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GEOLOGY OF THE AREA BETWEEN THE LUNA AND TORIO RIVERS, SOUTHERN CANTABRIAN MOUNTAINS, NW SPAIN

BY

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ABSTRACT

Precambrian clastic rocks, deposited under unstable conditions, were folded before in a relatively stable environment shallow marine sedimentation spread out over the whole area. Silico-clastic sediments were deposited from Cambrian to Devonian, except from the Lower-Middle Cambrian when carbonate deposition dominated.

With a hiatus in sedimentation during the Llanvirn to Llandovery the influence of a rising block, NNE of the present area, started. During the Silurian this rise resulted in development of clastic sequences trending to thin towards the N.

From the Devonian to Upper Carboniferous sedimentation circumstances became less stable. As a result an alternation of clastic and carbonate rocks developed.

Towards the end of the Devonian epeirogenetic uplift and tilting of the northern part of the area resulted in strong erosion and consequently the uppermost transgressive Devonian sandstone rests on a variety of older deposits.

The Sabero-Gordón line separates the uplifted area in the north from the area where continued subsidence and sedimentation took place during the Upper Devonian.

During the Lower Carboniferous differences in sedimentation circumstances were strongly reduced resulting in the deposition of the Alba Formation all over the area.

During the Namurian the Sabero-Gordón line renewed its function as a facies boundary between a northern and a southern area.

Together with the development of the progress of the maximal Carboniferous sedimentation towards the north the initial folding of the Hercynian orogenesis started south of it. After the orogenesis oblong coal basins developed during the Stephanian B along normal faults approximately parallel to the strike of the folding.

After folding of these coal basins a long period of non-deposition followed which ended in the Upper Cretaceous when sedimentation took place along the southern border of the folded Palaeozoic. The Tertiary morphogenetic uplift of the Cantabrian Mountains is accompanied by continental deposits forming the border of the Duero basin.

During the Hercynian orogenesis major deformation took place in the Leonides (Fig. 3). The Sabero-Gordón line separates the Leonides in a strongly folded area in the south and an area with thrust sheets north of it. The shape of folds and thrusts is mainly determined by the lithological properties of the Palaeozoic rocks. Table 2 shows the rocks units which are supposed to have their own tectonic-style.

In the southern area (Alba synclinorium) minor folding is an important feature. Based on a symmetry-concept most of these folds are parasitic folds. In some places minor folds in the folded area as well as in the thrust area show that deformation took place by gravity-stress.

The León line separating the Leonides from the Asturides seems to have no significance as a fundamental structural line in this area.

RESUMEN

Depósitos clásticos del Precámbrico, depositados en circunstancias inestables, se han plegado antes de que se esparcieran sedimentos marinos por toda la región. Sedimentos siliciclásticos fueron depositado durante el Cámbrico hasta el Devónico, a excepción del periodo entre el Cámbrico inferior y el Cámbrico medio en el cual dominaba la sedimentación de carbonatos.

La influencia de un levantamiento empezó con un hiato en la sedimentación durante el Llanvirniense y el Llandoveriense y dejando secuencias de menos espesor hacia el norte durante el Silurico.

Del Devónico hasta el Carbonífero superior las circunstancias de sedimentación eran menos estables, resultando en una alternación de rocas clásticos y rocas carbonatos.

Al final del Devónico el levantamiento epirogénico y la inclinación general de la parte del norte de la región originaron una fuerte erosión y como consecuencia los areniscos transgresivos del Devónico superior se han depositado sobre varios yacimientos de una edad más antigua.

La falla de Sabero-Gordón separa la región de levantamiento en el norte de una región de subsidencia y sedimentación continua en el

Durante el Carbonífero inferior las diferencias en circunstancias de sedimentación desaparecieron y se formó en toda la región la Formación de Alba.

Durante el Namuriense la falla de Sabero-Gordón se activó como limite entre distintos facies en el norte y el sur.

Al mismo tiempo en que la sedimentación máxima de depósitos carboníferos progresó hacia el norte, el plegamiento inicial de la orogénesis herciniana empezó al sur. Se desarollaron después de la orogénesis, durante el Estefaniense B, cuencas oblongas junto a fallas normales aproximadamente paralelas a la dirección del sistema orogénico.

Despues del plegamiento de estas cuencas terminó un periodo largo sin sedimentación con la sedimentación del Cretácico superior a lo largo de la zona de depósitos plegados del Paleozóico. El levantamiento morfogenético durante el Terciario acompañó la formación de depósitos continentales en la orilla de la Cuenca del Duero.

Durante la orogénesis herciniana la mayor deformación tuvo lugar en los Leonides (Fig. 3). La falla de Sabero-Gordón separa dentro de los Leonides una región de plegamiento fuerte en el sur de una región de corrimientos en el norte. La forma de pliegues y corrimientos se han determinado principálmente por las propiedades litológicas de los depósitos paleozóicos. Tabla 2 indica las unidades de tipos de rocas, cada una de las cuales tiene su estilo tectónico propio.

En la región de plegamiento (synclinorio de Alba) un segundo plegamiento es de importancia. Los pliegues son parasíticos cuando se les considera a base de propiedades simétricas perteneciendo a las estructuras primarias. En varias localidades de los Leonides los pliegues secondarios indican unas deformaciónes causadas por presión de gravitación.

La falla de León, separando los Leonides de los Asturides, no tiene importancia ninguna como falla fundamental en esta región.

CONTENTS

CHAPTER I

INTRODUCTION

General remarks

The present map is the result of a geological study of part of the southern slope of the Cantabrian Mountains, carried out as part of a programme of systematic mapping by the Department of Structural Geology of the University of Leiden. Evers (1967) and van den Bosch (1969) published the maps east and west of the present area. The surrounding areas have also been mapped by others and on various scales, Fig. 1. In 1962 de Sitter published a provisional map (1 : 100,000) of the area between the Pisuerga and Luna rivers. The present area is situated in the western part of that map in the province of León and the province of Oviedo. Its geographical limitations are: Latitudes 42°47′01″ and 43°01′38″N, longitudes 2°11′48″ and 1°50′00″W of Madrid, which correspond with ca 5°18′ and 5°41′W of Greenwich.

Fieldwork was carried out on topographic maps 1 : 25,000 enlarged and simplified from topographic maps (1 : 50,000) of the Instituto Geográfico y Catastral, Madrid. The present area is covered by the following sheets: 77 (La Plaza), 78 (La Pola de Lena), 102 (Barrios de Luna), 103 (La Pola de Gordón), 128 (Riello) and 129 (La Robla).

Mapping was facilitated by the availability of aereal photographs. The area is covered by seven runs of photographs with the following numbers: 45260–45271, 52976–52964, 43178–43167, 43330–43338, 43512–43504, 43722–43713 and 53596–53589. The northern area of Carboniferous rocks has been mainly interpreted from these aerial photographs. Field mapping was carried out on a scale of 1 : 25,000 and the photo-interpretation was adjusted to this scale with a 'Rocks Sketchmaster'.

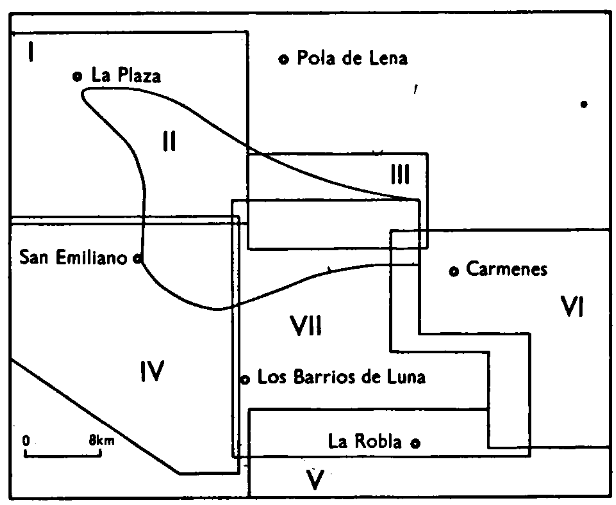

Fig. 1. Map showing existing geological maps in the SW part of the Cantabrian Mountains on different scales. I. García-Fuente (1959); II Marcos (1968); III Llopis Lladó (1955); IV Van den Bosch (1969); V Pastor Gómez (1963); VI Evers (1967); VII map presented with this thesis.

Hydrological remarks

In the present area the main watershed of the E-W running Cantabrian Mountains separates the province of León from the province of Oviedo. North of it drainage is to the Bay of Biscay by the Valgrande river and further north by the Pajares river. South of the watershed three rivers flow south towards the Meseta, where they feed into the Duero river system. From west to east they are the Luna, Bernesga and Torío rivers. The drainage systems run at approximately right angles to the mountain range. They are separated by N-S running secondary watersheds.

In the Luna river a dam has been built to form a reservoir for the water of the Lake Luna (Embalse de Luna) to be used for the electric plant near Mora de Luna. In the Páramo region on the Meseta the water of this reservoir is used for irrigation purposes.

In the Casares river, an affluent of the Bernesga river, two artificial lakes are planned, one in the Casares Valley, the other near Beberino, flooding the lower course of the Casares river. Both reservoirs are meant to serve as reserves for the electric plant which has been built for industrial purposes in the Bernesga river near La Robla.

Another plan now investigated is the building of an artificial lake in the Torío Valley for which the water of the upper Bernesga river would have to be diverted from Villanueva de la Tercia to the Torío Valley.

Lithology greatly affects the courses of the main rivers. Nevertheless the four principal rivers Luna, Bernesga, Torío and Valgrande are consequent rivers on the southern and northern slopes of the Cantabrian Mountains. The affluents are mainly subsequent rivers, running generally in an E-W direction parallel to the general strike of the folded Palaeozoic.

Most watersheds are rounded except for the areas where limestones are the separating rocks.

The main riversystems are often asymmetric and dendritic, especially in their upper courses.

There is a great difference in length between the Valgrande river and the others. Within 100 km the Valgrande river reaches the Bay of Biscay, whereas the southbound rivers reach the Atlantic Ocean after more than 500 km. These differences in length have caused the upper Bernesga course to be cut off by the Valgrande riversystem. The main watershed became deformed and now shows an E-W split. Llopis Lladó (1954a) already mentioned the capturing of the upper Bernesga, supposing that the original Bernesga had its course up to Telledo at a level of 1300–1400 m. This seems to be an exaggeration.

The Valgrande precipices are steep and have caused a number of small landslides and V-shaped valleys. The south draining rivers generally have rather U-shaped valleys.

Drinking-water problems do not exist in this area since the mountain hills supply abundant water for the villages around.

Remarks on glacial phenomena

North of Celleros a cirque with a moraine bow is shown on the geological map. Throughout the northern area small cirques can be found, many of which have been deformed by further erosion and scree-building.

North of Casares a moraine, mainly composed of Barrios quartzite boulders and alternating cross-bedded sandy shales point to a fluvio-glacial origin. The lowest level of these deposits is about 1,350 m above mean sealevel. A small tributary of the Casares river now has cut off its own accumulation terrace. West of Folledo comparable phenomena have been found. A small river cut through the moraine material, mostly consisting of San Pedro boulders. Furthermore accumulation terraces occur along the Bernesga river and its tributaries towards Rodiezmo. The terraces have a thickness here of 3–7 m.

River terraces also occur along the southern border of the mountains in the Luna, Bernesga and Torío river valleys. Vegetation and agricultural cultivation, however have often obliterated heights and the kinds of terraces actually occurring. Where noticeable they are thinner than 2 m.

Acknowledgments

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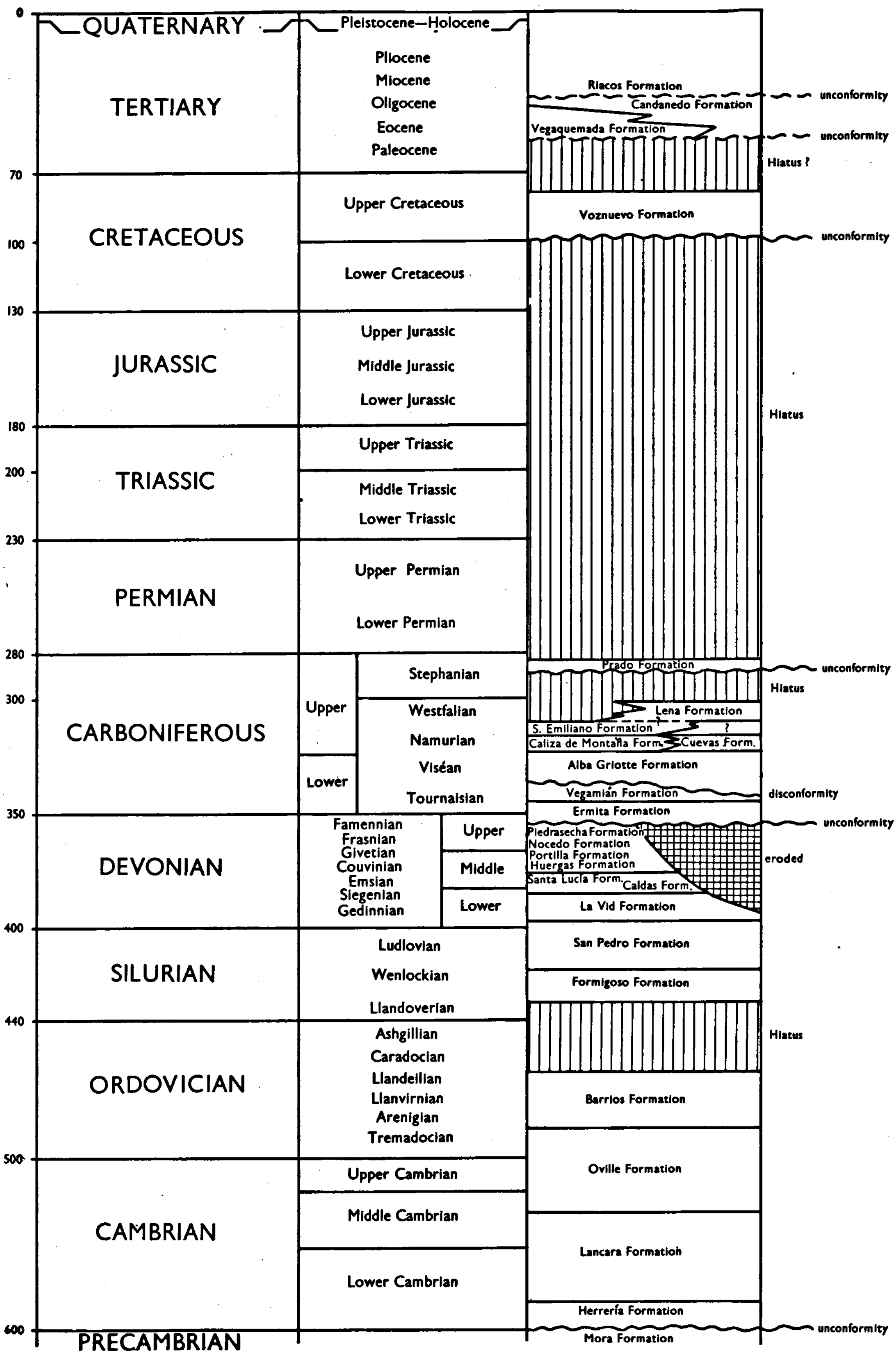

Fig. 2. Correlation between chronostratigraphic units and lithostratigraphic units in the Luna-Torío area.

CHAPTER II

STRATIGRAPHY

INTRODUCTION

The rocks exposed in the area of the present map range in age from Precambrian to Tertiary. These rocks have been subdivided into formations as listed in Fig. 2. The succession does not represent a continuous sedimentation. Angular unconformities, disconformities and hiatuses occur.

The following major breaks in sedimentation have been observed:

a) an angular unconformity between the Mora Formation and the Herrería Formation;
b) a hiatus between the Barrios Formation and the Formigoso Formation. Comte (1959) assumed that during part of the Llanvirnian, the Caradocian, the Ashgillian and part of the Llandoverian no sedimentation took place.
c) an unconformity between the chiefly Famennian Ermita Formation and older Devonian rocks. From south to north the eroded part of the older Devonian rocks increase with a maximum in the Bodón unit (Fig. 3) where the Ermita Formation covers the La Vid Formation which is possibly of Siegenian age there.
d) an angular unconformity between the Stephanian rocks of the Prado Formation and older Palaeozoic rocks.
e) an unconformity between the Upper Cretaceous Voz-

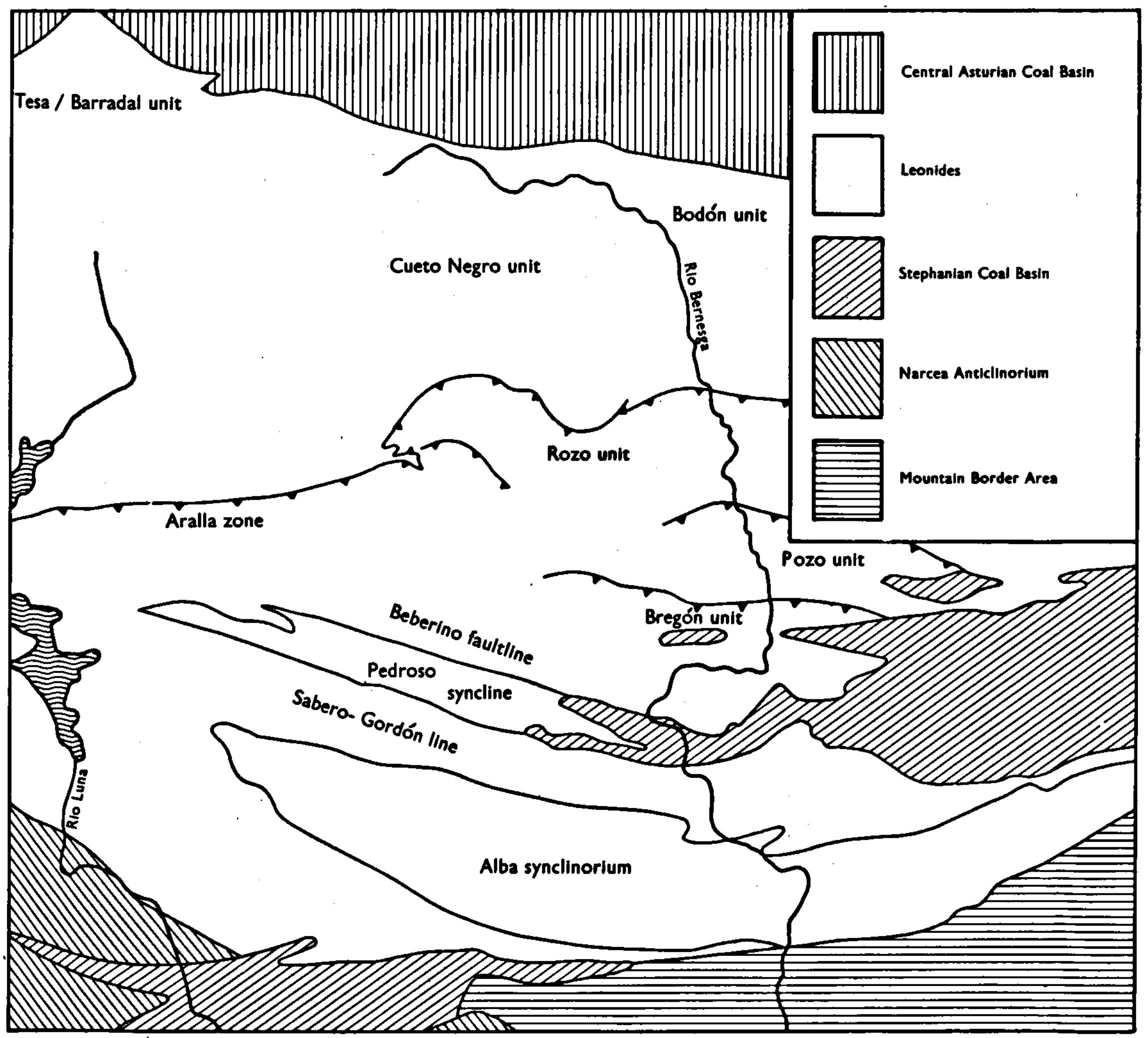

Fig. 3. Major structural elements in the Luna-Torío area and subdivision of the Leonides into structural units.

nuevo Formation and Palaeozoic rocks, giving evidence of the absence of almost the entire Mesozoic.

f) an assumed unconformity between the Vegaquemada Formation (Eocene-Oligocene) and the Upper Cretaceous Voznuevo Formation.

The angular unconformities of (a) and (d) were caused by previous folding. The unconformities of (e) and (f) are chiefly due to tilting before deposition, which also holds for the disconformity of (c). The origin of the hiatus of (b) is questionable.

Previous work

Many French, Spanish and German geologists studied the southern Cantabrian Mountains before Comte de-

Comte's subdivision (1959):	Thickness in meters	Age	Formations
Conglomérats			Riacos
de Castille		Miocène	Candanedo
Argiles		– (lacune?) –	Vegaquemada
Marnes de Boñar			
		Crétacé	
Graviers et Sables kaoliniques	± 200		Voznuevo
		– lacune –	
Schistes houillers et Grès de Tineo			Prado
		Stephanien	
Schistes houillers de Sabero	> 150		
		– lacune –	
Grès et Schistes de Sama			
		Westphalien	
Calcaire et Schistes de Lena	>1000		Lena
Calcaire et Schistes de Villanueva			San Emiliano
			Cuevas
Calcaire des Cañons	200–800	Namurien(?)	'Caliza de Montaña'
Griotte de Puente de Alba	25–40	Viséen	Alba
		– (lacune?) –	
'Couches de Vegamián'	± 15	Tournaisien(?)	Vegamián
		– lacune –	
Grès de l'Ermitage	0–1000	Strunien	Ermita
		– lacune –	
Schistes de Fueyo	± 100		Piedrasecha
		Famennien	
Grès de Nocedo (Calcaires de Valdoré)	± 500		Nocedo
		– (lacune?) –	
		Frasnien	
Calcaires de la Portilla	50–80		Portilla
		Givetien	
Grès et Schistes de Huergas	220–300		Huergas
		Eifelien	Caldas
Calcaires de Santa Lucia	100–250		Santa Lucía
		Emsien	
Calcschistes et Calcaires de La Vid	180–500		La Vid
		Siegenien	
		Gedinnien	
Grès rouges de San Pedro	70–170		San Pedro
		Silurien	
Schistes du Formigoso	50–100		Formigoso
		– lacune –	
		Arenig	
Quartzites de Barrios	160–480		Barrios
Schistes et Grès d'Oville	120–240	Tremadoc & Potsdamien	Oville
		– (lacune?) –	
Griotte de Lancara	12–25		
		Acadien	
Dolomites de Lancara	45–80		Láncara
		Géorgien	
Grès de La Herreria	>1400		Herrería
		– (lacune?) –	Mora
		Précambrien	

Table 1. Slightly modified after Evers (1967).

scribed this area. Most of them did so in view of the mineral resources and rich fossil localities. Their publications include: C. de Prado (1850, 1852); de Vernieul (1850); Barrois (1882); Mallada (1887); Oehlert (1897); Kegel (1929); P. H. Sampelayo (1934); Ciry (1939); Delépine (1943) and Almela (1949).

For the lithostratigraphic subdivision of the Palaeozoic deposits of the present area Comte (1959) chose a type-section along the Bernesga river between Busdongo and La Robla. Although the chosen rockunits (Table 1) coincide to a great extent with formations in the sense of Hedberg (1961, 1972) Comte's nomenclature differs from Hedberg's. In the type-section it was Comte who defined the Palaeozoic formations, but the names of the pre-Devonian formations and those of the Devonian Portilla Formation were derived from reference sections. The names of these formations have now been accepted on the basis of priority rules. Later publications either complete the lithostratigraphic subdivision of this area or detail the pre-existing units.

The Precambrian rocks in this area remained unknown for a long time. De Sitter (1962) first described an angular unconformity in what had until then been called the Herrería Formation. He suggested the rocks underlying the unconformity to be of Precambrian age and called them Mora slates and greywackes, now formally named Mora Formation.

Oele (1964) studied the sedimentology of the Herrería, Láncara, Oville and Barrios Formations. Van der Meer Mohr (1969) investigated the Láncara Formation on the basis of carbonate sedimentology.

Bäcker (1959) studied the San Pedro Formation with reference to the Silurian-Devonian boundary in the Asturo-Leonese area.

Brouwer (1962, 1964, 1967, 1968a and b) paid special attention to the development of Devonian stratigraphy in the Leonese facies area. Van Adrichem Boogaert (1967) reviewed the Devonian stratigraphy in general and the Upper Devonian and Lower Carboniferous in particular.

Graduate students from the Department of Structural Geology of the University of Leiden produced a number of map sheets on a scale of 1 : 50,000 contributing to the regional stratigraphy of the southern Cantabrian Mountains. These include Kanis (1956), Nederlof (1959), Koopmans (1962), de Sitter & Boschma (1966), Frets (1965) and Boschma (1968) in the Palentian area. In the Leonese area map sheets have been published by Rupke (1965), Helmig (1965), Sjerp (1967), van Veen (1965), Evers (1967), Savage (1967), de Sitter & van den Bosch (1968) and van den Bosch (1969).

The official Spanish edition of the geological memoir and map sheet La Robla as prepared by Pastor Gómez (1963) also deals with the Palaeozoic stratigraphy of the present area.

Smits (1965) described an aberrant development of Couvinian deposits in the Caldas de Luna area (see map) and named this sequence the Caldas Formation. De Coo et al. (1971) studied the palaeogeography and the carbonate facies of the Santa Lucía Formation and concluded that the Caldas Formation should be included in the Santa Lucía Formation.

Reyers (1972) and Mohanti (1972) made a detailed study of the Portilla Formation.

Van Ginkel (1965) redefined the Alba and Vegamián Formations and described the San Emiliano Formation. He also designated the Lena Formation.

Boschma & van Staalduinen (1968) introduced the Cuevas Formation and the Prado Formation.

Brouwer & van Ginkel (1964) introduced the Escapa Formation instead of the Caliza de Montaña Formation and other synonyms. Winkler Prins (1968) made a detailed study of the Escapa Formation and its brachiopod fauna.

Van Adrichem Boogaert (1967) reviewed the Lower Carboniferous stratigraphy and studied the conodonts of the Vegamián and Alba Formations.

Llopis Lladó (1954b) described the Lena Formation of the southern border of the 'Cuenca Central' of Asturias.

The Ciñera-Matallana and Magdalena Coal Basins have been intensively studied by Wagner (1962-1964). Van Amerom & van Dillewijn (1963) described and mapped the western part of the Ciñera-Matallana Basin.

Evers (1967) defined the post-Palaeozoic rocks of the adjacent area.

Biostratigraphic data have been published by Lotze & Sdzuy (1961), who studied tribobites of the Láncara Formation and part of the Oville Formation.

Cramer (1964, 1966) studied micro-organisms of the Formigoso Formation, the San Pedro Formation and the La Vid Formation.

Devonian fossils have been studied by Krans (1964) (spirifers); Westbroek (1967) (Rhynchonellida); and Sleumer (1969) (stromatoporoids).

The Lower Carboniferous biostratigraphy has been dealt with by Kullmann, 1961, 1963; Wagner-Gentis, 1963; Higgens et al, 1963 (cephalopods); van Adrichem Boogaert et al., 1963; Budinger & Kullmann, 1964; van Adrichem Boogaert, 1967 (conodonts).

To Upper Carboniferous stratigraphy contributed studies on fusulinids (van Ginkel, 1965), bivalves (Winkler Prins, 1968), algae (Rácz, 1964) and land plants (Jongmans, 1951; Wagner, 1962-1964; Gomez de Llarena, 1950 and Stockmans & Willière, 1965).

LITHOLOGIC CHARACTERISTICS OF FORMATIONS

Mora Formation. – The Mora Formation contains mainly greenish slates, greywackes, quartzites and shales. The green slates have only been observed south of Vega de los Caballeros, with laminar stratification and small-scale cross-bedding. Sandy layers generally grade upwards into pelites which are foliated with a slaty cleavage. Towards the north and probably upwards in the sequence the grain size increases and greywackes and some quartzites have been deposited. In the coarse part of the formation slaty cleavage is absent, but a slightly developed fracture cleavage may occur.

Near the unconformity, (Fig. 4) the rocks show little evidence of metamorphism, which is partly due to the

Fig. 4. Unconformity between the Precambrian Mora Formation (below left) and the Herrería Formation near Irede.

absence of pelites and possibly to a diminishing grade of metamorphism in this direction (Matte, 1968) as well.

The uppermost part of the Mora Formation is usually red coloured and the stratification has been obliterated due to weathering before the deposition of the Herrería Formation.

Herrería Formation (700–1000 m). – In the Luna area the sequence starts with up to 1 m of quartz-pebble conglomerate followed by brown and green shales alternating with quartz sandstones. The shales and sandstones are overlain by the bulk of the formation with very coarse sandstones and gravels.

Cross-bedding and graded bedding in parallel bedded rocks are the most striking sedimentary structures in this part of the formation. Near the top of the formation green and brown shale beds reappear.

In the Luna area in the lower part of the shale and sandstone sequence a cleavage has developed locally. The cleavage of the Mora Formation here seems to continue in the same direction into the Herrería shales.

Near Mora de Luna brown dolomites occur in the shales at about 100 m above the base. Dolomite beds have also been found in the northern area in a sandstone sequence. These beds were mostly disturbed by synsedimentary slumping.

Láncara Formation (50–180 m). All authors subdivide the formation into three parts, a lower dolomite member, a middle limestone member and an upper griotte member. In the present area the three members could be recognized south of the Lake Luna and in the Rozo unit (App. 2). In the areas north of the Rozo unit and the Aralla zone (Fig. 3) no limestones occur between the dolomites and the griottes. The dolomitization of the limestone member in the Rozo unit is partly of secondary origin because here dolomitization is irregular and not parallel to stratification.

The griotte member is important for its outstanding red colour and the limestone nodules, which serve as markers for mapping and stratigraphic interpretation.

In areas where the thrust-plane follows the basal part of the dolomite member, e.g. in the Rozo unit, ores have accumulated within the formation: barite, together with azurite and malachite. In the area of the Sierra del Cueto Negro the dolomite is often intraformationally brecciated.

Oville Formation (150–400 m). – The formation can be subdivided into two parts: a lower shale-siltstone member and an upper sandstone member.

The lower member is composed of glauconitic shales and siltstones with limestone nodules in its lower part. Upwards the number of limestone nodules gradually decreases. The shales in this member are fossiliferous.

The upper member is mainly composed of green sandstones and white quartzites alternating with green shale beds. Weathered orange coloured glauconite grains differentiate the quartzites of the Oville Formation from the overlying Barrios quartzites.

In the northern Bodón unit intrusives occur in contrast to the rest of the area. They are found in both the Oville and the overlying Barrios Formations. On the map only the larger ones have been indicated. Apart from these dolerite sills tuffite beds parallel to general bedding are present in this area.

The thinnest development of the formation occurs in the Aralla zone, where the upper member was deposited under conditions which allow erosion, as is shown by several eroded bedding planes.

Barrios Formation (200–400 m). – The lithology of the formation is rather monotonous and composed of white to pink coloured quartzites, and thin beds of green to black shales in between. Van den Bosch (1969) based a subdivision into five members upon the occurrence of two shale beds between the quartzites in the southern part of this area. In the present area this subdivision is not useful. In some sections, however, there are one or two often thin (20–50 cm) carbonaceous black shale beds between the quartzites. These shale beds may be comparable to the shale members referred to by van den Bosch (1969).

Van den Bosch also stated that the upper part of the formation is not developed in the Aralla zone.

Formigoso Formation (90–200 m). – The formation can be subdivided into two members, a basal black shale member and an upper shale/sandstone member.

The black shales at the base are the most indicative sediments of the formation. There are, however, very few exposures where these shales are not covered by scree from the Barrios Formation. In the well-exposed sections north of Villasimpliz (App. 2, Pajares section), and north of Geras de Gordón in the Rozo section the shales overlie the Barrios quartzites with a sharp contact. In both sections after about 1 m of shale coarser deposits occur, mainly composed of quartzites alternating with thin mudstone beds. The latter always have a high concentration of Fe. Maybe this sandy part in the shale member represents the transitional beds, mentioned by van den Bosch (1969). In general this sandy sequence does not exceed a thickness of 15 m.

The black shales are rather fossiliferous, especially in the lower parts with the finest grainsizes.

The shale member becomes more sandy towards the top. The transition to the upper shale/sandstone member is very gradual. The sandstone beds in this member dominate the shale beds. In general these sandstones layers are very irregularly-bedded and disturbed by bioturbation.

San Pedro Formation (100–170 m). – Van den Bosch (1969) and Bäcker (1959) investigated the San Pedro sandstones in detail and subdivided the formation into three members which could also be recognized in the present area. These members are:

a) a basal member, composed of thick red channelling, sandstone beds, often with hematite ooids,
b) a middle member, composed of an alternation of green shales and red and greenish sandstones with hematite and chamosite ooids resp.,
c) an upper member, composed of an alternation of white quartzites and black shales.

The lower member often contains a high percentage of hematite. White to yellowish coloured phosphorite pebbles are often concentrated in beds of the coarsest ferruginous sandstones. Alternating shales and sandstones as well as quartzites of the upper member have no significant hematite content. Brown and black shales become more abundant near the top of the member. Shales occupy nearly 70% of the whole upper part, and the cement in the coarser parts grows more calcareous. Due to gradual increase of the often dolomitic calcarous components there is a transitional zone between the San Pedro Formation and the overlying La Vid limestones.

Sedimentary structures are abundant especially in the middle member. Current ripples, burrows, linsen and flaser structures as well as loadcast and slump structures have been frequently recognized.

In general the formation has developed rather uniformly throughout the area. In the Aralla zone, however, erosion removed parts of the upper and middle member.

In general the thickness of the San Pedro Formation as well as that of the underlying Formigoso Formation decreases towards the north.

La Vid Formation (130–400 m). – In the present area the La Vid Formation could be subdivided into two members: a lower limestone member and an upper shale/limestone member.

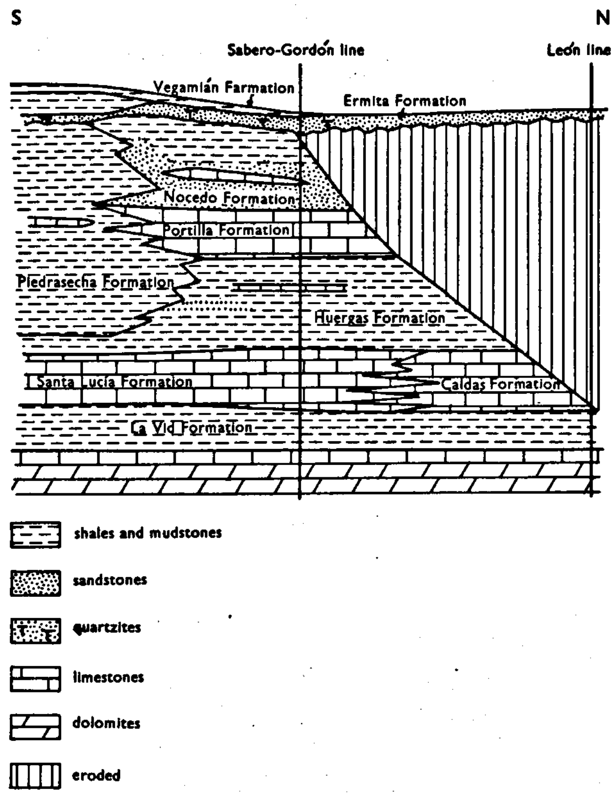

Fig. 5. Stratigraphic position and lithology of the Devonian formations in the Luna-Torío area.

The transition of sandstones of the San Pedro Formation into limestones of the La Vid Formation is gradual. At the top of the San Pedro Formation the sandstone percentage decreases, while the carbonate percentage (cement) increases. The La Vid Formation starts arbitrarily, where the first carbonate bed appears (Brouwer, 1968). The lower part of the limestone member has for the greater part developed as dolomite beds (Fig. 5). Upwards blue nodular limestones and fossiliferous (mainly brachiopods) dark gray limestones occur. In general the limestones change abruptly into carbonaceous shales. On top of a shale sequence, which is often disturbed by faulting and folding, thin limestones occur. These limestones are mainly composed of reddish crinoid stems. The transition into the overlying Santa Lucía Formation is abrupt in many locations. In other places, e.g. N of Mirantes, there is a gradual transition from red crinoidal La Vid limestones into more massive limestones of the Santa Lucía Formation.

In the southern part of the Alba syncline the lower limestone member is poorly developed. N of Los Barrios de Luna a thickness of 50 m has been measured. Eastwards the lower member becomes even thinner. In the northern areas the formation becomes thinner. Comte (1959) assumed that part of the upper shale member was eroded. This may be true for the Bernesga section near

Camplongo, where the Ermita Formation unconformably covers the La Vid Formation, but north of Caldas de Luna the upper shale member has the same thickness without an unconformable contact at the top. Therefore it is still questionable if erosion strongly affected the thickness of the formation in the Bernesga valley.

Santa Lucía Formation (75–240 m). – The Santa Lucía Formation is almost entirely composed of limestones and dolomites. Therefore it is often difficult to subdivide the sequence. Microscopically a subdivision of seven units could be made in the reference section (de Coo, in prep.). A subdivision based upon morphological features gives three members (Fig. 6): a lower well-bedded limestone member, a middle massive limestone member, and an upper well-bedded limestone member with red coarse crinoidal limestone beds in its uppermost part.

Although the shale percentage in all sections is very small, it shows a remarkable increase from S to N. In the present area no noticeable variations could be observed from W to E.

Caldas Formation (0–360 m). – Two members could be recognized in this formation: a lower argillaceous limestone member, characterized by an alternation of grey limestones and calcareous shales; a limestone member containing dark grey limestones with two marker beds of red nodular limestone.

The most complete sections occur in the western part of the area. To the E the Caldas Formation has been

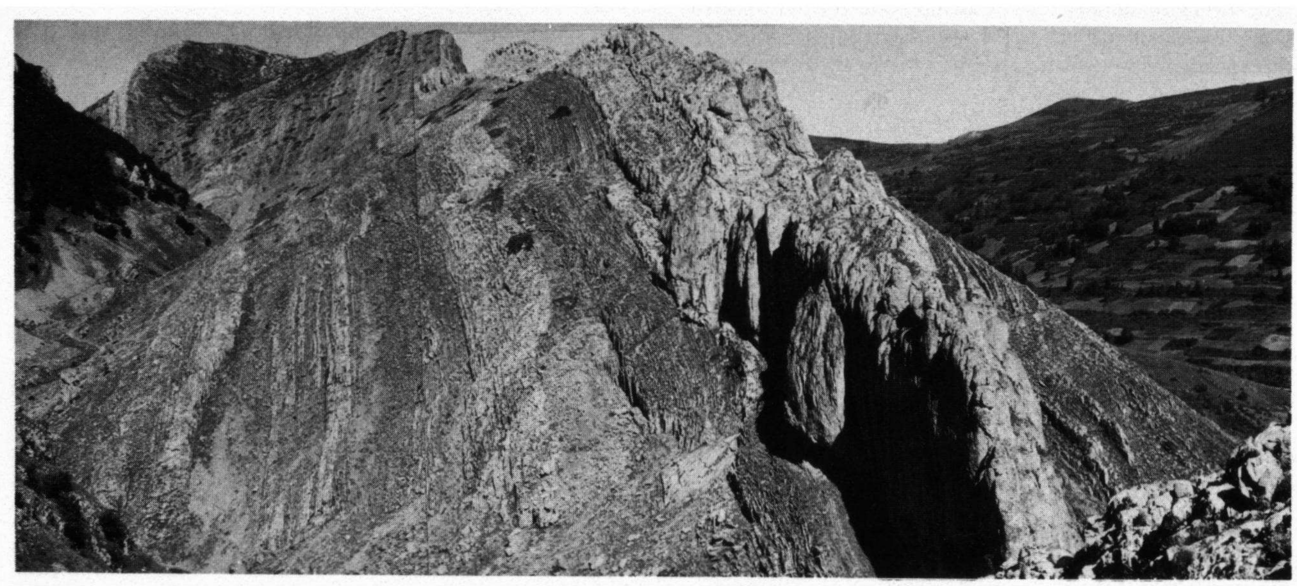

Fig. 6. Santa Lucía limestones, south of Geras de Gordón in vertical and slightly overturned position.

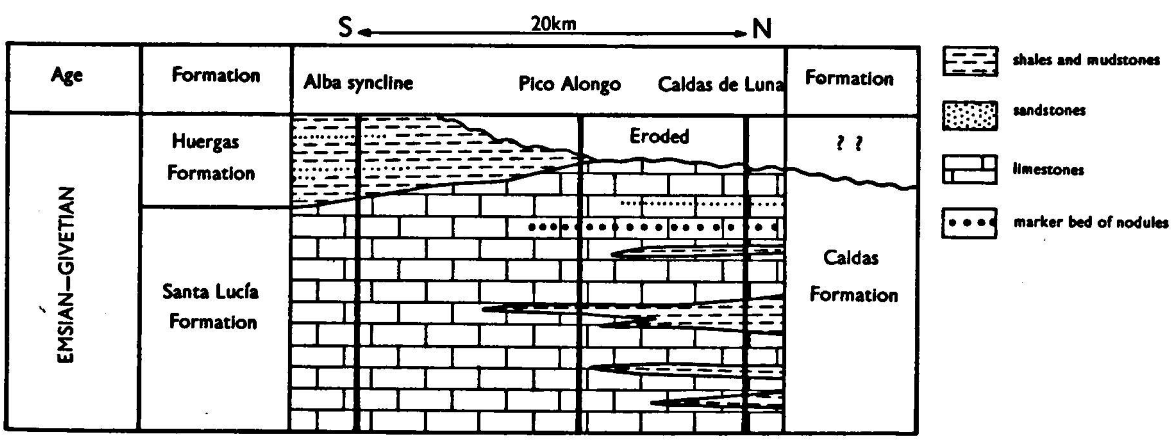

Fig. 7. Relations between the Santa Lucía and Caldas Formations and their differences in lithologic properties.

eroded. Therefore near Casares de Arbas only the lower member occurs. Further east even the lower member has been eroded.

The southernmost outcrop of the formation has been observed near the Collada de Alongo (Fig. 7). Here the shale content is very small. The argillaceous limestone member could not be recognized there. Due to intensive folding it is impossible to reconstruct the original position of the sequence, although it is significant that it consists of a limestone sequence. For this reason it is more likely to belong to the Santa Lucía Formation than to any other section of the Caldas Formation. The typical subdivision of members of the Santa Lucía is missing and therefore this outcrop is classified with the Caldas Formation.

Compared with the type-locality (N of Caldas de Luna, Smits (1965)) the shale content of the argillaceous limestone member along the northern boundary of the Tesa syncline is greatest. Even thin sandstone beds occur in this sequence. The limestone member too is thicker, suggesting that erosion here had not such a great influence as it had farther east.

Huergas Formation (200–300 m). – The bulk of the sediments of the Huergas Formation consists of micaceous and decalcified silty shales. Sandstone beds, however, occur everywhere in the upper part of the formation and in the lower part of the sequences in the northernmost sections. Near Ciñera the formation is almost entirely composed of brown weathered sandstones. In the area between Vega de Gordón and N of Mirantes de Luna shales are strongly predominant with small limestone concretions, often with specimens of *Buchiola* spec. in their centre. E and S of Mirantes de Luna even limestone lenses are intercalated with the shales. These limestone lenses often have a coral fauna resembling that of the overlying Portilla Formation.

Along the southern limb of the Alba syncline the shales of the formation are thinly developed in the area W of Saguera; E of Saguera the identity of the Huergas Formation becomes vague owing to the absence of overlying limestones. After the deposition of the Santa Lucía Formation here a clastic series developed during the rest of the Devonian, which is called the Piedrasecha Formation.

Portilla Formation (0–200 m). – The Portilla Formation consists of three different parts:

a. a lower massive, sometimes platy, limestone member;
b. a middle member, composed of an admixture of thin-layered limestones, shales and even sandstone beds;
c. an upper massive limestone member.

In the steeply inclined sections the middle member forms a depression between the limestone members (Fig. 8). The subdivision into three members could be recognized in nearly all sections W of the Bernesga river. E of the Bernesga and S of the Matallana Coal Basin this subdivision is difficult to maintain.

In the Beberino section the upper limestone member shows knoll-like features where it borders the overlying Nocedo Formation (Fig. 8). The shales of the Nocedo Formation are regularly layered upon the massive lime-

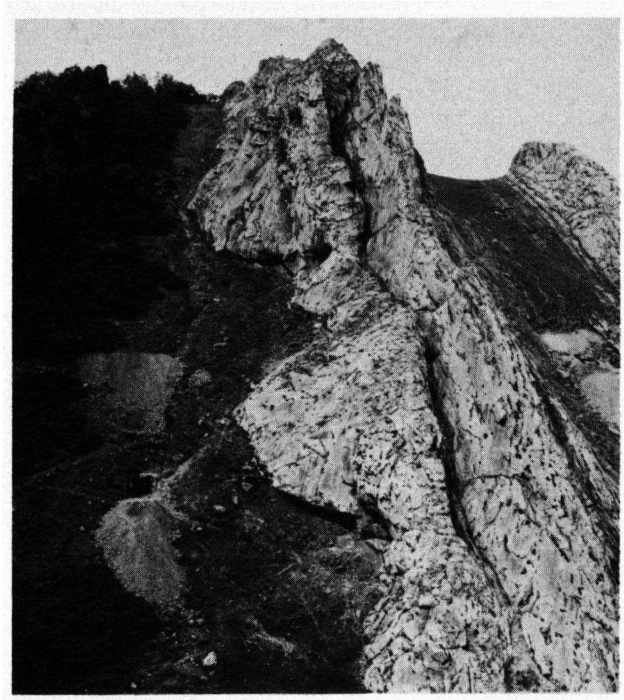

Fig. 8. Portilla Formation near Beberino; a lower limestone member (right-side of photograph), a middle shale/limestone member and an upper limestone member with knoll-like structures at the top.

stones and there are no indications of talus deposits at the sides of the limestones. Therefore it is assumed that erosion caused the formation of knolls.

Along the southern flank of the Alba syncline the limestones thin out to the E. Near Saguera only small remnants of the Portilla limestones have been found.

In the hinge of the Alba syncline the limestones are extraordinarily thick.

Mohanti in Struve and Mohanti (1970) regards the facies of the formation as: "a 'reef' facies with developments of 'lagoonal' facies, biostromal 'reef' barrier facies. Slow and prolonged differential subsidence of the depositional basin with possible emergence of marginal areas is believed to be one of the main factors in controlling the thickness and facies changes in the Portilla Formation of the Alba syncline." Reyers (1972) confirms this statement.

Nocedo Formation (30–600 m). – Calcareous sandstones are the main components of the basal part of the formation. A limestone lens occurs in the type-locality but one km E and W of that locality no limestones are present. The limestones are followed by a sandstone/shale sequence, changing upwards into a black shale unit with sandstone lenses. This unit represents the 'Schistes de Fueyo' of Comte. Its small lateral extension causes the latter unit now to be included as the upper part of the Nocedo Formation.

The general aspect of the Nocedo is that of a calcareous sandstone sequence with diminishing grain size towards the top of the formation. Cross-laminations are the most striking sedimentary structures.

With respect to the sandstone/mudstone ratio of the Nocedo Formation Pieters (1967, int. rep.) observed a constant 1:1 ratio in the northern limb of the Alba syncline (Fig. 9).

In the southern bank of the Alba syncline the Nocedo Formation could not be recognized. The gradual lateral change into the Piedrasecha Formation there is to be seen around the nose of the Alba syncline and an arbitrary cut is proposed along the axis of this structure, separating the Nocedo from the Piedrasecha Formation.

Piedrasecha Formation (0–600 m). – The Piedrasecha Formation is introduced here for the clastic sequence of mainly Devonian rocks deposited on top of the Santa Lucía Formation along the southern limb of the Alba syncline. This clastic sequence is present from Portilla de Luna to the Bernesga valley. The western delimitation of Piedrasecha, Nocedo and Ermita Formations is purely arbitrary. It is assumed on the map that there exists a lateral facies change between these three formations.

The section near Piedrasecha (App. 3), starting S of the village up to the Alba Formation N of the village, has been chosen as type-locality. The thickness decreases to the E as can be observed from map and sections. Intense faulting, especially near the contact with the Santa Lucía Formation in the area between Piedrasecha and Santiago de las Villas causes repetition of the basal beds.

The sandstone/mudstone ratio in the type-section is 1 : 4 (Fig. 9). Compared with the ratio in the area north of the Alba syncline, where Nocedo and Ermita Formations occur, this ratio is much smaller. This diminishing tendency continues from Piedrasecha eastwards.

The lithology of the type-section near Piedrasecha and of a section NE of Santiago de las Villas is compared with the type-section of the Nocedo and Ermita Formations in the Bernesga valley (App. 3).

The type-section starts with about 200 m of brown shales in which concretions of claystone and limestone are common. The limestone concretions are often fossiliferous, especially in the basal part of the sequence. Apart from the concretions thin ferruginous beds occur in this lower part of the formation. Then follows a monotonous series of ca 160 m of dark brown to black shales with thin ferruginous sandstone beds. The following sequence of about 100 m is composed of thin-layered sandstones alternating with shale beds. In this part of the sequence animal tracks, loadcasts and slump struc-

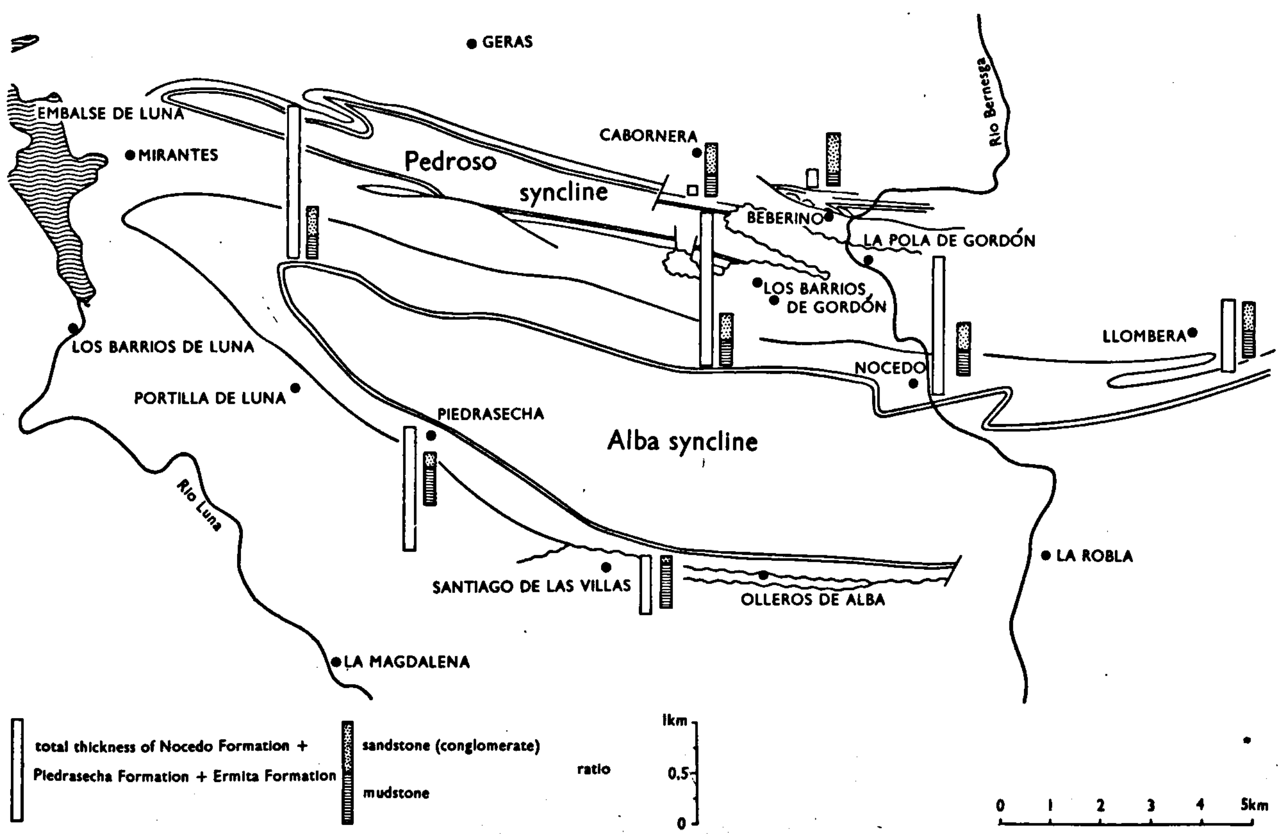

Fig. 9. Sandstone/mudstone ratio and diagrams showing the thickness of the Piedrasecha, Nocedo and Ermita Formations.

tures could be recognized. The upper 20 m of this sequence also contains concretions. White quartzites and sandstones form the following 10 m of the section, and 30 m of badly exposed brown shales cover the quartzites and sandstones. The upper 50 m of the sequence consists of alternating brown sandstones and mudstones with cross-bedding in the sandstones of the uppermost 10 m. Casts of brachiopods have been found in the mudstone beds.

Compared with the type-section (App. 3) of the clastic Upper Devonian in the Bernesga valley the Piedrasecha Formation shows more shales and mudstones. The lower part of the section with the fossiliferous limestone concretions is likely to be the equivalent of the Huergas Formation elsewhere. The upper part of the formation, showing more sandy layers, should perhaps be correlated with the Ermita Formation of the Bernesga section. The lower limit of this Ermita-like sequence should be the contact between the quartzitic sandstones and the underlying sandstone/shale sequence. However, a sharp contact could not be observed in contrast to a comparable sequence near Santiago de las Villas (App. 3). The brown shales which have been deposited upon the sandstones there, are very different from the coarse transgressive sandstones of the Ermita Formation in the Bernesga section.

Ermita Formation (1–140 m). – The base of the formation is often typified by microconglomerates which grade upwards into coarse decalcified sandstones and ferruginous quartz sandstones. This part of the formation often has a reddidh colour, mainly caused by the high iron content of the sandstones.

The upper part of the formation is often composed of arenaceous limestones. There is a gradual transition from the underlying sandstones to the limestones.

The Sabero-Gordón line separates in this area a thick development of the Ermita Formation in the S from a thin sequence north of it. South of the Sabero-Gordón line a gradual contact between the Nocedo Formation and the Ermita Formation makes the boundary an arbitrary one. North of the line the transgressive nature of the Ermita Formation is marked by thin microconglomerates deposited upon Devonian deposits, which are older when situated more to the N. Near Villar del Puerto the Ermita sandstones overlie Huergas shales and sandstones. Near Camplongo the Ermita Formation covers la Vid shales, showing an increasing gap below the Ermita Formation (Fig. 5).

Vegamián Formation (5–10 m). – The formation is mainly composed of black cherty shales with phosphatic and limestone nodules at several levels. Thin-bedded cross-laminated sandstones occur in the middle part. Generally the uppermost part of the formation is composed of reddish-brown mudstones alternating with thin limestone beds.

The contact with the underlying Devonian rocks is generally sharp. In the section near Santiago de las Villas the base of the formation is composed of a thin conglomeratic sandstone layer (Wagner, 1963, p. 41) resting with a sharp contact on the Piedrasecha Formation.

The contact with the overlying Alba Formation is sharp in the Santiago and Olleros sections, but in general the contact is gradual.

Alba Formation (5–40 m). – The Alba Formation is composed of red and green nodular limestones with fine crystalline wavy limestones in the basal part, red cherty shales in the middle part, and red shales between the nodules in the upper part of the formation. Winkler Prins (1968) defined these subdivisions as members, naming them Gete Member, Valdehuesa Member and La Venta Member.

The contact with the underlying Vegamián Formation or the Ermita Formation is sometimes sharp (disconformable with the Ermita Formation) but generally gradual with the Vegamián Formation. North of the Sabero-Gordón line the Alba Formation is overlain by the Caliza de Montaña Formation and south of that line by the Cuevas Formation.

Caliza de Montaña Formation (50–750 m). – The Caliza de Montaña Formation can be divided into two parts: a lower part, mainly composed of thin-bedded fetid limestone in which calcite veins are common phenomena, and an upper part, consisting of more competent often reefoid limestones, which have been dolomitized in several places. Winkler Prins (1968) named the two parts Vegacervera Micrite Member and Valdeteja Biosparite Member respectively.

In the area of the Pedroso syncline only the lower part of the formation has been developed. North of the Rozo unit and the Aralla zone both members are present.

The thickness of the lower part varies between 200 and 400 m, whereas the thickness of the upper part is difficult to establish as the top of the formation interfingers with the overlying San Emiliano Formation.

Cuevas Formation (> 400 m). – The formation lies conformably upon the Alba Formation and can be subdivided into two members; a shale/greywacke member and a limestone member (Fig. 10).

In the lower shale/greywacke member the beds are alternating shales and greywackes. The sequence has at its base first thinly laminated shales in which some limestone nodules still occur. Higher in the sequence the limestone nodules disappear and thin sandstone beds occur. The sandstone beds show sharp basal contacts and gradation to their top. Some of these sandstone beds are cross-bedded. Upwards the greywacke beds become thicker and coarser. Gradation is less pronounced and internal structures have not been found. In the valley of Santiago de las Villas loadcasts and groove castst are present incidentally at the base of greywacke beds. North of Cuevas a nice slump structure is exposed in the upper part of this member.

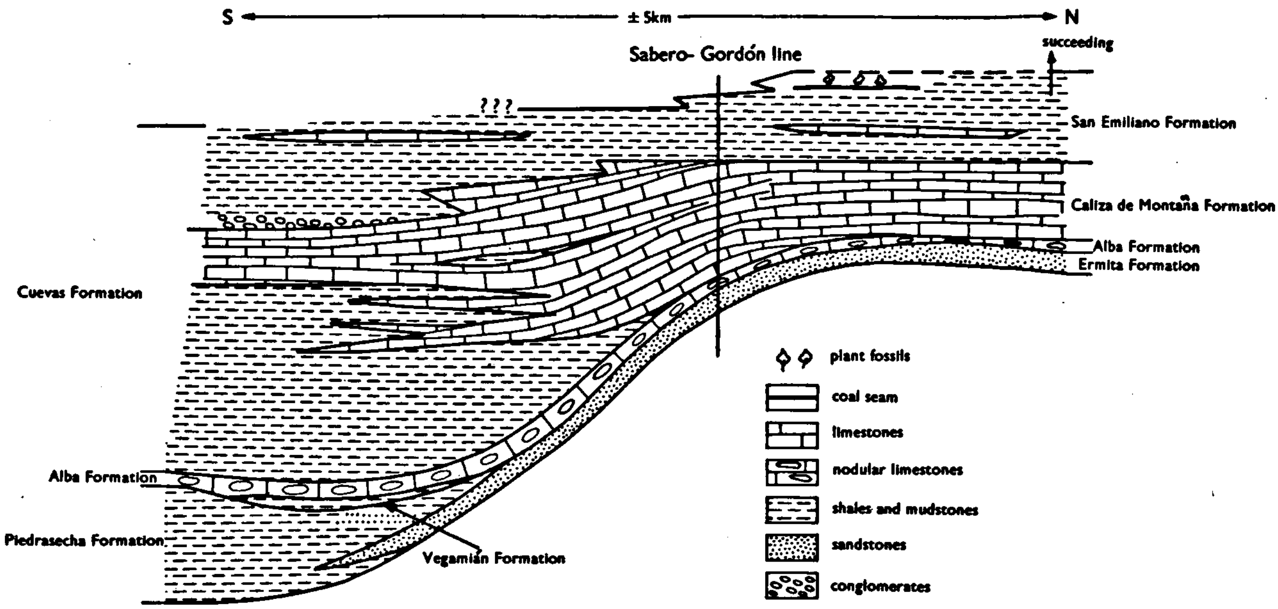

Fig. 10. Stratigraphic position and lithology of Carboniferous formations in the Leonides (Prado Formation excluded).

There is a gradual contact between the shale/greywacke member and the limestone member, i.e. thin beds of laminated limestones between the shales and greywackes are overlain by a vast deposit of limestones. The limestone member consists of thin-bedded, dark-grey, slightly fetid limestones in which calcite veins are abundant.

The limestones are strongly deformed by folding, making it difficult to establish their original thickness. In the section as given by Boschma & Van Staalduinen (1968) their thickness seems therefore to be somewhat exaggerated.

The limestone member strongly resembles the lower member of the Caliza de Montaña Formation (p. 179).

On top of the limestone member another clastic sequence was deposited. This upper part of the Cuevas Formation generally starts with a fine-grained conglomerate with subrounded grains. The grains are chiefly composed of quartz with angular chert fragments. The conglomerates are followed by a shale/greywacke sequence with thin limestone lenses. This upper part shows a striking resemblance with greywacke/shale series below the limestone member, and is considered to be part of the same clastic member.

The sedimentological structures found in the shale/greywacke member point to turbiditic depositional conditions. Typically deltaic features such as cross-bedding, ripple marks and bioturbation as well as fossils are absent in this member.

For the limestone member euxinix conditions, possibly in an area that was sheltered from the open sea, are suggested.

The upper clastic series of the Cuevas Formation, however, were deposited in a turbiditic environment. The conglomerates are likely to point to the rapid change in depositional conditions and reflect the initial movements of the basin, resulting in a shift of shore line and the return to predominantly clastic sedimentation.

San Emiliano Formation (500–2500 m). – The lower part of the formation is mainly composed of brownish grey shales and mudstones alternating with thin subgreywacke beds. In the central part mudstones alternate with argillaceous and fossiliferous pure limestones and a few beds of thick (10–15 m) subgreywackes. In the upper part greywackes, mudstones and carbonaceous dark-coloured shales are the dominant rock types.

In the Pedroso and Tesa synclines only the lower part of the formation is exposed. E of the Bernesga in the Bodón unit the limestone beds mainly occur in the upper part. Greywackes here are restricted to the central part. Thin coal beds have only been found in the section N of Villamanin.

Lena Formation (1400–3400 m). – The formation is composed of shales which are often carbonaceous, mudstones, sandstones, greywackes, thin limestone beds, thin conglomerate beds and some poor coal-seams. Evers (1967) recognized in the adjacent area three members: a basal marine to paralic mudstone member, a paralic limestone and wacke member, a paralic to limnic greywacke member.

These subdivisions could not be mapped in the whole of the present area. The oldest part of the Lena Formation in this area is present in the Corallón anticline, which on the map has been erroneously indicated as a syncline. The sequence becomes younger to the west with its top in the clastic deposits north of Busdongo.

According to van Ginkel (pers. comm.) the Corallón limestones belong to the *Profusulinella* subzone B and the coal seams north of Busdongo to the uppermost Westfalian C (Wagner, 1962). These data lead to the conclusion that the Corallón limestones belong to the limestone and wacke member and the rocks north of Busdongo to the upper greywacke member. Further west, however, it is difficult to ascertain which member the rocks belong to. It seems appropriate to place the limestones near San Miguel del Rio within the limestone and wacke member.

Prado Formation (> 1000 m). – The main characteristics of the formation are: a basal part composed of conglomerates; a younger part composed of subgreywackes, mudstones and coal seams. Lenticular conglomerates occur locally. In the Ciñera-Matallana Basin the first unconformable deposits are limestone conglomerates with very coarse and badly sorted angular boulders and smaller better rounded quartzite elements. The limestone conglomerates, especially the limestone boulders, seem to derive from the Palaeozoic limestones in the immediate environs, whereas the quartzite boulders may have been transported over a long distance.

The contact with the underlying Palaeozoic is irregular. In some places red earth occurs just below the conglomerates, pointing to subaqueous weathering before deposition.

The basal conglomerates are followed by a cyclic series of greywackes, siltstones and coal seams. In the western part of the basin the coal seams are thicker than in the eastern part. Coal seams are more frequent in the eastern and southern parts of the basin.

The Matallana Basin is an asymmetric basin as is illustrated by the occurence of conglomerates throughout the basin. In most places the original southern border is now obscured by faulting.

In the Magdalena Basin sedimentation started on a weathered surface of lower Palaeozoic or Precambrian rocks. Along the Luna river north of Garaño this weathered surface is red coloured. The basal deposits in this basin are well-rounded quartzite conglomerates.

The thickness of these conglomerates never exceeds 50 m. The conglomerates are followed by a series of greywackes, siltstones and coal seams as in the Matallana Basin. There are fewer coal seams in this basin and the seams are poorer. As they are less productive than in the Ciñera-Matallana Basin actual coal mining has been stopped.

Voznuevo Formation (0–400 m). – The rocksequence starts unconformably upon the Cuevas Formation (Fig. 11) with thin fluvial sandstone beds alternating with clay beds. Consequently linsen and flaser structures are the most striking sedimentary structures. In some cases the flasers have been deformed to clayballs. Most sandstones are white but upwards they become multicoloured. Within the sandstones quartz pebbles of > 2 cm are common. In the upper part of the formation lignitic clay bands and gravel beds are also present. The sandstones are mainly composed of well-rounded quartz grains coated with hematite, micaflakes and felspars. Due to chemical weathering most of the felspars have changed into kaolinite. Contacts are indicative of a source rock in which quartz grains and felspars are the main constituents. Only once a felspar content of 11 percent has been observed in the Herrería Formation (Oele, 1964). None of the other Palaeozoic formations has a felspar content of more than 4 percent. Consequently the Palaeozoic rocks did not supply the material for the Voznuevo Formation. Due to previous peneplanisation of the Palaeozoic (Evers, 1967) hardly any clastics could have supplied to the Voznuevo Formation.

The numerous granites as described by Pannekoek (1968) from the Castillian-Galician zone have the proper composition. The assumption of a western source area is supported by the occurence of inclined stratifications which suggest currents running from west to east.

In the upper part of the formation a small band of arenaceous limestones occurs especially near Brugos de Fenar and south of Candanedo and Solana de Fenar. These limestones represent the Boñar Formation as named by Evers (1967) east of this area. They are yellow and are more resistant to weathering than the accompanying sandstones.

Vegaquemada Formation. – Reddish brown argillaceous sandstones, thin conglomerate beds, claybeds and micaflakes all over the sequence are the most characteristic features of the formation. The lower part of the formation is mainly composed of argillaceous sandstones. Only when unconformable contacts with the underlying Voznuevo Formation or the limestones of the Boñar Formation are present could the sandstones be distinguished from the sandstones of the Voznuevo Formation. The upper part of the formation is characterized by increasing grainsize of the sandstones and channelling of coarse grained sandstones in finer sandstones. Channels cut with erosive base through well-sorted sandstones. The gravels of the channel fill do not show any sorting. The thickness decreases from the border of the Tertiary basin towards the centre of the basin and varies due to the lateral interfingering with the Candanedo Formation. The sandstones and the conglomerate material have presumably been supplied by two manners of transport. The fine sandstones, which generally show a good sorting, were transported by rivers. The poorly rounded conglomerate beds of the upper part of the formation may represent mountain flood debris, supplied by vertical mountain erosion.

Mabesoone (1959) investigated the conglomerates of the Cuevas facies in Palencia, an equivalent of the Vegaquemada Formation. From the pebble analysis he too could distinguish two types of pebbles, one derived from slopes, another of fluviatile origin.

Candanedo Formation. – The Candanedo Formation is composed of limestone conglomerate and coarse sand-

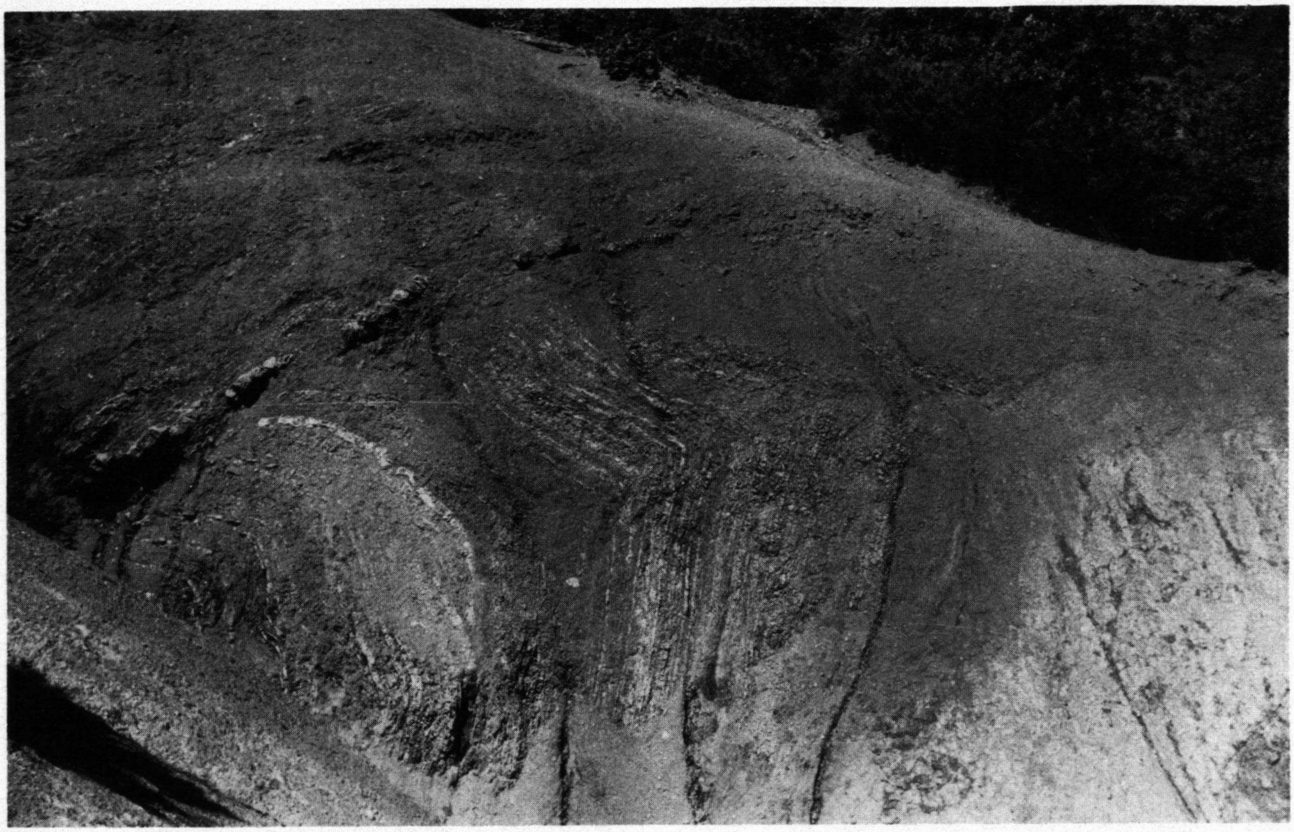

Fig. 11. Unconformity between the Carboniferous Cuevàs Formation (upper side left) and the Cretaceous Voznuevo Formation, north of Brugos de Fenar.

stones with calcite cement. At the base of the formation the conglomerate cobbles have a mean diameter of 15 cm. In general they are subrounded. Upwards in the sequence the grainsize diminishes and in between the conglomerate beds thin layers of coarse sandstone occur. The cement in these sandstone layers also contains calcite though less than in the conglomerate beds. Near the top the proportion of limestone cobbles decreases rapidly to less than 10 percent of the total composition. Most cobbles here are well rounded quartzites. The limestone cobbles are the typical black bituminous limestones with a high content of calcite veins, from the Carboniferous limestones of the Cuevas or from the Caliza de Montaña Formations.

With regard to the origin of the Candanedo Formation Evers (1967) suggests that intermittent cloudbursts in a semi-arid climate caused heavily loaded flash floods which carried the limestone cobbles from the mountains. East of the present area the cobbles at the base of the formation were derived from the Cretaceous limestones, near the top also Palaeozoic quartzites were deposited. This 'inverted' sedimentation was also noticed in the present area. The composition of the limestone conglomerates in this area reflect the outcrops in the adjoining mountains (e.g. the Cuevas Formation). The uppermost part of the formation contains quartzite cobbles which may have derived from older Palaeozoic conglomeratic rocks.

The present author is not entirely satisfied with the suggested modes of transport. In his opinion the rounded quartzite cobbles together with small channel features suggest river transport rather than flash floods.

Riacos Formation. – Coarse boulders of quartzites alternating with red sandstones and sandy clays are the most characteristic rocks of the Riacos Formation. The beds lie in an almost horizontal position resting slightly unconformable upon the underlying rocks. The unconformity is largest between the Riacos Formation and the Mora Formation in the west of the map sheet.

The sandy clays and the red sandstones dominate in the areas outside the river valleys, whereas the quartzite cobbles and boulders are abundant in the valleys. The conglomerates and the clays here represent the latest stage of the filling up of the Duero Basin. The sediments are continental and were transported by rivers and possibly by floods. The flash floods carried the conglomerates which lay distributed over the Duero Basin, depositing them fan-wise around the river mouths at the border of the mountain chain.

CHAPTER III

STRUCTURAL GEOLOGY

INTRODUCTION

Lotze and Sdzuy (1961) subdivided the Iberian Hercynian Orogene into a Galician-Castillian axial zone, a West-Asturian-Leonese internal zone and a Cantabrian external zone, the last being roughly the present morphological mountain chain. This mountain chain owes its morphological prominence to Alpine movements rather than to any special effect of the Hercynian orogenesis (de Sitter, 1964, p. 432).

The Luna-Bernesga-Torío area is situated in the SW part of the Cantabrian Mountain chain, thus forming part of the external zone of the Hercynian orogenesis. However, a small part of the Precambrian rocks in the SW of our area is regarded as belonging to the West-Asturian-Leonese internal zone. The rocks form part of the Narcea Anticlinorium as mentioned by Julivert and Martínez García (1967) and Matte (1968), Fig. 3.

De Sitter (1964, p. 432) distinguished two subzones within the external zone; a central subzone, the Asturides, and a southern subzone, the Leonides. These subzones are supposed to have been separated by a sharp line, the León line, which acted as a structural as well as a facies boundary. In the present area this line separates the Carboniferous rocks of the Lena Formation, which belong to the Central Asturian Coal Basin (Fig. 3), from the Lower Palaeozoic rocks of the Leonides.

The present study describes structures in the

Lena Formation San Emiliano Formation	Heterogenous sequence with shales, sandstones and limestones. Irregular folding specially governed by the rapid lateral facies change of these sediments
Limestone member of the Cuevas Formation Caliza de Montaña Formation	Platy limestones, concentric folding, largely parasitic folds
Shale member of the Cuevas Formation Alba Formation Vegamián Formation Piedrasecha Formation	Incompetent beds, disharmonic concentric folding
Ermita Formation Nocedo Formation Portilla Formation	Competent sandstones Competent limestones — concentric folding
Huergas Formation	Incompetent sequence, small-scale folding. Sequence causing disharmony between Santa Lucía limestones and overlying Portilla limestones
Santa Lucía/Caldas Formation	Competent limestones, strike-faulting, large-scale box-folding, often with thrust anticlines
Shale member of the La Vid Formation	Incompetent and mobile shales causing considerable disharmony between underlying and overlying Palaeozoic rocks, small-scale folding, fracture cleavage, core of box-folds
Limestone member of the La Vid Formation San Pedro Formation	Competent sequence, intensive minor folding, long strike-faults between the limestones and the San Pedro sandstones
Formigoso Formation	Black shales, incompetent sequence, minor folding allowing disharmony between the Barrios and San Pedro Formations
Barrios Formation Upper part of the Oville Formation	Competent sandstones and quartzites, controlling shape and width of most folds, characteristic cross-faulting
Lower part of the Oville Formation Lancara Formation	Shales and limestones, incompetent sequence, base of the Láncara Formation served as 'detachment' plane of thrusts
Herrería Formation	Competent sandstones, not involved in thrust structures, causing faults in core of anticlines, functioning as 'basement' to overlying thrust structures
Mora Formation	Slates and greywackes, slaty cleavage and fracture cleavage, intensely folded

Table 2.

Leonides and the interpretation of the León line as a structural phenomenon. The Leonides are typically thin skinned supra-structures as described by de Sitter (1962, 1965), Evers (1967), Rupke (1965) and van den Bosch (1969). The dominant folds are concentric and are controlled by a number of competent horizons (Table 2) as have been specially worked out by van den Bosch (1969, p. 191).

Incompetent formations allow disharmony to develop between structures of various parts of the stratigraphic sequence. Complete decollement developing into thrusting is a typical feature of the Leonide structures. Other faults also developed in relation to folding. Their detailed behaviour is clearly dictated by the competence of the individual strata cut.

The major geological regions – the Narcea Anticlinorium, Central Asturian Coal Basin, Stephanian Coal Basins and Leonides – are shown in Fig. 3, which also shows the subdivision within the Leonides, used in this study.

Tectonic activity ranging from Precambrian to Tertiary, has been proved here. The various orogenetic phases classified according to the system of Stille (1924) are shown in Table 3. Further details are given in the discussion of the appropriate units.

Three coloured sections at the left side of the geological map reflect the general structural pattern of the investigated area. In App. 4, (back pocket) 13 cross-sections show the relations between the folded structures of the Carboniferous rocks in the southern map area.

Major Periods	Stille classification	Results
Alpine Period	Savian morphogenetic phase	The Cantabrian Mts. obtained their present configuration, started in the Palaeogene.
	Saalic phase	Folding of the Stephanian Coal Basins. Activity in Post-Stephanian and Pré-Triassic times. Angular unconformity between Triassic and Palaeozoic rocks. Gentle folding.
Hercynian Period	Asturian orogenetic phase	Folding in the Leonides, started at the end of the Westfalian and continued into the earlier Upper Stephanian. Angular unconformity of overlying Prado rocks of Stephanian B age. Main folding.
	Sudetic initial folding phase	Thrusting and folding of the Leonides. It started in the Namurian and continued in the Westfalian.
	Bretonnic epeirogenetic phase	Uplifts started at the end of the Devonian and continued in the Lower-Carboniferous (base of the Viséan). A structural stratigraphical hiatus between the Ermita Formation and older Palaeozoic rocks is the result of these uplifts.
Precambrian Period		Orogenetic phase which folded Precambrian rocks of the Narcea Anticlinorium before the deposition of Cambrian sediments. Angular unconformity between Mora rocks and Herrería sandstones.

Table 3.

NARCEA ANTICLINORIUM

The strongly folded Precambrian rocks of the Mora Formation which occur in the southwestern part of the map area, S of Vega de los Caballeros, are part of an arc of Precambrian rocks, starting at the Asturian coast at Cudillero, trending S through Cangas de Narcea, then bending E near Villablino, continuing to the E. The Palaeozoic strata also follow this arc on either side of the Precambrian, giving rise to the concept of the Narcea Anticlinorium (Julivert-Martínez-García (1967) and Matte (1968).

In the map area Precambrian rocks form the most easterly part of the Narcea Anticlinorium, here partly covered by Carboniferous and Tertiary deposits. Generally the outcrops of the anticlinorium disappear to the east under a cover of Tertiary deposits at the border of the Meseta (Miocene-Pliocene deposits according to Pastor Gómez (1963). The cover of Carboniferous rocks in the map is bounded by faults to be discussed under the heading Mountain Border Area (p. 199).

According to Matte (1968) the Cambrian in the southern flank of the anticlinorium rests unconformably on the Precambrian rocks, but Julivert and Martínez-García (1967) consider the contact to be conformable. In the northern flank of the E-W trending Narcea Anticlinorium the contact is cleary unconformable (de Sitter, (1963), van den Bosch (1969).

About one km east of Vega de los Caballeros near the main road along the Luna river an affluent from Portilla de Luna runs into the main river. There E-W folded

Precambrian deposits are abruptly cut off by microconglomerates at the base of the Herrería deposits. The overlying Herrería deposits, here slightly overturned, dip steeply to the south. The angle between Mora and Herrería deposits is different due to folding of the Mora Formation near the unconformity.

Near Vega de los Caballeros the unconformity is also exposed. Here it occurs between slates of the Mora Formation and the Herrerían shales. The beds of the Herrería Formation are subvertical, whereas the Mora beds dip about 20° to the north. In the slates a cleavage occurs with a southward dip of 15–30°. This cleavage with a WSW strike occurs not only in the Mora deposits but is also present with equal direction and inclination in the lowermost shaly beds of the Herrería Formation.

Matte (1968) made observations in the area west of the river Sil which yielded the same results. His conclusion was that the cleavage is of Hercynian age. In the present area, however, the cleavage generally continues to about 30 m above the base of the Herrería Formation. Therefore the present author would suggest the cleavage in the Herrería beds to be a reactivation of a pre-Hercynian cleavage during the Hercynian deformation.

A comparable phenomenon has been observed in the Central Pyrenees, where Devonian cleavage seems to continue into unconformable Triassic. Mey (1969) demonstrated that post-Triassic folding reactivated movements along pre-Triassic cleavages planes, resulting in parallelism of non-contemporaneous structures.

Just outside the present map area, near Irede, lies the best known exposure of the unconformable contact (de Sitter, 1963; Julivert y Martínez-García, 1967 and van den Bosch, 1969), Fig. 4.

In the three localities where the unconformities are exposed, the angle between Mora and Herrería beds is different, suggesting that orogenetic forces folded the Mora deposits before the deposition of the rocks of the Herrería Formation. More to the west, however, the angle between the two formations is more or less invariable (van den Bosch, 1969) affording proof of tilting movements rather than folding.

In the southern part of the area, near Otero de las Dueñas, an E-W trending fault separates Carboniferous rocks of the Prado Formation from the Mora deposits south of them. This fault extends to the west over a long distance (Pastor Gómez, 1963, van den Bosch, 1969) and forms the southern border of the Magdalena Coal Basin. The fault-plane dips steeply to the south and has thrust the Mora deposits onto the Prado beds.

South of Vega de los Caballeros another E-W trending fault also forms the boundary of the Carboniferous rocks. This time the northern block has been upthrown. Both faults are strike faults in the Narcea Anticlinorium, and caused depressions within the Mora Formation, which afterwards were filled during the Stephanian by continental, coal bearing deposits. At a later stage reactivation of the fault movements caused overthrusting of Mora beds upon the Prado Formation. This may have occurred during a late deformation stage which may also have caused the kinkbanding in the Mora slates along the southern thrustfault.

LEONIDES

Alba synclinorium

The Alba synclinorium (Fig. 3) forms the southernmost Palaeozoic structure of the Cantabrian Mountain slope here and is succeeded farther south by the unconformable Mesozoic sequence of the Duero Basin.

In the Alba synclinorium the north flank is formed by Devonian to Namurian formations, whereas in the south flank the sequence is complete from Namurian as far down as the Precambrian basement.

The Lower Palaeozoic rocks along the south flank show different structural elements as compared with the structures of the Carboniferous rocks in the core of the synclinorium. They will therefore be dealt with separately.

The pre-Carboniferous southern flank. – The southern flank shows a system of faults clearly related to the development of the major structure. Strike faults are common, repeating considerable parts of the stratigraphic succession, e.g. south of Barrios de Luna and south of Portilla de Luna (section 2, App. 4). These SE-NW running faults have downthrown the sequences at its southern side.

Near to and north of Portilla de Luna the Santa Lucía limestone is cut several times by faults slightly oblique to the strike but retaining the same sense of downthrow to the south. They cannot be traced further in the incompetent shales above or below this formation. The fault-planes of the strike-faults and the oblique strike-faults are vertical or dip steeply north.

Cross-faults also show a very peculiar development especially in the zone between Los Barrios de Luna and Vinayo. The fault-planes appear to be more or less vertical, trending approximately NE-SW. Most of these faults extend to, though apparently not through, the competent upper Herrería, contrary to what might be expected in view of their very clear expression in the other competent formations such as Barrios and Oville. The NE-SW trend of the fault-planes (mean: N35E) roughly indicates the stress direction. Assuming that the direction perpendicular to the axial plane of the main syncline in this area (30–35°) also points to the main deformation stress direction, the suggestion is that the cross-faults and the folds originate from the same stressfield.

Structural elements in the Carboniferous part of the Alba synclinorium. – Although the central core of the structure looks rather simple it is composed of a number of smaller synclines and anticlines. The complexity of the structure is probably connected with the irregular sedimentation pattern as mentioned earlier (p. 180).

The average trend of the syncline varies from WNW-ESE in the west to WSW-ENE east of Santiago de las Villas. As shown in the cross-sections belonging to the southern map area (App. 4) the western part of the main structure has no irregular pattern (sections 1–4). It was here and especially in section 4 that by means of construction methods for concentric folds (de Sitter, 1941) the general model of the Alba syncline s.s. was obtained. This simple pattern only holds for the general model because the limestones of the Cuevas Formation here have been strongly affected by minor folding. The structure evidently is a single major syncline up to Cuevas, but to the east secondary structures develop on the southern flank en echelon with the main syncline (sections 5–8). Up to the Bernesga river the southern structures become dominant and the former major syncline SE of Nocedo is even lost in two minor synclines and anticlines.

East of the Bernesga river the folded structure is exposed on a lower stratigraphic level, e.g. the clastic lower member of the Cuevas Formation, (sections 10–13). Due to the rapid lateral change of facies at this stratigraphic level the structures are irregular and disharmonic.

In the area north of La Robla two faults are present. The fault running from about 1 km east of Sorribos de Alba to the NE acts partly as a wrench fault in that it exposes the Piedrasecha Formation more to the N along the eastern side of the fault-plane, partly as a normal fault, lowering the stratigraphic level on that same side of the fault-plane. This fault is possibly comparable with the oblique strike-faults in the pre-Carboniferous rocks along the southern flank of the Alba syncline s.s. Its presence affects in the same way the shape and the stratigraphic level of the exposed major syncline.

The fault running from Alcedo to the east is a strike-fault expressed at the Alba griotte level. It indicates a broken anticlinal structure possibly related to the Southern Border Fault, which will be discussed on p. 199.

East of the Bernesga the north side of the exposed Carboniferous rocks have been deformed by normal faulting. This change actually occurs east of Llombera, where the Matallana Basin is nearest to the Alba syncline. Here the generally competent Nocedo Formation is also thinning. It is supposed that both these factors were involved in the cross-faulting in this part of the area.

So far the outcrop pattern of the external deformation as a whole has been discussed. The limestones and the greywackes, however, also react differently to the folding. This internal deformation will be dealt with now. In these sediments it is also clearly demonstrated that the manner in which individual beds are layered, governs the size of folds, resulting in disharmonic structures at various levels.

Minor folds north of Piedrasecha. – The thin-layered rocks, limestones in particular, were thrown into a large number of minor folds in the core of the Alba syncline. Many of these folds are asymmetrical to the major structure and thus related to it. Most of them could be determined as parasitic folds in the sense of de Sitter

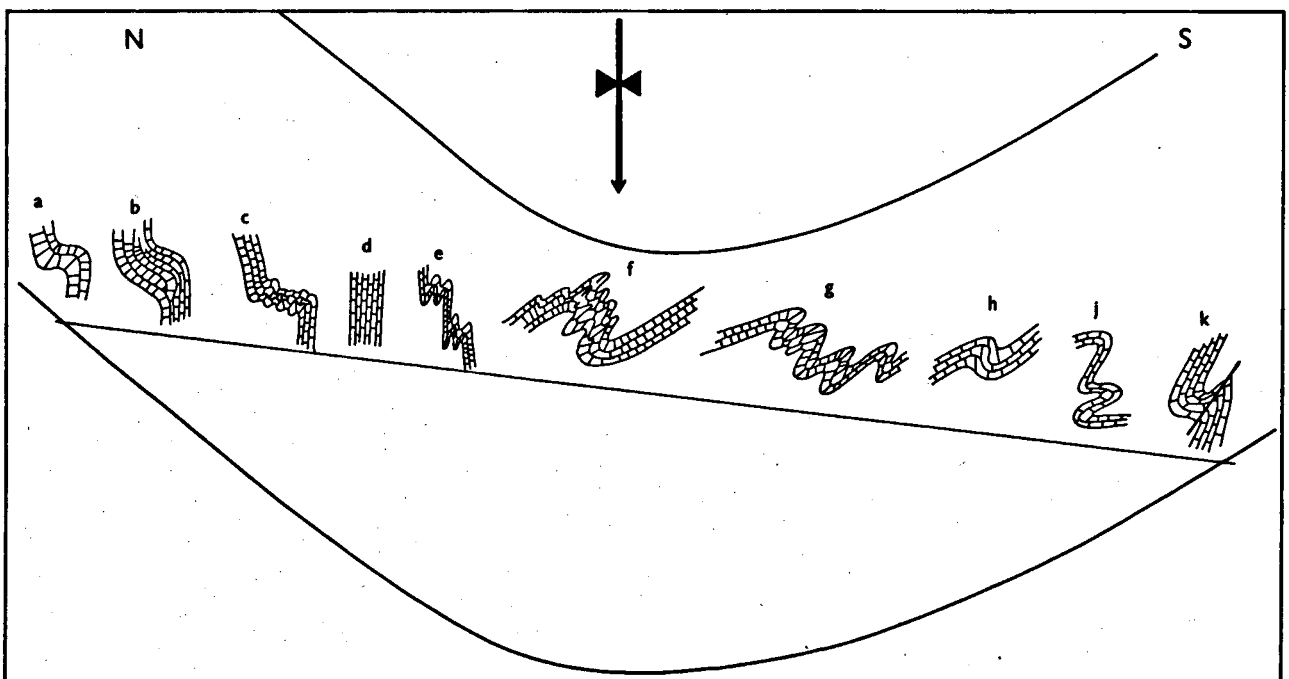

Fig. 12. Synclinal model of Cuevas limestones north of Piedrasecha with types of minor folds as observed from north to south.

(1958). In the present author's view other minor folds seem to be due to gravity rather than to a tectonic stressfield, and hence should indeed be regarded as cascade folds (de Sitter, 1964, p. 240). The cascade folds, belonging to one of the four classes of collapse structures as defined by Harrison and Falcon (1934), develop by the gliding of competent beds into incompetent deposits.

A good impression of the minor folds in the core of the Carboniferous rocks within the Alba syncline is given by a section through the limestone member of the Cuevas Formation north of Piedrasecha, Fig. 12.

The first type of minor folds is illustrated in Fig. 12a. These folds occur in a rather uniform limestone sequence, dipping 60–70° to the south. The limestones are composed of 40–60 cm thick beds alternating with 5 cm beds, apparently of the same composition. Although differences in competency between folded and unfolded beds are not noticeable the difference in thickness of the limestones point to a small difference in competency. However, there is not enough difference to cause flattening in the foldflanks or stretching in the foldhinges. The southern limbs of antiforms are long flanks in comparison with the almost horizontal northern flanks. The axial plane dips ca 60°N. The fold-type described can be regarded as parasitic to the major syncline because:

a. the direction of the axial plane fits with the concept of a parasitic fold belonging to the northern limb of the Alba syncline;

b. the southern limb is many times longer than the northern limb, which would be expected of flankfolds of the major syncline.

Southwards the aspects of the minor folds change slowly. Fig. 12b shows one of this new type of folds. The folds have been found in a limestone sequence comparable to that described above. There is no apparent difference in competence between the different layers. Antiforms of these folds have flat south-dipping northern flanks and nearly vertical southern flanks. The axial planes of these folds are flatter than those described above and foldaxes plunge 5°E. These folds the present author considers to be intermediate between parasitic and cascade folds because they have the expected asymmetry of parasitic folds, though some of the hinges show small scale faulting, causing the hinges to be downthrown to the inner part of the syncline.

Further southwards a composite set of minor folds is found. There is a type of folds with long vertical limbs alternating with small nearly horizontal limbs. The axial planes dip 50–60°N. This type corresponds to the description given of the parasitic folds of Fig. 12a. The horizontal limbs of these folds are here strongly folded. These small folds show an irregular pattern but nearly all the axial planes are vertical. All foldaxes plunge 5–10°E. The small faults have straight flanks and sharply curved hinges and are therefore accordeon folds (de Sitter, 1964, p. 294). The geometric properties, such as vertical axial planes and plunging axes agreeing with the calculated plunge of the major syncline, make it plausible to include these accordeon folds with parasitic folds. The relation existing between the two sets of minor folds (pl and p2) and the major Alba syncline is given in Fig. 13.

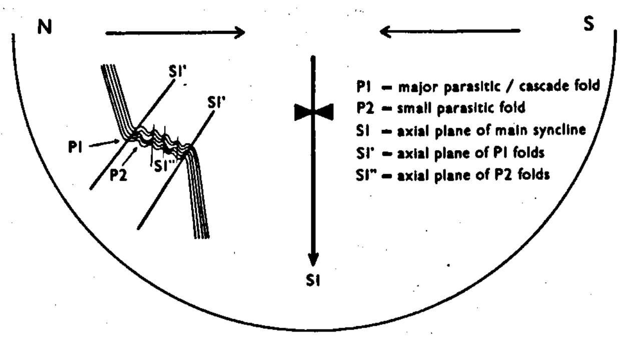

Fig. 13. Position of two sets of minor folds (pl and p2) along the northern limb of the limestone syncline north of Piedrasecha.

Near the centre of the main syncline a sequence of about 70 m of unfolded limestone beds occur (Fig. 12d). The beds are nearly vertical in this part of the syncline and the limestones themselves are very thin-bedded (5–10 cm).

Further south but still on the northern side of the major syncline again minor folds have been observed in still vertically bedded limestones (Fig. 12e). The axial planes are also almost vertical or dip steeply northwards, the fold-axes plunging 1–5°E. These folds have sharp hinges and are often broken in the cores, without loosing, however, the connection between beds along both sides of the fault-plane. The asymmetry of these folds suggests that they are parasitic.

Further south and close to the axial trace of the major syncline minor folds have the following properties (Fig. 12f):

a. northern flanks of antiforms are flat and north-dipping (10–30°);

b. the northern flanks are long in comparison with the southern folded limbs;

c. the secondery set of folds of the southern limb have axial planes, dipping steeply north (60–80°). Fig. 14 shows some of these folds. Near the bottom of the photograph starts a simple concentric fold-hinge which higher in the sequence is continued as two anticlinal accordeon folds. This proves that lack of space, evidently occuring when the folding is only concentric, is anticipated by the development of two accordeon folds on the hinge of the original fold. There is no sign of thickening in the hinges of the folds. The axial planes of the first and second set of minor folds here are parallel and dip northwards.

It is difficult to decide as to what type of fold is found here. The asymmetry of the flanks and the resemblance of these folds with parasitic folds described earlier suggest that they belong to them. However, their asymmetry indicates that these folds belong to the

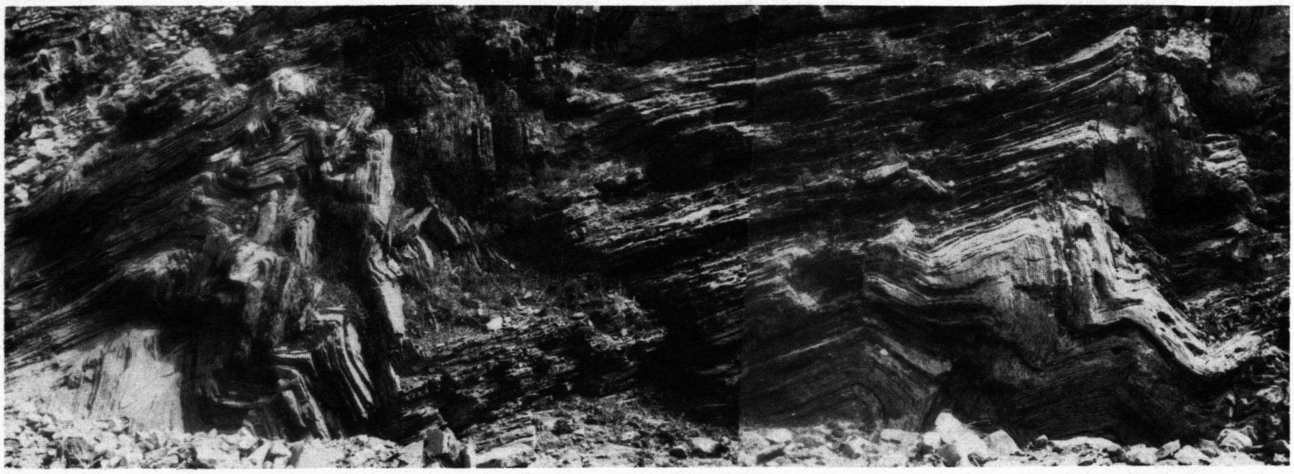

Fig. 14. Minor folds in the core of the limestone syncline north of Piedrasecha (model f. of Fig. 12).

southern limb of the main structure, whereas they still occur in the hinge of the syncline. This may be an indication that during folding the major synclinal axis migrated towards the south.

South of the major synclinal hinge nearly all minor folds have north-dipping axial planes. (Fig. 12g). The fold-plunges are low ($< 5°$) both eastwards and westwards. The minor folds are parasitic because:

a. the axial planes dip more steeply than the strata;

b. the axial planes are slightly fanning out from the centre of the main syncline;

c. the limbs are asymmetric, the long limbs coinciding with the general 40–60°N dip and the short limbs dipping 60–80°N. The short limbs of some of the folds are comparable with those in the core of the syncline. The axial planes of a secondary set of folds, occuring in the short limbs of the first set of minor folds, are parallel to the axial planes of the first set of minor folds. The secondary set of folds, however, have flanks of equal length.

Further southwards a first and second set of folds cannot be distinguished. The minor folds occuring here are intermediate in size between the two sets of folds described above. Their axial planes dip to the north (70–90°) with axes gently plunging to the west. The minor folds here are asymmetric, the longer limbs of antiforms dipping gently to the north and the short limbs dipping steeply south (Fig. 12h). The short limb is now straight in contrast to the folds in the centre of the syncline. These folds are regarded as parasitic to the major syncline.

Further south an irregular minor fold pattern occurs (Fig. 12j, k). Folds with horizontal axial planes change into folds with nearly vertical axial planes and vice versa. There seems to be no regular pattern of axial plane dips. Some of the folds show an asymmetry coinciding with that of parasitic folds as described earlier. Most of the fold forms are irregular with generally flat axial planes dipping towards the centre of the major syncline. These folds occur in the slightly coarser limestones which are in contact with more thinly bedded limestones. Their irregular forms are possibly due to the small differences in competency between the limestones. Therefore it is assumed that in this part of the synclinal structure parasitic as well as cascade folds occur.

Near the southern border of the limestone member of the Cuevas Formation the fold pattern becomes simple (Fig. 12k). The limestones are still thin-bedded and are almost vertical or dipping steeply to the north. The fold-hinges of synforms are often rounded and directed to the central part of the major syncline. Antiforms are often broken in the hinges. The synforms have long northern limbs dipping steeply north and generally flat lying north dipping southern limbs. This gives an asymmetry that characterizes cascade folds rather than parasitic folds.

Minor folds north of Olleros de Alba. – In limestones of the core of the southern syncline many minor folds have been observed. Especially north of Olleros de Alba both synclinal flanks are folded by minor folds. On the southern flank the folds generally have steeply south-dipping axial planes, and plunge gently west and east (Fig. 15, uppermost diagrams). The small anticlines have short southern flanks which are flat and south-dipping. The longer northern flanks dip steeply north. The hinges of the folds are sharp and show no stretching of any importance in the flanks. The style of folding is accordeon folding and resembles that of the Piedrasecha section. The geometric properties of most of the folds in relation to the main syncline classify them as parasitic folds. However, a conjugate set of minor folds has also been observed in this part of the syncline (Fig. 16). The fold-system in the photograph indicates extension in a horizontal sense and shortening in a vertical sense. The strain ellipsoid for this deformation then has a long horizontal axis and a shorter vertical axis. This implies that the main stress here has been a vertical one.

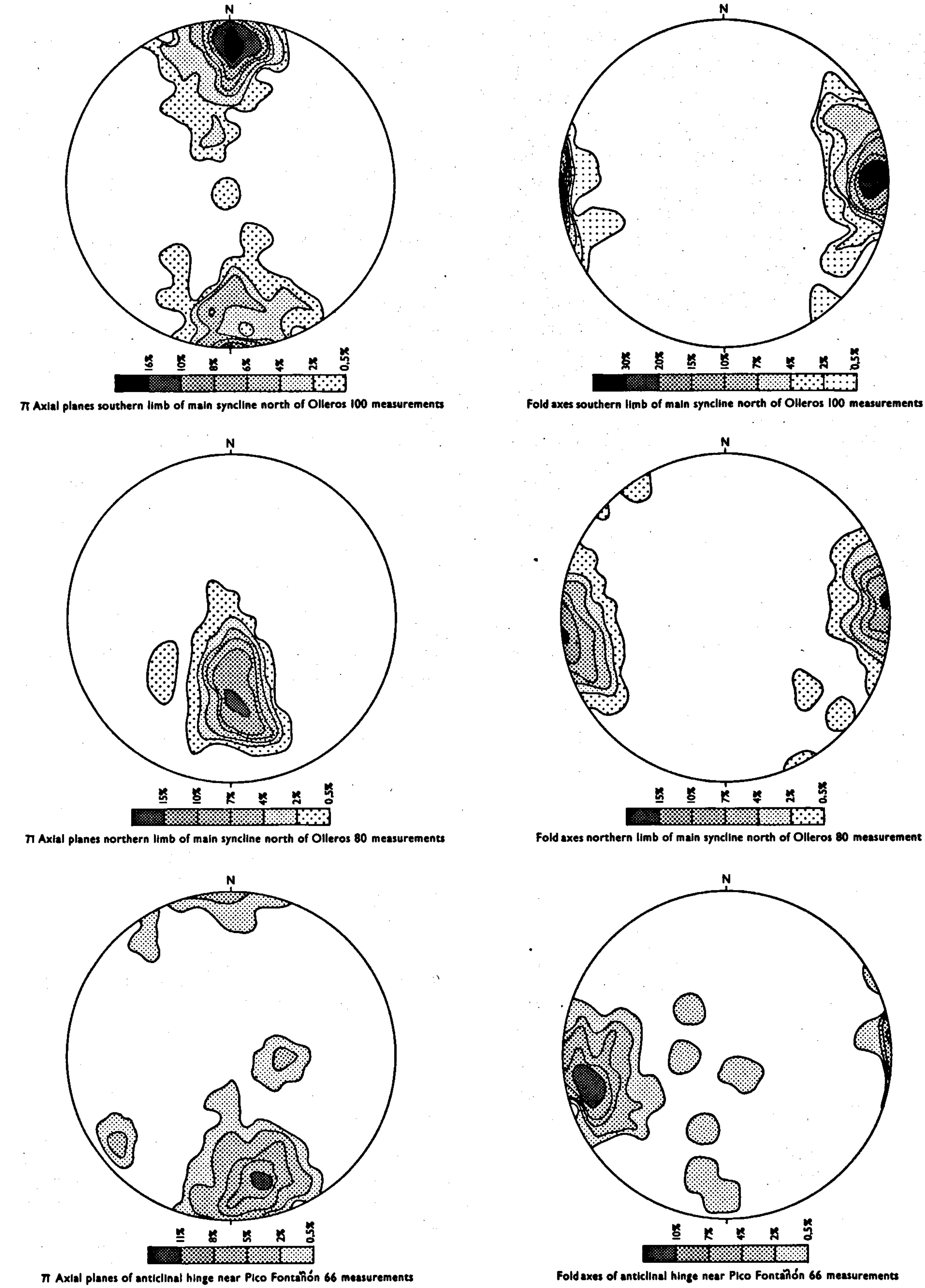

Fig. 15. Stereograms of minor folds in the Alba Syncline north of Olleros de Alba and near Pico Fontañon.

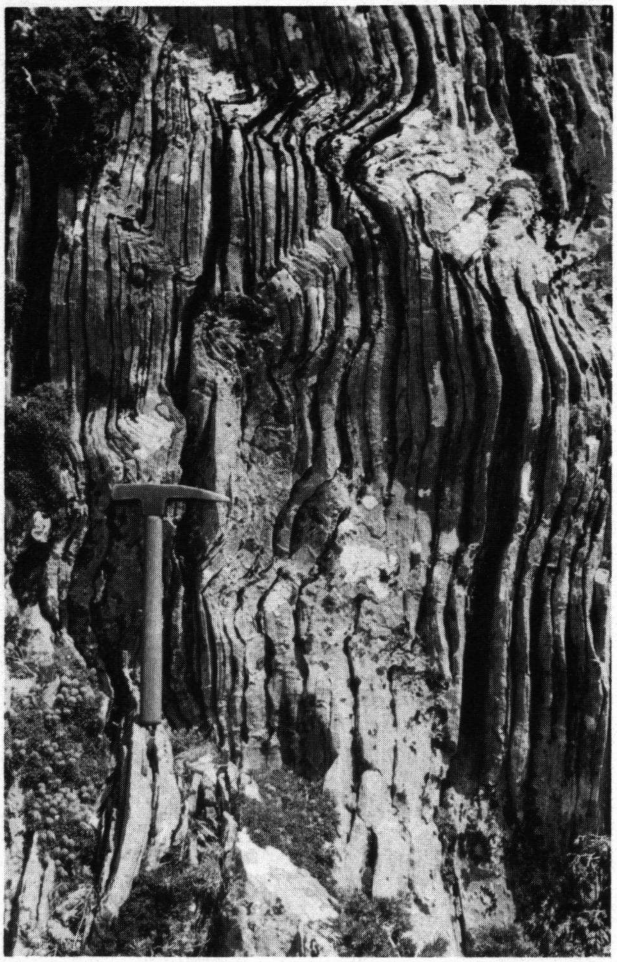

Fig. 16. Conjugate fold set of minor folds north of Olleros de Alba.

Therefore I suggest that these folds originated from a late stage of the folding, when the synclinal structure already existed.

Along the northern flank of the syncline the minor folds also show E-W trending axial planes. They dip northwards and are notably uniform, thus showing a clear point maximum in the stereograms (in the middle of Fig. 15). The folds are symmetric but symmetry is often disturbed by small faults. Most of them seem to fit with the geometric properties of parasitic folds for this part of the syncline. The fold-axes plunge with a bend (5°) in an easterly as well as westerly direction (Fig. 15).

Minor folds near Pico Fontañón. – In the flat anticlinal hinge of the limestones near Pico Fontañón, in the geological map erroneously indicated as Fontún, minor folds have axial planes, dipping steeply (60–80°) north, while the fold-axes plunge gently (5–10°) west (Fig. 15). The folds here are again of the accordeon type of folding but most of them are twice as large as the minor folds in the synclines north of Olleros and Piedrasecha. The folds are almost symmetrical, matching very well with the form expected for minor folds on hinges of anticlines (de Sitter, 1964, p. 282).

Minor folds north of Alcedo may be called typical of the synclinal hinge. The folds are accordeon folds. The axial planes dip steeply north (75–85°) along the northern synclinal flank. Fold-axes plunge east with a dip of 15–25°.

South of the synclinal axis minor folds occur in a limestone sequence that is interbedded between thick greywacke layers. The fold-pattern is irregular as the axial planes have various dips from horizontal to vertical. They are perhaps comparable to the conjugate folds as described above. In addition there occur parasitic and cascade folds as well.

Plunge pattern of minor folds in the Alba syncline (Fig. 17). – The plunges of axes of minor folds in the entire Alba synclinorium are generally about 5 to 10° west or east. They are parallel to the general strike, trending SE from Portilla de Luna to Santiago de las Villas, EW from Santiago to the Bernesga and WSW east of the Bernesga. In the western and southern parts of the synclinorium the folds plunge E, whereas east of the Fontañón most folds plunge W. The southern syncline seems to have a horizontal axis because there is no preference for any special plunge direction. In a small area west of the fault-contact in the valley of Santiago de las Villas fold-axes, plunging 50–70°E, occur.

Sabero-Gordón line

Rupke (1965, p. 39) and de Sitter (1965, p. 376) mentioned the Sabero-Gordón line as an important fault which represents a facies boundary during the Upper Devonian. Rupke (1965, fig. 14) traced the line from Valderrueda in the east to La Vecilla in the west and mentioned its western prolongation into the Bernesga region where it is considered to border the asymmetric Matallana basin to the south.

In the area south of the fault-zone a complete Devonian succession is present whereas north of it the uppermost Devonian Ermita Formation covers the Portilla Formation but leaving a hiatus. The difference in thickness is mainly due to the absence of the Nocedo Formation in the northern area (Fig. 5). This absence could have been caused by differential subsidence. This assumption is strengthened by the fact that near Beberino on top of the Portilla Formation indications of erosion have been found (Fig. 8). The limestones there show gulley-like structures. The erosion remnants out of the gullies formerly used to be interpreted as reef structures (Evers, 1967, p. 96). Sleumer (1969, p. 8) already recognized the limestone humps as the results of erosion, stating furthermore: "They cannot be interpreted as bioherms as they consist of normal biomicrodites and not of rigid coral and stromatoporoid skeletons".

The development of the stratigraphic sequence upon

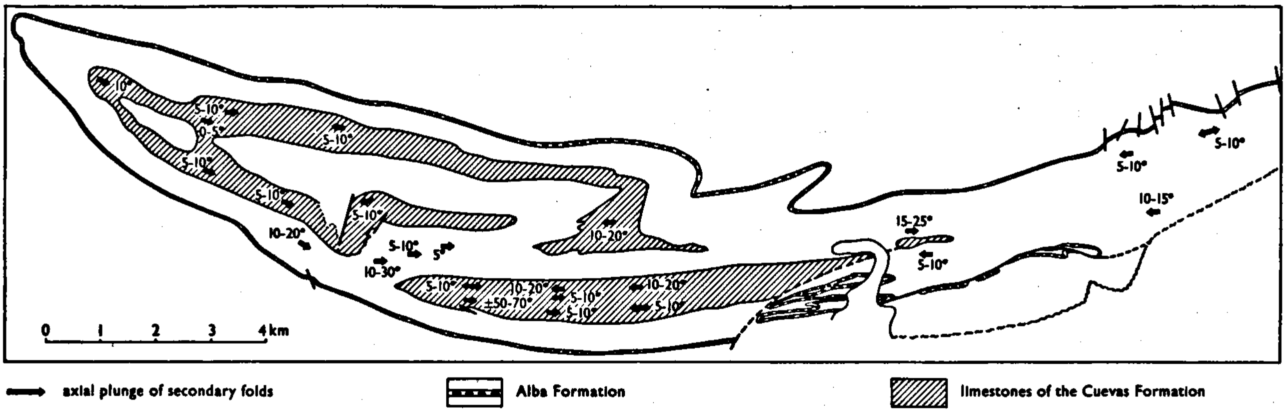

Fig. 17. Alba synclinorium with axial plunges of secondary folds; SE plunging axes in the western part of the structures and west plunging axes in the eastern part.

the Alba Formation also varies on the two sides of the fault-zone (Fig. 10) so that a facies boundary seems to exist.

The exposed structures in this fault-zone are strongly related to the lithologic properties of the formations exposed. The Devonian formations are much broken up by strike-faults. The multiple faults are quite clear in the west; east of Los Barrios de Gordón, however, only a single fault can be traced along the contact with the Prado Formation. Further east even this fault disappears, probably below the Prado rocks (sections 1–13, App. 4). A similar behaviour for this fault-zone has been recorded in the areas to the east by Evers (1967, p. 127) and Rupke (1965, p. 55). The differences between the zone in the west and the single fault in the east are mainly due to the different levels reached by erosion. In the west the zone is exposed where it is a fault-contact to the Santa Lucía Formation, a competent limestone mass, rather rigid in comparison to the surrounding shales of the Huergas Formation. In the east the fault-zone cuts the topographic surface at the level where the very incompetent La Vid Formation reacted to faulting by small scale folding, especially of the upper shale member. Near Lake Luna, just north of Mirantes, the Santa Lucía Formation is also strongly folded. These folds may have developed above the main Sabero-Gordón fault and although no faults have been drawn on the map, small movements may have taken place within the vertically dipping and isoclinally folded Santa Lucía Formation.

East of the Bernesga river the fault is exposed at the contact of La Vid shales and Prado conglomerates. Part of the Prado Formation as presented on the map west of San Mateo is now considered to be a marine succession of older Carboniferous age, comparable with the facies of the Cuevas and the San Emiliano Formations. The

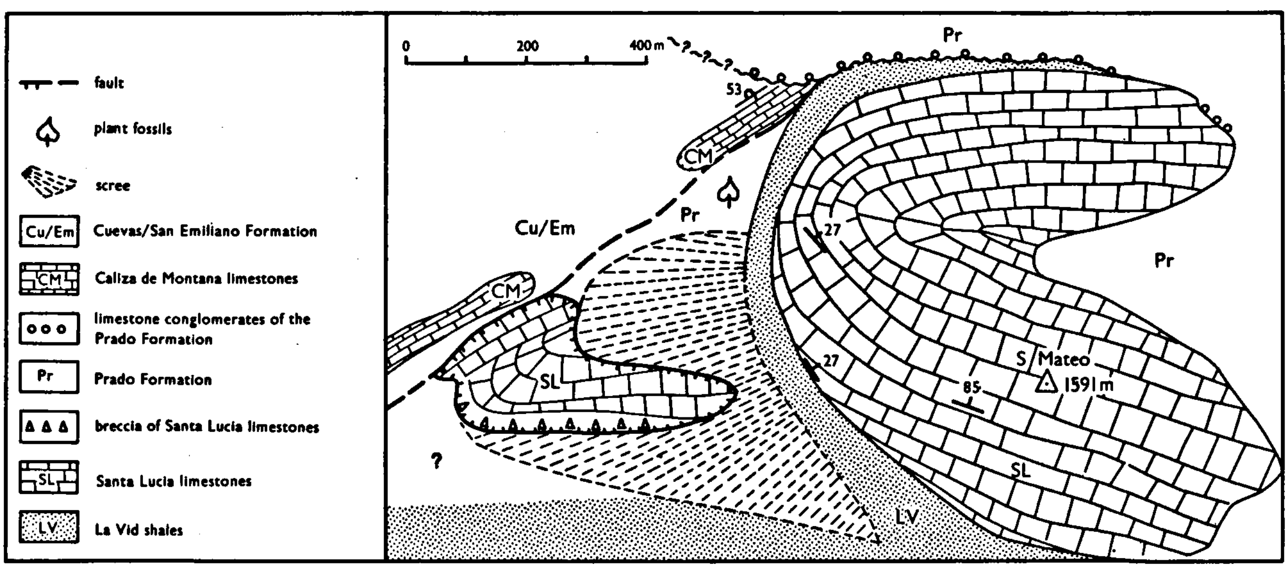

Fig. 18. Map-detail of the San Mateo structure.

map as given by Wagner and Artieda (1970, p. 32) gives an impression of the formations exposed in this area. The area around San Mateo is also reinterpreted as shown by Fig. 18. La Vid shales have been observed at the base of the Santa Lucía limestones in the hinge of the synclinal structure. In the north the limestone conglomerate of the Padro Formation covers La Vid shales.

The Santa Lucía lense-shaped exposure on the geological map is now looked upon as Caliza de Montaña limestones or even Cuevas limestones. The presence of the Carboniferous limestones suggests that the Sabero-Gordón fault runs along the north side of the San Mateo limestones. The San Mateo structure itself represents an asymmetrical syncline in the crest of a box-fold (see section 9, App. 4) with La Vid shales in its core. To the west of the limestone structure a mass of limestones have slid down to the valley. The sliding has resulted in a layer of brecciated limestones around the glided mass. On the geological map this breccia has led to the misinterpretation as being the fault-contact of the Sabero-Gordón fault. South of the SW-NE fault (Fig. 18) also Prado rocks are exposed. It is supposed that these rocks and the slid mass are related to each other in that they are of the same age.

Conclusions about the Sabero-Gordón line are now as follows: The zone has acted as a facies boundary during parts of the Devonian and Carboniferous sedimentation period. The fault-zone is therefore a fundamental line as mentioned by Rupke (1965) and Evers (1967). The structures in the present area are related to the lithologic properties of the formations exposed and not directly influenced by a fundamental fault in the depth. The place of the Sabero-Gordón line could therefore be anywhere in the area between the Alba syncline and the Pedroso syncline.

Pedroso syncline

The Pedroso syncline is bordered in the south by the Sabero-Gordón faults and in the north by the Beberino fault. Between Lake Luna and the Pedroso the nose of the syncline is clearly expressed by the Ermita, Alba and Caliza de Montaña Formations. North of Los Barrios de Gordón the synclinal structure is partly covered by the westernmost deposits of the Prado Formation in the Ciñera-Matallana Coal Basin and most probably loses its identity in the complicated folding between La Pola de Gordón and San Mateo Mountain.

The asymmetric syncline has an overturned or vertical northern limb. The asymmetry is accentuated by differences of thickness in the Ermita and Caliza de Montaña Formations (sections 1, 2 and 5, App. 4) in both flanks.

East of the Pedroso a small isoclinal anticline is developed along the steeply north-dipping north flank (App. 4, sections 3 and 4). The anticlinal axis trends almost SE-NW and forms an angle with the major ESE-WNW Pedroso syncline. The smaller folds plunge to the west so that a closure of key horizons such as the Alba Griotte can clearly be seen near the Beberino fault. This fold seems to have developed in connection with the swing in strike noticeable north of the Pedroso syncline.

Beberino fault

The Beberino fault is the E-W strike-fault that runs along the northern limb of the Pedroso syncline from Lake Luna to Beberino near the Bernesga valley and continues further east along the north of the Santa Lucía limestones of the San Mateo Mountain into the Ciñera-Matallana Basin. The Beberino fault is strongly controlled by the presence of the Santa Lucía Formation. Its structural aspects change from west to east. Near the Lake Luna the fault is hidden to the eye but two kilometres farther east the Huergas shales are in fault-contact with the calcareous shales of the upper member of the La Vid Formation. Irregulary north-dipping La Vid shales are part of the upthrown block whereas moderately south-dipping Huergas shales form the downthrown block. East of the Pedroso Mountain the apparent throw becomes zero. It increases again to the east, however, now with a downthrown northern part.

Further east the Beberino fault maintains its course along the Santa Lucía Formation cutting it obliquely from the base to the top of the limestones at a distance of a few kilometres from Pedroso to a point about one kilometre east of Cabornera. From that point onwards the northern part becomes the downthrown part and the Huergas shales of the southern block eastwards contact the Portilla, the Ermita, Alba and Caliza de Montaña Formations respectively. The fault attains its maximum displacement east of Beberino where Huergas shales are in contact with Caliza de Montaña limestones. East of the Bernesga the fault becomes involved in the complex structures around the San Mateo Mountain. A continuation of the Beberino fault is postulated below the unconformable Prado Formation of the Ciñera-Matallana Basin and further east it probably joins the Sabero-Gordón line.

Bregón unit

The Bregón unit comprises the southernmost thrust-unit of the Leonides. It trends E-W starting one km E of Ciñera and continues westwards to Lake Luna. Rocks ranging from the Láncara to the Santa Lucía Formations are incorporated in this structural unit. East of the Bernesga river it is obliquely cut off by the northern boundary-fault along the Ciñera-Matallana Coal Basin. Evers (1967, p. 128) assumes that east of this basin the unit reappears in the Montuerto syncline from below the Prado beds.

In the present area the thrust-unit has its greatest displacement in the east, where Cambrian limestones of the Láncara Formation contact with Devonian shales of the Huergas Formation. To the west the displacement along the fault is diminishing rapidly, partly through the upward cutting of the thrust through Láncara limestones to San Pedro sandstones and partly through the fact that westwards the covered sequences become older when changing from the Huergas Formation to the La Vid Formation.

From the Bregón Mountain to the area west of Lake Luna, where the fault enters the strongly faulted northern flank of the Abelgas syncline (van den Bosch, 1969), the fault is structurally unimportant near the surface.

In the railway cutting near Ciñera minor folds are exposed in the Oville and Barrios Formations just south of the main thrust (Fig. 19). Despite the proximity of the thrust-front the majority of these folds do not seem to have been actively involved in the thrusting. The typical parasitic symmetry to a thrust originating from a fold should have given minor folds with flat lying short limbs and steeply dipping long limbs. The exposed folds, however, have short flanks dipping steeply north and long flanks dipping gently south. In my opinion the folds as shown on Fig. 19 are due to gravity at a stage when the stratigraphic sequence already had an inclined position. The present nearly horizontal position of the axial planes could have been caused by refolding of the original south-dipping beds.

North of Geras de Gordón in the valley of the river Casares minor folds have also been observed. They only occur in the immediate neighbourhood of the Bregón fault-contact, where they have deformed the San Pedro sandstones. They do not show any favourite orientation in their axial planes.

Pozo unit

The Pozo unit comprises the thrust anticlinal structure which has its anticlinal nose in the Barrios quartzites of Pico del Pozo, west of the Bernesga river between Villasimpliz and La Vid. On the map the core of the anticline is erroneously indicated as the San Pedro Formation, this should be the Oville Formation (Evers, 1967). Rocks from Láncara to Huergas Formations have been found in this unit.

The front of the Pozo unit is a thrust-fault with the Oville sandstones as the oldest sequence against the fault-contact in the Bernesga valley as far as Villar del Puerto. From this point eastwards the Láncara dolomites are exposed along the fault-contact. In the Bernesga valley the thrust sequences are almost vertical but from Villar del Puerto to the east they are strongly overturned to the north.

From Pico del Pozo onwards in a westerly direction the structure can be recognized as far as the La Vid shales. There it disappears in the shales which are exposed north of Buiza.

The main feature of the unit is an anticlinal structure plunging slightly to the east. The shape of the structure is controlled by the folding of the Barrios Formation.

Fig. 19. Minor folds in the Oville Formation near Ciñera in the Bernesga valley.

East of Villar del Puerto small secondary strike-faults run almost parallel to the main thrust-fault. Just north of Pico del Pozo a small secondary syncline developed in the Barrios and Formigoso Formations. Its axial plane trends to the north and the fold-axis plunges strongly northwards (45°) and disappears near the thrust-front. Along the southern side of the unit synclinal and anticlinal folds occur in the competent Santa Lucía Formation (section 3). These folds disappear as soon as they reach into the incompetent La Vid Formation near Buiza. Generally the strata in this unit are north-dipping and overturned with the exeption of the noses of the fold structures.

Rozo unit

The Rozo unit, situated north of the Pozo unit in the east and north of the Bregón unit further west, consists of quite complicated structures west of the Bernesga. The sequence generally gets younger to the south. It has been thrust northwards over the youngest rocks of the San Emiliano Formation. East of Rodiezmo two thrusts developed, the Rozo thrust to the north and the Correcilla thrust cutting obliquely through the Rozo sequence in a SE direction (cf. Evers, 1967). The complications between these thrusts form an extension of the intricately folded Correcilla unit of Evers (1967). These are most striking in the Láncara Formation south of Rodiezmo (Fig. 20).

The cross-sections of Fig. 21 give an impression of the change in the structures from east to west. The thrusts,

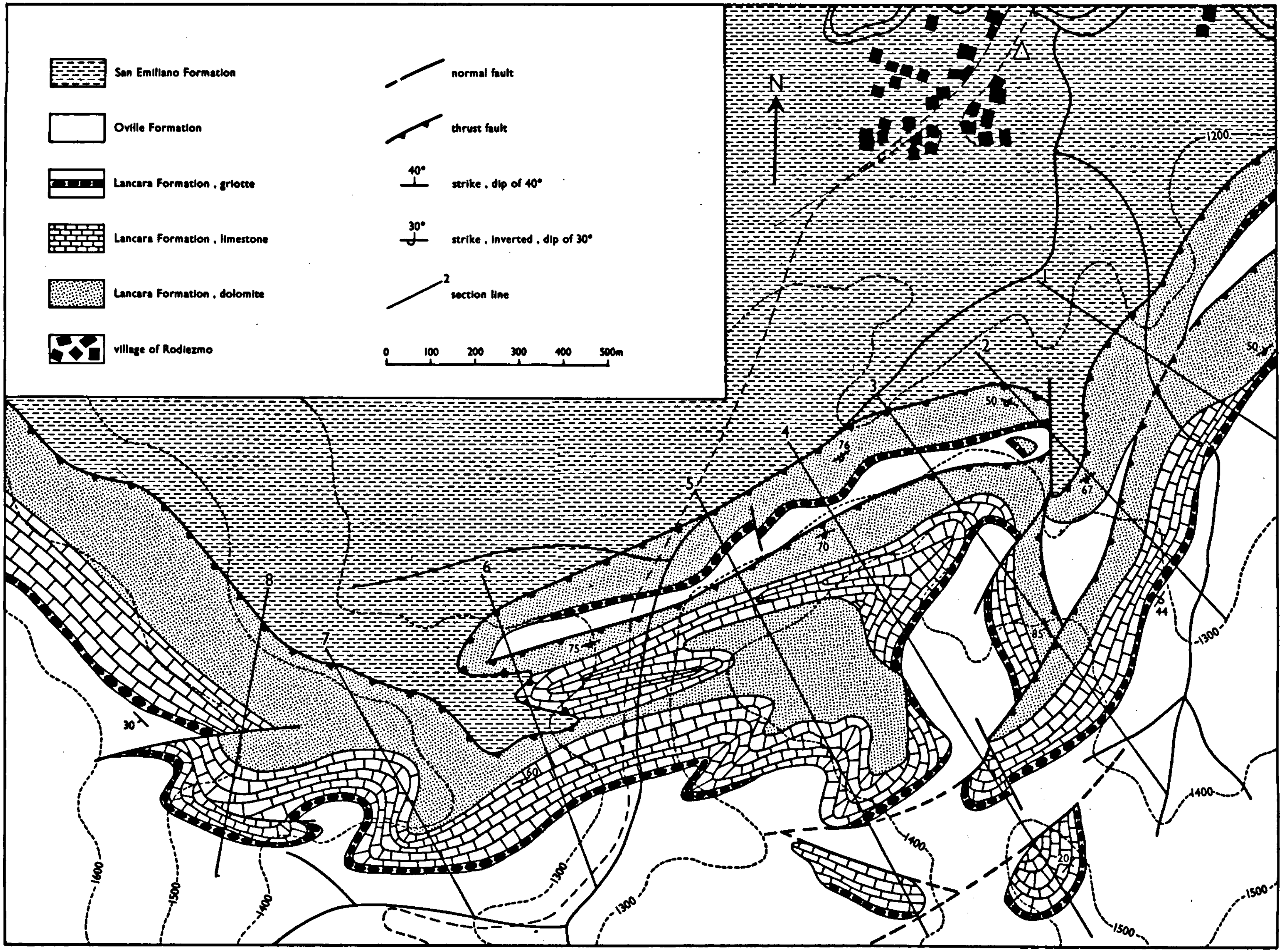

Fig. 20. Map-detail of the Rozo thrust unit, south of Rodiezmo.

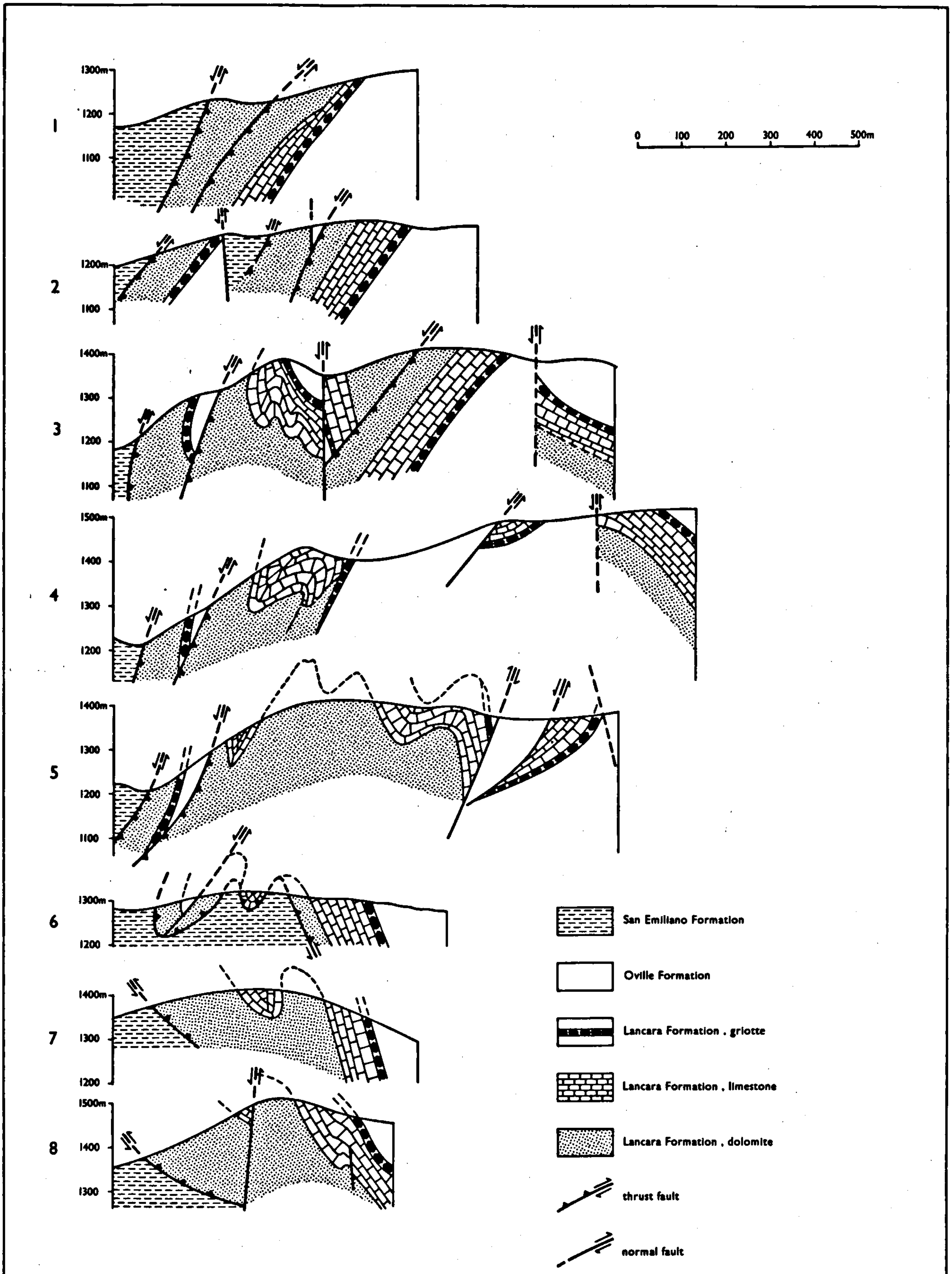

Fig. 21. Cross-sections of the Rozo thrust unit south of Rodiezmo; sections are indicated on the map of fig. 20.

originally probably south-dipping, became overturned at an advanced stage of the orogenesis and are now generally north-dipping. In the east the north-dipping thrust-front is composed of faulted and folded deposits of the Láncara and Oville Formations. In the west the same thrust-front is south-dipping without complicated structures. It can be followed as far as the Casares river. The two major thrusts can be followed through sections 1–6 (Fig. 21). The complicated structures south of the front are illustrated in Fig. 20 and in my opinion it is evident that they must be interpreted as refolding and simultaneous faulting of already existing thrust structures. This faulting has also obliterated the original thrust phenomena in some places. The N-S cross-fault (Fig. 20) between the older Palaeozoic and the Carboniferous San Emiliano deposits is perhaps connected with the secondary folding and faulting, but the antiform indicated on the geological map in the San Emiliano deposits is now considered to be nothing else but small scale crumpling along the fault-contact.

It is obvious that the northern blocks along the thrust-faults are downthrown. This agrees quite well with the idea that upthrusting from the south was the first tectonic movement, followed by folding and accompanying faulting. Faulting probably took place contemporaneously with folding.

It is difficult to account for the southernmost isolated Láncara outcrops in the Oville Formation. In my opinion these outcrops may be accounted for by assuming the occurrence of faults originating from refolding. It is quite possible for one of these faults to be the eastern prolongation of the fault which broke the Barrios quartzites more to the southwest (see geological map). It was impossible, however, to follow the fault-line through the Oville deposits between the Barrios quartzites and the isolated Láncara outcrops.

A third thrust level has been recognized in the Rozo Unit to the south of the others. The displacement is smaller; Láncara deposits are in contact with Barrios quartzites.

The geological pattern along the western boundary of the Rozo unit along the Casares river is similar to the pattern south of Rodiezmo (compare section 2, geological map and section 6, Fig. 21). Here the two phenomena, thrusting and refaulting of the thrust-front, are clearly expressed.

Aralla zone

The Aralla zone is the thrust strip north of the Bregón unit in the western part of the area, forming in essence the western prolongation of the Correcilla, Rozo and Pozo units together. They are represented by a thin thrust slice, about 2 km wide, less than half the thickness of the thrusts in the Beresga valley.

The Aralla zone was first mentioned by de Sitter (1962, p. 261), who included all deposits between the thrust-front and the Pedroso syncline. Now its southern part is considered to be part of the Bregón unit. The La Vid limestones and shales form the southernmost youngest rocks of the Aralla zone. The thrust-front is as a rule composed of Láncara dolomites but in some places the Láncara Formation is entirely absent and Oville shales form the contact with the San Emiliano rocks. Dips on either side of the thrust are steep, trending north. It is supposed that present occurrences of the deposits of the Peña de la Plaza are the result of downfaulting of the original thrust-front. This idea is strengthened by the fact that the Láncara deposits in this area show a very irregular folding pattern, whereas such irregular folding is completely absent in the original thrust-front. The area north of Aralla is also strongly faulted. The Láncara Formation is imbricated so that the dolomite member is repeated three times in its contact with Carboniferous deposits.

The igneous rocks in the Carboniferous just north of the thrust-front also indicate that the shales were broken to an extent enough to permit intrusions here. Moreover the inlier of Devonian deposits in the Collada de Alonga can only be explained by assuming faulting.

The Barrios Formation also plays its special role in controlling the structures in the Aralla zone. In the western part of the area, where the formation is relatively thin it conforms to the shape of the structure in the other formations. East of Peña de la Plaza the formation becomes thicker and cross-faulting testifies to the relative rigidity of the quartzites. Most of the cross-faults die out in the La Vid Formation. In Pico del Cueto the thick Barrios is also folded but even here two faults have developed to enable the much more extreme convolutions of the neighbouring beds to be accommodated.

Bodón unit

The Bodón unit was first described by de Sitter (1961) as a thrust-faulted sequence in the area north of Carmenes in the Torío valley. Evers (1967, p. 131) used the name for the eastern extension of that area, while the Bodón unit in this area is the western extension of the original.

East of the Bernesga it includes all formations from the Herrería as far as and including the San Emiliano Formation, bounded in the south by the Rozo thrust-fault and in the north by the fault that separates Leonides from Asturides. West of the Bernesga the unit loses its significance as a single unit; the area there will be discussed later under the headings Cueto Negro unit and Tesa-Barradal unit.

The front of the unit coincides with the Léon line as described by de Sitter (1961) and is here a fault contacting the Herrería Formation against the Lena Formation. A displacement of about 3.5 km may be estimated, this being the thickness of the stratigraphic sequence here. The fault-line between the Herrería and Lena Formations is badly exposed except for a few places where isolated measurements suggest a steeply north-dipping fault-plane. The front of the Bodón unit is further complicated by small secondary folds near Millaró and longitudinal faults near Busdongo.

The San Emiliano Formation is folded in E-W trending infraformational folds which are parallel to the main trend. Near Fontún and Barrio small faults have been found cutting off thin limestone beds. Some are normal faults and others longitudinal faults, possibly originating from the Gayo thrust-front (Evers, 1967, p. 132).

Cueto Negro unit

The Cueto Negro unit comprises the Celleros anticlinorium of Herrería deposits west of the main road between Busdongo and Puerto Pajares. Just north of the Sierra del Cueto Negro, Láncara, Oville and Barrios rocks are exposed which are interpreted as faulted outliers. The faults in the centre of the unit could not be traced throughout the area due to poor exposures. The faults are extensions of the faults in the Babia Baja unit (van den Bosch, 1969, p. 208), e.g. the core of the Robledo anticline and the eastern prolongation of the Grajos fault in the Villasecino anticline, (see also section 1, geological map). In the Cueto Negro Unit the Robledo fault is splayed into three or more E-W running longitudinal faults. From the Serronal syncline only outliers of the youngest deposits in the Herrería Formation are left.

The section through the Celleros area (geological map, section 2) has been constructed from measurements along a S-N line running across Los Celleros as well as from measurements in the railway tunnel east of it. In contrast to the concept presented here, Marcos (1968) assumes that the Láncara, Oville and Barrios Formations in the centre of the Cueto Negro unit are autochthonous and exposed as a tectonic window. Thus Herrería deposits are assumed to be part of a nappe thrust from south to north over the Cueto Negro unit. In this way he derives a horizontal thrust-movement over a distance of nearly 10 km, which is exceptional for thrust-faults occurring in the Leonides. The largest ever recorded have a displacement of not more than 4 km. Marcos' concept fails to provide an explanation for the connection between this nappe and the probable autochthonous nature of the Babia Baja unit.

Tesa-Barradal unit

The Tesa-Barradal unit is the most northwesterly unit of the Leonides and consists of two synclines and two anticlines (geological map and section 1). Its eastern boundary is the Herrería anticlinorium, its southern border the Rozo thrust-fault and its northern border the fault between the Láncara and Lena Formations. The structural elements have been described in their continuation to the west by van den Bosch (1969).

The Robledo anticline in the south is symmetrical with a WSW-ENE trending axial plane. The axial Robledo fault has clearly been caused by lack of space in the core of the anticline. To the north the Serronal syncline also trends ESE-WNW and plunges west. Its axial plane starts to crumple from the Formigoso nose onwards to the east. The axial plane is north-dipping. North of the Serronal syncline (section 1) lies the Villasecino anticline which is broken in its core by the Grajos fault, originally a left lateral wrench fault trending WSW (van den Bosch, 1960, p. 210). The northern flank of the anticline apparently has been upthrust by the Grajos fault. This may have occurred under the influence of the differences in thickness of the Barrios Formation on either flank.

The Tesa syncline is the most complicated of the four structures. Its irregular shape is partly due to stratigraphic differences on either side of the structure. In the southern limb the Caldas Formation is missing and La Vid shales directly underlie the Alba griotte and the Ermita sandstones. On the northern limb, however, a well-developed Caldas Formation is exposed. The Barrios Formation, too, is thicker on the northern flank of the syncline, and the Caliza de Montaña Formation shows great differences in thickness. The unusual thickness of the La Vid shales here are likely to be due to minor folding and crumpling. Normal faulting has also deformed the syncline as it was displaced by these faults. This faulting might have accompanied the small-scale secondary folding which took place on the northern limb of the structure, north of Pico Tesa. The axial planes of these folds are N-S orientated.

In the southern limb of the structure small-scale folding is found in the limestones of the Caliza de Montaña Formation. The asymmetry of these folds on the limb of the structure points to parasitic folds.

León line

De Sitter (1962, p. 255) defined the León line as a lineament separating the Asturides from the Leonides. The most characteristic function seems to be that of the divide between the distinctive stratigraphic sequences of the two blocks. In de Sitter's opinion the present fault between the Lena Formation and Lower Palaeozoic rocks represents the León line in this area and we have to consider whether the definition is fulfilled.

The fault separates Westfalian C deposits of the Asturian Coal Basin from Lower Palaeozoic deposits in the southern part. Therefore this fault has a minimum throw equal to the thickness of the stratigraphic sequence between the Herrería Formation and the Lena Formation, estimated at about 3.5 km.

In the field the fault was often difficult to follow except where the Láncara Formation occurs next to it. In the area between Millaró and Puerto Pajares the fault between the Lena and Herrería Formations was always found to occur in small topographic depressions in the ridges between the N-S valleys. The mapping indicated that the fault-plane is vertical nearly everywhere near the surface (geological map, sections 1–3). Since the fault is nearly always found between the Lena and Láncara Formations the structure is similar to that found in all the thrusts of the Leonides and it would seem reasonable to include it with them.

In the Leonides the thrust blocks were in most cases related to broken anticlinal structures which developed during folding. This relation seems to be confirmed here (van den Bosch, 1969, p. 217).

Marcos, (1968, p. 82) noticed that the fault was not only a thrust-fault but was later active as a normal fault as well. Near Millaró the fault cuts through the Lower Palaeozoic rocks as far as the Barrios Formation. Here it would seem that the fault was active even after the folding of the Lena rocks. The movements caused by the normal faulting possibly led to this upcutting near Millaró and may even have caused small-scale upthrusting from the Carboniferous deposits onto the Lower Palaeozoic succession. If the fault is a thrust-fault the age of thrusting can be derived by comparing the ages of the youngest rocks on either side of the fault. In the thrust block the youngest deposits are those of the San Emiliano Formation in the area west of Villamanín. Moore et al. (1971, p. 307) concluded that the age of the youngest San Emiliano Formation is Westfalian C. Llopis Lladó (1955) and Wagner (1962) on the other hand decided upon a Westfalian C and possibly Westfalian D age for the Lena rocks just north of the fault-contact. This implies that thrusting took place after the Westfalian C or even D.

The normal faulting cuts off the folded structures of the Carboniferous in the area between Millaró and Camplongo which must have taken place after the folding of these rocks or at least at the same time. This folding is generally considered to be a result of the Asturian folding phase, which is of pre-Stephanian B time in this area.

The faulting along the León line has also been thought of to relate to the presence of the Stephanian B-C basins in the areas to the east (Evers, 1967 and Rupke, 1965).

In the stratigraphic successions in the northern Leonides major differences in the facies of the San Emiliano and the Lena Formations do not exist. A small difference is that in some places in the Lena Formation the facies is locally continental as is proved by the presence of coal measures north of Arbas del Puerto, whereas in the San Emiliano Formation all deposits are marine. In the older Palaeozoic rocks there is a sudden change in the development of the Caldas Formation into the Tesa-Barradal unit. South of the Villasecino anticline and in the northern limb of the Tesa syncline the Caldas rocks are well-developed, whereas between the two structures these rocks are absent. This could be an indication that there used to be a facies boundary due to differences in the sedimentary basin. North and south of the Herrería anticlinorium there are no indications of facies differences.

The stratigraphic and structural indications seem to suggest that the León line in this area is of little significance. If present, it may run through the Villasecino anticline.

CENTRAL ASTURIAN COAL BASIN

The Central Asturian Coal Basin lies mainly north of the present map area but a small section of the southernmost part is represented by the rocks of the Lena Formation. In the eastern part of this area the E-W trending Corallón anticline takes a central place. The smaller structures change from an E-W direction in the east to a NW-SE direction between Millaró and Camplongo, while further west near Pajares two interfingering folds could be found with N-S and E-W trends (Fig. 22).

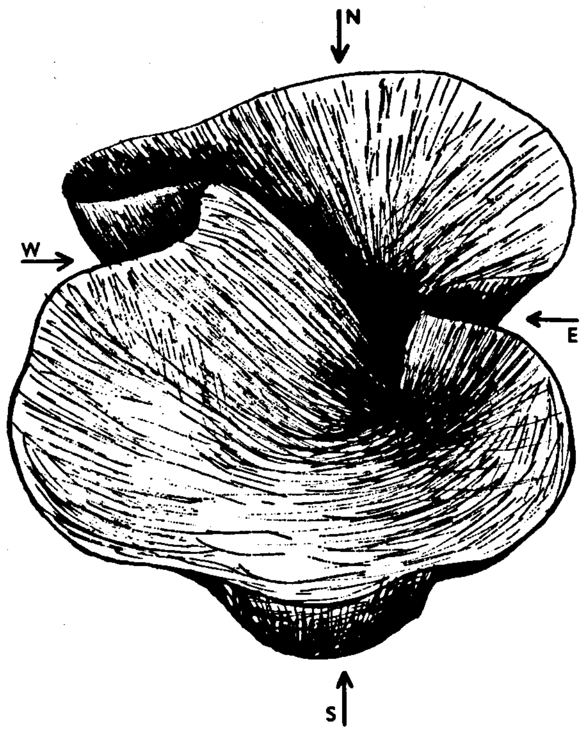

Fig. 22. Model of interfingering structures in limestones of the Lena Formation near San Miguel del Rio.

STEPHANIAN COAL BASINS

The Stephanian Coal Basins occur in two segments of the southern part of the map area. They are the Ciñera-Matallana Coal Basin in the east and the Magdalena Coal Basin in the west. The western part of the former is represented by the Prado deposits roughly between the Bernesga and Torío rivers, but also includes isolated outcrops near Valle de Vegacervera, north of Vega de Gordón and between Pola de Gordón and Barrios de Gordón. An intensive study of the stratigraphy and the detailed structural pattern of this coal basin has been made by Wagner (1963 a and b, 1964, 1972) and Wagner & Artieda (1970). Van Amerom & van Dillewijn (1965) and Evers (1967) also dealt with the structural pattern of this coal basin.

The coal basin lies unconformably upon a folded substratum of pre-Westfalian rocks. It was gently folded during a post-Stephanian and pre-Triassic (Table 3) orogenetic phase. The coal basin is an E-W trending synclinorium of which the synclines are well-developed, whereas the anticlines are strongly deformed due to faulting. The synclines occur in the depressions of the pre-Stephanian surface which developed after the folding of the Leonides. These structures are shown in the sections 10–13 of App. 4. Details of such N-S sections are presented by Wagner & Artieda (1970, figs. 34–38).

The northern boundary is often an unconformity followed by a mass of quartzite and limestone conglomerates except for the part of La Campa Mountain and the river Bernesga, where a fault-contact is undeniable. Along the southern border no conglomerates have been observed. This may be an indication that the supply of coarse sediments came from the north, where the conglomerates have been found, whereas in the south fine material was deposited. The fining of the material to the south may indicate that there was a certain asymmetry in the basin.

The structural pattern within the basin is one of concentrically folded synclines with E-W trending subvertical axial planes. The shape and width of the synclinal structures are governed by the depressions in the pre-Stephanian surface which on their part depend on the rock type exposed and the faults present in the Lower Palaeozoic rocks. It is supposed that these main faults of the substratum continue into the Prado rocks at places where the present anticlinal structures are exposed.

During the detailed mapping of Wagner and Artieda (1970) another set of faults was recognized. These generally south-dipping, small thrust-faults are only significant near the present surface. Apparently they do not reach the base of the coal basin (Wagner & Artieda, 1970, cf. figs. 34–38).

The Magdalena Coal Basin is largely comparable with the Ciñera-Matallana Coal Basin. An unconformable contact has been found in many places along the northern border and conglomerates are usually the basal sediments of the Prado rocks. From Olleros on to the east conglomerates are the only remnants of the Prado rocks. West of the Luna river the nature of the northern contact is questionable but a fault along an original unconformable contact seems to be the most appropriate explanation. A small part of the southern border is exposed south of Carrocera which is undeniably faulted. Mora deposits can be seen to have been thrust upon the Prado rocks. The basin itself is a syncline with a subvertical axial plane in the area east of the Luna river. West of the Luna river the basin is split up into two subbasins separated by Mora rocks. The subbasins show properties symmetric to the Mora rocks in the centre. Both basins have an unconformable contact with the Mora rocks in the centre and both basins have a faulted border at their outer side. On both occasions the Mora rocks have been thrust upon the Prado rocks due to that faulting (geological map, sections 1 and 2).

THE MOUNTAIN BORDER AREA

The Mountain Border Area includes the post-Carboniferous sediments in the south. The structural pattern is simple; no complicated folding and faulting has been observed. In the extreme south beds are horizontal but to the north this changes into vertical and even overturned beds against the Palaeozoic rocks.

In the eastern part of the area between the Bernesga and Torío rivers the contact is partly a fault and partly an unconformity. The fault occurs between Robledo and Candanedo de Fenar with a subvertical to steeply north-dipping fault-plane, (sections 12 and 13, App. 4). Near Solana the vertical fault-plane throws north-dipping Cuevas rocks against south-dipping Voznuevo beds. The straight contact between north- and south-dipping strata on both sides of the contact-plane suggests a fault-plane. The dips in the Voznuevo beds strongly varying between overturned north dips and 40° south dips over a short distance, also suggest that faulting was not restricted to one movement only but that post-Cretaceous movements took place probably during several periods.

Between Candanedo and Brugos de Fenar the contact is supposed to be an unconformity (Fig. 11), which has been proved by the observations of a fossil weathered zone at the top of the Carboniferous deposits immediately near the contact. The present shape of the structure near Brugos de Fenar is supposed originally to lie as a fold-thrust uplift system (Berg, 1962). Here the Carboniferous deposits were thrust upon the Voznuevo Formation, the latter was rotated into an overturned position by the same event.

The straight contact to the west of Brugos de Fenar between the Vegaquemada Formation and the Palaeozoic is considered to be a fault. The contact between conglomerates of the Prado Formation and Voznuevo beds west of Llanos de Alba is probably the original unconformity which was faulted later on. The relationship between the two faults which strike into the Bernesga valley from east and west is not clear, since there is no obvious connection by means of their trends. Since the structures in the Carboniferous and Devonian rocks in the La Robla-Alcedo-Llanos de Alba triangle are complex it can only be expected that other complexities must lie concealed beneath the recent deposits of the Bernesga river. To the west as far as Santiago de las Villas the Southern Border Fault is well-exposed in various places. In the absence of Cainozoic sediments further west no definite evidence for this type of faulting could be expected. In fact no faulting at all could be detected in the Prado Formation and it is thought that the relative fault movement was taken up again along the thrust-fault between the Prado and Mora rocks more than one kilometre to the south. The movement of this thrust has lifted the southern block upwards in contrast to the movements further east.

Other evidence as to the nature of the faulting has been gathered from geophysical observations. A gravity survey carried out by a team from the Geological Institute of the University of Leiden has yielded interpretable results owing to the density contrast between the Palaeozoic on the one hand and younger rocks on the other (int. rep. 1967). The interpretation implies a structural hinge near Carrocera, east of which fault blocks tended to drop down to the south, while west of it fault blocks dropped down to the north. The surface of the Precambrian rocks is supposed to dip quite strongly (20–30°?) to the east. In the Bernesga valley the interpretation yields the major fault near Alcedo with a fault-plane dipping 35° to the north. This is a further confirmation of the fold-thrust type of structure suggested from the area of Brugos and Candanedo (p. 177). The calculated fault-plane in the Torío valley dips much more steeply (± 60°) and the fault consequently outcrops in almost the same place as is given by the gravity interpretation (Matallana Estación, Evers, 1969, p. 146). The fault between Llanos and Brugos is therefore considered a minor feature, perhaps a splay or synthetic fault to the major border fault somewhat to the north.

CHAPTER IV

GEOLOGIC HISTORY

The Precambrian Mora rocks were deposited under unstable conditions (van den Bosch, 1969, p. 143) and are composed of erosion products of the Galician Precambrian (Matte, 1968b). The structural pattern of the Mora rocks differs from the structures of the Palaeozoic sequence. An angular unconformity between the two sequences demonstrates that the Mora rocks were folded before the deposition of the Palaeozoic rocks.

During the Lower Palaeozoic sedimentation took place under shallow marine conditions throughout the whole area. Silico-clastic material formed the Herrería, Oville and Barrios Formations, and only during the deposition of the Láncara Formation (Lower-Middle Cambrian) did carbonate deposition dominate. In the Leonides the stratigraphic successions do not show great differences in thickness.

After the deposition of the Formigoso Formation (Middle Silurian) differences in the stratigraphic development can be detected in that thicknesses of individual units vary trending to thin towards the N. This probably means that the rate of sedimentation in the S was greater due to more rapid subsidence. Deposition could still match the subsidence so that sedimentation could take place under shallow marine conditions.

During the deposition of the Santa Lucía Formation the differences between the southern and northern areas became more accentuated. In the south the Santa Lucía limestones were deposited under shallow marine conditions. In the north, however, the proximity of a coastline resulted in an increased supply of clastic material forming the Caldas Formation, which is partly the equivalent of the Santa Lucía Formation. There was even an area of non-deposition in the NE part.

The marine conditions did not change during the following part of the Devonian Period. Alternations of the supply of silico-clastic material and carbonates continued to exist, resulting in the deposition of the Huergas shales and the Portilla limestones. Later structural movements caused the absence of these rocks in the northern part of the area. Huergas rocks occur as north as the Correcilla unit (NE of Villar del Puerto). The Portilla Formation is exposed as far north as Beberino.

South of the Alba syncline the facies of the Huergas and the Portilla Formations varies from west to east, which makes it difficult to recognize their presence. A comparable change affected the deposition of the Nocedo Formation, so that the clastic sequence resting unconformably upon the Santa Lucía Formation in this area has been renamed Piedrasecha Formation.

Along the north flank of the Alba syncline the Nocedo Formation is fully developed but north of the Sabero-Gordón line it is absent. This means that at least at the time of deposition of the Nocedo Formation (approximately Frasnian-Famennian) the northern area was uplifted. At the end of the Devonian the irregularities causes by differential subsidence had been levelled off by the deposits. The sea then transgressed again, flooding the entire area and allowing deposition of the Ermita Formation upon the Nocedo Formation in the south, upon the Portilla Formation further north near Pola de Gordón and Beberino, upon the Huergas Formation north of Villar and upon the Caldas and La Vid Formations in the northern Leonides.

While north of the Sabero-Gordón line differential subsidence followed by a transgression and sedimentation of the Ermita Formation provides an adequate explanation of the variations in the stratigraphic succession, there remains a problem when this area is compared with adjacent regions. Evers (1967) and Rupke (1965) described the deposition of the transgressive Ermita Formation not only upon Devonian deposits but also on older Palaeozoic rocks, e.g. the Oville Formation. As the facies and the thickness of these older rocks do not differ from south to north, differential subsidence could not be the cause of the disconformity in this case. Therefore it is suggested that uplift and erosion of the older Palaeozoic had taken place in the north before the deposition of the Ermita Formation. De Sitter (1962) called this uplift the Bretonic uplift, suggesting a relation with the uplift at the end of the Devonian in France. It is suggested here that both phenomena, differential subsidence and emergence, took place in this area.

The Sabero-Gordón line apparently acted as a hingeline during these movements, as no gaps in the stratigraphic succession can be found further south. As we shall see below, this basical structural line also strongly influenced the kinds of structure that were to develop at a later stage of deformation.

In the Tournaisian the Vegamián Formation was deposited in depressions under shallow marine conditions as a condensed sequence with synsedimentary erosion and chert deposition.

During the Viséan the Alba Formation was deposited and the sedimentation conditions were the same all over the area. In the Namurian, however, again differences in facies developed and once more the Sabero-Gordón line acted as a hinge-line.

South of it the Cuevas Formation begins with a clastic sequence, followed by black bituminous, thin-layered limestones. After the deposition of the black limestones, which may point to quiet conditions, coarse clastic detritus prove the onset of unstable conditions, possibly due to oncoming tectonic activity. The succession of the Cuevas Formation ends with deposits comparable with the basal clastic part.

North of the Sabero-Gordón line black limestones comparable with those of the Cuevas Formation overlie the Alba Formation. They are followed by a more massive and coarser limestone sequence. All these limestones are included in the Caliza de Montaña Formation. Clastic sediments, considered to belong to the San Emiliano Formation, are found on top of the limestones.

It was impossible to establish the age of the various clastic sequences upon the limestones in the Alba and Pedroso synclines. As they are the youngest deposits in these structures their original thickness is unknown but a minimum of 200 m could be measured.

North of the Rozo unit the San Emiliano Formation provides more details. The age of the youngest clastic rocks here is established by Moore et al. (1971) as Westfalian C, whereas the top of the Caliza de Montaña limestones is of Namurian B age. During the Westfalian A these authors suggest a break in the sedimentation.

The relation of deposition-time between Namurian B and Westfalian C and the tickness of about 2,500 m of the San Emiliano Formation gives us a sedimentation rate of approximately 0.25 mm/year. This could only occur in a subsiding basin in which the unstable conditions governed the kind of sediments deposited. Rapidly changing conditions are also proved by the great variety in fossil assemblages. This subsiding basin in the northern part of the area ends the sedimentary history of the Leonides.

In the Asturides the Central Asturian Coal Basin had already started to develop from the Namurian C onwards (Sjerp, 1967; Julivert, 1957). Part of the deposits in this basin have been named Lena Formation. The rate of sedimentation increased rapidly after the Westfalian A, comparable to what happened in the Leonides. In both areas a minor hiatus represents the lower part of the Westfalian, and the greatest thickness developed afterwards. The deposition of the Lena Formation, which, as far as the present area is concerned, continued till the end of the Westfalian D; a period of non-deposition followed in which the Leonides and the Asturides were folded (Asturian folding phase).

After this folding continental intramontane coal basins developed during Stephanian B time. The strike of these basins agrees with the general trend of the folded belt, e.g. an EW orientated longitudinal axis of the basins. With a continental facies of the Prado Formation, resting unconformably upon older Palaeozoic, sedimentation came to an end in this area. A folding phase after the Stephanian B and before the Triassic folded the Prado rocks into regular concentric folds.

At the end of the Upper Cretaceous a marine transgression approached this area along the folded belt from the east. The sandstones of the Voznuevo Formation form the continental counterparts of marine deposits, successively younger towards the west. In this area only small marine limestone beds have been found exposed as far west as Brugos de Fenar. After sedimentation of the sandstones and thin limestone a subsequent regression changed the deposition conditions eventually resulting in the continental facies of the Vegaquemada Formation.

At that time the morphogenetic uplift of the Cantabrian Mountains started. This uplift and the consequent Southern Border Fault resulted in the Vegaquemada, Candanedo and Riacos Formations due to fluvial and torrential activity along the hilly southern border.

CHAPTER V

CONCLUSIONS

Most of the structures in the area of the present map sheet originated in the Hercynian orogenic Period (Table 1), which mainly caused the fold belt of the Palaeozoic rocks in this area. The Cantabrian Mountains which now expose this folded belt are due to a morphogenetic uplift during the Alpine Period.

The Narcea Anticlinorium is one of the Hercynian fold-structures in which we find the Precambrian Mora rocks in the core of the anticlinorium separated by an unconformity from the possibly Palaeozoic rocks of the Herrería. The fold structures in the Mora deposits are cut off by this unconformity in several places. Thus a pre-Hercynian deformation must have caused these structures, which moreover differ from the Hercynian deformation in that they are accompanied by slaty cleavage in the fine grained sequences and by fracture cleavage in the greywackes. In most cases the cleavage is an axial plane cleavage belonging to folds cut off by the unconformity. The folds in our part of the area mainly trend E-W parallel to the Palaeozoic structures, giving no indication as to their Hercynian or pre-Hercynian origin. Near the unconformity greywackes are mostly exposed, often showing fracture cleavage. The same direction of cleavage has been observed in the Herrería sediments on the other side of the unconformity. Within 30 m from the unconformable contact the fracture cleavage of the Herrería Formation generally ends. The fracture cleavage in the Herrería could therefore be regarded as a reactivation of the pre-Hercynian cleavage during the Hercynian deformation, especially because the cleavage has only been observed near the unconformable contact.

The Hercynian deformation is mainly expressed in the structures of the Leonides. Within the Leonides different structural patterns can be distinguished. The Sabero-Gordón line separates intensely folded rocks in the south from folded thrust-structures in the north.

South of the Sabero-Gordón line a subsiding basin was filled up during the Palaeozoic up to the Namurian or even Westfalian. A well-developed syncline, the Alba syncline, resulted from the Hercynian orogenesis. In the core of the syncline Cuevas limestones have been deformed by intensive minor folding. From the asymmetry of these minor folds to the main structure most of them seem to be parasitic, though not parasitic in the sense as given by de Sitter (1964, p. 282) as they originate from steeply dipping (>45°) strata.

The core of the synclinal limestone structure north of Piedrasecha contains minor folds which show an asymmetry expected for folds south of the core. This may imply that the synclinal axis migrated to the south during folding. The migration of the fold-axis could have been caused through activity of the fundamental Sabero-Gordón line at the northern border of the basin. This line already caused differential subsidence during the Upper Devonian and even during the deposition of the Cuevas Formation. Renewed activity during the folding could easily have been the cause of tilting of the northern border of the basin and the consequent migration of the fold-axis to the south as well. North of Olleros de Alba a conjugate fold set (Fig. 16) indicates that during the folding minor folds developed which are obviously due to a vertical stress-field, e.g. gravity.

The migrating synclinal axis of the major syncline and the occurrence of conjugate fold sets prove it to be dangerous for the conclusion of folds being parasitic to be based on the symmetry concept.

The structures of the Sabero-Gordón fault-zone do not show any influence of movements along a fundamental line. On the contrary the fold structures and the faulting here are strongly related to the lithologic properties of the formations exposed, resulting in box-folding of the Santa Lucía Formation (e.g. San Mateo structure) and strike-faulting along the limestones of the Portilla and Santa Lucía Formations. Due to their incompetency the la Vid shales almost everywhere form the core of the box-folds.

The deformation of the Leonides north of the Sabero-Gordón line started with the thrusting of the units discussed above during the Sudetic initial folding phase. The main folding during the Asturian folding phase brought the thrust units in their generally overturned steep north-dipping position. In the present area there are no indications that the two phases worked separately. On the contrary it seems quite likely that thrusting and folding are progressing developments, making it possible for them to be put together in one particular orogenetic phase. Post-thrusting structures are observed in the minor folds within the thrust units. They probably developed due to gravity-stress. Gravity also influenced the deformation of the thrust-front of the Aralla Zone, giving rise to imbricated structures of the Lower Palaeozoic rocks with shales and greywackes of the San Emiliano Formation.

The fault that separated the Lena Formation from the Lower Palaeozoic in the northern part of the area is considered to be the boundary between Leonides and Asturides. The structural aspect of this fault offers no reason to relate it to the fundamental León line as defined by de Sitter (1962).

The structural pattern of the Asturian Coal Basin shows an interference of E-W trending and N-S trending structures (Fig. 22), which is probably due to a general bend of the Palaeozoic basin from an E-W direction to a N-S direction west of the present map sheet area (Asturian Knee).

Structures in the Leonides trending N-S (e.g. the fold structures in the Rozo unit, Fig. 20) are probably also due to the bending of this basin.

Intramontane coal basins developed in the lower parts of the folded Leonides and are folded concentrically during the post-Stephanian-pre-Triassic (Saalic) phase.

The present configuration of the Cantabrian Mountains was finally obtained during the morphogenetic Savian phase, which uplifted the Palaeozoic belt. Along the Southern Border Fault unconformable Cainozoic rocks are partly thrust during this uplift.

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