The León Line (Enclosure 2, Index Map) as given by de Sitter runs more or less between the present area and the area to the S studied by Koopmans (1962). The strikingly different lithosequences of the two facies areas in all but the lowest Devonian and especially in the Carboniferous demonstrates that the León Line was an important facies boundary during those times of deposition. Moreover it became clear that each facies area is characterised by its own type of deformation. Thus the León line separates also two tecto-facies areas called the Leonide and Asturide facies areas (de Sitter 1962, b). Brouwer (1962) proposed alternative terms respectively Asturo-Leonian and Palencian facies. He proposed to substitute the term Asturo-Leonian for Leonide because of the very probable presence of the same facies in Asturias. This facies, is however, fully represented and defined in the province of León and for simplicity's sake we still prefer to stick to the original term Leonide. The term Palentian is very useful because it refers well to the very different northern facies, which is found in its most complete and typical development in the province of Palencia.

In the present area a line comparable with the León Line is expressed which runs in a ESE-WNW direction and meets the León Line in the SE of the present area. This facies boundary marked by the belt of thick Carboniferous limestones in the southern part of the present area has been named Cardaño Line, since it passes between the two villages Cardaño de Arriba and Cardaño de Abajo.

The Cardaño Line is clearly a fundamental feature in the sense of de Sitter (1964) because it can be traced for to the NW perhaps as far as Oviedo. Consequently it may be compared to the León Line and by inference thought to have a similar deep-seated origin. The role of this structure has been by no means simple, varying considerably in time and also along the strike. As with the León Line the Cardaño Line cannot be traced as a sharp linear feature on the ground. The surface expression of the activity of these deep-seated structures often seems to have affected a wide zone but due to the rapidly changing nature of these effects terms such as "ridge" and "zone" tend to be misleading. The term line is meant to refer to the underlying linear fundamental structure which it is not felt should necessarily feature as a single line in the complicated suprastructures.

The geological map (Enclosure 1) shows that a large part of the present area is covered by Carboniferous rocks, the Devonian gradually disappearing below them to the W. The full sequence of Devonian of the Rio Carrion in the E is gradually lost, plunging to the W so that in the most western part only uppermost Devonian rocks are exposed. To the E this area is separated from the important Carboniferous developments of the Cervera Basin by the NNW-trending Polentinos Fault.

The rocks have only incidentally yielded any abundant faunas. Useful results have been obtained from identification of conodonts, brachiopods, goniatites, trilobites and lamellibranchs. Gastropods, crinoids, bryozoa, fusulinids and solitary corals have not yet proved useful for the biostratigraphy of the present area.

The difference of facies north and south of the León Line is also expressed by the fossil fauna development. The type fauna of brachiopods and corals published by Koopmans, 1962 (brachiopods, corals) from the Leonide facies has not been found in the Palentian facies.

Good floras have been collected from shaly intercalations in the Curavacas conglomerates. Stockmans (1965) determined a Westphalian A age for a collection from the very basal part of the Curavacas conglomerates north of the village of Cardaño de Arriba.

1. STRATIGRAPHY OF THE DEVONIAN

Carazo Formation

The Carazo Formation typically developed on the Carazo Mountain is composed largely of interbedded sandstones and shales with two massive bands of quartzite in the middle which allowed a threefold subdivision, Enclosure 1 Section Ia. The lower boundary of this formation has not yet been observed, the uppermost layers become quite calcareous and clearly presage a passage into the overlying Lebanza Formation limestones. The lower part of the Carazo Formation has been dated through microfauna as uppermost Silurian by Cramer (1964). The upper part of the sequence has been dated on macrofauna (brachiopods) as Lower Gedinnian by Binnenkamp (1965). The maximum thickness observed is 320 m.

The basal Member a of the Carazo Formation consists of alternating sandstone and shale beds, which are thin bedded and significantly highly micaceous. The sandstones are mostly brownish and middle grey and sometimes greenish. They are fine-grained, argillaceous and contain very often worm burrows. The sandstones are ferruginous and at some intervals approach an ironstone. Bottom structures such as load casts, groove casts and occasionally flute casts have observed. The shales are darkgrey to blackish, weathering brownish and grey and predominantly develop in the basal part of the formation. This shaly basal part reflects presumably the latest Silurian shale deposition cycle. The maximum thickness observed is 65 m.

The middle part of the sequence which we call Member b of the Carazo Formation is composed of thick quartzitic sandstone sequence with interbedded sandstone shale sequence between them. On Carazo Mountain, for example, we find from base to top respectively 60 m quartzitic sandstone, 35 m interbedded sandstone-shale and 90 m quartzitic sandstones. The thickness and number of the quartzitic sandstones however are very variable; towards the east (Peñas Negras) we find several quartzites only 10—20 m thick, whereas to the north and northwest only one quartzitic horizon of about 12 m thick has been found. The quartzitic sandstones are yellowish brown weathering rocks, rather well sorted fine to coarse grained, average to thick bedded and locally ferruginous. Clay pebbles occur frequently and thin shale interbeds can be found between the quartzitic sandstones. The fauna in this member is very poor and only a few brachiopods, trilobites and gastropods have been collected, but not yet determined.

The upper Member c contains darkgray, yellowishbrown weathering, well stratified, thin to medium bedded fine grained sandstones and brown weathering, splintery, silty shales. This part is much less micaceous than the lower sequence, is predominantly shaly and is moreover characterized by a rich fauna of brachiopods, lamellibranchs, ostracods, trilobites, pteropods (tentaculites). These criteria made the two sequences easily distinguishable from each other.

The sequence becomes clearly calcareous in its upper part and forms a transitional sediment towards the overlying limestones of the Lebanza Formation. Like Member c, this part of the Carazo Formation decreases considerably in thickness towards the north and northwest. Its maximum thickness is about 70 m, its minimum thickness about 40 m. Reviewing we note that the Carazo Formation forms a sedimentary cycle; it starts with a shaly sequence, grades into quartzitic sandstones and terminates again with a shale series.

Lebanza Formation

The Carazo Formation is conformably overlain by a limestone sequence which is known as the Lebanza Formation. The sequence has been mentioned by Mallada (1885) in his paper on a fauna collected by Casiano de Prado. Alvaredo and Sampelayo (1945) introduced in the name Lebanza after the village of Lebanza. The thickness is variable between 20—100 m according to Binnekamp (1965), who states that the age of the Lebanza Formation ranges from Upper Gedinnian to Middle Siegenian.

The Lebanza Formation exposed in the River Arauz (see Enclosure 2, section Ib), consists of light grey weathering, medium grey, well bedded, mainly thin to average bedded, limestones with an abundant fauna of brachiopods, crinoids, pteropods, goniatites, trilobites, corals, bryozoans, stromatopores, etc. The best developed sequence of the formation consists of four intervals. It is then composed, from base to top, of thin to average bedded, very argillaceous limestones with shale interbeds; thin bedded to average bedded limestones; thick bedded to very thick bedded limestones which again are followed by thin to average bedded limestones. Crossbedding and minor slumping is commonly found in the basal interval and occasionally in the second interval. The thicker bedded third interval is clearly less fossiliferous than the other three intervals. The Lebanza Formation forms a distinctive and extensive stratigraphic marker-horizon, although its lateral development is rather variable. Lateral passage of the limestones into shales can be observed. The average thickness of the formation decreases clearly towards the north and the northwest. In these areas the formation is mainly represented by the basal argillaceous limestone-shale interval overlain by thick bedded limestones, resembling lithologically the first and third interval mentioned above. From the second and fourth interval a brachiopod fauna has been collected and determined by Binnekamp (1965). This study points respectively to Upper Gedinnian and Lower Siegenian for the second and Middle Siegenian for the fourth interval. The formation is easily distinguishable from younger limestone intervals by its thickness, abundant macrofauna and lack of shale interbeds in the upper intervals.

Abadia Formation

The Abadia Formation consists of a thick shale sequence in which two limestone members will be distinguished. This formation has been described previously by Binnekamp (1965). The lower limestone member will be indicated as the Requejada Limestone Member and the upper limestone as the Polentinos Limestone Member named respectively after the Requejada Lake and the village of Polentinos. The thickness of the Abadia Formation is 150—220 m.

The best developed sequence of the Abadia Formation is exposed in the Arauz River (see Enclosure 2, Section Ib). It starts with about 80 m darkgrey silty shales being in its basal development locally sandy and in its upper development calcareous with thin argillaceous limestone interbeds. Only a poor fauna of a few trilobites has been collected in this upper part of the shales.

The Requejada Limestone Member which gradually succeeds the lower shale sequence is about 15 m thick It is mostly built up of 3 limestone horizons, very hard and recrystallized, some dolomitized, of about 3 m thick separated by medium grey calcareous shales. The limestones are lightgrey weathering, medium grey, well bedded (5—30 cm) and argillaceous with thin shaly laminations and interbeds which locally may result in a wavy to nodular bedding. These limestones are fairly fossiliferous containing brachiopods, lamellibranchs, goniatites, ostracods, corals, pteropods (tentaculites) etc.

The Requejada Limestone Member is succeeded by a dark ferruginous shale interval of about 30 m thick with a few limestone interbeds of 5—15 cm thick. No fossils have been found in these shales. They are overlain by yellowbrown weathering, lightgrey, argillaceous, poorly sorted and fine to coarse grained sandstones being slightly ferruginous. This sandstone interval is only a few meters thick (maximum 8 m) but it is quite persistent throughout the present area. Upon this sandstone a greybrown shale of 10—30 m thick is found. This shale is fossiliferous, containing mainly trilobites and goniatites. In its upper sequence this shale interval has some argillaceous limestone interbeds which reveal the gradual transition into the overlying Polentinos Limestone Member.

The Polentinos Limestone Member is rich in shale near the base effecting nodular and wavy bedding of the limestones which often has reddish purple weathering colours. This lithological aspect may resemble strikingly that of the Upper Devonian nodular limestones. This basal development however, is not always present. The Polentinos Limestone Member is 10—30 m thick and consists of dark grey to black limestones which weather whitish grey and are very well bedded, platy and argillaceous with thin shale laminations and intercalations. These limestones are poor in fossils and only a few solitary corals and trilobites have been found, making them easily distinguishable from the older Requejada Limestone Member.

The Polentinos Limestone Member west of Cardaño de Arriba is rather strongly recrystallized with a conchoidal fracture and usually deformed by cleavage. In this area this limestone is in contact with ferruginous blackish grey platy and massive slates which are almost hornfels rocks. Contact metamorphism by granite intrusions has given rise to garnet and andalusite crystalloblasts in these slates. The slates are highly resistant to weathering and form a strong positive feature in the terrain, frequently without any vegetation cover. The location of these slates with the Polentinos Limestones in a structurally very complicated and compressed zone make their stratigraphic relationship very difficult to unravel; from west to east the Polentinos Limestones disappear and only dark slates are found at their stratigraphic level. Small sandstone outcrops, however, at the base of the slates can be correlated with the sandstone sequence of the Abadia Formation. For these reasons we prefer to explain the disappearance of the Polentinos limestones by tectonic cutout and not by facies change into slates; the slates are therefore the equivalent of the Abadia shales below the Polentinos Limestone Member.

The age of the Abadia Formation is not clearly established. Only the brachiopod fauna of the Requejada Limestone Member has been dated and reveals an Emsian age (Binnenkamp 1965). The formation is overlain by the Gustalapiedra Formation for which a Givetian age has been stated see p. 56; for the time being the Abadia Formation is thought to have been deposited during Emsian times only.

Gustalapiedra formation

The Gustalapiedra Formation conformably overlies the Abadia Formation. It consists of a sequence of 50—70 m thickness of alternating dark grey and black slates and argillaceous black limestones. The Gustalapiedra Formation passes gradually into a nodular limestone sequence (Cardaño Formation) and represents uninterrupted sedimentation through Upper Givetian time. Very good exposures occur in the valley of the Gustalapiedra River, west of the Murcia Mountain, after which the formation has been named.

The type section of the Gustalapiedra Formation is shown in Enclosure 2, Section Ic. The black slates and the argillaceous material in the limestones are strongly recrystallized; chloritoid and sericite are abundant, the former showing striking crystalloblastic textures. The thicker beds of slate tend to have more strongly developed crystalloblastic textures whereas in thinner bands this texture becomes less marked and irregular shaly parting is developed.

Fig. 2. 'Diapiric-like' intrusions of slates into limestones, belonging to the Gustalapiedra Formation.

The black limestone is medium crystalline and pyritic. It usually includes thin lamellae of hard shale, more resistant to weathering, which have complicated patterns due to tectonic deformation. The limestone often shows coarse fracture cleavage which is sometimes associated with 'diapiric-like' intrusions of slate into the limestone (fig. 2).

Occasionally some thin lenticular sandstone beds can be found interbedded with the shales and limestones.

The sequence shows a distinct but gradual lateral variation in lithology. Around Cardaño de Arriba the limestone beds may be up to 1.5 m thick including the shale lamellae; the overall shale-limestone ratio of the sequence is about 2 : 1. To the W and to the E, however, shale deposits largely prevail over the limestone deposits. The latter range up to only 30 cm, showing very fine shale lamellae, whereas the shale-limestone ratio is about 10 : 1 or even higher (E of the River Carrion).

The slates have yielded a specialized fauna containing many examples of only a few trilobites, goniatites, lamellibranchs and gastropods. The goniatites determined from the type section include specimens of *Agoniatites* cf. *costulatus* (d'Archiac & de Verneuil 1842), *Agoniatites* cf. *vanuxemi* (Hall, 1879) and *Tornoceras* spec. A locality north of Aguasalio Mountain yielded *Agoniatites* cf. *costulatus* (d'Archiac & de Ver-

Fig. 3. Natural exposure of nodular limestones belonging to the Cardaño Formation.

neuil 1842), *Cyrtobactrites* spec.? and *Subanarcestes* spec. These goniatites indicate an Upper Givetian age (Dr. J. Kullmann, Tübingen, written communication). Preliminary determinations of the trilobites merely exclude the possibility of a Lower Devonian age. (Dr. W. Struve and Prof. Erben, verbal communications).

The limestones have yielded a poor fauna of conodonts which point to Upper Givetian and perhaps Lower Frasnian age (van Adrichem Boogaert 1965.)

Cardaño Formation

Very representative outcrops of the Cardaño Formation are found a few hundred meters N of the village of Cardaño de Arriba after which the formation has been named. The Cardaño Formation is commonly formed by medium grey to beige nodular limestones interbedded with darker coloured shale beds. Differential weathering (fig. 3 & 4) gives a very typical appearance in the natural exposure. These rocks rest conformably upon the Gustalapiedra Formation, persist for only 25—40 m and then pass upwards into the overlying Murcia Formation. The Cardaño Formation has yielded a rich conodont fauna of Middle and mainly Upper Frasnian age. No evidence of an hiatus has been found.

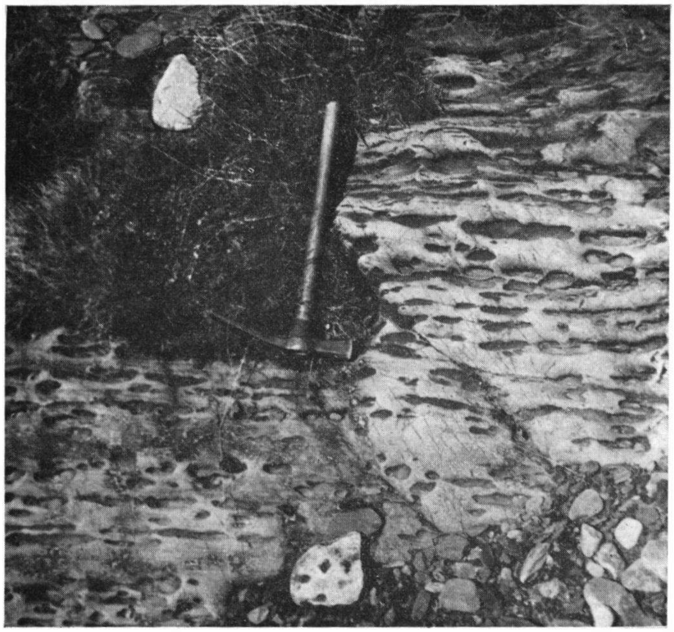

Fig. 4. Typical appearance of differentially weathered nodular limestones of the Cardaño Formation.

The section in the South ridge of Murcia Mountain is shown in Enclosure 2, Section Id. The limestone is fine grained, but gives a coarser textured weathered patina than the shaly material which is strongly recrystallized showing again abundant chloritoid porphyroblasts. The limestone is strongly recrystallized and has a conchoidal fracture. The limestone is finer grained than the Polentinos and the Gustalapiedra limestones. Generally we find well developed micro-structures in the shaly alternations which curve round the limestone nodules and are wrinkled into microfolds often developing into fracture cleavage. The proportions of the shale and limestone vary considerably. The most typical extreme (shale prevailing largely over limestone) shows isolated strings of nodules (few cm diameter) set apart in the shale (fig. 3). Flattening of the limestone nodules parallel to the cleavage planes is obvious in many exposures. No examples could be found which exhibit a preferred direction of elongation in the plane of flattening. The extreme ratio, where limestone dominates over shale, show essentially limestone with irregular shaly partings (fig. 5). In the thicker limestone beds with the shaly partings the cleavage-bedding relation is much more noticeable than in the thinner bedded exposures. This results often in a picture comparable to that found in the thicker limestone beds of the Gustalapiedra Formation. The Cardaño limestones, however, are easily distinguishable from the older formation by their brighter colour and its very rapid alternation of shales and limestones.

Lateral variation here is not so distinct as in the older formations, but a decrease of the limestone content to the E can be detected.

The overall limestone-shale ratio approximates to 1 : 1.

Fig. 5. Limestone with irregular shaly partings belonging to the Cardaño Formation of nodular limestone.

The abundant conodont fauna has been determined as of Middle and mainly Upper Frasnian age (van Adrichem Boogaert, 1965) Poorly preserved goniatites and shell fragments constitute the only other observed faunal elements. The goniatite fauna includes specimens of *Manticoceras* and *Ponticoceras*, which permit the conclusion of a Frasnian age (upper Manticoceras Stage), (written communication, Dr. J. Kullmann). A few thin stems of wood have been found. They are 5—10 cm long and 0.5—1 cm broad and are sculptured with scars like *Lepidodendron* stems.

Murcia Quartzite Formation

The Murcia Quartzite Formation has been named after the impressive Murcia Mountain that is formed by this quartzite.

The Murcia Quartzite Formation consists of dark-coloured, well bedded, quartzitic sandstones to quartzites alternating with thinner dark-coloured shales. The formation grades conformably out of the uppermost beds of the Cardaño Formation and above passes into the overlying Vidrieros Formation of Middle and Upper Famennian age. Hence the Murcia Quartzite represents a continuous sequence from the highest part of the Frasnian into the Middle Famennian. A thickness of 60—200 m of sediments is assumed.

The Murcia Quartzite Formation as shown in the type section illustrated in Enclosure 2, Section Id, is usually found as a dark-grey, well-sorted, fine to medium grained, quartzitic sandstone (fig. 6), locally light grey or brownish, and often as a real quartzite. The average thickness of individual beds is about 50 cm; thickness above 150 cm are exceptional. The quartzites in the Southern Area between the limestones along the Cardaño Line are markedly thinner and lighter coloured. They are light gray to white well-rounded orthoquartzites which resemble very much the Ordovician Barrios Quartzites of the Leonide facies.

Fig. 6. Quarzitic sandstone belonging to the Murcia Quarzite Formation, with specimen of *Buchiola* (× 10).

The interbedded slates are dark gray to black in colour and often have very thin interbeds of quartzitic sandstone. Sometimes the slates are strongly folded and cleaved in beautiful minor disharmonic structures that do not affect the thicker beds of quartzitic sandstone. Some rather badly preserved lamellibranchs have been found, but these have not been determined yet.

The basal part of the Murcia Quartzite Formation (indicated on the Geological Map) shows predominantly very thin beds of alternating light gray, fine grained, quartzitic sandstone and dark grey shales. Crossbedding and load casts are common features; wormtracks also occur. Sometimes graded bedding is found (fig. 7). The basal part is moreover characterized by the abundant occurrence of the small, finely ornamented lamellibranch *Buchiola* (fig. 6). Preliminary determinations of these lamellibranchs as *Buchiola palmata* and *B. angulifera*, suggest an Upper Frasnian age. Lamellibranchs of *Mytilus* and *Posidonia* forms have also been found.

The thin sandstone interbeds have an average thickness of under 5 mm and are often lenticular with small scale crossbedding. The alternating shales which are usually recrystallized into slate have an average thickness of 10 mm. Locally quartzitic sandstone intervals of the normal Murcia type can be found. This basal development becomes much thicker towards the E where increased interbeds of the normal thick Murcia Quartzites have also been found. This basal development is indicated on the map by a double stripe symbol.

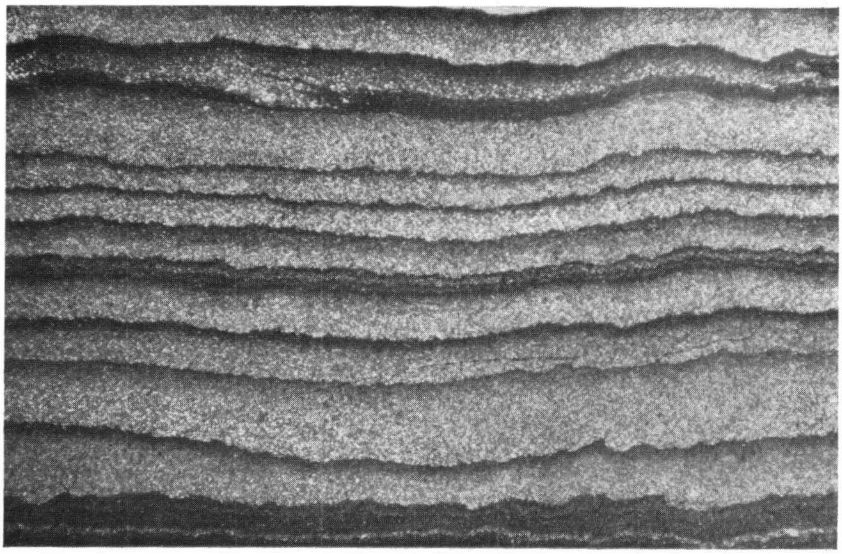

Fig. 7. Graded bedding in sandstones of Murcia Quarzite Formation (× 10).

In the eastern part of the area (around the village of Vidrieros) the average Murcia beds are markedly coarser. Also the thickness increases in this direction to about 200 m. A distinct grading upwards from microconglomeratic sandstone into finegrained sandstone has been found in several outcrops. Slumping phenomena and bottom structures (groove casts especially) occur rather frequently. Load casts occur in all outcrops of the Murcia Formation.

Vidrieros Formation

The Vidrieros Formation is strikingly similar to the Cardaño Formation in the field. It consists of thin to medium bedded limestone lenses or thinbedded nodular limestone beds both interbedded with shales. The Vidrieros Formation follows gradually upon the underlying Murcia Quartzite Formation, the upper part of which consists of thin quartzites and dark gray shales. The formation which is on average 20 m thick seems to pass gradually in the overlying black shale sequence of the Vegamián Formation. Excellent faunal evidence dates the rocks of the formation as Middle and Upper Famennian. The formation is very well exposed around the village of Vidrieros after which it has been named.

The Vidrieros Formation is well exposed in the north ridge of Murcia Mountain and is illustrated by the measured sequence in Enclosure 2, Section Id. The limestone is mostly finely recrystallized, varicoloured; light gray to dark gray, light greenish and purple reddish colours are shown. A medium gray limestone which weathers light gray is usually found. Fracture cleavage is strongly developed here. In hand specimen the cleavage can be seen to cut through both limestone and slate although there are less cleavage planes visible in the limestone and the angle between the cleavage and bedding is usually smaller in the limestone than in the shale. As in the Caradaño Formation the field appearance of the formation is mainly controlled by the strong tectonic deformation (c.f. fig. 30). Pyrite crystals up to 1 cm diameter can be found in the limestone.

The interbedded shales or slates also are varicoloured; white gray to dark gray, purple reddish and black colours have been observed. Especially the purple reddish exposures can easily be mistaken for the Visean Alba Griotte. Often we find only a shale with hollows from which the limestone has been dissolved.

A light gray to yellow brownish colour is mostly seen on the weathered surfaces. Locally the shales predominant largely over the limestone; north of Ojeda Mountain the first 10 m of the sequence of the Vidrieros Formation consists almost exclusively of dark gray to black shales while above it many thin limestone beds of an average thickness of 10 cm are intercalated in the remaining 12 m of this formation. Very often the shales are converted into real chloritoid schists.

Trilobites, goniatites and lamellibranchs are rather rare and poorly preserved in the slates due to tectonization. Conodants and goniatites have been found in the limestone. The conodonts are generally abundant and indicate a Middle and Upper Famennian age (van Adrichem Boogaert, 1965). Goniatites are locally very numerous, specimens of *Cheiloceras* (?) and *Sporadoceras* have been identified which point to a Famennian age (Dr. J. Kullmann, written communication).

Review of Devonian Stratigraphy

Strong deformation of the strata is found in the present area. Isoclinal folding thrust faults, cleavages and partial metamorphism make it very difficult to define the stratigraphic sequence of the Devonian in detail. The fossils often show some deformation, but the fossil fauna has not been completely destroyed by the deformation of the area.

The sequence consists predominantly of finegrained clastics which reflect that the influence of the coastal area was not very strong during the deposition. Only the oldest and youngest Devonian show important sand deposits (Carazo and Murcia Formations).

The Devonian is significantly thinner than that in the Leonide facies south of the León Line. From the top of the Carazo Formation to the youngest Devonian about 500 m and respectively 1000 m section is found moreover an important uppermost Devonian hiatus is detected in the Leonide section.

The Carazo and the Lebanza Formations are rather well comparable with the San Pedro and the La Vid Formations of the Leonide facies. The younger beds, however, are strikingly different, showing a much more shaly development in the Palentian facies. Also the faunal associations are different. A deeper water fauna of conodonts, goniatites and trilobites is found in the Palentian facies whereas a more shallow environment is indicated by the coelenterates, stromatoporoids and brachiopods of the Leonide facies.

The Palentian and the Leonide facies can be compared respectively with the Bohemian and the Rhenian facies (Brouwer 1962).

In the eastern part of the area the Lower Devonian becomes thinner, and more argillaceous to the north which reflects the deepening of the Lower Devonian basin in that direction. The Upper Devonian Murcia Quartzite becomes thicker and older to the NW (Montó area) whereas also the shaly interbeds disappear. These data are still too scant to draw a valuable conclusion for the Upper Devonian basin development along the Cardaño Line.

In chapter 3 the Palentian facies is compared with the neighbouring areas.

In general it can be stated that the Devonian sequence of the Palentian facies has been deposited in a quiet and stable shelf environment ($\pm$ 50—150 m) outside the coastal reef belt. Conditions of restricted circulation may be reflected in the dark coloured pyritic sediments (e.g. Gustalapiedra Formation) and their specialized faunas.

2. STRATIGRAPHY OF THE CARBONIFEROUS

Vegamián Formation

The Vegamián Formation has been named from exposures in the Bernesga area by Comte (1959, page 330). The Vegamián Formation consists of dense hard black slates. Both the base and the top of the formation show contacts which presumably suggest uninterrupted sedimentation. The average thickness of the formation is about 30 m. Tournaisian and Lower Visean stages may be represented.

The Vegamián Formation is illustrated by two measured sections in Enclosure 2, Sections IIa and IIb. The basal part of the formation consists of splintery slates, while higher in the sequence more massive and relatively softer beds occur. Pyrite crystals are abundant especially in this higher part of the formation. The weathered surfaces of the rocks are usually reddishbrown (due to the alteration of the Fe min-

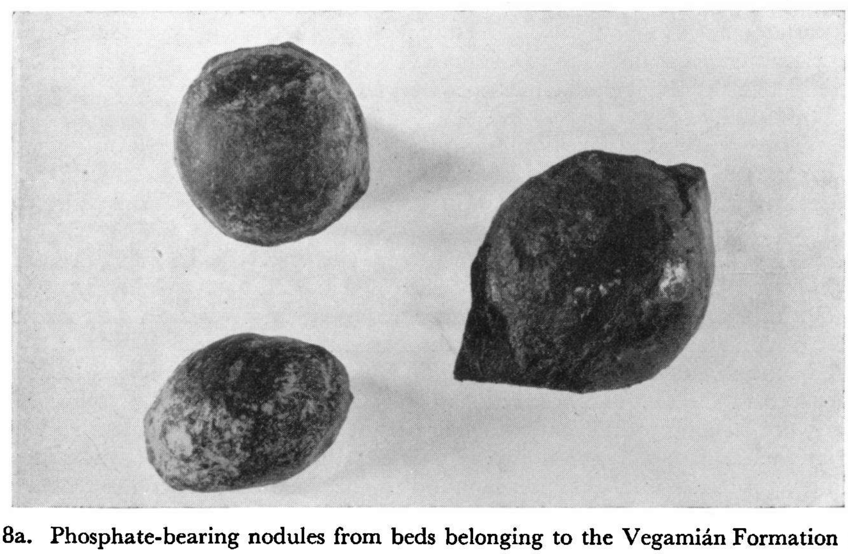

Fig. 8a. Phosphate-bearing nodules from beds belonging to the Vegamián Formation (x 1).

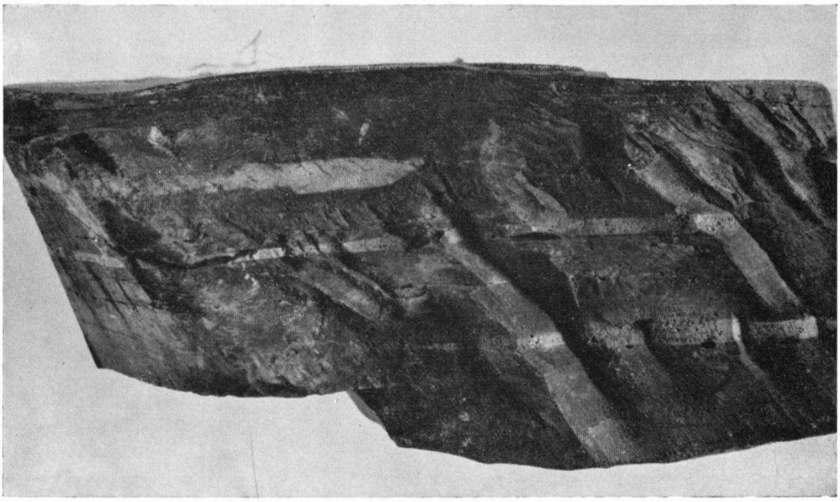

Fig. 8b. Phosphate-bearing, white-weathering, black lenticular laminae from beds belonging to the Vegamián Formation (x 1).

erals) but locally yellow-white spots of sulphate are found on the weathered surface. In these slates phosphate and chert have been found. The phosphate has been found throughout the sequence in very hard bluish-black phosphate bearing nodules of few cm diameter (fig. 8a) and in white weathering, black lenticular laminae of few mm thickness (fig. 8b). The chert is predominantly black coloured and has been found in relatively rare small lenticular occurrences. However 4 km east of Vidrieros and 1 km east of Triollo respectively, 25 m and 1 meter thicknesses of finely laminated greenish fine bedded cherts have been found.

A chemical analysis brought out that the shales consist of about 75 % $Si\,0_2$, together with 0.12 % P_20_5; whereas the nodules may contain about 30 % P_2O_5.

On two levels subrounded illsorted scattered quartzite pebbles (2—10 cm diameter) have been found lensing out laterally within distances of 1.5 m and 10 m respectively. They are presumably slumped in from the nearly source area.

Thin black very argillaceous and dense limestone lenses have been found in black shales on the very lowest and the uppermost parts of the formation. This stresses the probability of the postulated transitional sedimentation. The Vegamián Formation is a very important lithostratigraphic marker because of its unique lithology and very small thickness over a very wide area of León, Palencia and Asturias. No diagnostic faunas have been found in it but Higgins *et al* (1964) have recently demonstrated that the sequence near Olleros de Alba must belong to the Upper Tournaisian and/or the lowermost Visean stages.

It is remarkable that an extremely similar formation in the Pyrenees, made up of black chert and phosphate nodules, also overlies nodular limestones which have yielded Famennian and Tournaisian conodonts (Ziegler, 1959 and Mellema & van Adrichem Boogaert, internal rept. Leiden). This implies that the chert has been deposited over large distances at the same time. Restricted circulation and/or subaerial exposure characterize this type of sequence. The present Sahara gives many examples of the latter case where siliceous crust has often been found on unconformities. We assume euxinic conditions to explain the black colouring and the phosphate content, whereas silicification is thought to reflect temporary uplifts and exposure.

Alba Griotte Formation

The Alba Formation has been named by Comte (1959) and has an even wider regional extent than the underlying Vegamián Formation. It is described as a red fossiliferous limestone with rose and reddish limestone bands in shale.

In the present area the limestone is a light gray rock being found as large lenses built up of poorly bedded and massive recrystallized calcarenite and argillaceous limestones.

Measured sections of the Alba Formation are illustrated in Enclosure 2, Sections IIa and III. Small white, rarely rose, spots and bands are often associated with common styliolithic and cataclastic structures caused by recrystallization. In some exposures the limestone passes upwards into a very thin-bedded alternation of slates and fine crystalline limestone which are often deformed by folding and fracture into a nodular limestone (fig. 9). In this limestone beautiful isoclinal folds are developed in which a distinct post- or late-folding fracture cleavage has been observed, which is clearly not parallel to their axial planes (fig. 33).

Fig. 9. Deformation by folding and fracture of slates and limestones into a nodular limestone; Alba Griotte Formation.

Goniatites, conodonts and crinoid stems (often very large) have been found. Goniatites have been identified from two localities. One locality yielded specimens of *Goniatites* (*Goniatites*) *stenumbilicatus stenumbilicatus* Kullmann 1961 and *Goniatites* (*Goniatites*) ex gr. *crenistria*. They belong to the lower or middle Goniatites Stage, Upper Visean. Another locality yielded specimens of *Goniatites* (*Goniatites*) *granosus* Portlock 1843 (subspec. indeterm.), *Goniatites* (*Goniatites*) ex gr. *granosus* and *Goniatites* (*Goniatites*) cf. *granosus*. They belong to the upper Goniatites Stage, Upper Visean

(Dr. J. Kullmann, written communication). The conodonts also point to Upper Visean age (van Adrichem Boogaert, 1965).

In the western part of the area the limestones are sometimes partly more strongly coloured by rose and reddish spots and resemble much more closely the typical Alba Griotte. This colour change is still stronger in the neighbouring area to the west, whereas in the Montó area (Kullmann 1962; Mr. J. A. Kutterink, personal communication) the reddish brown spotted limestones follow on top of a grey one. The Alba Formation in the present area is not continuous but lateral passage into shales has been observed NW of Cardaño de Arriba. In the normal sequence the discontinuity is probably due to the lateral variations and sedimentation, whereas below the Curavacas unconformity the local absence of the Alba Formation may be due to erosion.

Caliza de Montaña Formation

The Caliza de Montaña has mainly to be correlated as the generally unfossiliferous limestones almost invariably resting upon the griottes of the Alba Formation. Like all Carboniferous limestones in this region they are mainly recrystallized and while considerable variations in colour, bedding and internal structures can be found none have been found exclusive to this sequence nor capable of subdividing it. The lack of satisfactory described sections emphasises the uncertainty that there is at present in respect of those rocks. In this situation it is felt that the substitution of a new name, the Escapa Formation as proposed by Brouwer & van Ginkel (1964), does not give any positive help. It is quite clear that some very careful work will be required to solve the problems posed by these rocks and until this is carried out it would seem best to retain a non-commital name established as long ago as Paillete (1855) and still in current use (Julivert 1961, Martinez 1962, Frets 1965).

In the present area the Caliza de Montaña Formation forms a broad SE-NW trending chain of limestone mountains of which the 2350 m high Espiguete mountain is the most prominent. It is illustrated by the sequence measured west of Triollo in Enclosure 2, Section III. The thickness in the SE, NW of the village of Santibañez de Resoba, amounts about 200—350 m whereas in the NW only 30—80 m thick limestones are found. North of this chain very thick equivalent limestones have been found in the Picos de Europe being 600—700 m thick according to Delépine (1943) and Barrois (1882). S and SW of the present area an about 800 m thick Caliza de Montaña sequence is found in the Leonide facies (Comte 1959, Koopmans 1962, Rupke 1965 etc.). This thickness decreases however to the E being in the Sierra del Brezo only about 300 m (Kanis 1956). The thickness has been reduced to only a few meters or even nothing in the area of the León Line between the thick Leonide facies and the present area. The Caliza de Montaña Formation has been deposited in this region during the uppermost Visean and Lower Namurian time (Frets 1965, van Ginkel 1965). SW of our area the Caliza de Montaña can be subdivided in a basal dark bluish black, platy and well layered, bituminous limestone without a fauna and a grey medium grained, thick bedded, massive limestone. The latter becomes at its top somewhat oolitic and fossiliferous; fragments of gastropods, foraminifers, crinoids, algae etc. have been collected in the generally recrystallized Caliza de Montaña. In our area however the limestone is predominantly found as white-grey weathering, bluish grey to dark grey, thick bedded or massive limestones. Jointing and recrystallization often obscures the original bedding. In the Sierra del Brezo east of the present area, however, well layered calcilutites (Frets 1965) occur as well as oolitic fossiliferous limestones in the higher part of the sequence. NE of

Santibañez the whole sequence becoemes more and more fossiliferous yielding foraminifers, algae, bryozoans, corals, crinoids and a few brachiopods and ostracods. Veins of azurite, malachite and chalcopyrite have been found in many areas. The Caliza de Montaña Formation passes gradually both laterally and vertically into the shales and sandstones of the Cervera Formation.

Cervera Formation

The Cervera Formation has been named in the area around Cervera de Pisuerga by Brouwer and van Ginkel (1964) and mapped as far west as Triollo by Frets (1965).

The Cervera Formation comprises the clastic deposits which pass laterally and vertically from the Caliza de Montaña facies. The name Cervera Formation forms a substitute (Brouwer and van Ginkel, 1964) for the Culm facies, The thickness varies from nothing up to several hundreds of meters. It contains shales and argillaceous, often graded, sandstones.

The Cervera Formation N of Cardaño de Arriba, as shown in Enclosure 2, Section IIa, consists of about 40 m dark gray shales alternating with thin graded sand laminae followed by about 8 m limestone-shale interval which is again overlain by a sandstone-shale sequence about about 35 m thick. The basal shaly part follows upwards gradually on the Alba Formation and probably passes in the same way into the overlying limestone-shale horizon.

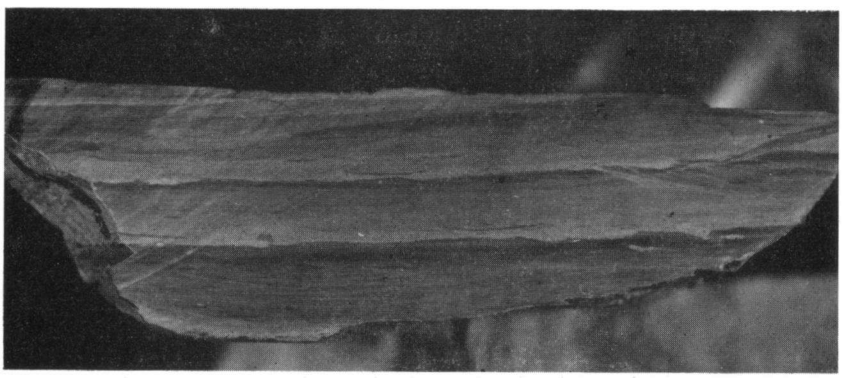

Fig. 10. Well-developed graded bedding in sandstone of Cervera Formation (x 1).

In the basal sequence the sandy intercalations only average about 0.5 cm thick and are much thinner than the shale beds. The sandstone is mostly light gray, mainly finegrained to silty showing well developed graded bedding (fig. 10). The shales are dark gray, slaty, usually thin bedded (2—3 cm) and well cleaved. The cleavage planes are undulated in responses to the grading of the layer (fig. 32). Only some distorted lamellibranchs, goniatites and plant remains have been found. Some isolated limestone lenses have been found in this part of the formation. These lenses consist of dense and very argillaceous, dark gray to black calcilutites similar to those of the overlying limestone-shale horizon. The lenses are mostly thinner then 10 cm and less than 30 cm long, the largest observed is 90 cm thickness and extends less than 3 m a lateral sense. Often these lenses have been dissolved away and only the remnant hollows betray their existence. The limestone-shale horizon comprises about 8 m thinbedded and bluish black calcilutites with interbedded dark grey shales (fig. 35). This part is strongly folded disharmonically in relation to both over and underlying beds so that its precise stratigraphic contacts are somewhat doubtful. The rock itself is recrystallized and partly dolomitic, locally showing rose and white spots and bands. The beautiful cleavage in this horizon is illustrated in fig. 35. A nodular limestone which again appears to have been developed by strong deformation from a thin bedded alternation of shale and limestone can also be found. The conodont fauna points to a Namurian age (verbal communication van Adrichem Boogaert).

The overlying sandstone-shale sequence is thin to medium bedded in which no fossils have been found. The shales are again dark gray and silty. The sandstones are mostly whitish grey to lightgrey, micaceous and show much small-scale cross-bedding and occasionally graded bedding.

More to the south however facies changes are considerable and this formation is mainly represented by a mixed limestone-quartzite conglomerate here called the Triollo Member. This unit usually rests unconformably upon older rocks and is itself frequently unconformably overlain by rocks of the Yuso Group.

The Triollo Limestone-Conglomerate Member is typically developed in the valley west of the village of Triollo after which it is named and which is shown in Enclosure 2, Section III. It consists of a polymict conglomerate of limestone and quartzite pebbles in a sparse sandy matrix in a sequence up to 120 m thick. Usually these rocks rest unconformably upon older formations down to the Murcia Quartzites. Upwards they are most often unconformably overlain by the Curavacas Conglomerates but, particularly in the W, the interbedded sandstone-shale deposits of the Cervera Formation seem to replace the conglomerates vertically as well as laterally.

Fig. 11. Unconformable contact of Triollo Limestone Conglomerate Member upon Caliza de Montaña Formation.

In the conglomerates the limestone components are bluish grey calcilutites or whitish grey fine to medium grained limestones. They have presumably been derived from the Caliza de Montaña limestone. An unconformable contact with the Caliza de Montaña is shown in Fig. 11. Interbedding with fine quartzite conglomerates occurs, e.g. also near Caldares de los Bilboas in the W of the present area. Black chert fragments are often found in these conglomerates especially in the outcrops 4 km SSE of Cardaño de Arriba and E of Barniedo de la Reina.

No fossils have been found in the Triollo Member but the lithology is strikingly similar to the limestone conglomerate near Puerto de Monte Viejo (Prioro beds, van Ginkel, 1965). Those rocks have yielded foraminifers of Profusalinella subzone B (van Ginkel 1965) and algae of Algal zone III (Rácz, 1964), which may correspond to a Namurian C-Westfalian A level.

Yuso Group

The Yuso Group comprises the mainly clastic Upper Carboniferous sequence in the valley of the river Yuso. Within this area an intricate pattern of interbedded and interfingering conglomerates, sandstones, shales and even limestones have been developed. Koopmans (1962) previously proposed the term Yuso Group for these sediments although it was not based on a subdivision into formations.

In the present area two main lithofacies could be mapped which are called the Curavacas Conglomerate Formation and the Lechada Formation. The proposal that all these sediments should be under a single Curavacas Formation (van Ginkel 1965) is not followed since it ignores the most fundamental subdivision in the lithology of this sequence.

The Curavacas Conglomerate Formation (Kanis, 1956) has been named after the conglomerates of Curavacas mountain (2500 m) and comprises the quartzite conglomerate facies of the Yuso Group. The name Lechada was informally introduced by Savage (1961) after the Lechada river flowing in the north of our area in the main exposures of this facies. The Lechada Formation comprises the sandstone-shale facies. A limestone member, the El Ves Limestone Member, has been distinguished in the Lechada Formation.

The two formations unconformably overlie the older strata. Their thickness varies between about 300—1200 m (Lechada Formation) and 0—800 m (Curavacas Formation). The Yuso Group is characterized by frequent interfingering and changes of the two facies represented by these two formations.

Curavacas Conglomerate Formation

The Curavacas Conglomerate in the present area consists of a thick sequence of rather well rounded but badly sorted pebbles, cobbles and boulders of quartzites. The highly argillaceous sandy matrix (graywacke type) is often very abundantly represented. Lateral and vertical variations are characteristic of this formation. The bedding is mostly poorly visible but locally it can be well bedded and flat pebbles can be found as well as sandy or shaly interbeds, which mostly are thinner than 1 meter, which are also indicative for the bedding attitude. The normal appearance of the Curavacas Formation is illustrated in fig. 12, 13.

North of Cardaño de Arriba the Curavacas Formation is present as two thick conglomerate horizons between which the Lechada beds have been intercalated. The basal horizon is very variable in lithology and contains many greywacke type sandstones and shaly interbeds. The upper one is almost purely a conglomerate in

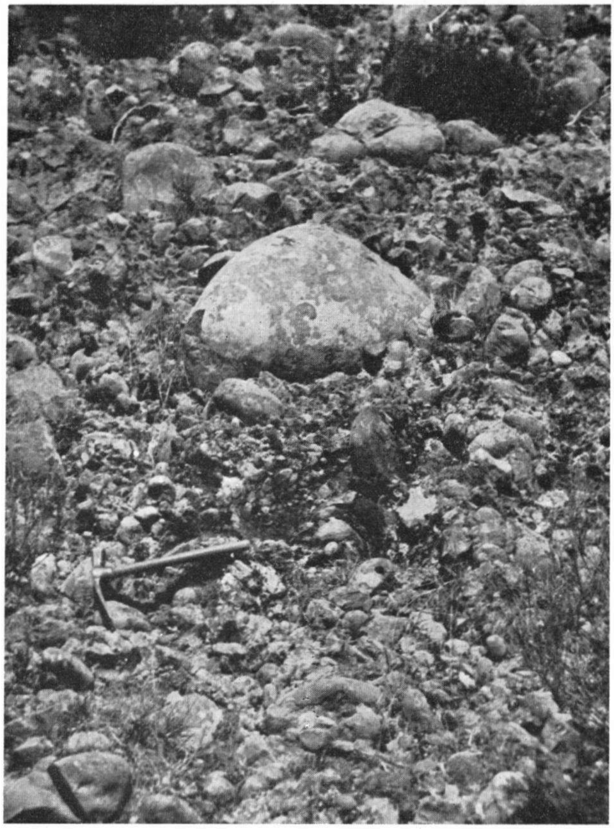

Fig. 12. Exposure of Curavacas Conglomerate Formation. Note poor sorting of constituent elements.

Fig. 13. Exposure of Curavacas Conglomerate. Note abundant matrix.

that section. More to the west (Peñas Matas) also this horizon starts interfingering with the Lechada facies. From the basal horizon a good flora could be collected which definitely points to a Westphalian A (Stockmans, written communication, study in press). The same holds true for a basal sandstone-shale sequence north of the village of Triollo.

The conglomerates are often impressively jointed. The joints are marked by thin sheets of recrystallized quartz cutting straight pebbles and matrix. This is especially the case in the rather compressed conglomerates of the Aguasalio Mountain.

To the south the conglomerates are found unconformably on the thick Caliza de Montaña limestones. Generally the sorting is better in this region which may reflect a northerly position of the source area.

Lechada Formation

The Lechada Formation contains rapidly alternating sandstones and shales both in lateral and vertical sense. The beds are mostly lenticular in shape which reflect the deposition in a rather unstable environment. Slumping and graded bedding occur locally.

The sandstones are best comparable with a subgreywacke as defined by Pettijohn (1957). They are light to medium grey, crossbedded and fine to coarse grained. The shales usually are dark gray and mostly silty.

In the Curavacas syncline the sandstone-shale ratio increases from S to N and from base to top. For instance in the south flank only about 100 m of shales occur below the El Ves Limestone Member whereas in the north flank a sequence of sandstones and shales about 50 m thick has been found below this member. In the NE part of the Curavacas syncline a rather clean quartzitic sandstone is exposed on the Alto de Calderon, in the highest part of the section. This horizon seems traceable along the Lechada river to the west. Near the Hoyos de Vargas, in the NW part of the Curavacas Syncline, much interfingering with the obviously strongly thinning Curavacas Conglomerates occurs. The regional distribution of the conglomerates designates this area as a relatively rather high area during the deposition, situated along the Peña Prieta line (fig. 22). Fossils are almost absent; some poorly reserved brachiopod fragments and a few worm tracks have been found.

The El Ves Limestone Member shows bluish-gray whitish-gray weathering, very recrystallized limestones in which only small fragments of crinoids, bryozoans and some foraminifers have been found, none of which have yet been dated.

An upper limit to the age of the Lechada Formation is supplied by the dating of the Panda Limestone Member which forms the top part of the Lechada Formation in the NW (van Ginkel, 1965). These limestones have yielded foraminifers which have been determined as belonging to Fusulinella subzone B, (upper Moscovian) probably equivalent to a Westfalian C-D level.

The geological map illustrates that the Curavacas conglomerates wedge out to the west, being replaced by the Lechada Formation facies. In the west conglomerates near the base of the formation with an algal reef intercalation (near Vejacerneja) point to a fully marine environment whereas in the east near Cardaño de Arriba the plant fauna in the conglomerates reflect a more paralic environment. Datings from these two fossil groups indicate the younging of the sequence towards the E. Together with the increasing conglomerate thickness it stresses the transgressive onlap of the Yuso Group sequence in a W-E direction whereas supply took place in the opposite sense. The whole sedimentation history has been regulated by the gradual subsidence of the Lechada basin (fig. 16, 39) in the west and its upheaval in the northeast Arauz Area (fig. 16).

Intrusives

The most important intrusion in the present area is that in the north, some 1 to 2 km southwest of Peña Prieta under which name it has previously been referred to. Although its surface outcrop is quite large this intrusion seems to be a fairly thin, sheet-like body owing the expanse of outcrop to having been stripped down a dip slope. A number of stringer dykes could be mapped connected to the intrusive body and other minor intrusions, mainly sills, have been found elsewhere, especially in the Devonian rocks.

From thin sections the large intrusion appears to be a hypabyssal granite of a granodiorite porphyrite composition. Contact metamorphism is very clear in the country rock in which andalusite porphyroblasts with well developed chiastolite crosses, cordierite patches, muscovite, reddish brown biotite, tourmaline and garnet (almandine or spessartine) have all been identified. Chloritoid is also common in the shaly beds of the Murcia Anticlinorium especially those of the Gustalapiedra and Vidrieros Formations. The latter has even been converted in places to a true chloritoid mica-schist of a white-grey colour sometimes purplish or reddish in the presence of biotite.

The occurrence of the main intrusion in the core of the Lechada syncline has been related to the folding of this structure (fig. 15) which would mean that the intrusion should be syntectonic. This conclusion is supported by the discovery of a sill folded in one of the minor folds of the Lechada syncline. Unfortunately it has not been possible to study metamorphism adequately so that it is still not clear what its exact role in the deformation has been.

Review of Carboniferous Stratigraphy

The sedimentation pattern during the Carboniferous is almost not comparable with that of the Devonian. Very instable conditions of deposition are indicated by the big lateral and vertical changes in thickness and lithology.

A gradual transition from the uppermost Vidrieros deposits into the lowermost black shale deposits of the Vegamián Formation is found. Conodonts and brachiopods (*Lingula*) are found in both sequences. Also in the area to the west it is found that almost uninterrupted sedimentation took place in the Famennian and the Tournaisian; only temporary and local subaerial exposure might be indicated (Higgins *et al.*, 1964).

The Alba Griotte is found as grey limestone lenses in which the characteristics of true griotte are not always recognizable.

The Caliza de Montaña facies shows clearly transition into the clastic facies of the Cervera Formation to the north. This transition zone is also marked by a belt of unconformable limestone conglomerates of the Triollo Member. This belt is aligned along the Cardaño Line (fig. 16). Graded bedding and major slumping and limestone conglomerates in the Cervera Formation emphasize the instable conditions during this part of the Carboniferous.

The thick molasse deposits of the Yuso group overlie unconformably the older strata. Its deposition pattern with the very thick coarse nature of the clastics was conditioned by major tectonic events. In the present area the Yuso deposition is presumably related with the tilting of the Cardaño block between the Cardaño Line, the Peña Prieta Line and the Polentinos fault; the Lines are expressed at the surface by Espiguete, Rasa, Peñas Matas and Rio Frio faults (fig. 16). Full marine and paralic environments are both represented.

The southern part of the area shows a pre-Westphalian lithosequence which is partly comparable with the Leonide facies (Enclosure 2, Sections 4 & 5; Table 1). The abundant Westphalian deposition deposition is almost not represented in the Leonide facies. The total thickness of the Yuso Group can easily exceed the total thickness of Devonian and older Carboniferous of both areas.

Interruption of the lithosequence by an hiatus is possibly present at small scale in the Vegamián Formation, but certainly present at the base of the Triollo Limestone Conglomerate and the Yuso deposits.

3. RELATIONS BETWEEN TECTOFACIES, LITHOFACIES AND BIOFACIES

Lithofacies

In Table 1 a comparison is given between the Palentian and the Leonide lithofacies.

TABLE 1 *Comparative Summary of the Differences between the two principal lithofacies.*

South of León Line (Leonide facies)	*North of León Line (Palentian facies)*
1. Thick Devonian reef limestones	usually thin argillaceous limestones
2. Thick alternation of strata	thin alternation of strata
3. Thick Upper Devonian Quartzite of whitish colours	dark and relatively thin Upper Devonian Quartzite with shale interbeds
4. Thick Caliza de Montaña facies	thin or no Caliza de Montaña facies, but the clastic facies of the Cervera Formation is well represented
5. Light coloured rocks	dark coloured rocks
6. Brachiopods, corals and bryozoans dominant	goniatites, trilobites and conodonts dominant
7. Reddish Alba Griotte Formation	dark gray Alba Griotte Formation
8. Hiatus in Middle and Upper Devonian and Lower Carboniferous (area to the S)	no hiatus indicated
9. Rare Westphalian	very thick Westphalian

N.B. The Carazo and Lebanza Formations are represented in both facies and are therefore not involved in this discussion.

This table emphasizes the very different sedimentological and tectonic history of the two areas. In the Leonide facies the deposition presumably took place in a rather stable shelf environment of generally shallow depth (reefs, bottomwellers) of $\pm$ 20—50 m. Block movements are reflected in the hiatus at the transition beween Devonian and Carboniferous.

In the Palentian facies the deposition presumably took place in a stable environment in which restricted circulation is reflected in several formations. Important tectonic disturbances cannot be derived from the lithosequence.

Enclosure 2 shows two partial sections of the Leonide facies, two partial sections of the Palentian facies and, partial section of the southern rim of the Central Pyrenees in which a basin development is found which is quite comparable to that in the present area.

Biofacies

The Palentian biofacies is characterized by the association of conodonts, goniatites and trilobites and by the absence of stromatoporoids, coelenterates and abundant lamellibranchs and gastropods. Lamellibranchs and gastropods are found in the Palentian facies, but they are mostly small tiny and bulbous forms with a diameter of a few millimetres; only in the Gustalapiedra Formation a bank of big *Posydonia* (3—8 cm) shells is found. In the shaly interbeds of the Murcia and Vidrieros Formations and also in the Vegamián Formation thin shelled *Mytilus* and *Lingula* forms can be found. The goniatites and trilobites occur in good numbers but only a few genera are represented which points to a specialisation being conditioned by the specific environment of deposition.

The only trace of terrestrial deposition is found in the beautiful fossil flora in shaly interbeds of the Yuso sequence. These plants reflect there a paralic environment. The main part of the Yuso Group, however, is thought to have been deposited as an onlap sequence in a marine environment (major slumping, local graded bedding and marine limestones, depositional pattern).

Resuming we might say that the pre-Yuso deposition of the Palentian facies took place in more deep and quiet water outside the coastal reef girdle and the main influence of the source areas (thinness and shalyness).

Tectofacies

Strong control of the sedimentation by the tectonics is reflected in the rock sequence of the Cardaño area.

1. Stable conditions during the Devonian allowed a quiet and regular sedimentation of low energy deposits. Higher energy deposits are only found in the lowermost and uppermost stages of the Devonian (Carazo and Murcia Formations). This might reflect the start and the termination of a major depositional cycle.

2. The Carboniferous sedimentation starts with a sequence in which the assumed local subaerial exposure might indicate that the long period of tectonic stability had started to become disturbed.

3. The striking facies changes between the Caliza de Montaña and the Cervera facies and the deposition of the accompanying Triollo Limestone Conglomerates form the first strong evidence of important tectonic disturbances. These disturbances culminate during the thick and coarse molassic Yuso deposition. The thick-

ness distribution and the occurrence of thick coarse limestone and quartzite conglomerates demand strong epeirogenetic upheavals in the source area and subsidence in the deposition area along faultzones (steep gradient, c.f. Pettijohn 1957). Two main tectonic lines are found in the present area; the Cardaño Line to the South and the Peña Prieta Line to the north (fig. 16). The geological map illustrates the obvious alignment of the Carboniferous sedimentation along these fundamental tectonic lines.

4. Strong folding of the whole sequence created complicated isoclinally folded and N-S upthrusted major structures in which the Yuso deposits took part (e.g. section E, geological map). The major folding phase therefore is Asturian whereas the Sudetic phase is mainly characterized by epeirogenetic movements. The influence of the Bretonic phase is hardly distinguishable. For detailed discussion of the tectonics we refer to Part II.

Figure 39 gives a tentative and schematic impression of the relation between major tectonic lines and sedimentation.

The very divergent rock sequences in the Leonide and the Palentian facies (being conditioned by their respective tectonic conditions during the sedimentation) point to a divergent palaeotectonic history south and north of the León Line.

Part II

TECTONICS

INTRODUCTION

The structural mapping of our area has revealed striking correlations between a variety of styles of tectonic deformation and different lithologic units. A strong control of the pattern of sedimentation has been exerted by a system of recurrently active lines of deep-seated origin. As can be anticipated irregular structural complications are to be found near these lines due to the changes in structural styles of different facies as well as the reactivation of the deep-seated structure itself.

The tectonic lines so evident in the sedimentary and structural patterns include the Peña Prieta Line which forms the northwestern limit to all our structures except the Lechada syncline which belongs exclusively to the northwest; the Polentinos and Curavacas faults that bound the Arauz Area to the east and west, respectively; and the Cardaño Line which underlies the complicated structures of the Central and Southern Areas.

The three major lithostratigraphic sequences which have already been distinguished in Part I are clearly identifiable with their own individual structural style so that the basic structural subdivision of our area corresponds to the broad lithostratigraphic distribution shown in the geological map and the sketch map of fig. 16. We have subdivided the region into four structural units; the Northern Area with sediments of the Yuso Group, the Central Area with Devonian rocks, the Arauz Area with a much more complete sequence of Devonian rocks, and the Southern Area dominated by the massive limestones of the Caliza de Montaña. The structural style of the Northern Area has been controlled by the thick developments of Curavacas Conglomerates, so typical of the Yuso Group here, resulting in broad, simple structures — the Lechada and Curavacas Synclines. The relatively thin-bedded Devonian rocks of the Central and Arauz Areas have developed intensive folding and thrusting in very narrow structures. The thick massive limestones of the Southern Area have been deformed into a complicated pattern of WNW — ESE and NNW — SSE isoclinal folds.

4. NORTHERN AREA

The Northern Area is mainly occupied by the broad belt of rocks of the Yuso Group. This belt shows a basically synclinal structure being named Curavacas syncline (de Sitter 1957) in its eastern part and Lechada syncline (Savage 1961) in its western extension. They are respectively named after the Curavacas Mountain and the Lechada river. The Curavacas syncline is separated from the Lechada syncline by a zone of disturbance. This zone is represented here by the Peñas Matas thrust fault and the Hoyos de Vargas anticline which are both aligned along the Peña Prieta Line (fig. 39).

Lechada syncline

The major structure of the Lechada syncline is clearly shown by the attitude of the Curavacas conglomerates. The syncline is moderately asymmetric with the south flank (about 60°) steeper than the north flank (about 40°). The main axis is distorted both in horizontal and vertical sense. The rocks of the Lechada Formation in this structure and especially those in its core have developed many minor folds which have been proved to bear a secondary parasitic relation to the major structure. Although the major and minor axes and directions of deformations are widely divergent it has been demonstrated by extensive measuring and plotting (Savage 1961) that they could all nevertheless have resulted from the application of a single major stress system (fig. 14).

At least two generations of cleavage, usually a slaty and a fracture cleavage are found (figs. 18, 24). The slaty cleavage is generally flatlying and shows the typical orientation of recrystallized mica and quartz almost entirely restricted to the argillaceous beds. At the junction between argillaceous and coarser clastic beds these cleavage planes curve into bedding planes. This cleavage bears a clear geometric relationship to the axis and axial plane of the concentrically folded bedding planes of the competent non-cleaved beds. Therefore the original description of this cleavage

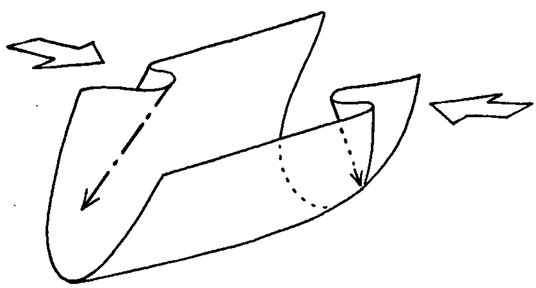

Fig. 14. Plunging minor folds resulting from stress oblique to the main structure.

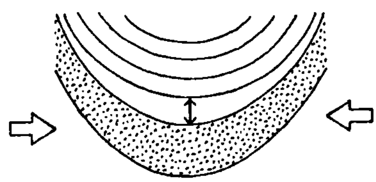

Fig. 15. Reduction of vertical stress in the axial zone of a syncline due to the folding.

as bedding schistosity was not satisfactory and the term concentric cleavage has been applied recently (de Sitter 1964). In this term the genetic relationship of this cleavage with the concentric folding stage is clearly expressed. The concentric cleavage has been deformed by the younger parasitic minor folding and crumpled during the development of an axial plane fracture cleavage to the parasitic folds (fig. 17). For extensive documentation of the cleavage-folding relations in the Lechada syncline by measurement and plotting and their interpretation we refer to Savage (in preparation).

The Hoyos de Vargas intrusion (Peña Prieta Granite) is clearly related to the major structure. It has a sheetlike shape with a complex system of crosscutting dykes and lies in the axial zone of the easternmost extension of the Lechada syncline that reaches as far as Peña Prieta Mountain. This intrusive activity presumably will be related to the deepseated underlying structure (Peña Prieta Line) but its precise position here was no doubt dictated by vertical tension in the axial zone of the fold (fig. 15). Comparable intrusions have been described from the Central African Shield (de Sitter 1964, p. 445).

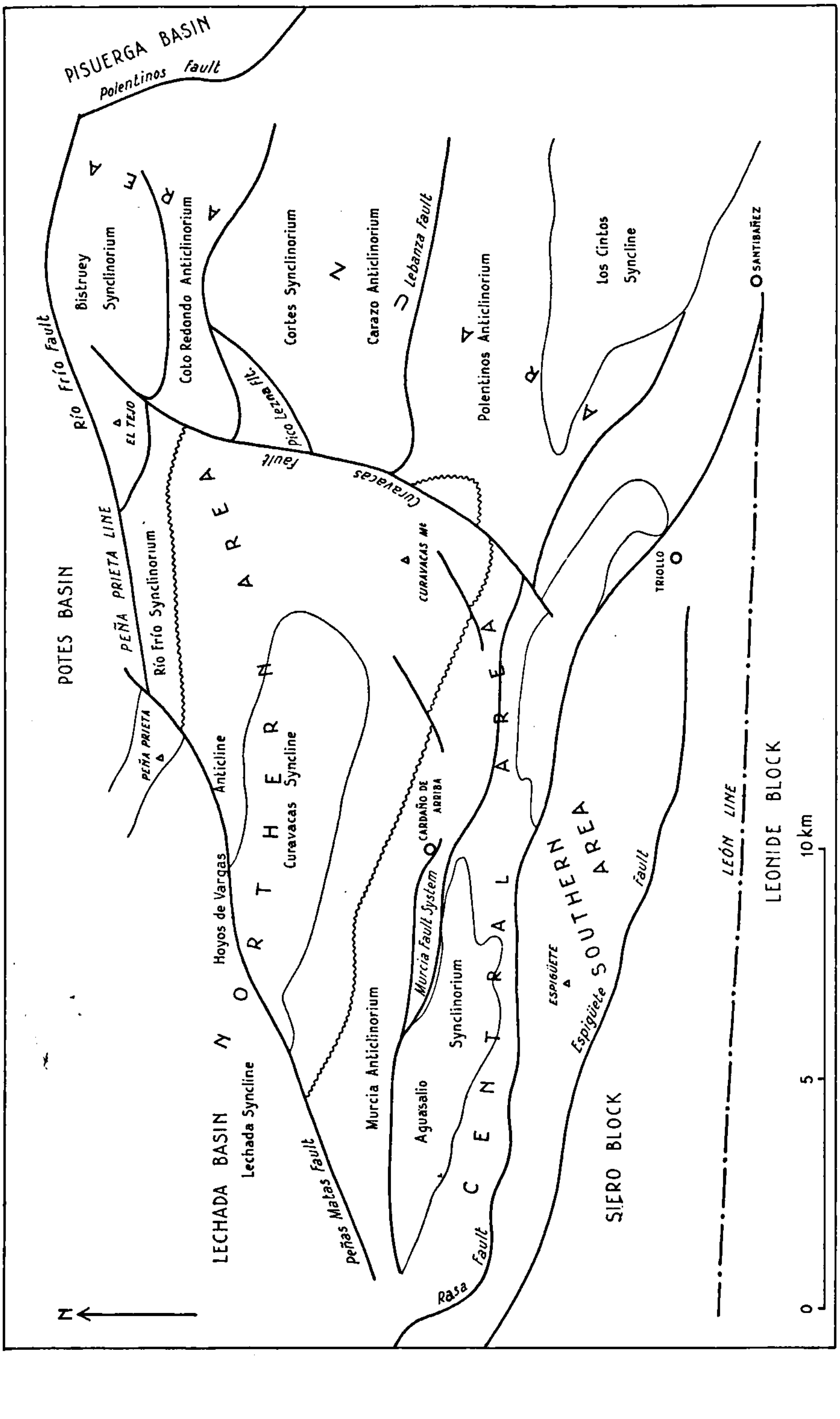

Fig. 16. Sketch map of the main structural trends and areas of the Cardaño Block.

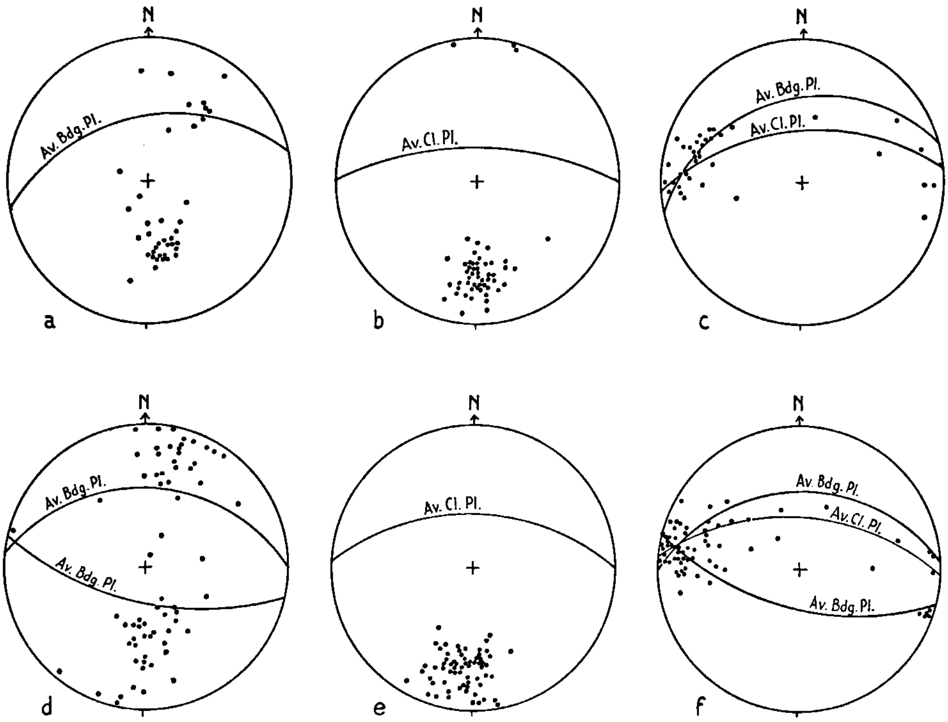

Fig. 17. Stereographic projections (Schmidt Net) of measurements from the Lechada Formation, Curavacas Syncline;

Eastern part:

a) Bedding plane poles (38); average bedding plane, 345/40.
b) Cleavage plane poles (46); average cleavage plane, 355/60.
c) Bedding/cleavage intersections (38); average intersections plunges, 25/285.

Western part:

d) Bedding plane poles (67); average bedding planes, 194/680:000/45
e) Cleavage plane poles(67); average cleavage plane, 004/60.
f) Bedding/cleavage intersections (67); average intersection plunges, 08/279.

Curavacas syncline

The Curavacas syncline shows a somewhat different structural pattern in its major and minor folding. The major structure has been strikingly dictated by the thick Curavacas conglomerates which have given it an asymmetric form with a steeper northern flank. In the east the dips are approximately 50° S and 25° N in the north and south flanks respectively, but to the west the structure becomes oversteepened due to the influence of the Peña Prieta Line so that the north flank is overturned and overthrusted while the south flank dips at about 50° N. Hence the asymmetry of the Curavacas syncline is just opposite to that of the Lechada syncline.

The asymmetric major structure of the conglomerates is often reflected in the minor structures found in the overlying Lechada rocks in the Curavacas syncline. Many minor structures are asymmetric having north dipping axial planes. This holds especially true for the zone south of the overthrusted north flank (section D). The minor structures can usually be seen to have developed a clear axial plane slaty cleavage in the more argillaceous beds. The average cleavage plane dips rather steeply to the north and has a ESE-WNW strike (fig. 17). This type of cleavage development in the Curavacas syncline is not obviously similar to that in the Lechada syncline. The axial plane cleavage frequently shows a convergence upwards into the crest of anticlines (fig. 19) in contrast to the usual picture (fig. 18, 20a). There seems

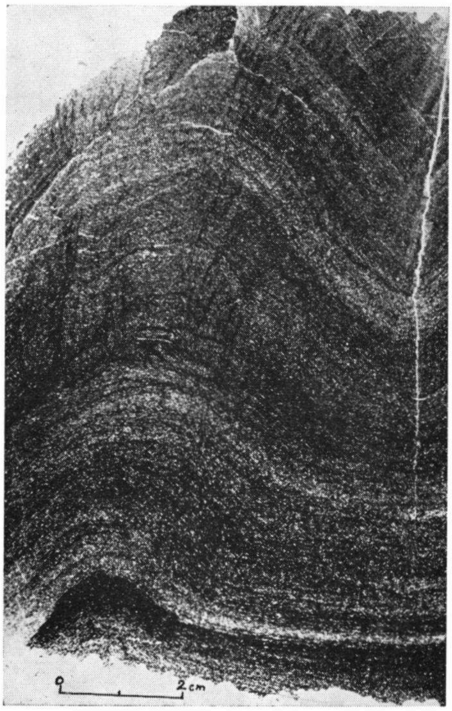

Fig. 18. Minor fold showing folding of concentric, slaty cleavage and development of axial plane, fracture cleavage; Lechada Formation, Curavacas syncline.

Fig. 19. Minor anticline showing axial plane, slaty cleavage converging towards the crest; Lechada Formation, Curavacas syncline.

to be a tendancy for this type of fanning to be associated with thicker bedded sequences (30—50 cms) of sandstone-shale alternations (fig. 19). In this case the dip of the cleavage approaches that of the bedding in either flank of the structure so that it may be regarded as an intermediate form between the most usual, upward-fanning, and the rare concentric cleavage (fig. 20). A connection between these two types seems quite probable since here they are both of a slaty type. However, in the south flank of the Curavacas syncline, as in the Lechada syncline, the slaty cleavage is of a concentric type, almost parallel to and indistinguishable from the bedding, associated with the development of the major fold; whereas in the core of the former structure it is an axial plane cleavage to the minor folds (fig. 19). In these cases the

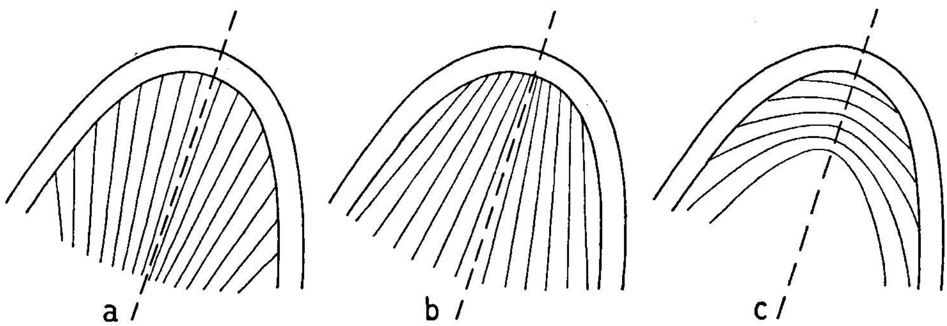

Fig. 20. Diagrams of cleavage relations in an anticline a) normal axial fan diverging upwards b) unusual axial fan converging upwards c) concentric cleavage.

slaty cleavage deviates quite widely from the bedding and sometimes a few planes can be seen penetrating silty and sandy layers forming a 'German' type of fracture cleavage. In this case the slaty cleavage of the argillaceous beds is represented by a fracture cleavage in the sandy interbeds.

It is quite evident in the field that the angle of refraction and the reduction

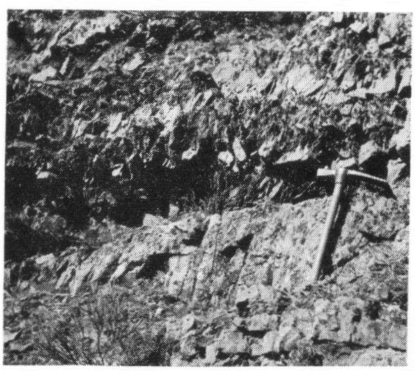

Fig. 21. Sandstone beds pulled apart and the segments rotated into the plane of the slaty cleavage; Lechada Formation, Curavacas syncline.

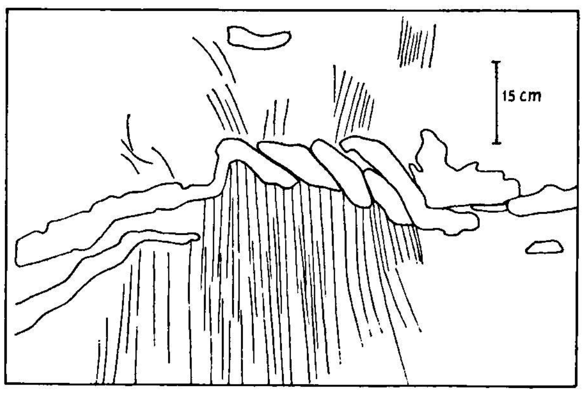

Fig. 22. Broken segments of sandstone beds shoved together by compression perpendicular to the slaty cleavage; Lechada syncline, Curavacas syncline (after photograph).

TABLE 2 *Comparative Summary of the Differences between the two Synclines.*

Lechada syncline	*Curavacas syncline*
1a Lechada Formation much thicker than Curavacas Formation; ratio > 10:1 Conglomerates occur in large lenses in the sandstone-shale facies of the Lechada Formation.	1b Lechada Formation thinner or equally as thick as the Curavacas Formation; interbedded as parallel layered units.
2a South of the Lechada syncline only a narrow faulted zone of pre-Yuso rocks is present along the Cardaño Line.	2b South of the Curavacas syncline a broad belt of strongly folded and/or N-S thrusted pre-Yuso is exposed along the Cardaño Line.
3a No flora-bearing layers have been observed in the conglomerate facies; however, foraminifers could be collected from an intercalated limestone lenticle.	3b Rich flora-bearing interbeds have been found in the basal sequence of the conglomerates north of Caradño de Arriba.
4a Great width of the structure; wavelength about 6 km.	4b Relative moderate width of the structure wavelength about 2 km.
5a Moderately asymmetric major structure with steepened south flank.	5b Moderately asymmetric structure with steeper north flank.
6a Main axis of the syncline is distorted, showing east plunge in its western part part near the Yuso river and west plunge in its eastern part near the Peñas Matas fault. Moreover, deflections to the north seen also in those parts.	6b Main axis without complications; general west plunge of the structure.
7a Initial concentric folding accompanied by a flat lying slaty cleavage of the concontric cleavage type.	7b Initial concentric folding not clearly connected with a flat lying concentric cleavage.
8a Continued folding created parasitic minor folds in which a clear axial plane fracture cleavage developed in its shaly parts	8b Continued folding created strongly asymmetric parasitic folds in which an axial plane slaty cleavage and a fracture cleavage developed; north dipping and ± E–W striking cleavage plane reflects main geometry of the syncline.
9a Very complicated minor structures throughout the syncline, especially in its central part.	9b Minor structures gradually develop and become complicated towards the western part of the syncline.

in frequency of the cleavage planes per unit are both directly related to the grainsize of a rocklayer; the angle with the bedding increases with increasing grainsize and the frequency of the cleavage planes per unit decreases. In the latter case the aspect of the planes also changes and they become more irregularly curved and rougher surfaced, being of the fracture cleavage type. For instance a slightly conchoidal cleavage plane in siltstones and an irregular hackely plane in fine grained sandstone can be found. Usually this fracture cleavage in the coarser grained interbeds is refracted towards the perpendicular to the bedding. The separated parts of a sandstone bed may be deformed or rotated between cleavage planes (fig. 21). Sometimes these segments have been pulled apart and separated by shales while in other placed the broken parts seem to have been shoved together to form an imbricate type of structure (fig. 22). Only a few exposures have been found in the Curavacas syncline where the fracture cleavage can be see clearly cutting across the earlier slaty cleavage as is so common in the Lechada syncline. In the latter structure initial recrystallisation can occasionally be seen on the fracture cleavage planes which gives rise to some difficulties in distinguishing them from the slaty cleavage in the field. Crenulation cleavage (Knill 1960) showing the same relation to the minor folds as the fracture cleavage in the Lechada syncline, crumpling the slaty cleavage into fine microfolds, has not been found in the Curavacas syncline.

Synthesis of Geological History

The very unstable regional conditions that existed during sedimentation are reflected in the large and variable thickness as well as the nature of the Yuso deposits. Thick sequences of shales, sandstones (many graywackes), very poorly sorted and extremely coarse conglomerates have been deposited in an environment where large-scale slumping and locally turbidite deposition occurred. This picture implies that vertical movements have to be evoked to explain the considerable gradients which induced the strong erosive activity and special conditions for the transport and deposition of the coarse conglomerates. Major block movements are though to have occurred along faults connected with fundamental lines such as the León Line, Cardaño Line, Peña Prieta Line and faults as the Polentinos fault. The first clear evidence of important crustal disturbances after the stable shalf conditions of the Devonian and the Lower Carboniferous have been revealed in the deposits of the Cervera Formation. Its thick limestone conglomerates (short-distance transport, mostly connected with strong vertical movement, Pettijohn, 1957, p. 252) and its turbidite and slumped sequence confirm the start of a period of strong epeirogenetic activity.

The main belt of Yuso sediments in the present area extends from the Curavacas Mountain in the central part towards the west and thickens considerably in that direction. In the eastern part of the belt, for instance north of Cardaño de Arriba, about 800 m of Curavacas conglomerates are overlain by about 400 m Lechada beds whereas 15 km to the west only about 50—100 m of Curavacas conglomerates but at least 1500—2000 m of Lechada beds have been measured. The basin in which the Yuso deposition took place will be called Lechada basin, it is named after the central Lechada river. The Lechada basin is limited to the south by the Cardaño Line. Eastwards a deeply eroded area of older Devonian outcrops limited by large faults occurs, which is called the Arauz Area, after the main stream. The limiting faults have been named the Curavacas, the Rio Frio and the Polentinos faults being situated respectively to the W, N and E of the Arauz Area. East of the Polentinos fault a second important Yuso basin is preserved known as the Pisuerga basin. A third big Yuso basin is found between the Cardaño Line and the León Line and is called the Siero basin after the village of Siero de la Reina. The position of these three major basins is illustrated by figure 39, which also brings out that their locations have been rigourously dictated by a few main tectonic lines.

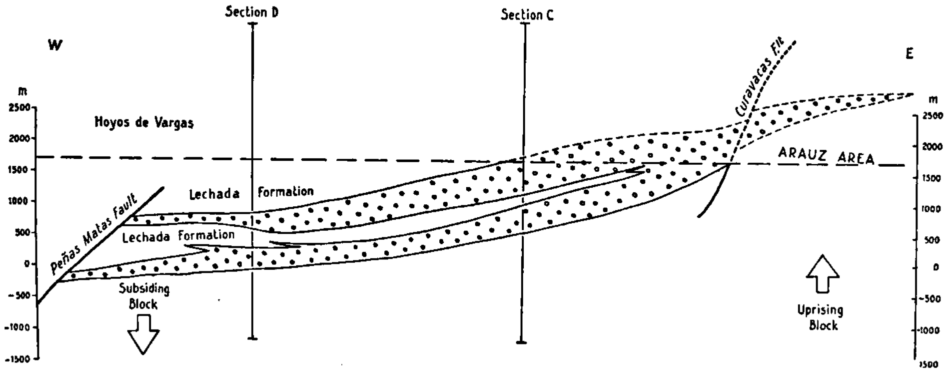

Fig. 23. Longitudinal section along the axis of the Curavacas syncline.

The sedimentary picture of increasing thickness of the Yuso deposits to the west concurrent with the pinching out of the Curavacas conglomerates is a strong indication that the basin was subsiding more towards the west (fig. 23). Such tilting of the Cardaño block is also demonstrated by the pre-Yuso unconformity surface which on the south flank of the area cuts down to the Cervera Formation in the west near Barniedo to the Upper Devonian in the east near Vidrieros; and along the north flank of the area where the contact is with the lowermost member of the Carazo Formation in the east but only the Lebanza Formation near Peña Prieta in the west. Furthermore the structural style developed in the Upper Devonian near Vidrieros must be expected to occur about 500 metres below the present structures exposed around Barniedo. It seems most consistent and simplest to consider that all this tilting was part of the same process immediately preceding and accompanying the development of the Curavacas and Lechada Basins.

The Curavacas conglomerates are somewhat thicker in the northern flank of the Curavacas syncline than in the south, as can be seen from Section C with the geological map. It seems probable that this difference could cause the asymmetric development of the structure with the steepened limb dipping in the direction of thinning (de Sitter 1964, p. 200). Further to the west beyond Peña Prieta the overturned and overthrust Hoyos de las Vargas anticline forms the north flank but this structure is so closely associated with the Peña Prieta Line that it is impossible to define the exact roles of the sedimentary variations and renewal of deep-seated tectonic activity in the formation of this structure.

In conclusion it must be reiterated that the tilting and subsidence of the Cardaño block along deep-seated structural lines is clearly indicated in the pattern of sedimentation found in the rocks of the Yuso; most especially the westward passage of the Curavacas conglomerates into the beds of the Lechada Formation. The gradual overlap of the plantbearing, near-coastal conglomerates by clearly marine deposits also emphasises such tilting of the basin border. The regular succession of younger rocks beneath the Yuso unconformity going towards the west seems to make it most likely that the pre-Yuso rocks were little if at all folded at the time of the block movements.

The later tectonic deformation of the Asturian phase was caused by a N-S directed stress which folded the Yuso rocks of the Curavacas basin into an asymmetric syncline with a steeper north limb because of the thickening of the competent Curavacas conglomerates there. The form of the minor folds in the Lechada Formation in the core of the Curavacas syncline has been dictated by the geometry of the major fold and hence also by the sedimentary pattern. A strong facies change may be expected across the Peña Prieta Line near Hoyos de Vargas but the area is so complicated that the exact picture cannot be drawn. Nevertheless it seems clear that the whole structural development within the rocks of the Yuso Group in this Northern Area has been predetermined by a system of deep-seated tectonic lines mainly through their influence on the pattern of sedimentation.

5. CENTRAL AREA

The structures of the Central Area are found between the Yuso rocks of the Northern Area and the belt of thick Carboniferous of the Southern Area. They are mainly composed of Devonian rocks and are generally inclined to the NNE with WNW-ESE trending structures. Strong deformation has resulted in isoclinal thrust-folds very tightly compressed (Geological Map and Sections B, C, D and E). The intense stress has also resulted in the development of beautiful cleavage, the metamorphism to a low-grade, green-schist facies and the intrusion of thin acid sheets.

The Central Area comprises two major structures, the Murcia anticlinorium and the Aguasalio synclinorium both of which plunge gently to the WNW. The strongest deformation took place west of Cardaño de Arriba (Sections C, D and E). West of Barniedo de la Reina the Devonian rocks are only found in small exposures along a ridge running further WNW (de Sitter 1962b, Geological Map). This ridge forms the continuation of the Cardaño Line although without the large development of Carboniferous limestones seen in the Cardaño area.

Murcia Anticlinorium

The Murcia anticlinorium, named after the dominant Murcia Mountain through which it runs, mainly involves Middle and Upper Devonian rocks although fair sections of older Carboniferous beds have been preserved on its northern flank. To the north the structure is flanked by Yuso deposits which unconformably overlie the Murcia Quartzite near Vidrieros (to the east) and the Cervera Formation near Cardaño de Arriba.

East of Vidrieros the Murcia structure shows a system of relatively simple, gentle folds in the Murcia Quartzite having wavelengths of about 1400 m (Section B). These structures become more complicated towards the west becoming steeper and intricately thrust-faulted. Near Cardaño de Arriba they form part of a large, clearly asymmetric, anticlinal structure with an overturned southern flank and the northern limb thrust over it by at least 500 m movement along a branch of the Murcia fault system. The width of the whole structure here has increased to about 1100 m but the core of Middle Devonian rocks have been crumpled and deformed by minor folding and faulting while the Murcia Quartzite takes part in subsidiary folds about 500 m in wavelength (Section D). West of Murcia Mountain the Middle Devonian rocks of the south flank of the Murcia anticlinorium have been thrust over the Yuso rocks to the south; the amount of overthrust is here greater being at least 650 m. At its most westerly extension the Murcia anticlinorium starts to plunge to the WNW, being overriden by the southward advancing Lechada syncline. The Murcia structure is still overturned to the south and is partly thrusted over the Curavacas Conglomerates of the bordering Aguasalio synclinorium. The thrust movement here, however, has decreased to only about 200 m and the main deformation is to be found in the intricately folded and faulted north flank which is aligned along the Peñas Matas fault. These latter complex folds seem to have wavelengths of less than 500 m.

The Geological Map and Sections illustrate that the maximum deformation of the Murcia anticlinorium took place around Murcia Mountain, where Upper Devonian rocks, for example, must have been thrust up to an altitude of at least 2500 m on the crest of the structure. Near Barniedo and Vidrieros, at the western and eastern extremities of the structure, respectively, the same rocks form the crest at altitudes of 1200 and 1500 m, respectively. It is remarkable that this maximum deformation of the Murcia structure should have taken place due south of the Hoyos de Vargas anticline where the greatest deformation of the Yuso beds is to be found (Section D). This correlation may be purely coincidental but it seems possible that this arrangement may have been caused by the way in which the Devonian rocks have been affected by the tectonic stresses transmitted onto them through the massive conglomerates to the north. Sections C and D seem to illustrate clearly that the massive conglomerates of the Curavacas syncline must have had an influence on the structures developing in the much less competent Devonian and older Carboniferous sequence to the south.

Minor structures occur abundantly in the Central Area. At a certain late stage of folding minor structures will be formed in that part of the core of a syncline or

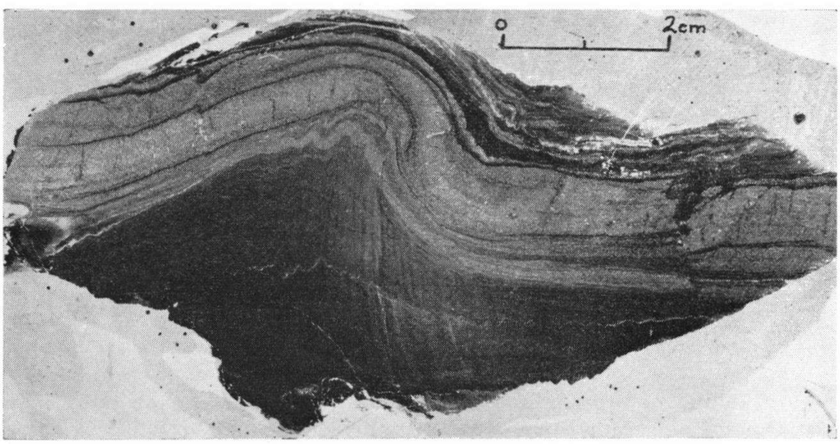

Fig. 24. Microfold in sandstone-shale laminations folding the subparallel slaty cleavage which is crumpled and broken by the fracture cleavage parallel to the axial plane of the fold; Lechada Formation, Lechada syncline.

anticline where lack of space demands special adaption to the compressive forms (fig. 24). At this level the major concentric folding pattern can be overtaken by a system of minor folds. Increasing minor folding therefore is a function of increasing compression of the structure. The tightness and strong compression of the structures in our area is well reflected in a rich collection of minor folding phenomena.

1. In the Gustalapiedra Formation near Cardaño de Arriba we find thick limestone beds with relatively thin intercalations of shale (fig. 2). The rocks are strongly recrystallized and the shales appear to be more competent and less influenced by weathering than the limestones. The shale intercalations show symmetric sinusoid-like or asymmetric structures. Their pattern is clearly conditioned by the position on the structure in which they occur. On a steeply folded flank we find the asymmetric shale structures with axial planes about parallel to the main axial plane. The same phenomenon can be found on a smaller scale in thin sections of very argillaceous parts of the limestone beds. These shale intercalations behave as parasitic folds with respect to the structure in which they are encountered; fig. 25 shows an example taken at the culmination of a structure; the shale intercalations have been stretched in the sense of the axial plane of this structure. They reflect in their

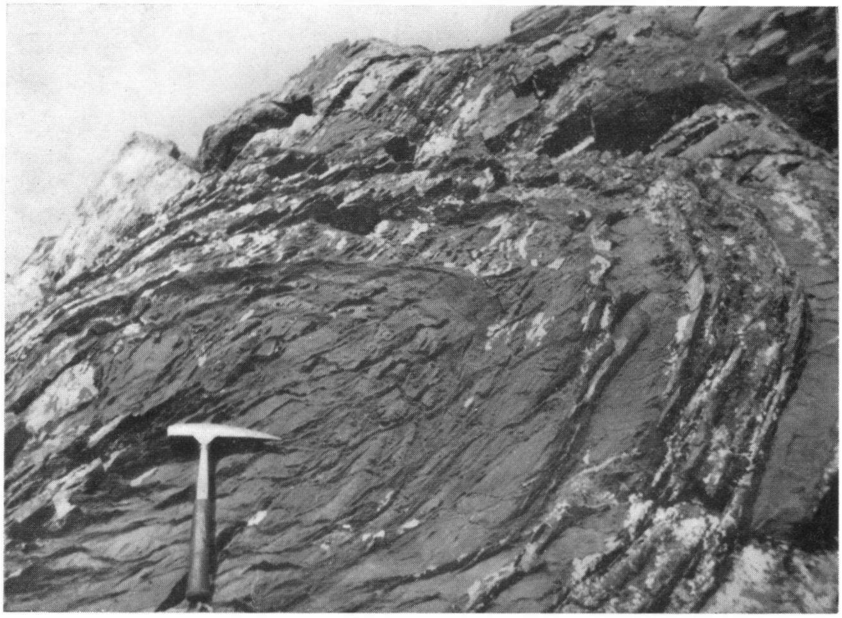

Fig. 25. Hinge of an isoclinal fold in limestone with shaly beds more resistent to weathering which have also stretched out parallel to the axial plane cleavage, Gustalapiedra Formation, Murcia Anticlinorium.

broken elongated structure again their parasitic relation to the structure in which they are found. The symmetrical shale structures with their regular sinusoid-like undulations can be found in the flatter parts of the sequence where there was no relative movement to exert a drag influence on the involved strata (different velocities at both sides of a bed during folding). Figure 26 sketches a reconstruction of these 3 types of shale structures. The axis of deformation and the cleavage-bedding intersections around Cardaño de Arriba show a generally eastwards plunge of about 30°. Exposures at other locations gave quite different values and therefore it became clear that the orientation of an axis is too much dependent on the position in the complicated structural pattern to give identical results throughout the main structure. For instance the thrusted Polentinos limestone, about 2 km to the west in the same structure (Murcia anticlinorium), show distinct west-plunging minor structures. In the Arauz Area good examples have been found of the indicated relation of the trend of folds and the orientation of the strata before their folding. Here we find at both sides of the Curavacas fault different plunge directions in the Rio Frio

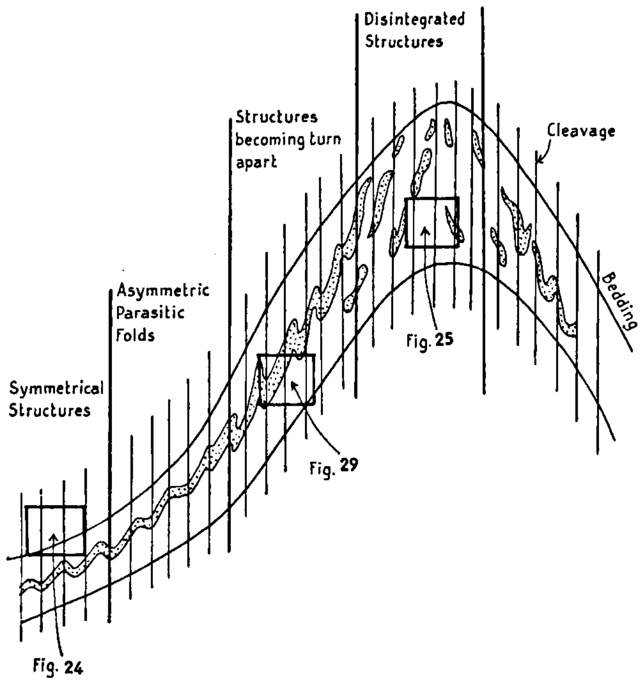

Fig. 26. The development of minor structures in different parts of a larger fold.

and Coto Redondo major structures, respectively. Eastwards tilting of the Arauz Area east of the Curavacas fault is expressed in the E dipping structures whereas the westwards dipping depositional inclination of the tilted Cardaño block is reflected in the westward plunging structures of the Rio Frio synclinorium. Figure 27 gives some idea of the relation of axial trends and prefolding orientation with respect to the stress field.

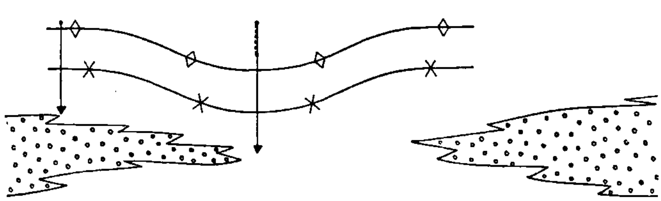

Fig. 27. The development of crosstrends of folding due to inhomogeneity in the stratigraphic sequence.

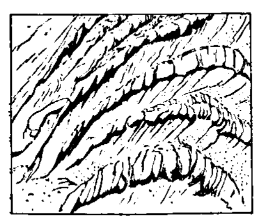

Fig. 28. Axial plane, slaty cleavage in fold of limestone (stippled) and slate; more perpendicular to bedding in slates: Cardaño Fm. After photo ($\times$ $^1/_{40}$).

2. The Cardaño and the Vidrieros Formation consist of thinly interbedded shales and limestones which are often strongly isoclinally folded. The shale intercalations formed also here the more competent components during the deformation. This is moreover shown in the steepening of the cleavage planes when cutting through the shale intercalations (fig. 28). Beautiful examples of parasitic folding can be found

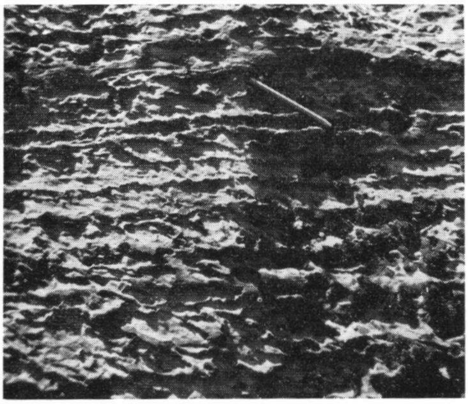

Fig. 29. Parasitic folds in thin shale partings in limestone being stretched out in the plane of the cleavage (parallel to pencil); Cardaño Formation, Murcia Anticlinorium.

Fig. 30. Strings of limestone nodules in shale formed from a thin bed (parallel to the light-coloured pencil) by deformation along the slaty cleavage (parallel to the dark-coloured pencil); Cardaño Formation, Murcia Anticlinorium.

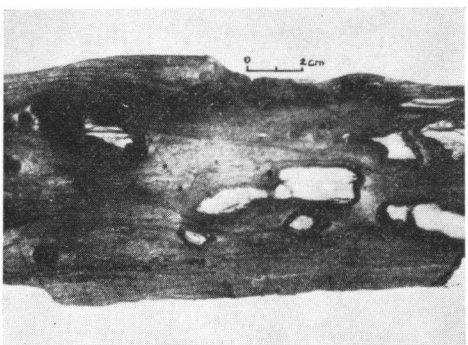

Fig. 31. Limestone fragments boudinaged in a chloritoid mica-schist; Cardaño Formation, Murcia Anticlinorium.

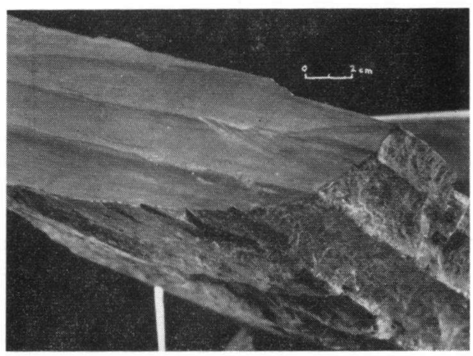

Fig. 32. Rhythmic graded beds with cleavage curving with the grading of each bed; Cervera Formation, Murcia Anticlinorium.

in these shale structures (fig. 29). The bedding can be destroyed and masked at places where the cleavage development is particularly strong. This destruction appears to be facilitated if the thin limestone intercalations occur between relatively thicker shales beds; the original limestone intercalations are found back as strings of small elongated nodules (fig. 30). A clear breaking of the limestone components under severe deformation conditions has been observed in the thin sections; limestone fragments float in the strongly recrystallized shale being mostly real mica-chloritoid schists (fig. 31).

3. Well developed minor folds have been found also in the basal thin bedded sequence of the Murcia Formation at about 5 km west of Cardaño de Arriba. In these folds however a cross cutting cleavage has been found which is only related to the geometry of a part of the minor folds; in many minor folds this cleavage has no

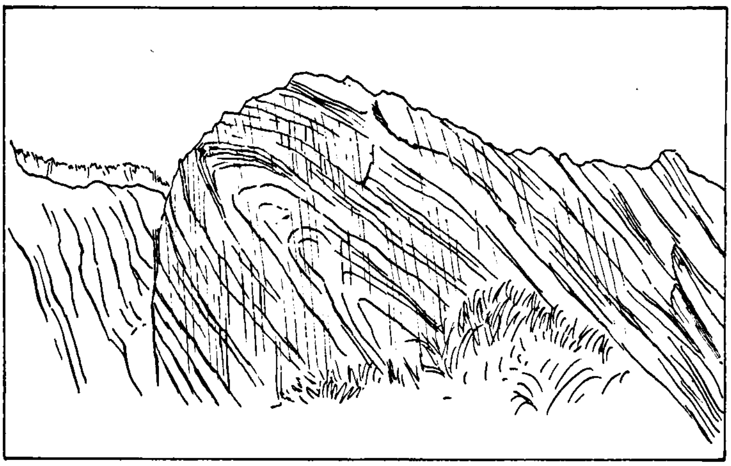

Fig. 33. Isoclinal hinge in Alba Griotte cut by a later throughgoing cleavage; Southern Area, (drawn from photograph, scale 1:50).

relation to their axial plane attitude. It appears that this cleavage developed only in an advanced stage of the minor folding process. The same phenomenon is found in the Alba Formation where isoclinal folded limestone shale beds have been intersected by a later cleavage (fig. 33).

4. A fine graded bed sequence has been found in the strip of rocks of the Cervera Formation north of Cardaño de Arriba. The sequence contains very finegrained thinbedded and mainly argillaceous sediments which resemble varve-type deposits. Cutting of the cleavage through the rhythmically changing lithology in vertical

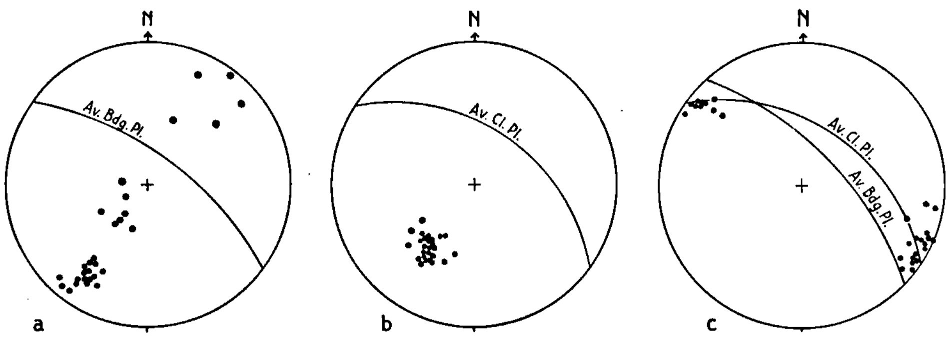

Fig. 34. Stereographic projections (Schmidt Net) of measurements from the Cervera Formation, Murcia Anticlinorium:

a) Bedding plane poles (28); average bedding plane 046/72.
b) Cleavage plane poles (27); average cleavage plane 034/54.
c) Bedding/cleavage intersection (27); Average intersection plunges 05/120.

sense, resulted in very regularly undulated surfaces. These cleavage planes have almost the appearance of that of a rippled sandstone surface (fig. 32); the sharp crests indicate the sharp bottoms of the graded sequences. The cleavage is clearly related to the minor folding in this formation (fig. 34).

5. Beautiful examples of pure axial plane slaty cleavage in tight and complicated minor folds are found in very argillaceous limestone layers of the Cervera Formation. These layers are interbedded with thin and more competent shale intercalations (fig. 35).

We hope that this selection of a few examples of minor folding gives an impression of the strong relation of the type of deformation with both the lithology and their position in the main structure. Finally we like to stress that comparable features are found in all sizes of structure. The same basic phenomena e.g. divergent axial plunges, thrusting of flanks, etc., can be detected in the *major* structures (e.g. cross trends in the Southern Area) in the *minor* structures (e.g. cross trends in Gustalapiedra and Lechada beds), and in the *micro* structures of handspecimen dimensions. Fig. 36 and 24 show an impressive example of the thrust development in a 'micro mountain' section.

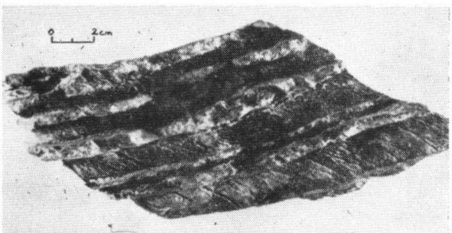

Fig. 35. Axial plane, 'slaty' cleavage in limestone bands between uncleaved shale; Cervera Formation, Murcia Anticlinorium.

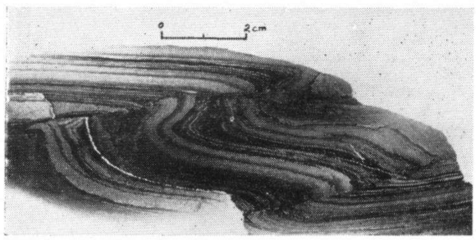

Fig. 36. Overthrust microfolds; Gustalapiedra Formation, Murcia Anticlinorium.

Aguasalio Synclinorium

The Agusalio synclinorium is marked by a long WNW-ESE strip of Yuso deposits forming a topographic ridge in the Central Area. The structure has been named after Aguasalio Mountain about 5 km west of Cardaño de Arriba. Here the massive Curavacas Conglomerates are relatively gently folded but around Aguasalio Mountain itself thrustfaults cut into them forming a type of imbricate structure although much coarser than that in the Murcia Quartzite of Murcia Mountain (Section D). The Curavacas Conglomerates clearly acted as competent slabs some of which may have moved laterally with respect to one another (e.g. just southeast of Cardaño de Arriba), but in such flat-bottomed structures axial trends are not very informative. The deep N-S valley through Cardaño de Arriba exposes the underlying Devonian rocks which show a considerable disharmony of structure with that of the Curavacas Conglomerates, having at least an anticline and two synclines within the limits of the broad Aguasalio syncline in the conglomerates. The Yuso beds transgress onto a number of older formations including the Cervera, Caliza de Montaña, Alba, Vegamián, Vidrieros and Murcia but no convincing outcrops of a discordance could be found and the attitude of the beds near the contacts never allows any strong angular unconformity to be present. This type of disconformity seems most probably to have resulted from local warping and faulting accompanying the movements effecting the subsidence of the Carboniferous basins (chapter 4). At all events there is no positive evidence for a strong folding phase immediately preceding Yuso deposition here.

The north flank of the Aguasalio synclinorium has been mostly overridden by the thrusted Murcia anticlinorium. The south flank is a complicated strip of mainly pre-Yuso rocks largely exposed in Rasa Mountain. The south flank is basically an overthrust zone along the Rasa fault system, made up of Devonian rocks now resting upon the Carboniferous of the Southern Area. This Carboniferous is of an unusual lithology for this region being of the Caliza de Montaña facies, and it seems that this structure has developed over the fundamental feature that formed the facies boundary, the Cardaño Line.

Both the Murcia and Rasa fault systems reflect the trend of the Cardaño Line and are therefore considered as higher level expressions of this fundamental tectonic structure. These two faults converge and die out in the area of Santibañez de Resoba at the convergence of the León and Cardaño Lines.

6. ARAUZ AREA

This large area forms the eastern segment of the Cardaño block separated from the western part by the Curavacas fault; to the north, east and south it extends to the limits of the block along the Peña Prieta Line, the Polentinos Fault and the Cardaño Line, respectively. At present the majority of the area is underlain by Devonian rocks but sufficient remnants of the Curavacas Conglomerates are to be found to suggest that this formation probably extended over most of the area. The largest outlier of the Curavacas Conglomerates occurs in the southeast in the Los Cintos syncline, together with a small amount of older Carboniferous rocks. This structure is flanked by mainly Upper Devonian rocks whereas only older Devonian formations are to be seen further to the north. As has been argued above (Chapt. 4) it is considered probable that this has been brought about by tilting and bevelling of the block before any extensive folding took place. The preservation of the Curavacas Conglomerate outlier in the south suggests the persistence of a southerly component in the tilting movement at least through the time of deposition of the Curavacas Conglomerates and possibly even later.

The major folds, often strongly thrusted and overturned to the south, trend basically E-W and are considered to have been formed mainly during the Asturian folding phase of the Hercynian orogeny. All the complex anticlinoria and synclinoria show a distinct plunge to the west must which clearly be due to a westerly component in the block tilting. This corresponds well to the evidence found in the western half of the Cardaño block where the palaeogeology of the pre-Yuso surface and the pattern of sedimentation in the deposits of the Yuso Group suggest a similar sense of movement (Chapt. 2). This plunge of the structures has allowed recent erosion to expose the deeper complexly folded cores of the anticlinoria to the east and has preserved the equally complicated cores of the synclines to the west. Some of these complications are the usual type of parasitic fold clearly related to the development of the major fold, others with oblique trends can be clearly related to the interference by the deep-seated structures along the edges of the block and some, seem to be entirely disharmonic developing in a diapiric fashion in response to extreme pressures in the cores of some structures.

A slaty cleavage is often seen in the shaly parts of the Murcia, Gustalapiedra, Abadia and Carazo Formations and even in the limestones of the Vidrieros and Cardaño Formations. This can most frequently be seen to be an axial plane cleavage to many subsidiary folds but as these are so often parallel to the major structure the cleavage is equally parallel to their axial planes as well. This cleavage never cuts all the rock units of a structure so that the folds are all essentially concentric or mixed except for the diapiric ones.

Rio Frio Synclinorium

The Rio Frio synclinorium is made up of the lowermost Devonian, Lebanza and Carazo Formations. It is about 7 km long and 1 km wide running E-W between the Rio Frio fault and the northern scarp of the Curavacas Conglomerates. This is essentially east of the line of the Curavacas fault so that the structure really belongs to the western part of the Cardaño block. However, the rock types and the style of folding are so similar to those of the Arauz Area proper that they will be described with them. The whole of the structure cannot be seen because the unconformable Curavacas conglomerates rest upon the Lebanza Formation of the south flank 1 km south of Peña Prieta and gradually cut across to the lowermost Member a of the Carazo Formation near the Curavacas fault to the east. It seems fairly clear that the pre-Yuso unconformity here cuts across a prexisting structure formed earlier than the Asturian phase. However, the Curavacas Conglomerates now dip about 50° to the south so that at least some post-Yuso deformation must be postulated.

Within the Río Frio synclinorium the Carazo and Lebanza Formations show very intricate subsidiary folds. Near Peña Prieta in the west, these seem to be aligned along and between small oblique, NE-SW faults which are splays of the main Río Frio fault. The Carazo quartzites rarely show more than gentle warping but they are broken by many small faults along some of which acid dykes have been intruded. These faults allow the quartzites to take part in the deformation which has a very complex form in the overlying shales and the limestones of the Lebanza Formation. Some seem to demonstrate the plastic role of the shales in forming diapiric-like folds. Towards the east the structures tend to run more E-W parallel to the Río Frio fault and are perhaps more regularly developed. However, the abrupt termination of some as well as other anomolous trends may well be due to the same mechanism.

The origin of this fold seems to be closely associated with the Río Frio Fault which forms the expression of the deep-seated Peña Prieta Line here. It may be that the pre-Yuso folding that took place here was purely local, related to the tectonic activity of the line. Whatever the date of the major part of the folding there can be no doubt that there has been a very strong influence exerted by this deep-seated structure.

The various structures are usually markedly restricted in the groups of rocks which they involve. These individual groups have developed their own structural styles clearly related to the nature of the lithologic sequence and there can be little doubt that many of the anomolous features of the folding patterns may have their origin in variations within the rock layers themselves. Consequently we see, on a different scale and in very much greater complexity, the control of the lithology upon the type of tectonic deformation.

Bistruey Synclinorium

The Bistruey synclinorium is mainly formed in rocks of the Lebanza and Abadia Formations with some rocks of the Carazo Formation involved in places. This complex of intricate structures plunges for about 6 km west from the northern end of the Polentinos fault to the line of the Curavacas fault; having a total width of about 2 km between the Río Frio fault to the north and the Coto Redondo anticlinorium to the south. The structure as a whole is overturned to the south although it may be said to owe its form more to the large number of subsidiary folds of about 0.7 km wavelength than to any large scale fold. In the west these structures are thrust upon the north flank of the Coto Redondo anticlinorium although this fault only brings Lebanza Formation into contact with the lower Member c of the Carazo Formation to the south, which would imply that the thrust has a lower dip than the bedding.

A very large number of tightly compressed subsidiary folds are present here and their number and complexity increases down the plunge to the west. In the area of Peña Bisruey they are isoclinal, overturned and overthrust upon one another towards the south. Still further west near the line of the Curavacas fault divergent WNW and ENE axial directions occur. These all seem to plunge to the east although that would only be expected on ENE folds subsidiary to a north-dipping isoclinal structure. Despite the extraordinary amount of compression no obvious cleavage has been noted.

A small area comprising the same Devonian Formations in very similar folds to those of Peña Bistruey occurs along the strike to the west of the line of the Curavacas fault in El Tejo Mountain. Although they are very similar in style and order to the folds to the east direct connection could be established between the two fold systems and a zone of disturbance corresponding to an extension of the Curavacas fault must pass through here.

The formations are considerably thinner here; the quartzites of Member b of the Carazo Formation are only 20 m thick and they have clearly only influenced the smallest folds. Hence the folding style has been dictated by the limestone-sandstone-shale sequence of the Lebanza and Abadia Formations. As is so common in the Arauz Area folds in these formations are extraordinarily irregular along the strike and crosstrends commonly develop. It may be that the very strong deformation here and the relatively consistent trends can be explained by the presence of the Río Frio fault immediately to the north which has curved to an almost E-W trend here.

Coto Redondo Anticlinorium

The Coto Redondo anticlinorium comprises mainly members of the Carazo Formation with, however, some rocks of the Lebanza Formation taking part in the easternmost extension. This structure trends basically E-W although the development of crosstrends mean that its largest dimension of 8 km is in a WNW-ESE direction. Divergent trends are to be seen to the west near the Curavacas fault and, much stronger, in the east near the Polentinos fault. Here the structure attains a width of 3 km whereas in its central zone it is only 0.7 km wide. The western part of the Coto Redondo anticlinorium is overturned towards the south and thrust over the north flank of the Cortes synclinorium (Section B). In the east the asymmetry is merely shown by the steeper south dips (Section A). The overall structure tends to plunge to the west but in the east a large crestal syncline plunging eastwards brings down rocks of the Lebanza Formation in which the complex cross-folding near the Polentinos fault are developed.

The Carazo Formation has thinned considerably here; the quartzites of Member b are only about 50 m thick in contrast to the 200 m in the Carazo anticlinorium. They are also more shaly but since the other two members of the formation have also thinned in proportion it seems quite probable that the quartzites still form the most competent horizon during the deformation of this structure. This would account for the much shorter wavelength of this structure in contrast to that of the Carazo Anticlinorium.

The subsidiary cross-folds near the Polentinos fault seem to have developed in the flanks of the eastward plunging, crestal syncline consequently the axes show trends almost perpendicular to one another on the map. These folds are so small that they cannot involve any other beds than the thin Carazo and Lebanza sequence seen here. The widening of the structure towards the west has also been caused by the development of a crestal syncline although this is much smaller and gentler than that to the east. Ths structure is not clearly terminated by the Curavacas fault and the Carazo Formation continues to the west forming part of the Río Frio synclinorium but there does not seen to be any continuity between the two structures.

A slaty cleavage is commonly seen in the shaly lower Member a of the Carazo Formation. It forms an axial plane cleavage to some of the subsidiary folds.

Cortes Synclinorium

The Cortes synclinorium is mainly made up of rocks of the Lebanza and Abadia Formations although some of the Carazo Formation is involved in the southern, thrusted flank. The structure is about 9 km long overall but its maximum width, in the centre of the block, reaches only 2 km (fig. 16), and it pinches out entirely to the east. This is due to the westward plunge so that as the axis pitches upwards the higher, Abadia Formation is lost. The axial direction also deviates slightly from the general E-W trend to the ESE against the curving Polentinos fault. The quartzites of the Carazo Formation do not continue much west of the Arauz river but the structure to the west includes a great deal of the Abadia Formation brought down by the westerly plunge. The major structure is more or less symmetrical in the east (Section A) although the strong thrusting to the south illustrates the dynamic asymmetry of the deformation. Towards the west the structure becomes isoclinal, overturned towards the south and still thrusted, although the swing of the axis to the SW has allowed the main thrust to cut into the core of the synclinorium. (Section B).

The Lebanza Formation in the eastern nose has crumpled into a few subsidiary folds between 200 and 500 metres in wavelength. They are either symmetrical, like the major structure, or show the typical parasitic asymmetry (Section A). Many more minor folds develop to the west especially in the Abadia Formation. They tend to have an isoclinal form like that of the major structure here and are very much smaller, of wavelengths of less than 100 m, so that they can only be represented schematically in the cross-section (Section B). All these minor folds have an axial plane, slaty cleavage developed in the shaly layers and this cleavage is also parallel to the axial planes of many even-smaller minor folds that have formed in individual thin beds of limestone as parasitic folds on the flanks of the larger subsidiary folds. Many slaty cleavage planes dip steeply south but this impression may be overemphasised by the fact that the south flank is much better displayed. Usual fanning of the axial plane, slaty cleavage (upwards into the crest of anticlines) would tend to produce this attitude in the south flank. In the west where the major axis curves to the southwest the majority of the subsidiary folds still trend E-W so that they cut obliquely across the flanks. Steeply plunging and intricately interfering folds have been seen in places and the full, detailed picture must be extremely complicated.

The folds displayed in the Cortes synclinorium have been clearly controlled in their development by the characteristics of the Lebanza and Abadia Formations.

This is all the more emphasised by the anomolous behaviour of the Carazo Formation which is only found along the southern thrust of this structure. The development of cleavage and subsidiary folds has been clearly dependant upon the basic pattern of the major structure as well as the characteristics of the smallest lithological units.

Carazo Anticlinorium

The Carazo anticlinorium is mainly formed by rocks of the Carazo Formation the distinctive members of which have allowed its intricate form to be mapped. As a whole the major structure trends E-W plunges and to the west. There is a slight deviation to the WNW as the axis approaches the Curavacas fault to the west and a marked subsidiary ENE cross-trend at Carazo Mountain in the centre. The structure can be traced more than 10 km along the axis and varies from 2 to 3 km in width. It is overturned strongly to the south and thrust over the north flank of the Polentinos anticlinorium along the important Lebanza thrust fault so that the upper Member c of the Carazo Formation rests upside down upon the Murcia Formation which is here the right way up (Section A).

The extreme western nose of the Carazo anticlinorium shows a decided swing near the Curavacas fault suggesting that this fault has influenced the structure and may separate it from any continuation to the west below the Curavacas Conglomerates. In the east the structure is practically dominated by the development of a very large subsidiary axial syncline. In Carazo Mountain the syncline has a very flat axial zone (Section A) which is seen in the the large slab of the quartzites of the middle Member b cropping out in the western slopes of the mountain. The axis here is cut by a NW-SE trending fault that probably has very little throw but which has detached the rocks to the west allowing them to slide down the mountainside, in a most spectacular manner. The resulting fault scarp terminates this part of the structure with a curious oblique, angular outcrop very striking on the geological map. To the east this syncline is cut another fault trending NE-SW and the folds further to the east have a considerably different form. To the WSW the axis of the syncline can be traced plunging into the south flank where a small hinge can be seen in the inverted quartzites of the southern limb.

The lowest Member a of the Carazo Formation exposed in the core of the structure has been crumpled into many minor folds often of a wavelength of about 50 m. These folds have axial planes parallel with that of the major structure and the shaly beds have developed a slaty cleavage parallel to them. The cleavage is not absolutely parallel and often shows the textbook type of fanning diverging upwards in the crests of anticlines (fig. 20a).

The style of this structure suggests strongly that middle Member b of the Carazo Formation which is here more than 200 m thick has been the competent horizon controlling the resultant form.

Polentinos Anticlinorium

The Polentinos anticlinorium is the only major structure of this area to involve the full sequence of Devonian rocks and it is very striking that the various groups retain their individual structural styles within this single structure. This anticlinorium plunges in general to the west over a distance of about 13 km and yet maintains a fairly constant width of about 2.5 km. The plunge varies from rather steep in the east where the oldest rocks are displayed in the core to almost horizontal in the western half where the thin Upper Devonian covers nearly 5 km of the axis. As can be seen in Sections A and B the Polentinos anticlinorium is asymmetric overturned to the south and thrust against and over the north flank of the Los Cintos syncline. The even more important Lebanza thrust (de Sitter 1957) brings the overturned south flank of the Corazo anticlinorium onto the north flank of the Polentinos anticlinorium.

In the western part of the Polentinos anticlinorium the Murcia Quartzite has clearly been the most competent bed involved in the folding as in the Central Area to the west. Cross-section B shows that the amplitude of the major fold is relatively small and that the main deformation has been effected by the formation of many subsidiary folds. Many of these have wavelengths of only about 100 m so that they cannot be accurately presented in the cross-sections but they show the typical parasitic form with reference to the major structure as is shown schematically. Such small folds of near-isoclinal form could not involve all of the Murcia Formation which may be about 80 m thick. It is quite clear only the uppermost, more quartzitic 30 m of the formation have been involved (fig. 37) and that the lower more shaly part of the formation has been detached. Similarly these folds are detached upwards by gliding in the nodular limestones of the Vidrieros Formation. The detachment of these folds has lead to the development of a number of small thrusts, some of which it has been possible to map.

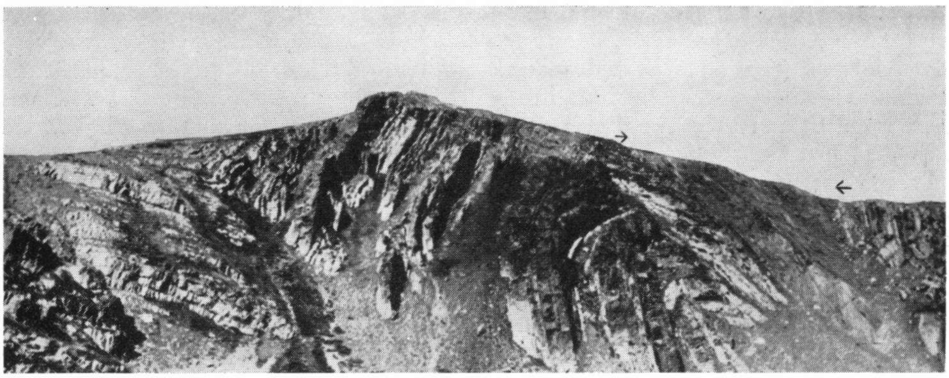

Fig. 37. Folds in Murcia Quartzite and Vidrieros Formation (between arrows), Monte Las Huelgas, Polentinos Anticlinorium, looking west.

In the south flank of the Polentinos anticlinorium west of the Río Carrion the subsidiary folds have the same asymmetry to the south but are much gentler with wavelengths up to 700 m. As the north flank is traced to the east of the Río Carrion the subsidiary folds there also become simpler and the thrusts die out. Here the folds in the competent quartzites are of an accordion type (fig. 37) though still showing the asymmetry typical for folds parasitic to the main structure. These folds probably involve even less of the total section of the Murcia Formation than the isoclinal folds to the west. They clearly owe their origin to the excellent parallel bedding of the quartzites and the shale partings between them. The Murcia Formation is absent from the eastern half of the Polentinos anticlinorium being cut off by the Lebanza thrust on the north flank and unconformably overlapped by the Curavacas Conglomerate to the south.

The core of the structure in the valley of El Clero, east of the Río Carrion the Cardaño Formation outlines a very regular, cylindrical major fold with only a very gentle plunge to the west. The south flank is still inverted and there are a considerable number of subsidiary folds in the Cardaño, Gustalapiedra and Abadia Formations here (Section A). These structures show the same asymmetry, overturned to the south, as the major structure and the development of even more complicated forms in the thin sandstone and limestone layers within the mainly shaly Abadia Formation, with wavelengths of only about 50 m.

Still further east where the Middle and Lower Devonian rocks form most of the structure the Polentinos Limestone Member takès part in structures equally as complicated as those of the Abadia. It seems that here, as elsewhere in the Arauz Area the shales of this part of the sequence have acted as a hydraulic medium causing the development of diapiric type structures.

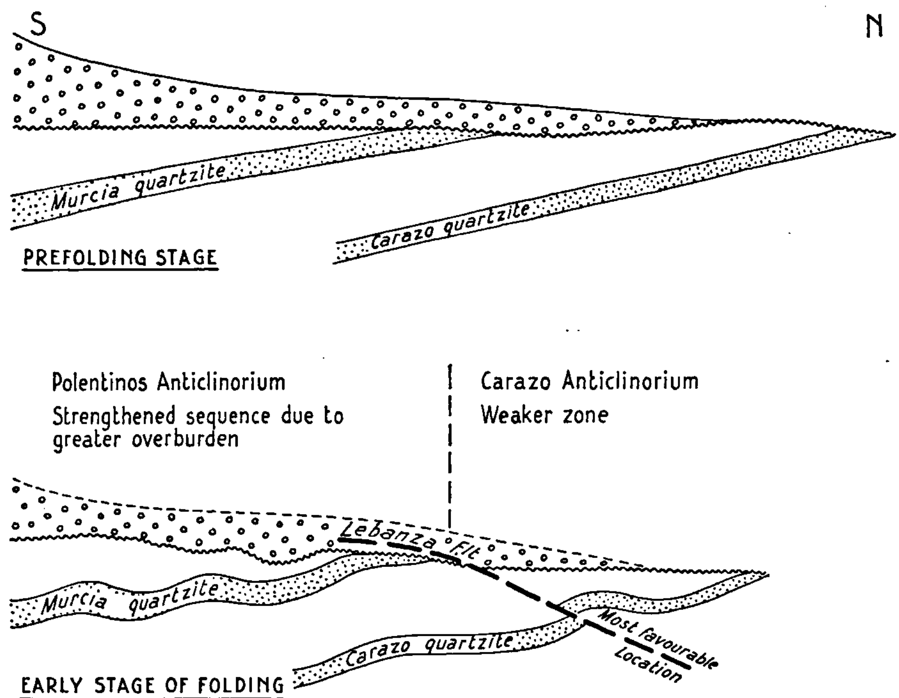

Fig. 38. The development of the Lebanza thrust at the pre-Yuso unconformity.

Cleavage is very common in the Polentinos anticlinorium, slaty cleavage occurring in the shales of the Murcia, Gustalapiedra and Abadia Formations. Since the major and minor structures are essentially parallel the cleavage is axial plane to both equally. The type of cleavage in the nodular limestones of the Cardaño and Vidrieros Formations is exactly the same as that described in more detail in the Central Area. The absence of cleavage from the very strongly deformed shales of the Carazo Formation which were faulted rather than folded seems to point again to the fluid role of these shales in the deformation here, meaning that no directional stress could be set up in them that would be capable of creating a cleavage.

It seems quite probable that the wedging out of the Murcia Quartzite on both flanks of the major structure at about the same longitude has a single cause. This is obviously pre-Curavacas erosion to the south so that it seems possible that the Lebanza thrust has taken advantage of an old erosion surface to ride up on (fig. 38).

This feature would fit the picture already built up of the bevelling of the tilting Cardaño block in the western part. It could well be that the rapid increase in plunge of the axis in the eastern half of the Polentinos anticlinorium may have been induced by the absence of the competent Murcia Quartzites from this part of the structure.

Los Cintos Syncline

The Los Cintos syncline basically consists of Curavacas Conglomerates forming their easternmost outlier resting unconformably upon older Carboniferous and almost the whole sequence of Devonian rocks. The structure stretches ESE out of the present area over a total of 12 km and reaches a maximum width of 4 km. In the map area the structure plunges down to the ESE between the previously described Polentinos and Murcia Anticlinoria. As has been noted above the Curavacas Conglomerates on the flanks of this structure continue their unconformable overstep over the underlying sequence and on the north flank procede regularly from the Upper Devonian, Murcia Quartzites as far as the Lower Devonian, Abadia Formation (Frets 1965). The detailed structure of the conglomerates has not been studied but is seems that they form a similar, if flatter fold, to one that must be present in the Devonian rocks below making the connection between the Polentinos and Murcia Anticlinoria.

7. SOUTHERN AREA

The Southern Area comprises the belt of thick Carboniferous limestones and unconformable clastics with small Devonian inliers stretching 20 km along the whole whole of the SW edge of the present area. The unusual facies of the Carboniferous, for this region, the Caliza de Montaña, is also accompanied by an unusual development of divergent trends. This is well displayed in Espiguete Mountain where the limestones are found up to 2450 m altitude and the usual WNW-ESE parallel to the structures of the Central and Northern Areas is strikingly interrupted by NNW-SSE crossfolds, marked by anticlinal cores of Devonian rocks shown on the geological map. To the east synclines of these crossfolds include cores of the unconformable Cervera and Curavacas Formations indicating that the crossfolding at least is post-Yuso in age.

The Southern Area is situated on the Cardaño Line and as has been described above (Chapt. 2) many different lithologies (shales, limestones, quartzites, conglomerates, etc.), are to be related to the tectonic activity during sedimentation. The various formations are extremely variable in thickness although rather difficult to show exactly due the complicated structures. The León Line which approaches the Cardaño Line near Triollo appears to have acted mainly as a positive feature during Yuso times (Helmig, 1965). Hence the Southern Area forming part of the westward-tilted Cardaño block may have been pushed partly over the León high by the strong Asturian forces. This may have facilitated local yielding of parts of the Southern Area leading to the formation and intensification of cross trends (fig. 39).

The rapid lateral changes in the sequence may well be responsible for the development of synchronous cross trends as has been worked out by Bhattacharji (1958). A clear example of this can be seen near Barniedo de la Reina where the massive Curavacas Conglomerates pinch out. This relatively weak zone has allowed the Lechada syncline to advance and ride over the west-plunging Devonian structures. This type of deformation may also have been facilitated by movements along

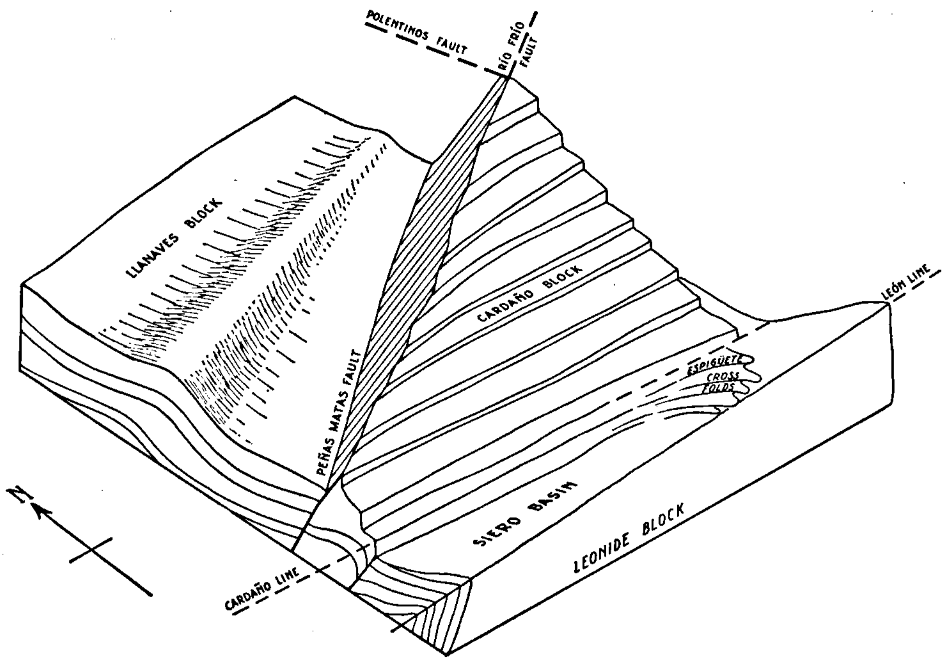

Fig. 39. Block diagram of the relations of the controlling tectonic lines and the Cardaño block.

the Peñas Matas fault. In any case it seems clear that a combination of lithologic and tectonic factors is here involved in the creation of cross trends in the subsequent structures. Hence the present author is strongly inclined to relate the crossing structural trends of this area to prexisting lithologic and tectonic framework.

8. DATING OF THE STRUCTURAL HISTORY

The structural pattern of the Cantabrian Mountains has been shaped during several folding phases which have been grouped by de Sitter (1962a) as below:

4. Saalic phase; post-Stephanian — pre-Triassic; trend E-W
3. Asturic phase; post-Westphalian — pre-Stephanian; variable trends
2. Sudetic phase; post-Westphalian A — pre-Westphalian B; trend E-W
1. Bretonic phase; Late Devonian uplift and erosion.

1. In the present area we find little evidence of changes within the Devonian sequence that might record the movements corresponding to the Bretonic phase. The Murcia Quartzites thicken and become much coarser towards the east, even developing a turbidite facies that might be taken to indicate instable conditions. It also seems probable that a post-Devonian hiatus or even erosive period took place before the Carboniferous. This follows from the fact that the Vegamián Formation can be seen resting directly upon the Murcia Quartzite at certain places.

The overlap by all of the younger formations onto the Murcia Quartzite at one place or another may have been at least facilitated by the pre-Carboniferous nonconformity. However, there is considerable evidence that discordances exist between all of Vegamián, Alba, Caliza de Montaña and Cervera Formations. Yet many of the contacts seen in the field are clearly gradational so that it seems most likely any movements at these times were essentially local. The type of deposition indicated by these rocks mostly took place under relatively quiet conditions, except for the turbidites and Triollo Limestone Conglomerate Member of the Cervera Formation. These indicate considerable instability; the limestone conglomerates being specific for local, active erosion. Some of these movements may well be regarded as precursors of the following Curavacas phase.

2. The Curavacas phase has been named here for the folding that preceded the deposition of the Curavacas conglomerates (Kanis 1956). Some clear unconformable contacts have been recorded but it has not been possible to map a major structure clearly unconformably cut off by this formation. The two best examples are: (a) the Río Frio synclinorium where the Curavacas Conglomerates overstep from the Lebanza Formation onto the basal Member a of the Carazo Formation and (b) the south flank of the Los Cintos syncline where the conglomerates cut down from the Cervera Formation across the Caliza de Montaña, Alba and Vegamián onto the Murcia Quartzite. These two localities, it should be noted however, are on or near one of the deep-seated structural lines—the Peña Prieta and Cardaño, respectively.

The direct assumption that these structures are wholly pre-Yuso involves considerable difficulties. The structures are continuous with many others, especially Los Cintos with the Central and Southern Areas, where the Curavacas Conglomerates are to be found taking intimate part with the same folds. This paradox can certainly only be resolved by assuming two periods of movements. The evidence of the Los Cintos syncline is of the overstep of a fairly gentle structure, while that of the Northern Area shows an even lower angle of discordance. Yet the Aguasalio synclinorium and the folds and cross-folds of the Southern Area demonstrate that the Curavacas conglomerates took part in the isoclinal folding here. This leads to the conclusion that pre-Yuso movements were relatively gentle, possibly block movements with some accentuated deformation along important hinge lines. This matches the evidence from throughout the Asturides (de Sitter 1962b).

3. The Asturian phase cannot be positively identified because no deposits of post-Westphalian age are found in the area. Still the strong post-Yuso deformation seen is clearly pre-Triassic from regional data (de Sitter 1962b). The severity of the deformation in the Cea Formation does not seem comparable with what is seen here but the nearest deposits are quite some distance away.

4. The Saalic phase seems even more unlikely to have had any influence upon the structures of this area following the analysis of its effects on the Cea Formation given by Helmig (1965).

This area shows very clearly the southward transport of the rocks during orogeny. This is typical for the Asturides and in direct opposition to the direction of transport in the Leonides (de Sitter 1962b). It is possible that the difference may lie in a difference in the time of folding so that the orogeny took place later in the Asturides than in the Leonides.

It is considered that no satisfactory evidence for the precise dating of folding phases has been found in this area. The main difficulty lies in the continued instability throughout Carboniferous times, that can be seen in the sedimentary pattern and which can be related to movements along certain deepseated structural lines. The correlation of structural style and trends with the rock sequence and its lateral variations do seem to be very clear. This correlation does not help structural analysis very much for it means that individual styles and trends may no longer be regarded as unique indicators for a single folding phase.

The rocks of the Cardaño Area have been laid down in a relatively active area, which gave rise to a number of variations in the sequence. Subsequent tectonic activity has brought about structures of many different styles, sizes and trends. Despite all these differences the present author considers that the simplest and most likely explanation is that they have been caused by a single compressive orogeny, probably during the Asturian phase.

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