LEIDSE GEOLOGISCHE MEDEDELINGEN, Vol. 44, p. 137–225, published separately 10-6-1969

GEOLOGY OF THE LUNA-SIL REGION, CANTABRIAN MOUNTAINS (NW SPAIN)

BY

W. J. VAN DEN BOSCH

ABSTRACT

After a late-Precambrian folding, clastic deposits, partly continental, spread out over the region during the Lower Cambrian; later, marine intercalations became more abundant when upper Lower Cambrian marine sediments were deposited. This sedimentation continued until the Upper Carboniferous and occurred on a generally stable shelf between a rise in the NNE (the Asturian Geanticline) and a curved geosyncline (present at least until the Devonian) in the SSW. Unstable conditions prevailed during the Upper Cambrian and the Couvinian-Givetian.

The Asturian Geanticline extended several times towards the SSW, which resulted in emergence, accompanied by erosion, of the Luna-Sil region during the Llanvirn-Llandovery, during the Gedinnian (only the NNE part) and at the end of the Devonian. The last epeirogenic uplifts resulted in strong erosion, especially in the NNE part of the region: Famennian-Tournaisian deposits unconformably overlie rocks of a Gedinnian age. These uplifts occurred along fundamental faults. Slight epeirogenic uplifts along the same faults during the Lower Viséan resulted in the erosion of the very thin black shales of a Tournaisian age.

With the beginning of the greywacke deposition during the Upper Namurian B, initial to the folding and thrusting of the region, the palaeogeographic pattern changed: the source area was now situated in the SSW, while a rapidly subsiding basin lay in the place of the former shelf. The line of maximum sedimentation, just as the folding front, was displaced during the Namurian and Westfalian from SSW to NNE. After the orogenesis, followed by strong erosion, oblong basins developed along normal faults approximately parallel to the strike of the orogene, in which thick coal-bearing sediments accumulated during the Stephanian B and C; these sediments were folded during the Lower Permian.

During the orogenesis thrusts were formed, generally parallel to the strike of the former curved basin, which developed from the breaking through of anticlines. These folds and thrusts have their vergence to the NNE. The shape and amplitude of the folds were mainly determined by a thick Ordovician quartzite formation. During the folding and thrusting a tectonic 'Stockwerk' was formed in the Palaeozoic rocks, each of the 11 levels in this 'Stockwerk' possessing its own tectonic style. During continued compression, movement along the thrust faults ceased and these faults were involved in the folding. This last stage of the folding has its vergence to the SSW.

One important and a number of less important WSW trending faults cross the folds and thrust faults which are generally parallel to the chiefly WNW strike of the former curved basin. Due to a reorientation of the stress field in a direction perpendicular to the WSW trending faults, deviation of the generally WNW trending structures took place in a direction parallel to these faults. This deviation is strongest near the ends and nil in the central parts of the WSW trending faults.

SUMARIO

Después de un plegamiento Precámbrico tardío, se esparcieron por la región durante el Cámbrico inferior, depósitos parcialmente continentales; más tarde, se hicieron mas abundantes intercalaciones marinas y desde las postrimerías del Cámbrico inferior se depositaron sedimentos marinos. Esta sedimentación continuó hasta el Carbonífero superior, teniendo lugar en un *shelf* generalmente estable situado entre un levantamiento en el NNE (el Geanticlinal Asturiano) y un geosinclinal curvado (al menos presente hasta el Devónico) en el SSW. Durante el Cámbrico superior y el Couviniense-Givetiense prevalecieron condiciones inestables.

El Geanticlinal Asturiano se extendió varias veces hacia el SSW, lo que condujo a una emergencia, acompañada de erosión, de la región del Luna y del Sil durante el Llanvirniense y el Llandoveriense, durante el Gediniense (solamente la parte NNE) y al final del Devónico. Los últimos levantamientos epirogénicos provocaron una fuerte erosión, especialmente en la parte NNE de la región, por lo cual los depósitos del Fameniense y del Tournasiense cubren con una discordancia a rocas de edad Gediniense. Estos levantamientos tuvieron lugar junto a fallas fundamentales. Pequeños levantamientos epirogénicos junto a las mismas fallas durante el Viseense inferior condujeron a la erosión de esquistos negros de poco espesor de edad Tournasiense.

Con el comienzo de la deposición de grauvacas durante el Namuriense B superior, anterior al plegamiento y corrimiento de la región, cambió el patrón paleogeográfico. La región de procedencia del material sedimentario estaba situado ahora en el SSW, mientras que una cuenca de subsidencia rápida ocupaba el lugar del *shelf* anterior. La línea de sedimentación máxima, igual como el frente de plegamiento, se desplazó durante el Namuriense y el Westfaliense del SSW al NNE. Después de la orogénesis, seguida de fuerte erosión, se desarollaron cuencas oblongas junto a fallas normales, aproximadamente paralelos al rumbo del sistema orogénico, acumulandose aquí sedimentos carboníferos durante el Estefaniense B y C; estos sedimentos fueron plegados durante el Pérmico inferior.

Durante la orogénesis se formaron corrimientos, generalmente paralelos al rumbo de la antigua cuenca curvada, que se originaron debido a la rotura de anticlinales. Estos plegamientos y corrimientos se inclinan hacia el NNE. La forma y la amplitud de estos plegamientos han sido principalmente determinados por una formación gruesa de cuarcitas Ordovícicas. Durante el plegamiento y corrimiento un *Stockwerk* tectónico se formó en rocas Paleozoicas, teniendo cada uno de sus once niveles un estilo tectónico propio. Mediante una compresión continua, el movimiento a lo largo de las fallas de corrimiento, cesó y estas fallas fueron envueltas en el plegamiento. La última fase del plegamiento tuvo su inclinación hacia el SSW.

Los plegamientos y las fallas de corrimiento que generalmente son paralelos al rumbo WNW de la antigua cuenca curvada se ven cruzados por una falla importante en dirección WSW y otras menos importantes en el mismo sentido. Debido a la reorientación del campo de tension, en una dirección perpendicular a la falla de orientación WSW, tuvo lugar una desviación de las estructuras de orientación generalmente WNW en una dirección paralela a estas fallas. Esta desviación es mas fuerte junto a los extremos y nula en las partes centrales de las fallas de orientación WSW.

CONTENTS

INTRODUCTION

General statement

The map presented with this study forms part of a mapping program which the Department of Structural Geology of Leiden University has been carrying out in the southern and central parts of the Cantabrian Mountains: de Sitter & Boschma (1966), Rupke (1965) and Helmig (1965), Sjerp (1967), Evers (1967) and van Staalduinen (1969).

Location and accessibility (fig. 1)

The region mapped by the author is situated mainly in the NW part of the province of León and, for a small part in the NE, in the province of Oviedo, between lat. 42°46′ and 43° N., and long. 2°11′ and 2°39′ W. of Madrid, which corresponds to 5°53′ and 6°21′ W. of Greenwich. The eastern limit coincides over a long distance with the Luna lake, the road La Magdalena-Villablino has been taken approximately as the southern limit, whereas in the north a straight line was drawn, generally just south of the watershed between the provinces of León and Oviedo. The following districts, all bearing old names, lie in the mapped region: Laciana, Babia, Luna, and Omaña.

The Cañada and the Yegüero, respectively 2163 and 2186 metres in height, are the highest summits. The lowest valleys are found in the NE in the province of Oviedo (1060 m), in the SE south of the Riello basin (1055 m), and in the west near Villablino (980 m).

A paved road runs from León, along the Luna lake, to Villablino; two side-roads, one going north via San Emiliano, the other via Meroy, have recently been paved. The road from León, via Riello, to Villablino has been paved as far as Guisatecha, just west of Pandorado. Villages are chiefly accessible along poor unpaved roads or tracks.

History

The oldest traces of civilization found in the Luna-Sil region have been left behind by the Romans. Near Mallo still lie the remains of a Roman road and bridge, now concealed by the lake. This road connected Asturias, via Puerto Ventana, with the rest of Spain (Teijón, 1946). Near Mena stand the remains of a fortification dating from the time that the Visigoths conquered the region from the autochtonous Celtic population; a sculptured stone in the church of Torrebarrio also dates from that period. After a short Moorish occupation, armies from Asturias freed this part of the Cantabrian Mountains and in the last battle Alfonso II defeated Hixem I north of Los Barrios de Luna. Starting from these mountains the whole of Spain was liberated. Alfonso III built a fortress near Los Barrios de Luna for the defense of the region; remains of it are still visible near the dam site. Smaller defense towers stand in Torrebarrio and Torre de Babia (fig. 2). There is a story relating that El Cid's famous horse, named Babieca, came from Torre de Babia. Quiet times followed. In Piedrafita monks of the Order of Paular de Segovia built a fortification; in Luna the Duke of Luna was the most important man.

During the early years of the 19th century French armies occupied the northern part of Spain. Again the Cantabrian Mountains were a base for the *guerrilleros* who attacked, and sometimes conquered, towns and fortresses on the Meseta, the vast plains south of the Cantabrian Mountains.

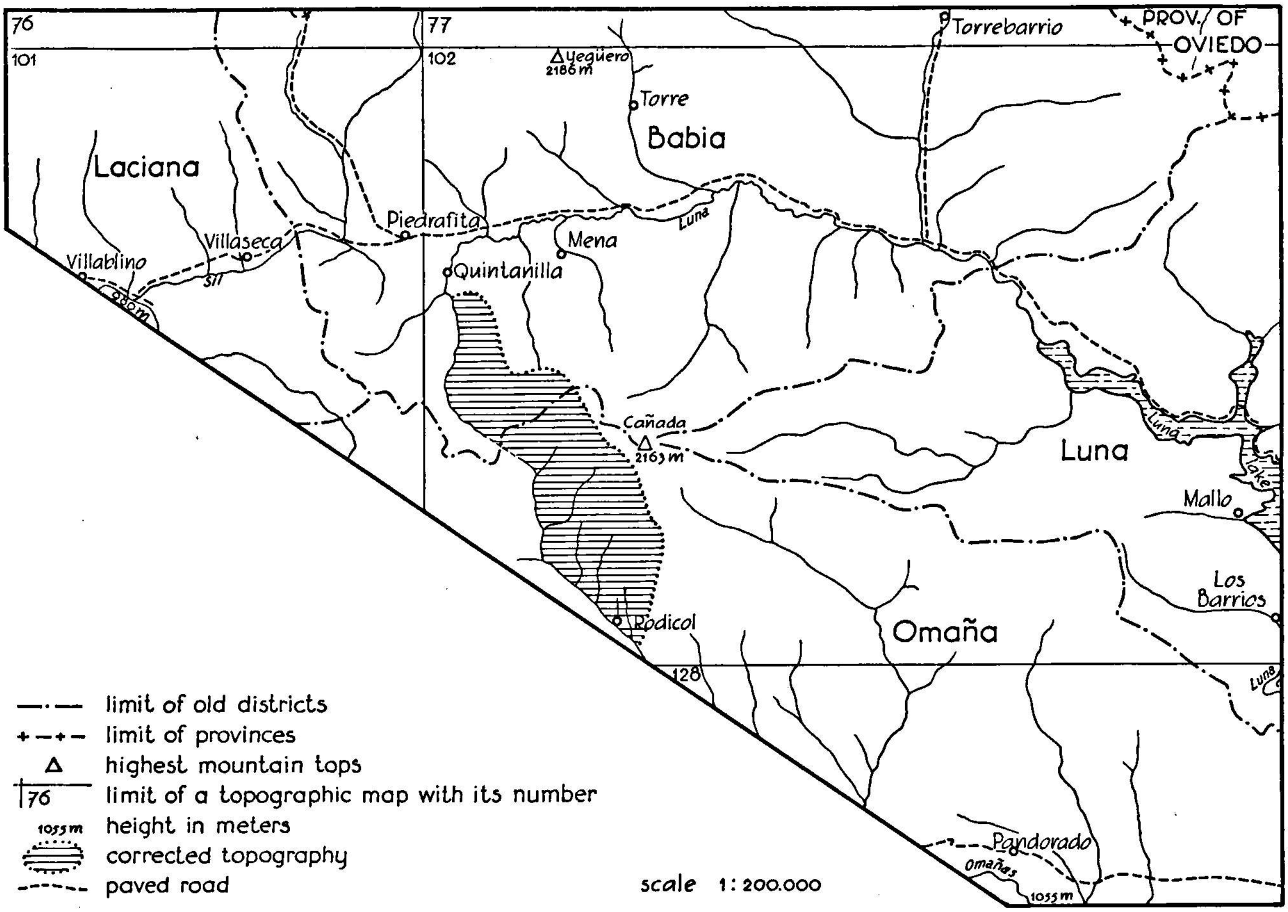

Fig. 1. Topographic features

During the Spanish civil war (1936–1939) many battles were fought in the passes connecting Oviedo and León.

There are expressions in the Spanish language referring to the region of Babia and its inhabitants: '*estar en Babia*' (to be in Babia) which means to be non-attentive, dreaming; a '*Babieca*' is an imbecile. But Babian people do not merit these qualifications; they are hard-working people, mainly living of rearing cattle, selling milk and making the famous Babian butter. Many men work in the coal mines of Quintanilla and Villaseca.

Reafforestation with pines took place on a small scale near Piedrafita and was started some years ago in Luna.

In regions far from the villages still live chamois, deer, capercailzies, wolves, boars and bears.

Hydrology

The Luna belongs to the drainage area of the Duero. This river, after passing the Spanish Meseta, discharges into the Atlantic Ocean near Oporto; the Sil flows directly to the Atlantic Ocean and, bearing the name Miño, constitutes the northern border of Portugal. The Sil is capturing the Luna.

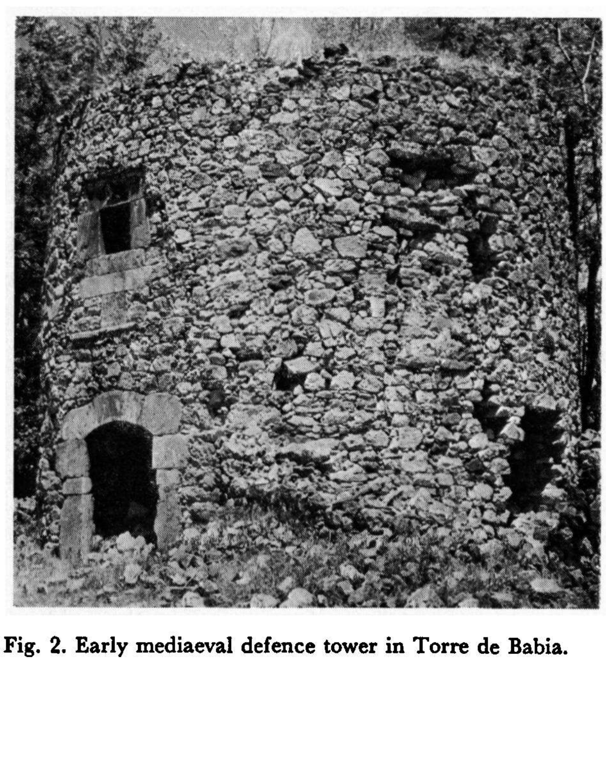

Fig. 2. Early mediaeval defence tower in Torre de Babia.

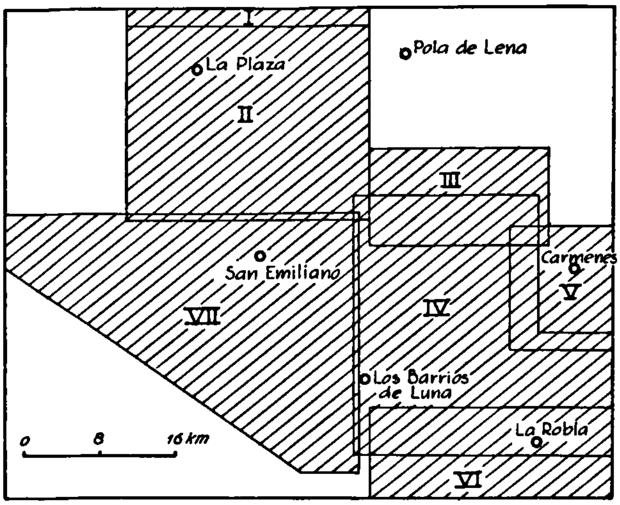

Fig. 3. Map showing existing geological maps in the SW part of the Cantabrian Mountains on a scale of 1:50,000 or less. I. Almela *et al.* (1956); II. García-Fuente (1959); III. Llopis Lladó (1955); IV. van Staalduinen (1969); V. Evers (1967); VI. Pastor Gómez (1963); VII. map presented with this thesis.

In the early fifties the Luna lake was created by the contruction of a 81.70 metres high gravity dam in the resistant Barrios quartzites near Los Barrios de Luna. The reservoir stores 308 million m^3 of water for the irrigation of 46,586 hectares in the Páramo region on the Meseta. Also an electric plant has been set up near Mora de Luna. The lake, covering 10.37 km^2, receives its water from an area of 501 km^2, with a total average annual precipitation of 470 million m^3 (Teijón, 1946).

Plans have been made for more artificial lakes in the upper course of the Río Sil; in 1967 a similar lake was completed just south of Villablino.

Climate

Due to the presence of the high watershed between the provinces of León and Oviedo the climate, being rather warm during the short summer but cold in winter, is scarcely influenced by the nearby Bay of Biscay. Extreme temperatures: −22°C in winter and +38°C in summer. In summer it is very dry, but during the remainder of the year there is much snow and rainfall, causing an annual precipitation of about 1250 mm (Teijón, 1946).

Fieldwork

The fieldwork was carried out during the summers of 1964 to 1968. All geological maps dealing with this region are on a scale of 1:400,000 or more, except a 1:100,000 scale sketch-map by Gómez de Llarena & Rodríguez Arango (1948). Existing geological maps of the neighbouring regions are indicated in fig. 3. De Sitter (1962b) and de Sitter & van den Bosch (1968) published two provisional maps on 1:100,000 scale of the southern Cantabrian Mountains. In the field use was made of 1:50,000 scale maps of the Instituto Geográfico y Catastral in Madrid, enlarged to a scale of 1:25,000 and in some cases even to 1:10,000.

The region mapped is covered by the sheets 76 (Pola de Somiedo), 77 (La Plaza), 101 (Palacios del Sil), 102 (Los Barrios de Luna) and 128 (Riello), see fig. 1. Aerial photographs were used mainly in regions where Devonian and Carboniferous crop out. In the regions with older strata they are of little advantage, due to the intense weathering and the complexity of the structures. Aerial photographs were also of great assistance in the field, especially in those parts where the topographic maps were rather bad, as in the area between Rodicol and Quintanilla. This portion of the map has been corrected (fig. 1).

Acknowledgements

The numerous inspiring discussions with members of the staff of the Geological and Mineralogical Institute, especially with Dr. D. Boschma, Dr. J. F. Savage and Mr. C. J. van Staalduinen, are greatfully acknowledged.

Skilful technical assistance received from Miss C. P. J. Roest, Mr. J. Bult and Mr. B. Henning (respectively drawing of the map, the text figures and the columnar sections) and Mr. W. C. Laurijssen (photographs) is greatly appreciated. I am also greatly indebted to Mr. A. V. Zimmermann who corrected the English text, and to Mr. and Mrs. B. van Hoorn for the Spanish translation of the summary. I wish to express my special gratitude to Mr. Luis Carlos García López,

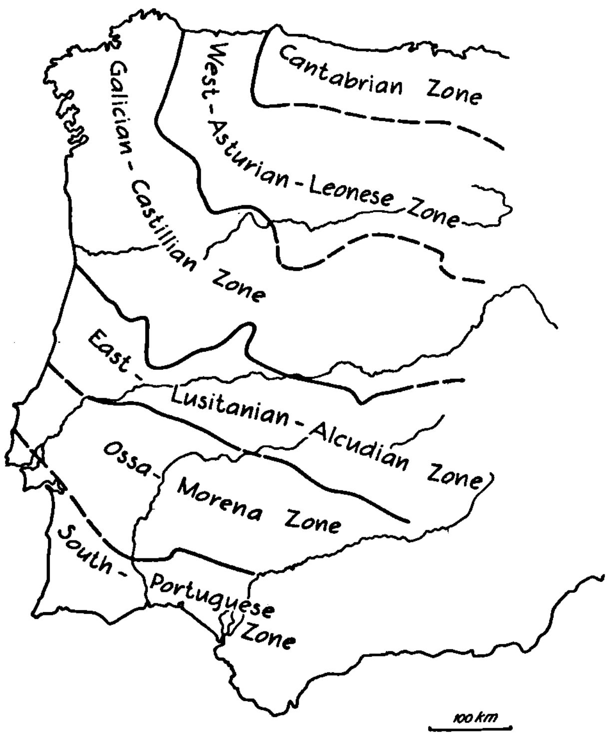

Fig. 4. Subdivision of the Iberian Hercynian Orogene, after Lotze & Sdzuy (1961).

Mr. Alejandro Melcón Arías, Mr. Leonel de Sousa and Mr. Francisco Cordero, and their families for their great hospitality and friendship during my stay in Villaseca, Huergas and Salce.

Geological setting

The Luna-Sil region is for the largest part situated in the SW corner of the Cantabrian Zone of Lotze's Iberian Hercynian orogene (Lotze, 1945, 1966, 1968; Lotze & Sdzuy, 1961); only the metamorphic Precambrian forms part of the West-Asturian-Leonese Zone (fig. 4). Lotze calls the Cantabrian Zone the external zone, the West-Asturian-Leonese Zone the internal zone and the Galician-Castilian Zone the axial zone of the orogene. Capdevila (1965, 1967) and Matte (1964, 1966, 1968b) demonstrated that respectively metamorphism and tectonic style in the West-Asturian-Leonese Zone are in accordance with its position in the internal zone. There is, however, no conclusive evidence of an eugeosynclinal development in the axial and internal zones of the orogene during most of the Devonian and Carboniferous, because strata of this age are almost absent in these regions (Drot & Matte, 1967). Moreover, the schistosity of the Precambrian is certainly not entirely of Hercynian age, because folds with axial-plane cleavage in the Precambrian are cut off by the Precambrian-Cambrian unconformity (de Sitter, 1968). De Sitter (1962b, 1964 p. 432, 1965), who subdivided the Cantabrian Zone into the Leonides and the Central Asturian Coal Basin, suggests that the Leonides possibly represent the external zone of the Hercynian orogene and that the Central Asturian Coal Basin represents the marginal basin of the Leonides. Only a very small portion of this basin is covered by the present map (fig. 65).

In the Leonides, of which the subject of the present study forms a part, a non-metamorphic sequence of Lower Cambrian up to Upper Carboniferous rocks more than 5000 metres in thickness has been folded during the Hercynian folding phases. The present configuration of the mountain chain is due to the late-Eocene and Miocene uplifts.

In general only studies of the last decades have been referred to. However, as early as 1781 A. de Prado Enriquez published the result of his geological investigations of the Cantabrian Mountains, followed by Casiano de Prado (1850), de Verneuil (1850), Barrois (1882), Delépine (1943) and many others. Many of their publications have been listed by Evers (1967), Rupke (1965) and Sjerp (1967).

CHAPTER I

STRATIGRAPHY

INTRODUCTION

In the Luna-Sil region, rocks of lowermost Cambrian to uppermost Carboniferous age, ranging in thickness up to more than 5000 metres, were deposited upon the folded Precambrian rocks. These sediments were deposited in a WNW to NNW trending part of the Cantabrian Zone which curves from WNW strikes in the SE part and NNW strikes in the SW part to ENE strikes in the northern part (fig. 23). The thickest succession was deposited in the central parts of the Luna and Babia Alta units, the thinnest in the Babia Baja unit (fig. 65).

The Cambrian to Devonian rocks can be subdivided into a chiefly clastic sequence of Lower Palaeozoic age (Luna Group), and a sequence of mainly carbonate rocks of Devonian age (Bernesga Group) (Brouwer, 1964b). The Carboniferous rocks have been subdivided into the Ruesga Group (deposited before the Sudetic folding phase), the Yuso Group (deposited between the Sudetic and Asturian folding phases) and the Cea Group (deposited after the Asturian folding phase) (Koopmans, 1962).

Mesozoic rocks are absent and only small outcrops of Tertiary deposits occur in the southern part of the region.

No important differences in lithology of the various rock units appear parallel to the axis of the basin: approximately the same rock units can be traced from the Esla region to the Asturian coast. Perpendicular to the basin axis, however, important differences in thickness and/or lithology occur.

The names of the various subareas to be used in the following description are stated in figure 65. Table *A* shows the stratigraphic rock succession and the ages and thicknesses of the different formations.

PRECAMBRIAN

Mora Formation

The Mora Formation crops out in a long arc, the Narcea anticlinorium, between the Asturian coast and La Magdalena (just east of the region mapped) where Tertiary deposits overlie it.

The name Mora Formation has been introduced by de Sitter (1962b); the type locality lies just east of this map, along the river Luna, south of the village of Mora de Luna.

A number of short articles on the Precambrian in the Narcea anticlinorium were published by Julivert & Martínez García (1967), Lotze (1960), Matte (1968a), Pastor Gómez (1962) and de Sitter (1962a).

The best-exposed sections occur north of Villabandín and in the N-S striking mountain chain west of the Río de San Miguel on the western limit of the present map.

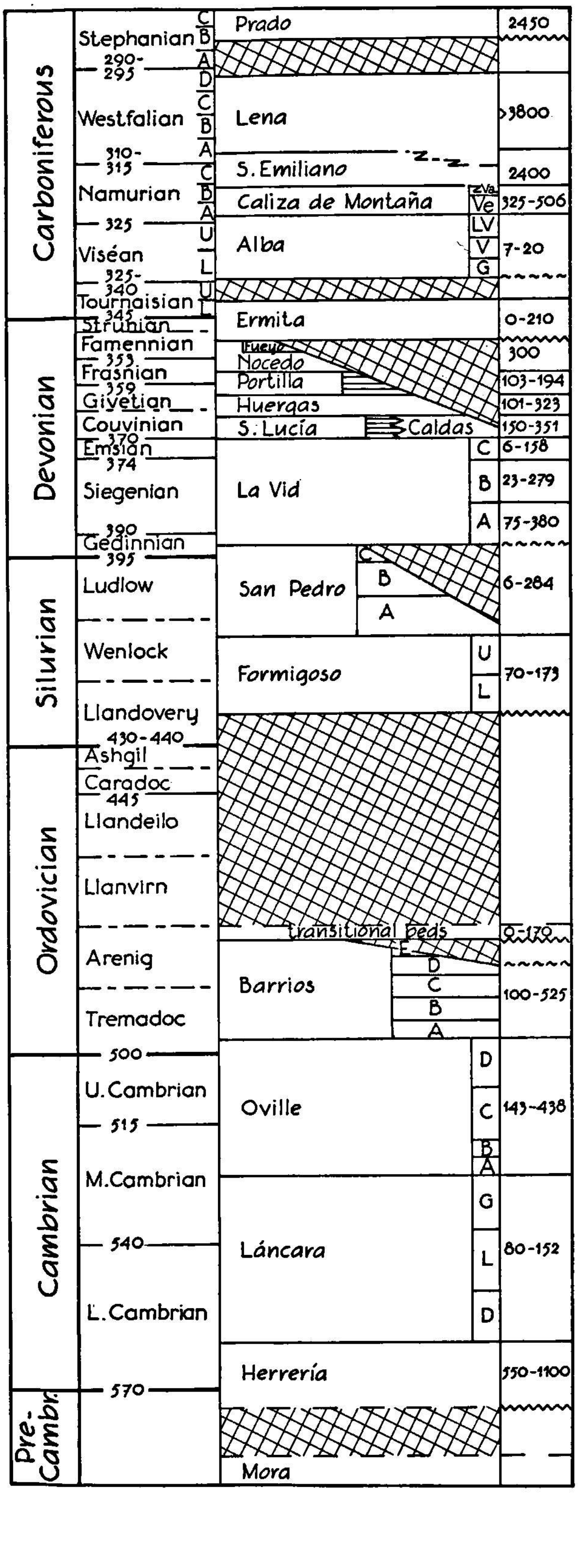

Table A. Lithostratigraphic units in the Luna-Sil region. Absolute ages after *Internationale Kommission für Geochronologie* (1968).

Age. – No fossils were encountered, but as the Herrería Formation of lowermost Cambrian, or even uppermost Precambrian age, overlies the Mora Formation unconformably, it must be of a Precambrian age. Matte (1968a), in a comparison with the Galician Precambrian, concluded that it must be the youngest Precambrian of NW Spain (fig. 7).

Lithology. – The Mora Formation consists mainly of very slightly metamorphic slates, greywackes and quartzites, all possessing a green colour. Several sets of cleavages and knick zones were found (fig. 107). The degree of metamorphism decreases towards the NE, and in the upper course of the Río de San Miguel, just NW of the area mapped, the rocks are non-metamorphic and of a dark grey colour. Notwithstanding the monotonous greenish colour, close observation reveals that sedimentary structures are rather abundant. Slumps (fig. 5 and 6), cross-bedding, graded bedding, flute-casts and load-casts were found. Graded sandstones with flute-casts at their base are exposed along the road south of Salce. Locally a conglomerate with pebbles approx. 1 cm in diameter was encountered. Locally rhyolite tuffs occur, especially in the western part of the region. Due to weathering before

Fig. 5. Slump level in the Mora Formation, west of Salce.

Fig. 6. Slump level in the Mora Formation (photograph by Dr. D. Boschma).

the deposition of the unconformable Herrería Formation, the upper 5 to 20 metres of the Mora below the unconformity have a characteristic red colour. This unconformity plane is very smooth (fig. 104).

Depositional environment. – The graded sandstones with flute-casts at their base may probably represent turbidites. The presence of these turbidites, greywackes and some conglomerates, the rapid lateral and vertical facies changes and the types of sedimentary structures suggest deposition during unstable conditions, with erosion of a landmass not too far away. This is in agreement with the conclusion of Matte (1968b), who stated that the Precambrian rocks in Galicia are older and were folded before the deposition of the Mora, the sediments of the Mora being composed of erosion products of the Galician Precambrian (fig. 7).

After folding of the Mora (p. 213) the region emerged and peneplanation took place before the deposition of the Herrería.

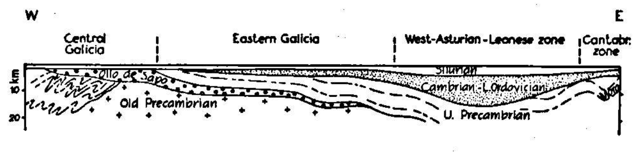

Fig. 7. Relation of the Precambrian and Lower Palaeozoic in the Luna-Sil region to that in the other regions in NW Spain, after Matte (1968b).

CAMBRO – ORDOVICIAN

Herrería Formation

The name 'Grès de Herrería' was introduced by Comte (1959), who chose the section near the hacienda of 'La Herrería' in the Porma valley (north of Cerrecedo) as the type section, although the lower part of the formation does not crop out there.

Previous work was done by Comte (1959), Lotze & Sdzuy (1961) and especially by Oele (1964).

The best exposures in this region occur north and south of Irede, south of Los Barrios de Luna, and along the Río de Sosas, north of Sosas de Laciana.

Age. – Lotze and Sdzuy found trilobites in the upper 55 metres of the formation which are characteristic of Lower Cambrian, hence the hundreds of metres of rocks below it may be lower Lower Cambrian, too, or partly belong to the uppermost Precambrian. Comte (1959) estimated the age to be the same, although he did not find any fossils.

Lithology. – The Herrería Formation overlies the Mora Formation unconformably. This unconformity is accentuated by a red weathering zone 10–25 metres thick on the Mora and by a conglomerate at the base of the Herrería, generally 0.5–3 metres thick. No real soil was formed, but the originally green rocks developed a vivid red colour while preserving their original properties. The conglomerate does not always rest directly upon the weathering zone but somewhat higher in places, though never more than 8 metres above the base. At places where the conglomerate is absent or does not lie directly on the Mora, the base of the Herrería is generally formed by a red silty clay, closely resembling the weathering zone of the Mora immediately below it, so that it is extremely difficult to draw the exact boundary. The conglomerate itself consists mainly of white rounded quartz pebbles 0.5–1.5 cm in diameter in a red silty matrix; in the western part of the region some rhyolite pebbles were found. The remaining part of the basal 10 metres consists of red and green silty shales and white and pink quartzites. At two places in the NNE trending Herrería west of Salce, slumps were found in the quartzites.

The overlying 30–70 metres are predominantly composed of violet, red and green silty shale beds (generally 1–2 metres, but locally even 5 metres thick) with a smaller amount of quartzites and sandstones. At places with a lack of coarse material, isolated sand ripples embedded in clay (*linsen*) were formed.

Yellow-weathering dolomite beds, alternating with shales and sandstones, come next. This dolomite horizon could be traced throughout the entire region.

The overlying succession, hundreds of metres in thickness, consists of medium to coarse-grained white and pink quartzites with a medium bed thickness of 60 cm, but also extremely thin and very thick beds (up to 1.5 metres) occur; drab-coloured sandstones and red and green silty shale layers are rather irregularly distributed over this succession with bed thicknesses

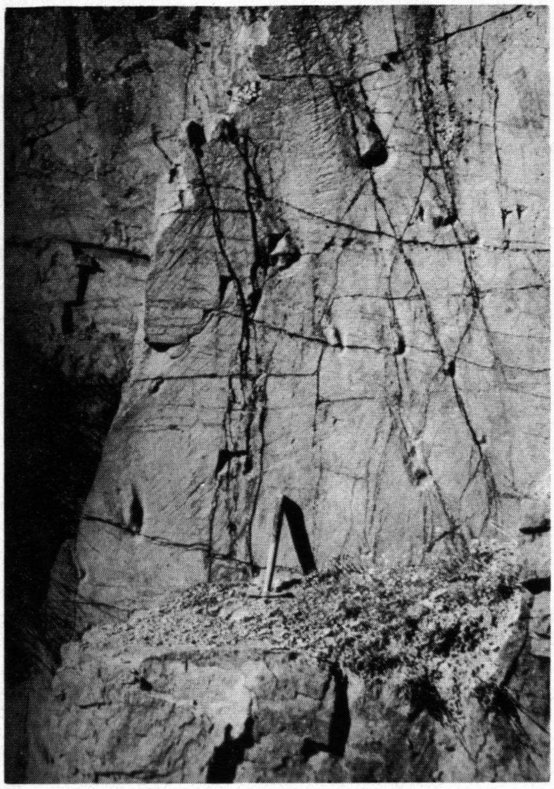

Fig. 8. Pseudo-crescent marks in the upper part of the Herrería Formation near the post office of Los Barrios de Luna (photograph by Dr. D. Boschma).

between 0.5 and 50 cm. In the western part of the region several dolomite beds are intercalated in this part of the formation. Bedding contacts between quartzites and silty shales are generally sharp, but undulating. Channels with locally thin conglomerates or clay pebbles at their base, grading, cross-bedding, ripples and load-casts were found in the quartzites. Small isolated quartz pebbles were also observed. Bed thicknesses in the quartzites laterally change rapidly. The particles are rounded to subrounded and, apart from the pebbles just mentioned, the sorting is good. Heavy minerals and feldspath grains may be mixed between the quartz grains.

In the upper 50 metres more silty shales are found than in the underlying part of the formation and most of the coarse beds consist of drab-coloured dolomitic sandstones; in the west dolomites are intercalated. Pseudo-crescent marks were found in this upper part of the formation near Los Barrios de Luna (fig. 8). The base of the Láncara was drawn at a level where more than 50% of dolomite occurs.

Thickness. – Approx. 850 metres south of Los Barrios de Luna; approx. 550 metres NW of Salce; approx. 1100 metres along the Río de Sosas.

Fossils. – Animal tracks 300 metres above the base, NW of Salce (fig. 9); *Astropolithon* (Seilacher, in Lotze & Sdzuy, 1961), trilobites and *Scyphomedusa* (van der Meer Mohr & Okulitch, 1967; fig. 10) in the upper 50 metres.

Depositional environment. – The sedimentary structures listed above, the often pink colour of the quartzites and the red colour of the silty shales often indicate fast flowing, shallow, oxidized water; the great thickness and the presence of load-casts indicate rapid sedimentation (Oele, 1964). The bimodal grain size distribution encountered locally is often an indication of fluvial sediments (Krumbein & Sloss, 1963, p. 162)

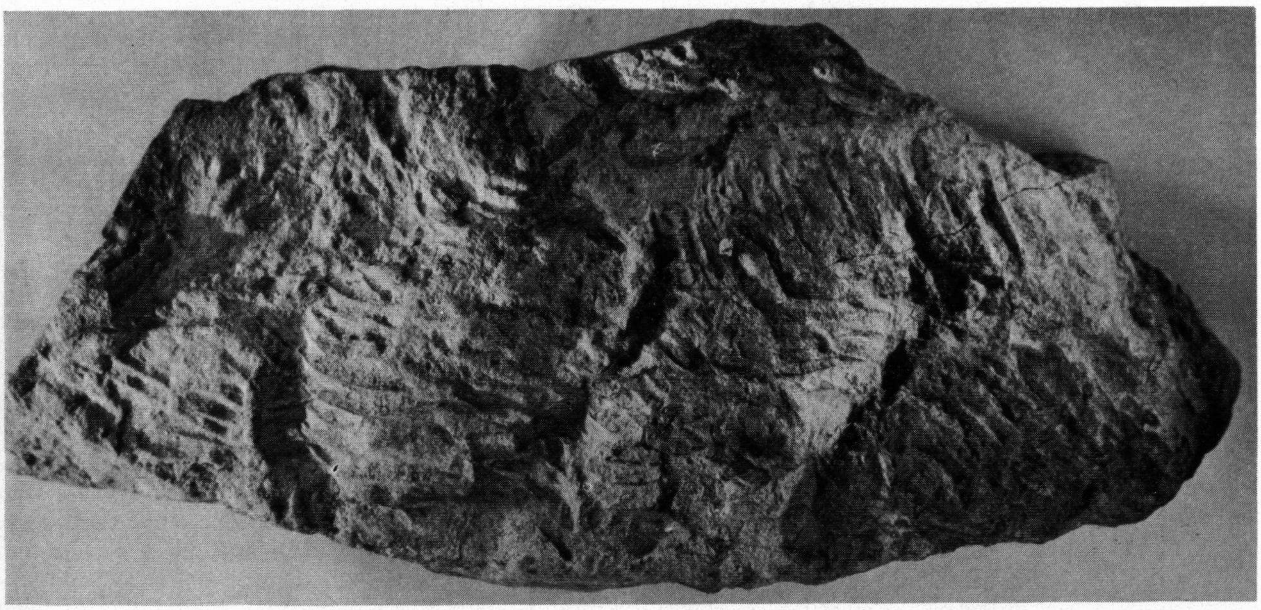

Fig. 9. Animal tracks from the middle part of the Herrería Formation, found in the valley of the Arroyo de Formigones.

Fig. 10. *Scyphomedusa* from one of the uppermost beds of the Herrería Formation, found along the path west of Los Barrios de Luna.

but the lateral extent of the formation seems too great for an exclusively fluvial origin of the sediments (Oele, 1964).

Van Houten (1948) concluded that the association of red and green mudstones with sandstone lenses, cross-bedded channel sands and conglomerates may be fluvial, and that these red-banded deposits may be derived from nearby elevated land. Due to the rapid decrease in the stream velocity the coarse debris were deposited in alluvial fans at the basin margin, and the sand and mud were carried into the basin to be deposited in stream channels and broad flood plains. Sedimentation took place during a warm and humid climate. Sediments of this type, deposited over a large region, are nowadays to be found in the extensive flat parts of Paraguay, Uruguay and northern Argentina (van Houten, 1948). Krumbein & Sloss (1963, p. 564) also state that these red beds are overwhelmingly more common in non-marine than in marine sedimentation.

The coarse cross-bedding, the frequent grading and the rapid lateral thickness changes might also be an indication of delta deposits; the silty shale intercalations would represent isolated quiet parts of the delta. The other sedimentary structures, too, may be encountered in a deltaic environment. Other indications of a deltaic environment (e.g. pro-delta clays) are, however, lacking, the presence of such an environment in the region mapped during the deposition of the Herrería thus remaining uncertain.

We may therefore conclude that the Herrería may have been deposited partly in a fluvial and, possibly, partly in a deltaic environment, with some marine intercalations (e.g. the dolomite beds). The upper 50 metres are at least partly marine, trilobites having been found there. Oele (1964) concluded the Herrería to be deposited in a very shallow-marine environment with fluvial influences.

The increase in the amount of conglomerates towards the NNE (Evers, 1967; Lotze & Sdzuy, 1961) and the increase in the amount of shales towards the SSW (Lotze & Sdzuy, 1961; Matte, 1968b), indicate that the source region was situated in the NNE.

Láncara Formation

Comte (1959) introduced the names 'Dolomies', 'Calcaires', and 'Griotte de Lancara', named after the village of Láncara de Luna in the present region, SE of Pobladura and now inundated except the church. The type section is inundated, too, and therefore a new type section was chosen along the new road. The basal part of the formation is not exposed there, however.

The formation is subdivided into the Dolomite Member, the Limestone Member and the Griotte Member. Previous work was done by Comte (1959), Lotze & Sdzuy (1961), Oele (1964), and especially by van der Meer Mohr & Schreuder (1967) and van der Meer Mohr (1969, in press).

From a structural point of view the Láncara Formation is of great importance, the most important thrust faults in the Leonides lying in the basal shales of the formation. The best-exposed complete sections lie east and west of Los Barrios de Luna, west of Salce and on the watershed between Asturias and León at the end of the valley of the Río de Sosas, just north of the region mapped. The Griotte Member is well exposed near the church of Láncara as is the whole formation minus the basal part of the Dolomite Member north of the Penouta (south of Riolago) and near La Majua.

Age. – Dolomite Member: middle Lower Cambrian; Limestone Member: upper Lower Cambrian and lower Middle Cambrian; Griotte Member: Middle Cambrian (Lotze & Sdzuy, 1961). Entire formation: lowermost Lower Cambrian to lowermost Middle Cambrian (Comte, 1959).

Lithology. – As the Láncara is of a rather uniform lithological composition in the region mapped, only the very well exposed section west of Los Barrios de Luna (section 1) will be described; only clearly different developments in other places will be dealt with.

The gradual transition from the Herrería to the Láncara is formed by dolomitic sands, arenaceous dolomites and dolomitic black shales. Next comes a dolomitic oolite, overlain by a thick-bedded dolomite containing some ooids and fossil fragments, and chert in the upper part. Authigenic quartz crystals are common in the oolites. After the overlying fine black shales comes a fine-grained dolomite of medium bed thickness, locally laminated and with small-scale cross-bedding; black shale beds about 10 cm thick are intercalated. Due to the different composition of the laminae, namely dolomite and limestone, the lamination is only visible on weathered surfaces. This alternation of dolomites and black shales is followed by a grey limestone 3.60 metres thick, completely composed of well-developed stromatolites, overlain by a laminated limestone 80 cm thick, with cross-bedding. On the following cross-bedded dolomite of medium thickness, lies a laminated argillaceous dolomite, overlain by dolomite beds with ripples, cross-bedding and intraformational breccias. Next comes an alternation of laminated black shales and dolomites with ripples and cross-bedding. These beds are overlain by 20 metres of thick-bedded dolomites, weathering to a typical orange-brown. East of Irede filled up cavities were found in the Dolomite Member.

The Limestone Member consists of thick-bedded grey limestones, weathering light grey, and internally often finely laminated. Bird's-eye structures and cavities, later filled, are frequent. One cavity, found in the Campo del Oro structures (fig. 65), was filled with red debris, mainly trilobite remains. The upper 4 metres are of a pink colour, and are rich in fossil debris. In the Campo del Oro structures, a dark red sandstone lens (subrounded quartz grains in a hematite cement), 0.5 metres thick and 30 metres long, was also found (fig. 11). North of Cospedal, the Limestone Member grows considerably thinner: near Genestosa it measures only 10 metres and still further to the north, no limestones were found at all.

The Griotte Member is composed of reddish nodular limestones and limestone nodules, with thin parallel bands of angular fossil fragments as well as well-preserved fossils, separated by red shales and hematitic material (fig. 12). At the contact with the Limestone Member many filled up cavities and irregular contacts were found locally. In the northern part of the region the griotte in places loses its red colour. There we locally find grey limestone beds separated by extremely thin curved seams of red shaly or hematitic material. Glauconite is locally abundant. The transition to the Oville shales is well exposed near the church of Láncara: between the last thick griotte bed and the olive green Oville siltstone lie 2 metres of greenish, red and black shales with green limestone concretions, weathering to an orange colour, and a few griotte beds were also found (section 5). Mineralizations near the thrust faults in the lower part of the Láncara were mainly found along the faults themselves or in the limestone-griotte contact.

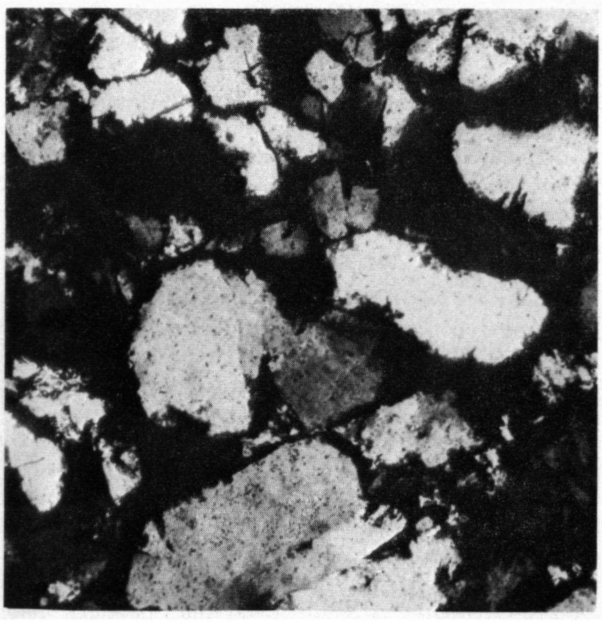

Fig. 11. Quartz sandstone with hematite cement, found in the Láncara Formation in the Campo del Oro structures (100x).

Fig. 12. Nodular limestones and shales of the Griotte Member (Láncara de Luna).

Thickness. – Thicknesses are given in the sections. East of Torre the Dolomite Member is more than 65 metres, the Limestone Member 50 metres, and the Griotte Member 15 metres thick. At the watershed north of the Río de Sosas the Limestone and Dolomite Members are together 110 metres and the Griotte Member 10 metres thick. The Limestone Member and the Griotte Member both grow thinner towards the north.

Fossils. – Stromatolites, trilobites, brachiopods, carpoids.

Depositional environment. – The oolites of the Dolomite Member were formed on wave-exposed shorelines, characterized by low rates of terrigenous sedimentation, by insolation, evaporation and agitation of carbonate-satured waters (Rusnak, 1960; van der Meer Mohr & Schreuder, 1967). Stromatolites grow in an intertidal environment. Filled up cavities, found locally, indicate subaerial solution. It may therefore be concluded that the sediments were deposited in very shallow water near the shore and were occasionally brought above sea level. Oder & Bumgarner (1961) concluded the same for comparable sediments. Because alternating limestone and dolomite laminae were found, it seems probable that a primary precipitation of dolomite has taken place which indicates a very high salinity of the water.

The internally laminated, detrital limestones of the Limestone Member, of medium bed thickness, with locally large numbers of fragments of marine fossils and stromatolites, indicate a shallow, quiet, marine environment. Bird's-eye structures and filled up cavities, especially at the top of this member, indicate subaerial solution. The sandstone with hematite cement, which is possibly a *sebkha* deposit (Mr. C. G. van der Meer Mohr, pers. comm.), is also in agreement with this conclusion.

Glauconite and large amounts of marine fossils in the Griotte Member indicate a marine environment. The deposition of clay and angular fossil debris in thin parallel layers indicate non-agitated water below wave level; the red colour of the clay shows that the water was well aerated. The nodular limestones and limestone nodules are most probably remains of parallel-bedded limestone beds which were attacked by submarine solution. The red shales are primary, but the hematitic components may be solution remains (Hollmann, 1962). An alternative possibility is that the primary sediment was deposited in an oxidizing environment and after burial, came to lie in a zone with a lower pH, so that the $CaCO_3$ could partly dissolve in the circulating water.

Oville Formation

Comte introduced the name 'Grès et Schistes d'Oville' and selected a type locality just south of the village of Oville, west of Valdecastillo in the Porma valley.

Previous work was done by Comte (1959), Lotze & Sdzuy (1961) and Oele (1964).

The Oville Formation is best exposed near Los Barrios de Luna, on the mountain north of the now inundated village of Campo de Luna (east of Láncara), west of Genestosa and north of the Penouta.

Age. – Upper Middle Cambrian B to uppermost Upper Cambrian (Lotze & Sdzuy, 1961); upper Middle Cambrian to Tremadoc (Comte, 1959).

Lithology. – The section near Los Barrios is very well exposed and will be used as the reference section (section 6). The subdivision of the formation into five members was proposed by Mr. G. Gietelink (written comm.).

There is a gradual transition between the Griotte Member of the Láncara Formation and the argillaceous siltstone of member A of the Oville Formation (p. 146; section 5). The 25 to 50 metres thick member consists of finely laminated, green or bluish argillaceous siltsones and silty shales with some thin, finely laminated, micaceous sandstone *linsen.* A grey sandstone bed, weathering yellow and several metres thick, with planar cross-bedding and often a burrowed top, was found a few metres above the base of the siltstones and can be traced throughout the entire region mapped. Thinner sandstone beds (10–20 cm thick) of this type were also locally encountered in the siltstones. Both the sandstones and the siltstones contain many trilobite fragments, weathering orange-brown, which are concentrated in certain layers, and glauconite grains of the same colour. These glauconite grains are typical of the entire formation.

After the basal 25 or 50 metres, gradually more sandstone beds are intercalated (member B). *Flaser* and *linsen* structures become more common. The sandstones are generally rather pure, with bed thicknesses of between 0.1 and 1.5 metres; the thickness of the silty shales varies between 0.2 and 30 cm. Shallow channels, some ripples and cross-bedding occur in the sandstones; some parallel bedding was, however, also found. Burrowing is generally restricted to the silty shale layers. This member is 25–40 metres thick.

Next comes an alternation of sandy siltstones, silty shales and drab-coloured impure sandstones (member C). The sandy siltstones contain many, relatively coarse, quartz grains and the sandstones have much clay matrix, generally about 15–20% but sometimes even 50% (Oele, 1964). The cement often has a high carbonate content. The quartz grains are generally subrounded. These sandstones are subgreywackes or quartz wackes (Krumbein & Sloss, p. 172). Micas are concentrated on the bedding-planes and often cause sharp contacts between the generally parallel-bedded sandstones and siltstones. This parallel bedding is typical of member C; both gradual and sharp contacts exist between the fine and the coarse beds; an internal lamination of the beds very often occurs. The bed thickness and the percentage of sandstones increase towards the top, and the sandstones become more pure. The bed thickness increases from 1–15 cm in the lower part to 5–50 cm in the upper part. Burrows are very frequent in the impure sandstone and in the siltstones, and also become less abundant towards the top; worm tracks are often concentrated at the sole of the beds. Ripples become more frequent in the upper part.

The transition to member D is gradual. This member is mainly composed of pure sandstones (often quartzitic), with furthermore important amounts of impure sandstones, sandy siltstones and argillaceous siltstones, the latter three all of the type described above. The quartzitic sandstones become even more abundant towards the top. Most of the sandstones show channelling, but striking parallel lamination was also found. Especially in the upper part, the channels have a clearly erosional base. *Flaser* and *linsen* structures, ripples, clay pebbles and many kinds of cross-bedding are frequent. Loading was found in the upper part. The top of the Oville was drawn below the first thick sequence of thick-bedded, cross-bedded quartzites (sections 6 and 7). From there on, glauconite grains are no longer found.

Tuffite beds and dolerite sills, parallel to the bedding and without contact-metamorphism, were found locally in the various parts of the formation. They are difficult to map because their weathering colour is the same as that of the surrounding rock and there may therefore be more of these igneous rocks than represented on the map. Mr. G. P. J. Blom (internal report) reported that these rocks are predominant in the Valgrande area, just NE of the Ubiña zone.

Thickness. – The thickness is very constant in the Abelgas syncline: 430 metres (sections 6 and 7). In the eastern part of the Aralla zone (on the mountain north of the inundated village de Campo) 143.40 metres; north of Puerto de la Cubilla 300 metres (Mr. G. P. J. Blom, internal report); west of La Majua 192 metres.

Fossils. – In the basal 50 metres trilobites (up to 10 cm long) and carpoids were found. Burrows and tracks were found throughout the entire formation.

Depositional environment. – The presence of glauconite throughout the entire formation indicates that the Oville Formation was deposited in a marine environment. The parallel lamination, the planar cross-bedding in the sandstone, the complete trilobite carapaces often found indicate that the fine-grained member A was deposited in a quiet environment with occasional currents with low velocities. This environment was not very suitable for brachiopods, which were so abundant in the Láncara.

The upwards increase of the sand percentage in member B indicates a regression. The sedimentary structures in this member indicate rapidly changing current velocities in shallow water, but seawards from the zone of breaking waves. The *flaser* and *linsen* structures may indicate that these sediments were deposited in an area influenced by tidal water movements.

The alternation of parallel-bedded, badly sorted argillaceous sandstones and sandy siltstones in member C indicates currents, loaded with sediments, in which the current velocity decreased suddenly, causing very badly sorted sediments to accumulate. The frequently observed grading from sandstone to siltstone is in agreement with this conclusion. The parallel bedding indicates that these currents had no erosional force. Burrows are very abundant in these sediments. All these facts indicate rapid sedimentation, so that no winnowing could take place, which may indicate that deposition took place in a rapidly subsiding basin; better winnowing occurred during the deposition of the coarser sediments of the upper part of member C, which may be an indication of a less rapidly subsiding basin and of a regression (Krumbein & Sloss, 1963, p. 173).

The increase of clean sandstones, accompanied by an increase in the number of ripples and channels, indicates that winnowing could occur more frequently at the transition to member D; the rate of burial of the sediments thus decreased further and hence probably also the rate of subsidence of the rather unstable basin. The thick parallel-bedded sandstone beds of member D of the Oville may represent beach deposits (Mr. G. Gietelink, written comm.). The clay pebbles encountered in the, often deep, channels with an erosional base show that erosion (sub-aerial or submarine) took place and *flaser* and *linsen* structures, often burrowed, indicate an influence of the tidal movements of the water. We may therefore conclude that member D has been deposited under stable conditions in a subtidal or intertidal environment, possibly a delta (Oele, 1964).

Thickness distribution (fig. 17) and the kind of environment (Lotze & Sdzuy, 1961), grain orientation (Oele, 1964) and the direction of cross-bedding (Mr. G. Gietelink, written comm.) indicate a source area in the NNE.

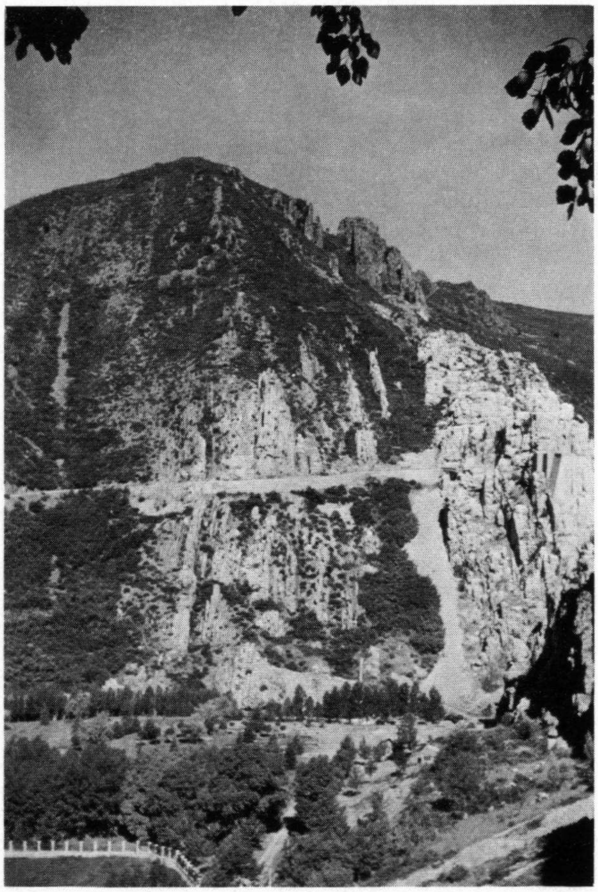

Fig. 13. Barrios Formation in the type locality near Los Barrios de Luna (looking westwards).

Fig. 14. Upper part of the Barrios Formation NNE of Torre de Babia (looking eastwards).

It may be concluded that the sediments of the Oville Formation were deposited on a shallow flat shelf (fig. 17) which was rather stable during the deposition of the members A, B and D, but became rather unstable during the deposition of member C, which occurred during the rapid subsidence in the West-Asturian-Leonese Zone. A regression took place from member A to member B and, after a slight transgression, another regressive sequence was deposited in members C and D.

Lotze & Sdzuy (1961) and Matte (1968b) found that sediments of the type encountered in the middle and upper part of the Oville were deposited in the whole of NW Spain but in very varying thicknesses (fig. 17 and 22).

Barrios Formation

Comte (1959), who introduced the name 'Quartzites de Barrios', selected the type locality near Los Barrios de Luna, in the present region mapped. Previous work was done by Comte (1959) and Oele (1964).

The formation is best exposed in the type section (section 8; fig. 13), north of the Penouta (section 9), NW of La Majua and just south of the Muxivén.

In the Abelgas syncline and the Babia Alta unit the formation could be subdivided into five members: three quartzite members separated by two silty shale members.

Age. – Due to the lack of fossils, the age of this formation is difficult to establish. Comte (1959) concluded that it is for the largest part of an Arenig age and that the age of the top lies between Upper Arenig and Lower or Middle Llandeilo. Nollau (1966) found the top of the *Cuarzitas Armoricanas* (of the same age and lithology as the Barrios) to be of an Upper Arenig age.

Lithology. – The lower boundary of the formation was drawn below the first thick succession of thick, cross-bedded quartzite beds and above the last major occurrence of glauconite. In the type section and other sections, in the SSW part of the Leonides, more fine-grained sediments were found between the quartzite banks than in other places.

Member A of the formation mainly consists of quartzite beds with parallel lamination or with very flat cross-bedding, separated by thin micaceous silty shale partings. Dark streaks of heavy minerals are often of great assistance in finding the bedding-plane where the rock has been intensively jointed. Channels and some steep cross-bedding are also present. Where the partings are thicker (in the southern sections) *flaser* and ripples were found. Bed thicknesses lie between 0.1 cm for the silty layers and some of the laminated quartzites and 5 metres for the thickest channelling quartzites. The colour of the quartzites is white or more or less pink (due to some hematite in the cement). The silty shales are generally green. Locally pyroclastic rocks or dolerites were found in this member.

The following 21 metres are not exposed in the type section, but in other places in the Abelgas syncline (section 9) 8.5 metres of black, laminated, silty shales and sandy silts, weathering brown or grey, are exposed at this level (member B). Poll (1963) found this member, and also the thick upper silty shale member, in the Belmonte region of Asturias, 50 kilometres to the NNW; these levels can therefore be traced over a great distance, parallel to the strike of the basin.

Next comes member C; it begins with an alternation of rock types of the type encountered in member A and pure quartzite beds with channels and steep cross-bedding (section 8). The upper part of this member is composed, in the type section, of an alternation of clean quartzites and green silty shales (about 15% of the rock). In this succession many burrows and tracks were found; one level with numerous vertical burrows is very characteristic and occurs in many other sections, too. The quartzites partly have a lamination or planar cross-bedding and are slightly channelling, but beds with steep cross-bedding in one direction and erosional channels with clay pebbles have also been observed (compare fig. 14). Load-casts, sand-balls, ripples and *flaser* and *linsen* structures are present in the silty beds in the type section. The maximum bed thickness is about 1.5 metres.

A weathering horizon with iron concentrations was formed in the type section on top of member C. On this surface lie, with a sharp contact, 12.5 metres of burrowed silty shales (member D).

These silty shales are overlain, with an erosional contact, by member E, completely composed of thick-bedded pure white and pink quartzites (fig. 13 and 14). At the base intensive loading took place. Large, steep cross-bedding and deep erosive channels occur frequently in these beds (fig. 14), but beds with parallel lamination or planar cross-bedding also occur. Bed thicknesses of more than 10 metres are not uncommon (fig. 14). The upper metres of member E near Torre are thin bedded and very micaceous (section 15). This member was not found in the eastern part of the Aralla zone.

A locally different development of the Barrios was found west of the Arroyo de Formigones (section 10) where an alternation of thin-bedded white quartzites, weathering brownish, quartzitic sandstones, green silty shales and sandy silts crop out. The quartzites and quartzitic sandstones have dark heavy mineral streaks, and no glauconite, and may therefore be included in the Barrios, but only the upper 9.5 metres, and 4.5 metres of quartzites 33 metres below the top, are thick bedded. The lower part is cut off by a fault. A comparable development occurs SE of Los Barrios de Luna.

The top of member E has undergone erosion in many places, and often iron concentrations or a thin layer of weathering clay were formed there. Burrows are also very frequent on top of the Barrios (fig. 15).

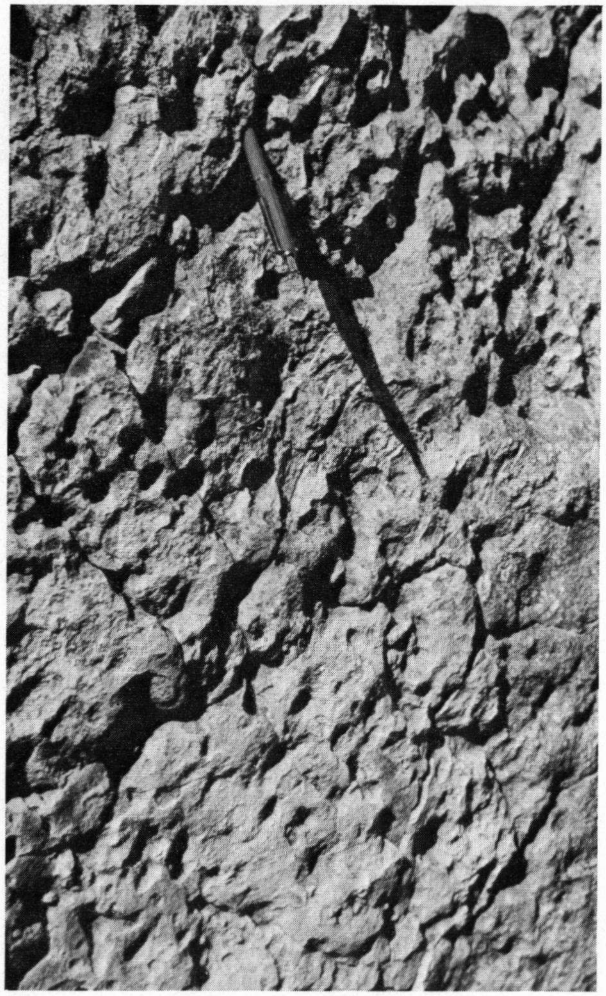

Fig. 15. Vertical burrows on top of the Barrios Formation (Los Barrios de Luna).

Thickness. – 301 metres in the type section; 259 metres in section 9; 243 metres south of Truébano; 525 metres in the Valgrande area, just NE of the region mapped (Mr. G. P. J. Blom, internal report); 143 metres north of the inundated village of Campo; 100 metres at 1.5 km further to the east (Mr. G. Gieteling, written comm.). Small thicknesses have also been encountered in the SW part of the Aralla zone.

Fossils. – Only burrows and tracks were found: *Cruziana* (fig. 16), *Scolites* (vertical), and *Glossifungites* (U-shaped, vertical) (Seilacher, 1967).

Depositional environment. – The pure quartzites of which the Barrios Formation has been mainly built up, were probably formed by an intensive winnowing of the sediments. The concentration of heavy minerals in thin streaks and also the presence of micaceous silt stone partings, especially in the parallel-bedded beds, are an indication of selection by waves.

The parallel, or nearly parallel-bedded quartzites with heavy mineral streaks of member A may be beach deposits. The intercalation of some beds with deep

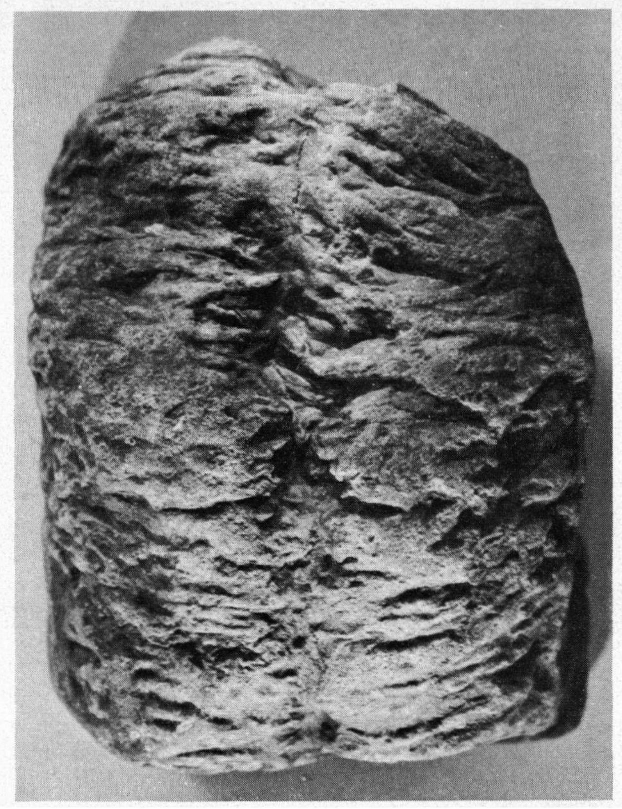

Fig. 16. *Cruziana* from the Barrios Formation NE of Torre de Babia.

channels and steep cross-bedding may represent tidal channels. This is confirmed by the presence of alternating cross-bedding with contrary dips.

The silty shales and sandy siltstones of member B may have been deposited in a badly aerated lagoon behind a barrier beach or bank (Krumbein & Sloss, 1963, p. 566).

The lower part of member C consists of an alternation of beach deposits and channels, with a decrease in the number of beds with beach deposits in an upwards direction. Some of the channels were formed in a tidal environment. The upper part of member C in the type section is of the same type as the upper part of the Oville. In other sections silty shales are less frequent. For the type section we may therefore conclude that these rocks were deposited in an area within the influence of tidal currents. The thick beds with steep thick cross-bedding over a long distance in one direction (fig. 14) may represent shifting banks.

After the deposition of these rocks large sections of the region emerged and erosion took place. The absence of the above-mentioned rocks in the section 1.5 km east of the inundated village of Campo, may have been caused by this erosion. The burrowed silty shales of member D were probably deposited under the same circumstances as member B. Member D was not found in the eastern part of the Aralla zone (Mr. G. Gietelink, written comm.).

The pure quartzites with an erosional base of member E were deposited upon these silty shales. In analogy with structures in the other parts of the Barrios already dealt with, these deposits have probably been deposited in an environment with beaches, shifting banks and tidal channels. Emersion and erosion occurred after their deposition. Member E is also absent in the eastern part of the Aralla zone. An erosion surface suggests that these deposits may have been eroded.

The absence of member D and E, and locally also of the upper part of member C in the eastern part of the Aralla zone, the increasing amount of fine material in the southern sections (sections 8 and 10) and transport directions (Mr. G. Gietelink, written comm.) indicate that a positive area existed in the eastern part of the Aralla zone (fig. 24). The rapid thickening of the rock type encountered in the Oville and the thinning of the winnowed sediments of the Barrios towards the SSW indicate that the main source area was situated in the NNE (fig. 17). It is quite possible that during the deposition of the Oville in the SSW, rocks of the type of the Barrios were already being deposited in the NNE. Oele (1964) concluded from grain orientations that in the Esla region, too, the source area should be situated in the north.

After the deposition of the upper part of the Oville a further regression took place, ending with the deposition of member B. A transgression again brought coarse material which was initially deposited in a basin of slightly increasing depth. Then the region emerged and erosion took place. After the deposition of member D, possibly in a lagoon, a sudden influx of coarse material occurred again and 30–65 metres of beach and tidal channel deposits were formed. Then the region emerged again and erosion could take place, especially in the Aralla zone where a portion of member C and members D and E were completely eroded.

It may be concluded that after the deposition of the middle part of the Oville, the subsidence of the basin decreased, which gave rise to a gradually intensifying winnowing of the sediments. The intensive winnowing of the sediments of the Barrios indicates that only a very slow rate of subsidence took place during their deposition.

Transitional beds

No fossils were found in the transitional beds between the white Barrios quartzites and the pure black shales of Formigoso. Their stratigraphic position suggests that they belong to the Luarca Formation (Poll, 1963;

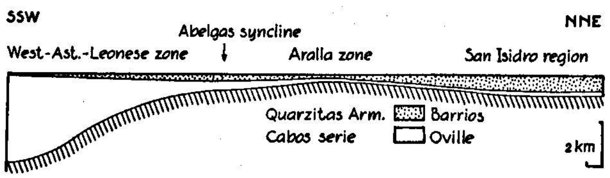

Fig. 17. Thickness distribution of the Oville and Barrios Formations in the Luna-Sil region and adjacent regions.

Radig, 1962; Riemer, 1963) but, as pure bluish dark shales, which constitute more than 90% of the Luarca Formation, are not very common in the transitional beds, these beds were not included in the Luarca Formation. The thickest transitional beds occur in the western part of the region (section 11) and near the electricity plant of Abelgas (section 12). The rocks in these sections are very similar to the rocks at the same stratigraphic level, described by Nollau (1966), in the region near Ponferrada (black silty shales, quartzites, iron concentrations). He also used the term 'transitional beds' (in Spanish). They are there overlain by the Luarca shales. It is, however, quite possible that the transitional beds laterally intertongue with the basal beds of the Luarca.

Previous work in the region NE of the Narcea anticlinorium was done by Bäcker (1959).

The best exposures were found north of the Muxivén, just east of the electricity plant of Abelgas, NNE of Torre, along the road east of Láncara de Luna and north of Caldas (just outside the region mapped). The transitional beds are generally less than 10 metres thick, absent, or not exposed. Therefore where they crop out, they are represented on the map by a dotted line; only in the west, where they are 170 metres thick, they have been given a colour of their own on the map.

Age. – Graptolites, found in the transitional beds near Ponferrada, indicate an Upper Arenig age (Oelmüller, in Nollau, 1966).

Lithology. – The thickest deposits were found in the western part of the region (section 11). On top of the Barrios lie 3 sequences consisting of black (often silty) shales which grade into dark grey siltstones, then into very thin bedded, platy, very fine grained sandstones with parallel lamination or very flat cross-bedding, these sandstones grading into pure white quartzites with channels or steep cross-bedding; the bed thickness, too, increases upwards in these sequences. The two thinner quartzites wedge out laterally within some hundreds of metres. The last quartzite is overlain by dark grey, burrowed siltstones containing 5 ferruginous layers in the lower part and overlain by a bed of thick, white, pure quartzite.

The upper metres are composed of burrowed ferruginous sands, siltstones and shales and a conglomerate bed. The individual beds often grow finer grained in an upwards direction.

In section 12, too, a great thickness was measured. More than 30 metres of black silty shales of the type encountered in section 11 and in members B and D of the Barrios, with three intercalated beds of burrowed sandstone, are covered by 10 metres of pure white quartzites with ferruginous weathered layers in their uppermost part. Upon the upper iron layer lies a sand-

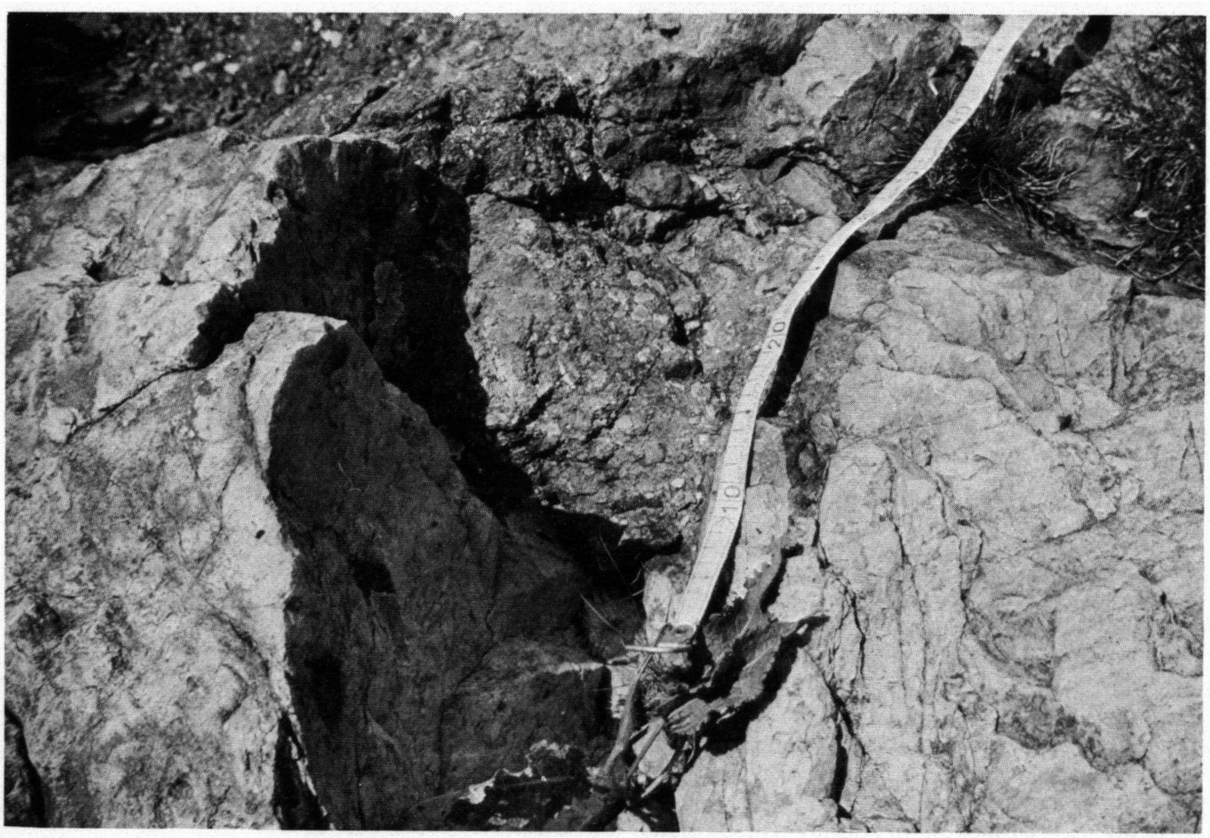

Fig. 18. Erosion channel, filled with a conglomerate with a very ferruginous matrix, in the transitional beds north of Caldas de Luna.

stone with a second iron layer upon it, overlain by a black shale bed, also with an iron layer upon it. A final sandstone bed with an iron layer upon it is followed by the black graptolite shales of the Formigoso.

Comte (1959) found a comparable situation in the section now concealed near Los Barrios. In the SW part of the nose of the Abelgas syncline, too, 10 metres of black shales and 15 metres of quartzites were found between the Barrios and the Formigoso.

In section 13, 11.25 metres of alternating light grey quartzites and light grey siltstones weathering dark grey, were deposited upon the white Barrios quartzites. Then follows a ferruginous shale, weathering brown, and overlain by burrowed quartzites with a thin intercalated layer of burrowed shale and thin ferruginous soft clay on top.

In section 14 small quartzite pebbles occur in the uppermost quartzites of the Barrios. A 40 cm thick lens of dark grey siltstones is followed by an alternation of ferruginous quartzite and siltstone lenses containing erosive channels filled with conglomerates of quartzite and igneous rock pebbles and a matrix very rich in iron (fig. 18). In sections north of Cospedal, too, up to 8 metres of quartzite pebbles, quartz and glauconite grains in a matrix very rich in iron crop out between the Barrios and the Formigoso (fig. 19).

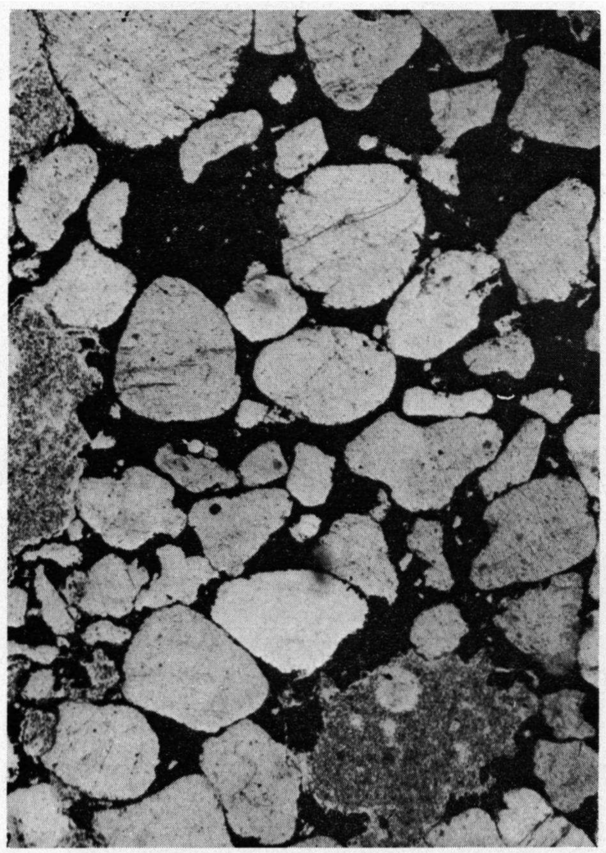

Fig. 19. Well-rounded quartz grains and partly replaced glauconite grains in a very ferruginous matrix; from the transitional beds north of Cospedal (25x).

Fig. 20. Glauconitic sandstone from the transitional beds NNE of Torre de Babia (40x); many of the grains partly replaced by hematite.

The Barrios in section 10 is overlain by a ferruginous weathering zone 0.5 metres thick.

On top of the Barrios in the Aralla zone, south of Riolago, lies 2 cm of markasite.

In many sections NNW of the Grajos fault and east of the Palomas syncline (sections 15, 16, 17, and 18), rocks of the same type crop out; especially impure, often burrowed, ferruginous sandstones and siltstones containing a large amount of glauconite are common (fig. 20). On top of the individual beds the glauconite is weathered, causing the rock to become particularly ferruginous there. Many of the more rounded quartz grains (fig. 19) of the sandstones are quartzite fragments; muscovite flakes, dolomite and tuff particles are also present. Bioturbate quartz wackes crop out in sections 16 and 18. The upper glauconitic sandstone bed of section 15 contains many dolomite fragments with a ferruginous crust. In section 16 a phosphate lens was found (fig. 21).

Thickness. – Generally 0–15 metres; 43 metres near the electricity plant of Abelgas; 25 metres in the southern part of the nose of the Abelgas syncline; 170 metres in the western part of the region mapped.

Fossils. – Many burrows and tracks only.

Depositional environment. – The alternation of black silty shales and white, often channelling, quartzite lenses or beds of the same type as the rocks in members D and E of the Barrios, respectively indicates that the rocks in the thickest sections (11, 12 and the section in the nose of the Abelgas syncline) were deposited partly in a lagoon and partly in stream channels. Below these rocks an emersion horizon was often

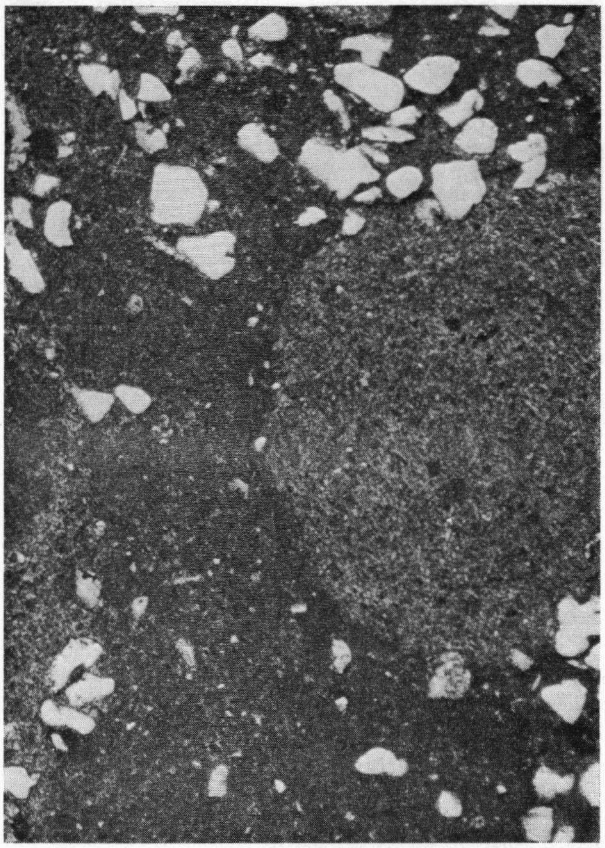

Fig. 21. Phosphate from the transitional beds NW of La Majua (25x).

found on top of the Barrios, frequently with a soft clay or iron concentrations. The predominance of silty shales indicates that the source area no longer delivered such large quantities of coarse material as during the deposition of the Barrios. Emersion and erosion occurred frequently. The silty shales may represent pro-delta clays, the quartzites barfinger sands. The other phenomena mentioned above are also indications of a deltaic environment. The channelling conglomerates with an iron cement of section 14 and the sections north of Cospedal, and the ferruginous shale of section 13 may have been formed in an oxidizing continental environment.

The sandstones containing very much glauconite and some phosphate nodules have been formed in a marine, reducing environment with slow sedimentation (Krumbein & Sloss, 1963, p. 185 and 186). The excellent rounding of all grains of this rock type indicates transport of very long duration. Many of the individual beds have been weathered in a very acid environment, the glauconite having been replaced by hematite, chert or authigenic clay and some of the quartz grains having been partly dissolved (Fairbridge, 1967). Bioturbate sediments are very common, too. Glauconite and phosphate nodules often occur at unconformities (Krumbein & Sloss, 1963, p. 185). It may be concluded that the glauconite and phosphate were formed in a quiet, reducing, marine environment with a low rate

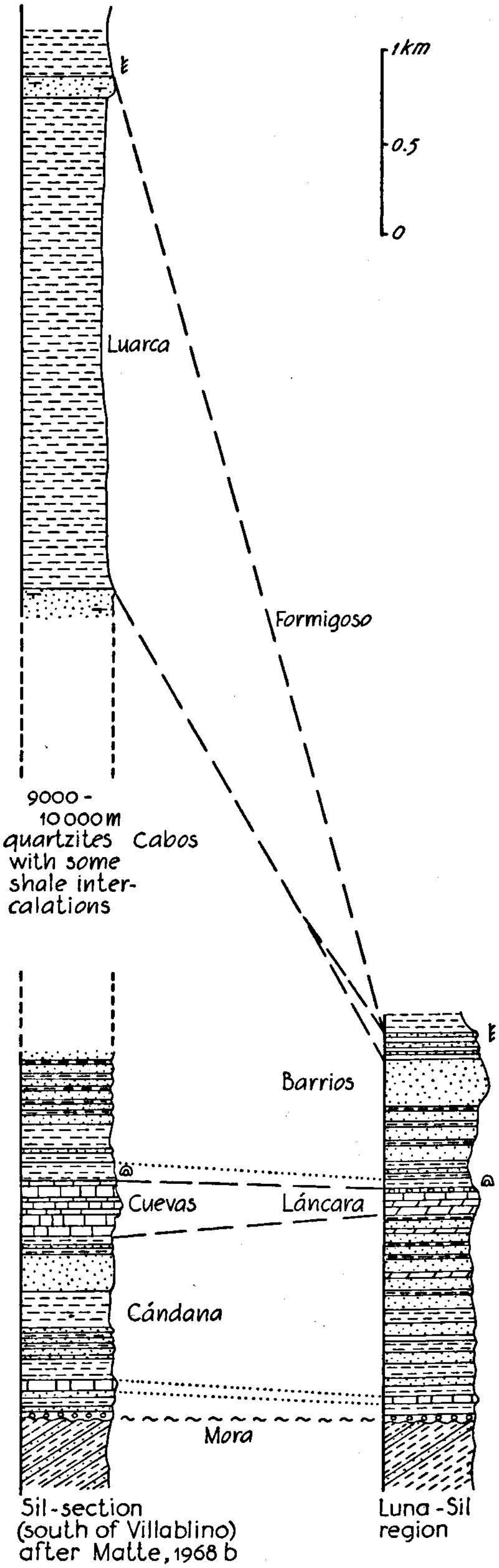

Fig. 22. Comparison of the Cambro-Ordovician rocks on either side of the Narcea anticlinorium.

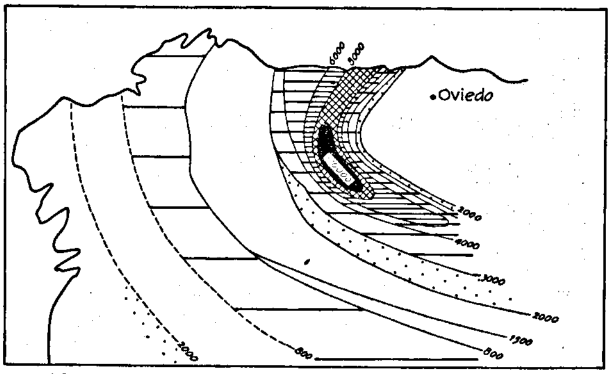

Fig. 23. Isopach lines of Lower Cambrian to Arenig rocks in NW Spain, after Matte (1968b).

of sedimentation (thus possibly in the more SW part of the basin, where black shales were deposited contemporaneously) and that these grains have been transported towards the NE.

The above-mentioned properties of the rocks in which these grains are embedded suggest deposition in a littoral environment which emerged frequently and for long periods, so that strong weathering could take place (Fairbridge, 1967). It becomes evident from fig. 25 that, after the erosion of the top of the Barrios, beach sand containing much glauconite and some phosphate and dolomite were deposited near the Asturian Geanticline, in the region NE of the Narcea anticlinorium; further to the SW was a deltaic environment: pro-delta clays with bar-finger sands. These bar-finger sands do not occur in the transitional beds near Belmonte (Poll, 1963). Still further to the SW pure bluish dark shales of the Luarca were deposited in a quiet marine reducing environment. The relationship between the rise in the eastern part of the Aralla zone and this facies pattern is not clear.

Palaeogeography and concluding remarks

After the folding of the Precambrian Mora Formation, mainly fluvial sediments of the Herrería Formation were deposited. Rocks of the same age and thickness spread out over large parts of NW Spain (Capdevila, 1965; Färber & Jaritz, 1964; Lotze & Sdzuy, 1961; Matte, 1968b; Nollau, 1966; Radig, 1962; Riemer, 1963; Walter, 1965; fig. 22). Marine intercalations became more abundant towards the upper part of the formation.

In a NNE direction more coarse quartzites and conglomerates and less shales were found; we may therefore conclude that the source region was situated in the NNE, in the location of the Asturian Geanticline (van Adrichem Boogaert, 1967). Matte found a possible second source area in Central Galicia (fig. 7 and 23), the core of the Precambrian orogene.

The shallow marine Láncara dolomites and limestones were occasionally brought above sealevel. The Láncara griotte was formed in a somewhat deeper marine environment. The source regions were still situated on the Asturian Geanticline and in Central Galicia (Matte, 1968b; Lotze & Sdzuy, 1961). Limestones of the same age and thickness were deposited in large parts of NW Spain (fig. 7 and 22).

Rocks of the same type and thickness as members A and B of the Oville Formation were also deposited

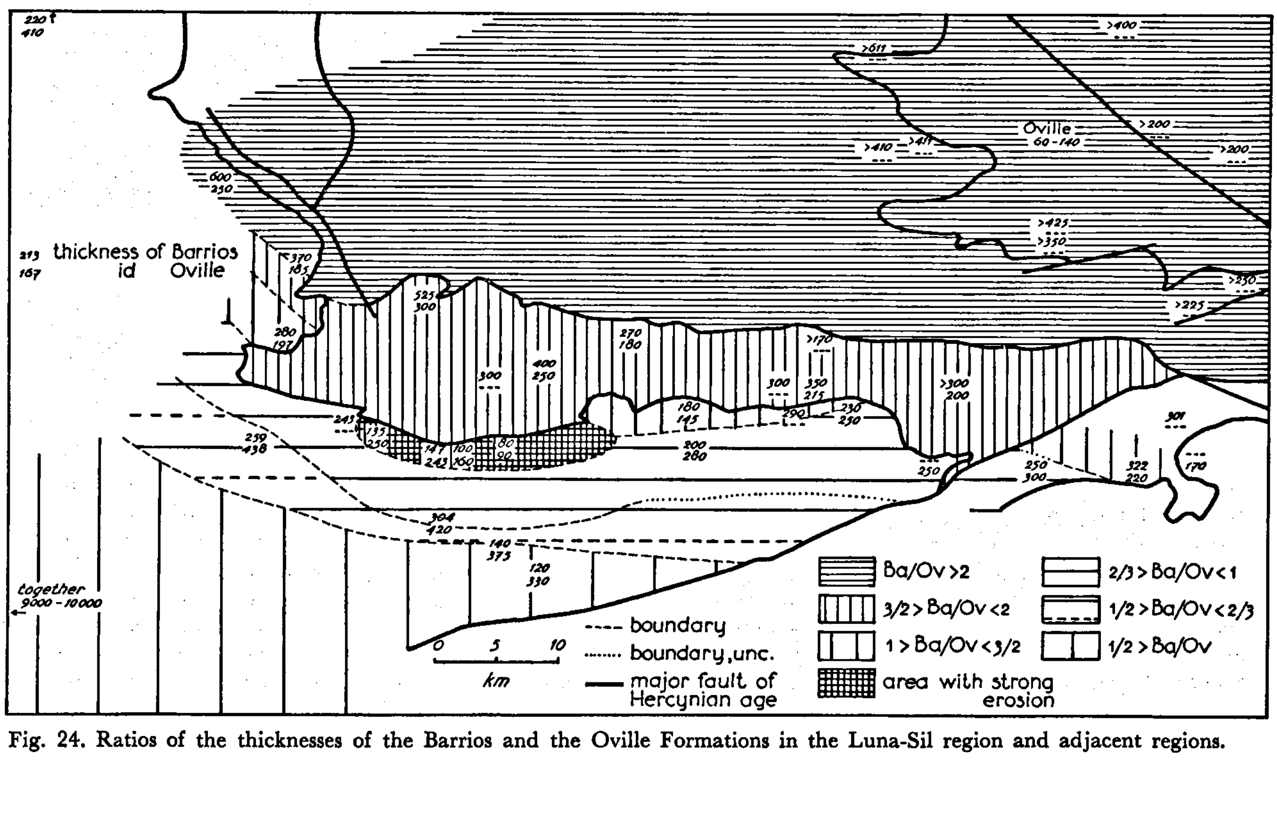

Fig. 24. Ratios of the thicknesses of the Barrios and the Oville Formations in the Luna-Sil region and adjacent regions.

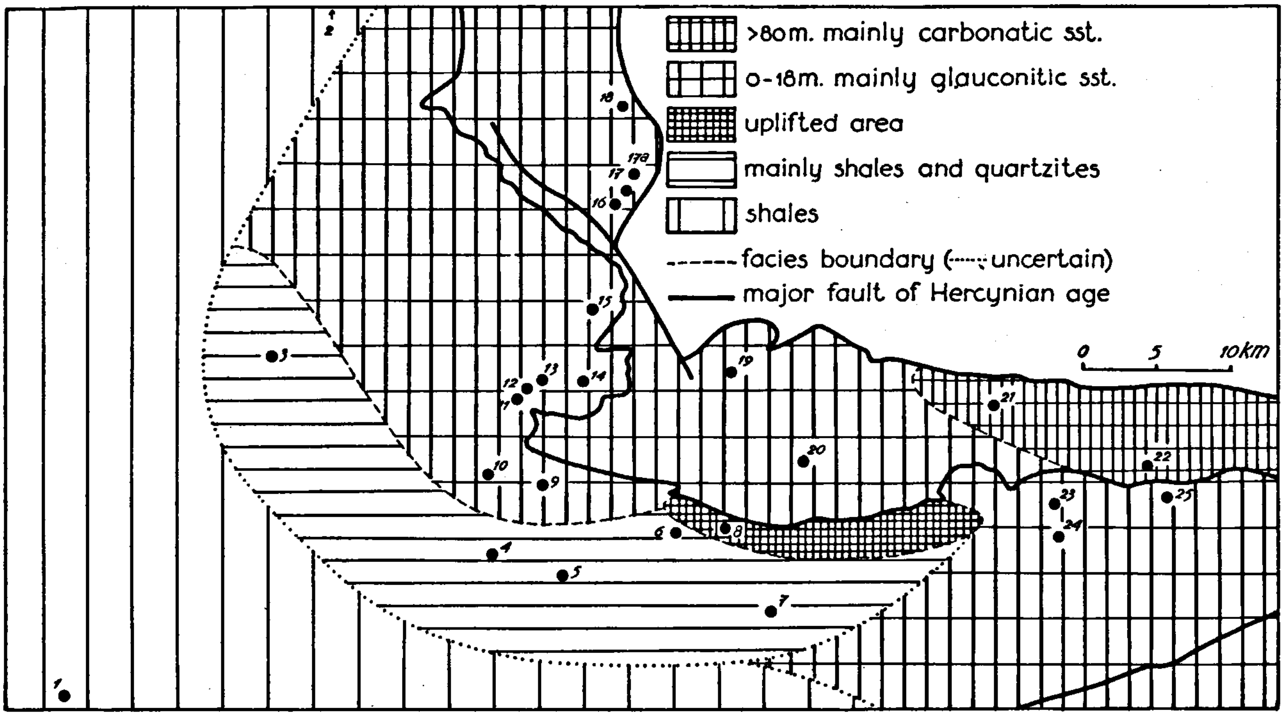

Fig. 25. Distribution of the transitional beds in the Luna-Sil region and adjacent regions.

1.	100 m	quartz.
	2000 m	sh (Luarca)
2.	60–70 m	sh. (90%) + sandy sh. + sst.
	0.4 m	Fe
3.	section 11	
4.	15 m	quartz.
	10 m	silty sh.
5.	0.5 m	ferrug. soil
6.	section 12	
7.		quartz. bed
	several m	sh.
	3–4 m	sandy sh.
8.	section 13	
9.	0.02 m	markasite
10.	2 m	glauc. sst.
11.	section 15	
12.	3 m	glauc. sst.
	2 m	cgl.
13.	8 m	cgl. + quartz. + Fe
14.	section 16	
15.	section 17	
16.	4 m	quartz.
	18 m	sst. + sh.
17.	3 m	glauc. sst. + Fe
17a.	5 m	glauc. sst. + Fe
18.	1.53 m	glauc. sst. + sh. + Fe
19.	section 18	
20.	section 14	
21.	1 m	Fe-ool.
	1 m	sh.
	0.6 m	calc. sst.
	8.40 m	calc. sst. + quartz.
	70 m	calc. sst.
22.	1 m	Fe sst.
	20 m	sst.
	90 m	calc. sst.
23.	1.2 m	sh.
	0.45 m	glauc. sst.
	0.15 m	sh.
24.	1.0 m	sh.
	0.95 m	sst. + sh.
	0.4 m	not exp.
25.	7 m	Fe sst.
	1.25 m	glauc. sst.

in large parts of NW Spain (fig. 7 and 22), but during the deposition of members C and D enormous differences in thickness developed (fig. 7, 22 and 23). The rock type and the enormous thickness in the West-Asturian-Leonese Zone indicate that a rapidly subsiding basin with an orthogeosynclinal character had formed there (Lotze & Sdzuy, 1961; fig. 17 and 23), while in the Cantabrian Zone the formerly stable shelf became unstable. The sediments were generally deposited in shallow water. During the deposition of the Barrios the rapid subsidence diminished and more stable conditions prevailed, giving rise to a better winnowing of the sediments in a more shallow environment. The decrease in the thickness of the Oville Formation towards the Asturian Geanticline and the increase in the thickness of the coarse deposits of the Barrios towards that region (fig. 17) indicate that the source region was still situated there; the source region in Central Galicia still existed, too, but during the deposition of rocks of the same type and age as the Barrios this region emerged (Matte, 1968b). After the deposition of the Barrios, the supply of coarse material stopped and the Asturian Geanticline extended towards the west and south: only beach and tidal deposits, generally very thin, were formed in the Cantabrian Zone during the Upper Arenig, while geosynclinal sediments were still being deposited in the West-Asturian-Leonese Zone (fig. 25). A secondary rise originated in the place of the Aralla zone (fig. 24 and 25). After the Arenig the entire Cantabrian Zone emerged, until the Formigoso black shales extended over this region during the Upper Llandovery. The former source region in Central Galicia remained a rise, but did not emerge (Matte, 1968b). In the West-Asturian-Leonese Zone the deposition of orthogeosynclinal sediments continued with the very thick Luarca shales.

SILURIAN

Formigoso Formation

Comte (1959) introduced the name 'Schistes du Formigoso'; the type locality lies in the Formigoso valley, SE of Villamanín in the Bernesga valley.

Previous work was done by Comte (1959) and Kegel (1929).

The best-exposed sections occur NNE of Torre and near the Churros; the basal black shales were intensively folded in all outcrops visited.

Age. – Lowermost Upper Llandovery to Upper Wenlock (Comte, 1959).

Lithology. – The lower part of the formation consists of pure, very finely laminated, carbonaceous black shales with graptolites. The main components of these shales are clay minerals, micas and some small detrital quartz grains.

In many places a sharp contact with the top of the Barrios is exposed. On this contact a thin markasite layer was found south of Riolago. In other places a few metres of sandy or silty sediments were intercalated between the Formigoso black shales and the white Barrios quartzites (the transitional beds, p. 151). A weathering clay or an iron concentration were always found at the base of the Formigoso. Markasite and pyrite were also found at higher levels in the black shales. These black shales are absent in the section west of the Arroyo de Formigones.

Determination of the thickness of the pure black shales was not possible, these shales always having been intensively folded into nearly isoclinal parasitic folds. The thickness of the folded black shales in the excellent exposure NNE of Torre is 35 metres, their real thickness will therefore be less. The same thickness was measured in other places in the southern and western part of the Cantabrian Zone.

In the black shales in the NE part of the region mapped, sandstone lenses were found.

Above the basal layers of pure black graptolite shales, sandstone lenses 0.5–3 cm in thickness are intercalated. These lenses become larger and more abundant towards the top, in the upper tens of metres sandstone beds constitute the most important rock type. Burrows and ripples are very frequent in these beds which have a rather constant thickness in a lateral direction. Locally well-developed load-casts occur (fig. 26). In the upper ten metres thick sandstone beds with cross-bedding were found. Chamosite and a few hema-

Fig. 26. Well-developed load-casts in the upper part of the Formigoso Formation NNE of Torre de Babia.

tite ooids occur in the upper tens of metres. The shales have at first a black or grey colour, but in the upper tens of metres they become more and more green.

Thickness. – NNE of Torre: folded black shales 35 metres, upper part 143 metres thick. North of the Penouta: upper part 140 metres thick (section 19). Smaller thicknesses occur in the eastern part of the Aralla zone: 70 metres in total.

Fossils. – Graptolites in the black shales; a few graptolites and a brachiopod in the upper part.

Depositional environment. – The presence of iron concentrations and erosion phenomena (p. 150) at the base of the black shales indicate a period of erosion before the deposition of the black shales. Fossil determinations (Comte, 1959; Nollau, 1966) indicate that a hiatus exists between the Upper Arenig and the Upper Llandovery. Where exposed, the erosion surface is rather flat, although slight irregularities exist enabling the black shale facies to spread out rapidly over this flat area without the development of a basal conglomerate.

The preservation of carbonaceous matter and the presence of markasite and pyrite in the pure shale, finely laminated and without coarse components, indicate a quiet-water, reducing environment. The depth can not be concluded from the lack of current activity or the lack of oxygen, but the alternation of coarse white quartzites with black shales in the transitional beds immediately below the Formigoso in the western part of the region mapped indicate a possible shallow environment with stagnant water. The presence of sandstone lenses in the black shales in the NE part of the region, indicates that this part of the region was situated nearer to the shore.

The presence of *flaser* and *linsen* structures in an alternation of shales and thin burrowed sandstone beds with a constant thickness in a lateral direction and of ripples in the upper part of the formation shows that this part of the formation was deposited in a flat shallow sea with rapidly changing stream velocities. The presence of chamosite ooids indicates that less reducing conditions prevailed than during the deposition of the black shales; these beds constitute the transition to the San Pedro.

San Pedro Formation

Comte (1959) introduced the name 'Grès de San Pedro' and selected the type locality near the village of San Pedro de Luna, now inundated by the Luna lake. A new, good exposure exists there along the new road, at long. 40°53′30″ N.

Previous work was done by Bäcker (1959), Comte (1959) and Mr. G. Kuipers (internal report).

The best exposures occur on the Penouta and NNE of Torre de Babia.

Age. – Upper Wenlock to Lower Gedinnian (Comte, 1959); top: Lower Gedinnian (Brouwer, 1964, 1967).

Lithology. – The San Pedro can be subdivided into three members:

A. The basal member, composed of thick red channelling sandstone beds with hematite ooids.

B. The middle member, composed of an alternation of green shales and red and greenish sandstones with hematite and chamosite ooids respectively.

C. The upper member, composed of an alternation of white quartzites and black shales.

Member A begins with the appearance of the first red sandstone bed, containing hematite ooids. The lowermost beds are generally thin and a rather high percentage of green sandstones, containing chamosite ooids, and greenish shales, such as encountered in the upper beds of the Formigoso, are intercalated. The entire remainder of this member consists chiefly of red, thick-bedded, channelling and cross-bedded sandstone beds, up to 4 metres thick, thin layers of green shales or silts and locally a thin greenish sandstone bed. The red sandstones are mainly composed of rounded or well-rounded quartz grains often with a hematite coating of varying thickness (the hematite ooids, fig. 27). The hematite has often replaced the original cement. At several levels coarse phosphorite grains and pebbles occur, frequently with a concretionary structure; they are often concentrated in badly sorted, coarse sandstones containing a large amount of hematite (up to 67%; Mr. G. Kuipers, internal report). Approximately 6–15 metres from the base of member A such a level, with coarse phosphorite pebbles up to 4 cm in diameter was encountered in

Fig. 27. San Pedro sandstone with hematite ooids, from section 22 (40x).

nearly all sections. This level also occurs in the Bernesga area (Bäcker, 1959).

Clay pebbles and flakes are common in the sandstones. Burrows were only found at the generally sharp sandstone-shale contact and are not as common as in member B. Often *linsen* were encountered in these shales. Bed-thicknesses of up to 4 metres occur in the red sandstones; in a lateral direction the bed thickness changes very rapidly. Some very thick beds show no internal structures, but most of the beds have channels (up to 1 metre deep) or cross-bedding, chiefly of the planar type. An increase as well as a decrease in the fineness of the grains occurs in the sandstone beds in an upwards direction. Ripples are not very frequent.

Member B is mainly composed of light green or grey oolitic sandstones, consisting of quartz grains, often with an iron silicate coating (iron silicate ooids), red oolitic sandstones, consisting of quartz grains, often with a hematite coating, and green shales. Replacement of the original cement by iron silicate (chlorite) is very common; this replacement of the original cement is stronger than in member A.

The bed thickness in the sands is generally 5–10 cm, with a maximum of 50 cm for the greenish sandstones, and of 1 metre for the red sandstones. Laterally the bed thickness does not change very much. Current ripples (fig. 28) are very abundant, as are *flaser* and *linsen* structures and burrows. An increase or a decrease in grain sizes was frequently found in an upwards direction. Locally slump-balls were found. Shell imprints are more common than in member A.

Member C consists mainly of laminated black shales and generally a minor quantity of white or yellowish quartzites, dolomitic quartzites and quartzitic sandstone beds. These beds usually show an internal lamination and sometimes also planar cross-bedding or ripples which have locally been disturbed by burrows. They were often found to contain levels of a high iron percentage and coarse rock fragments, phosphorite pebbles and fossil debris (brachiopods, bryozoans, crinoids). The presence of irregularly distributed ferruginous particles often gives the quartzites a speckled appearance.

The transition to the La Vid is generally gradual: the quartzite beds in the black shales contain more and more dolomitic cement. The base of the La Vid lies at the level where the beds adopt the appearance of a dolomite. These rocks, however, often contain 70% of quartz grains and only 30% of dolomite cement but resemble a carbonate rock (section 30; Mr. G. Kuipers, internal report). An interesting transition from the San Pedro to the La Vid exists west of

Fig. 28. Current ripples in member B of the San Pedro Formation, just south of the *viaducto* de Aralla.

Fig. 29. Crinoid fragments with a hematite coating, found in beds at the San Pedro-La Vid transition in the Arroyo de Formigones (40x).

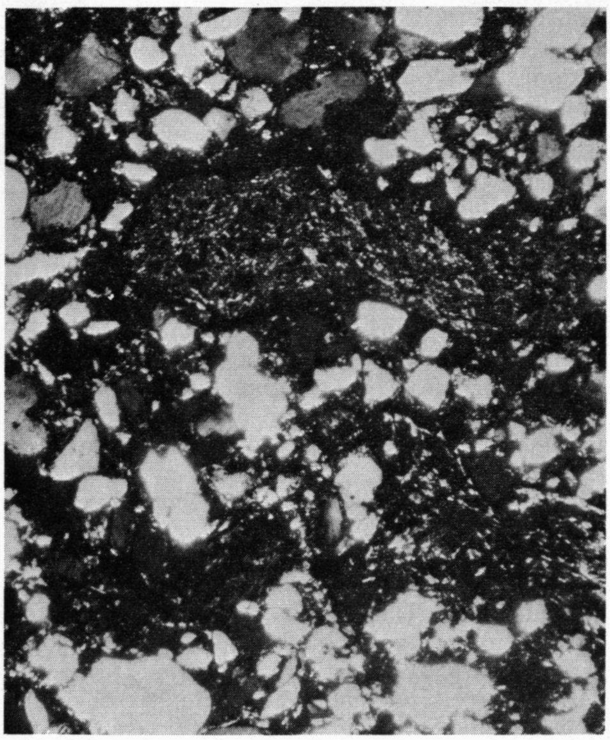

Fig. 30. Tuffaceous sandstone from section 26 (25x).

Fig. 31. Animal tracks, found in the San Pedro Formation NNW of Torre de Babia.

the Arroyo de Formigones, (section 29). A 31 metres thick succession of fossiliferous dolomites, limestones and shales is followed by a sandy succession 32 metres in thickness: mainly argillaceous and arenaceous dolomites with a white quartzite bed 9 metres in thickness and a bed, 2.5 metres in thickness, containing coarse crinoid fragments with a hematite coating or completely replaced by hematite (fig. 29). In the section south of the Yegüero, too, pure sandstone lenses occur in the La Vid. East of Caldas de Luna 30 metres of ferruginous arenaceous dolomites, sands, dolomitic sands etc. were found.

In the Babia Baja unit a very thin San Pedro occurs (sections 22–26). Levels in sections 22 and 23 can be correlated to levels in section 21. The rock type encountered in members B and C is absent in the Babia Baja unit; in section 24 a few beds of the type encountered in member B were, however, found. These facts show that only part of member A is represented there. In section 23, 2 metres of red ferruginous clay is exposed, and in most of the sections the above-mentioned phosphate and hematite level at the base were again found. Section 25 has a coarse breccia at the top, with siderite and phosphate pebbles, rock fragments and shell debris. The base of section 26 is formed by 18.5 metres of tuffaceous sandstone and tuff (palagonite tuff, fig. 30). The palagonite is often weathered, and formed an authigenous clay. The shale above it, of an immature composition, may be of the same origin (Mr. G. Kuipers, internal report).

Thickness. – The thicknesses of members A, B and C respectively are: NW of Torre 64 m, 66 m and 22 m; in Abelgas 110 m, 54 m and 30 m; on the Penouta 178 m, 48 m and 59 m; near the aquaducto de Aralla $\geqslant$ 59 m, $\geqslant$ 53 m, $>$ 13.5 m.

Fossils. – Brachiopods and shell imprints in members A and B; brachiopods, crinoids and bryozoans in member C; burrows and tracks (fig. 31) throughout the entire formation.

Depositional environment. – Berg (1944), Borchert (1953) and Braun (1964) studied the formation of iron oolites. Many of their conclusions will be incorporated in the following analysis. The iron, and also the silica of the iron silicate, dissolved in the sea water under weak acid conditions, presence of CO_2, decomposing organic material and the absence of oxygen (Braun, 1964). An important part of the iron in the San Pedro may, however, have derived from volcanic rocks. The thick palagonite deposits occurring in the Ubiña zone were rapidly altered into an authigenic clay; only a rapid accumulation, without reworking of the sediment, can explain why the palagonite has been preserved there.

Palagonite is a hydratation product of basaltic glass at a low or moderate temperature. It has the composition of basalt with an addition of about 20% of water, and occurs in vitreous basaltic tuffs formed under conditions of drastic cooling and subsequent prolonged saturation with water (Peacock, 1930).

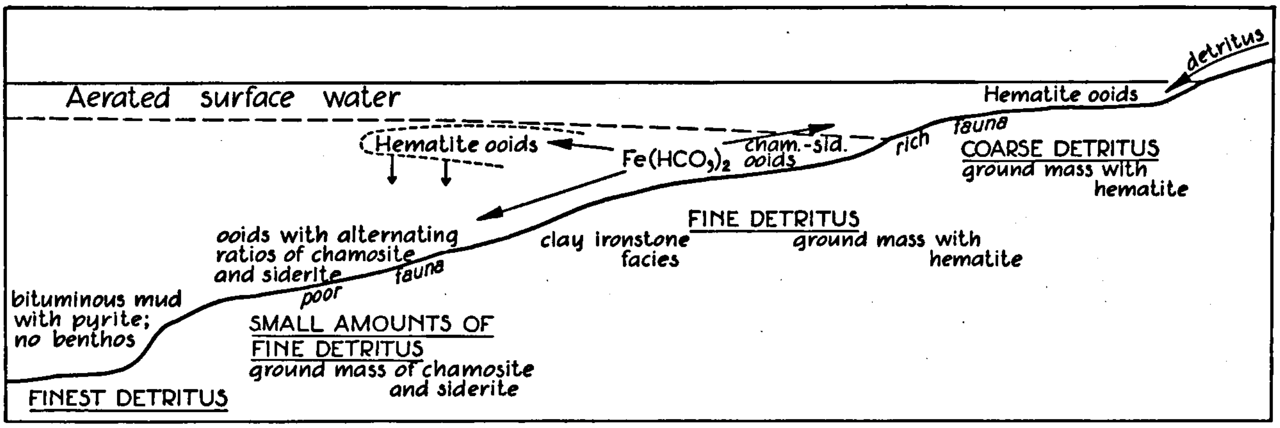

Fig. 32. Distribution of facies in a region with sedimentary iron; the increase in depth is not required for explaining the different facies. After Borchert (1953).

Peacock found its average composition to be: 35% SiO_2, 13.5% Fe_2O_3, 11.5% Al_2O_3 17% H_2O. While the clay was being formed, the iron could dissolve in the sea water, under the conditions already mentioned. It could not be established whether the iron in the San Pedro Formation was entirely of this origin, or derived partly from decomposing organic material, as proposed by Braun (1964).

The dissolved iron can precipitate on entering a well-aerated and more turbulent state of conditions. Chemical precipitation takes place first, then gels are formed on the sea bottom or the formation of ooids occurs (Braun, 1964). The sedimentary structures show that the most turbulent conditions reigned during the deposition of member A; in this very turbulent environment the hematite ooids could form in well-aerated water possibly in the zone of breaking waves and rapidly shifting coastal channels (fig. 32). The increase in depth sketched in figure 32 is, however, not neccessary in explaining the different facies (Borchert, 1953). The coarse, in places very coarse, detritus indicates that land was nearby. Under these conditions phosphorite, too, can be formed in the sea (Borchert, 1953); many of the phosphorite and hematite ooids have been broken by the strong currents.

The formation of the iron ooids is an exclusively marine process (Braun, 1964; Borchert, 1953). The iron of the ooids in thus primary, but most of the iron of the cement and matrix is secondary (Mr. G. Kuipers, internal report).

In the rocks of member B more *linsen* and *flaser* structures, burrows and ripples were found, and the bed thickness does not change very much in lateral direction. This indicates deposition in a less turbulent flat sea with rapidly changing stream velocities. The higher amount of fine detritus indicates that the land was further away. The iron silicate ooids occurring in this member also form under less turbulent and more reducing conditions (fig. 32). The alternation of hematite and iron silicate oolites in this member indicates that the environment was sometimes more oxidizing and sometimes more reducing. The hematite ooids can also be transported by surface currents towards the environment where iron silicate ooids were formed and sedimentate there (fig. 32). The most likely environment is one just within the reach of tidal currents, in a flat sea, seawards from the zone of breaking waves.

The transition from green shales with some kaolinite to black shales with some montmorillonite (Mr. G. Kuipers, internal report) at the transition from member B to member C indicates a transition to more reducing and restricted conditions (fig. 32). The lamination of the black shales, which form the largest part of the rock types in member C, indicates that these shales were deposited in a very quiet environment, possibly further from the coast than the sediments of member B (fig. 32). The coarse quartzites with a planar cross-bedding and some burrowing in places were probably deposited in stream channels. More towards the top, the quartzites receive a dolomite cement and still further upwards the detrital quartz grains gradually grow less abundant. Fossil remains were found only in the sandstones and may have derived from a less reducing environment. The chert encountered locally and the lack of fossils in the basal black shale and sandy dolomite beds of the La Vid are probably also indications of a reducing environment.

The absence of members B and C in the Babia Baja unit and the occurrence there of intraformational breccias, indicate that non-deposition and probably slight erosion took place between the deposition of the middle part of member A and the La Vid.

The red clay in section 23 indicates oxidation of a mud possibly on a coastal intertidal plain (Walker, 1967). A weathering surface was found on top of the San Pedro in section 25 and also in section 30 (fig. 33). The presence of very coarse grained palagonite tuff is an indication of nearby volcanoes. This volcanic region, emerging at times, was possibly the source area for the nearby deposits. Bäcker (1959), who studied the San Pedro in northern and central Asturias and in the

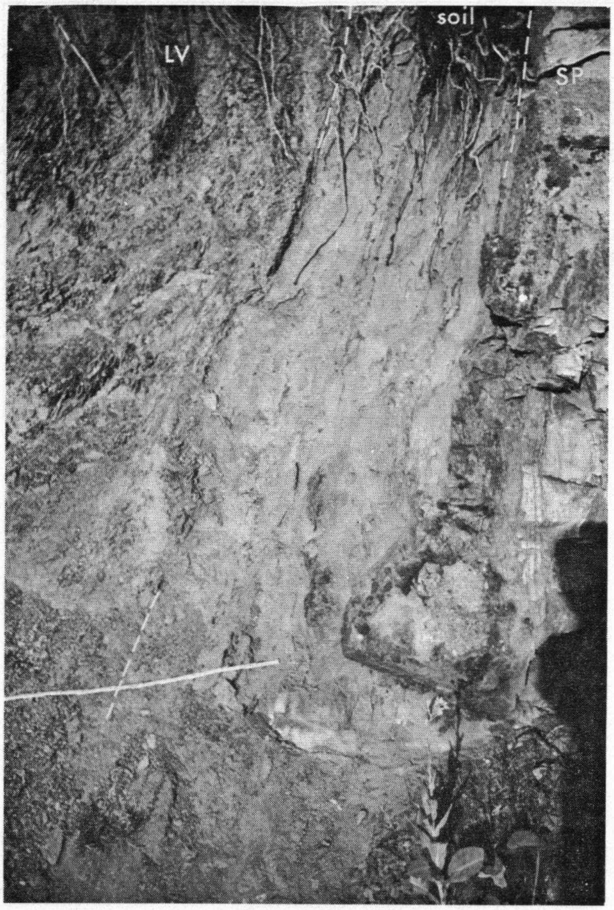

Fig. 33. Weathering surface on top of the San Pedro Formation north of Los Barrios de Luna.

Bernesga area, concluded that the source area for the sediments in those areas was situated in the same region.

Palaeogeography and concluding remarks

The deposition of the fine black shales of the Formigoso occurred in a very quiet reducing environment.

The upper part of this formation was deposited in a less reducing environment in a flat shallow sea with an increasing supply of coarse material.

Deposition of member A of the San Pedro occurred under oxidizing conditions in the zone of breaking waves with a large supply of detrital material. Member B was formed under approximately the same conditions as the upper part of the Formigoso but sometimes also in the environment of member A. Member C of the San Pedro was deposited under more reducing and quieter circumstances with an occasional influx of coarse material. We may therefore conclude that a regression took place from the Lower Formigoso to member A of the San Pedro while the environment changed from reducing to oxidizing, and that the reverse took place from member A to member C.

During the entire Silurian sedimentary iron could form. Braun (1964) stated that the formation of sedimentary iron took place in more or less restricted basins, comparable to the recent Black sea, with a humid climate. This humid climate is required for the formation of the relatively well-aerated fresh surface water supplied by rivers and rainfall. The iron cannot be supplied by rivers because iron would precipitate immediately on entering the fresh, very well aerated river water (p. 161), but derives for a large part from decomposing palagonite and possibly from decomposing organic matter.

The Formigoso black shales spread out over the whole of NW Spain attaining their greatest thickness in the West-Asturian-Leonese Zone and central Galicia. The Cantabrian Zone and eastern Galicia were rises which did not, however, emerge (fig. 7). The San Pedro (and its equivalent the Furada Zone in Asturias) was found only in the Cantabrian Zone (fig. 34); in the West-Asturian-Leonese Zone rocks of a Ludlow age and younger have been eroded (Bäcker, 1959; Capdevila, 1965; Färber & Jaritz, 1964; Kegel, 1929; Nollau, 1966; Radig, 1962; Riemer, 1963; Walter, 1965), except three outcrops near Ponferrada where limestones, shales and sandstones are exposed (Drot & Matte, 1967). Several erosion horizons in the Sobia unit (Bäcker, 1959) and Babia Baja unit indicate that the Asturian Geanticline was the source region (fig. 34).

DEVONIAN

La Vid Formation

Comte (1959) introduced the name 'Calcaires et Calcschistes de La Vid' and selected a type section east of the village of La Vid in the Bernesga valley.

The most important previous work was done by Comte (1959). The best exposures occur NNW of La Riera, north of Lumajo, north of the Cirbanal, west of Abelgas, and SW of the Churros (only the basal limestones and dolomites).

Age. – Middle Gedinnian to Middle Emsian (Comte, 1959); Middle Gedinnian to top Emsian (Brouwer, 1967).

Lithology. – The La Vid Formation has been subdivided into a lower Limestone Member (A), a middle Calcareous Shale Member (B) and an upper Crinoidal Limestone Member (C).

A generally very gradual transition takes place from member C of the San Pedro to member A of the La Vid. Only at several places in the Babia Baja unit and in section 30 has an erosional contact been found (fig. 33). Member A is very well developed in section 31, which will be used as the reference section for this member. A gradual transition takes place there from the San Pedro to the La Vid: the thin, white, dolomitic and slightly cross-bedded quartzite beds in the predominant black shales become more and more dolomitic; the boundary has been drawn at the level where they have the appearance of a dolomite (p. 159). Near

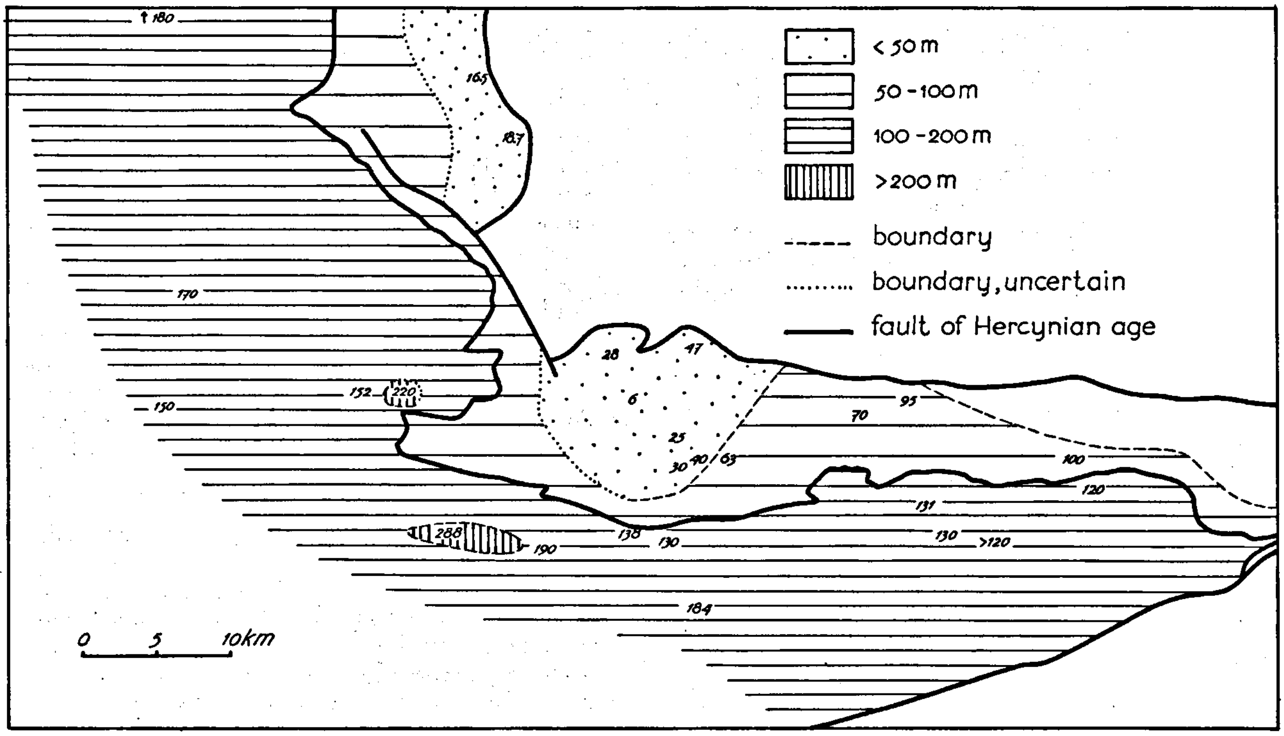

Fig. 34. Thickness distribution of the San Pedro Formation in the Luna-Sil region and adjacent regions.

Lumajo these transitional beds are about 20 metres thick and consist, for more than 90%, of black shales. Lateral differences occur in the composition of the dolomitic sandstones and arenaceous dolomites, so that the lower boundary of the La Vid does not always lie at the same level.

Next come dolomites (bed thickness about 10 cm) with black shale intercalations, overlain by a sequence of limestone beds, mainly composed of brachiopod shells. Thick-bedded, resistant dolomites lie on top of these limestones (bed thicknesses of 10–60 cm). A comparable development occurs in section 30, but the beds there have undulating bedding-planes. This is followed in section 31 by a thick succession of chiefly limestones and in section 30 by only lens-shaped, locally cherty and fetid, dolomite beds and black shales with some carbonaceous matter, which form the remainder of member A. At other places mainly thick-bedded dolomites constitute the lower part of member A. In some dolomites quartz grains, mica flakes (largely replaced by dolomite) and some pyrite grains have been observed. The different development of the basal part of member A in section 29 has been described on p. 160. The upper part of member A is formed by limestone beds, often with much fossil detritus or intact fossils, separated in places by thin shale beds. The bedding-planes are often undulating and locally even nodular limestones developed. Some of the limestones are slightly dolomitic. In the upper part, a generally gradual transition takes place to member B. A level with vertical burrows 8 cm in lenght crops out in the reference section. An increase and a decrease in the grain fineness has frequently been encountered in an upwards direction. More details have been given in the sections. In large parts of the Babia Baja and Luna units several metres of calcareous black shales, locally with thick or thin detrital, very fossiliferous, limestone beds (sections 31 and 34), occur between the last thick limestones and the green shales of member B. Being of the same type as the shales in member A, they have been included in that member.

Member B consists mainly of very fine, often splintery, olive-green calcareous shales with several thin, very fossiliferous beds. Especially north of Lumajo and in the nose of the Abelgas syncline these beds are very well exposed. They consist mainly of generally intact brachiopod shells, cemented by argillaceous limestone. The shales often contain small pyrite grains and iron silicate grains; the latter probably give these shales their green colour. North of Lumajo a sandstone bed was found in this member (section 33). Due to tectonic disturbance (chapter on Structures) the thickness of this member is generally difficult to establish; Sections 30, 32, 33, and 34, however, seem relatively undisturbed.

Member C is mainly composed of red and pink, coarse detrital limestones with a high percentage of crinoid debris; furthermore greenish limestones of the same type and green and red calcareous shales occur. This member is very fossiliferous, especially near Cacabelos. Its thickness clearly increases from SSW to NNE (compare sections 30 and 36).

The transition to the Santa Lucía or to the Caldas Formation is very gradual and there seems to be an intertonguing with the basal part of the Caldas, fossils of La Vid age having been found in the Caldas of

Puerto de la Cubilla (Comte, 1959) and fossils of a Caldas age in the very thick member C of the La Vid in the Sobia unit (Mr. A. de Hoop, pers. comm.).

Thickness. – The thickness of the members varies considerably: Member A: from more than 380 metres in section 31 to 75 metres in section 33; south of Lumajo this member even grows considerably thinner. Member B: from 23 metres in section 37 to 279 metres in section 33. Member C: from 6 metres in section 30 to 158 metres in section 36. The other thicknesses are given in the sections.

Fossils. – Brachiopods, crinoids, bryozoans and corals in member A; brachiopods, trilobites and crinoids in member B; crinoids, corals and brachiopods in member C.

Depositional environment. – The absence of detrital material of any importance indicates that the deposition of the La Vid Formation could take place under very quiet tectonic conditions and possibly at some distance from the coast.

The basal black shales and dolomites may have been deposited in a slightly reducing, quiet environment into which gradually more and more carbonate detritus was carried.

The presence of intact brachiopods and corals in the limestones indicates that they have been deposited in well-aerated and well-lighted water without strong currents. The coarse detrital material in section 29, however, must have been transported by strong currents. Some terrigenous detritus has also been encountered in the basal beds of section 34. The presence in section 30 of lens-shaped black, cherty dolomites and locally carbonaceous black shales, and of black shales in some other sections, indicates that locally the environment was of a slightly reducing nature during the deposition of the upper part of member A.

The monotonous green, often slightly pyritic shales with brachiopod beds of member B suggest deposition in a slightly reducing, very quiet environment suited for the development of brachiopod biostromes. Only in section 33 have sandstone beds been found.

The presence of coarse red detrital limestones in member C indicates that more turbulent and more oxidizing conditions prevailed during the deposition of that member. The absence of terrigenous detritus suggests that still no tectonic activity occurred. It appears that the lower and especially the upper beds of the La Vid intertongue with respectively the San Pedro and Santa Lucía or Caldas Formations. The differences in thickness of the members also suggest an intertonguing between the different members.

After the regression and local erosion before the deposition of the La Vid, the limestone facies spread out over the whole region mapped, reaching its greatest thickness in the northern part of the Abelgas syncline; in a SSW and NNE direction the thickness of member A diminishes. The thickness of member B was often difficult to establish. Member C apparently attained its greatest thickness in the Babia Baja unit.

Santa Lucía Formation

Comte (1959) introduced the name 'Calcaires de Santa Lucía' and selected a type section near Santa Lucía in the Bernesga valley. The most important work was carried out by the same author.

This formation was mapped only in the Babia Baja and Luna units. Rocks in the Babia Baja unit lie in the same stratigraphic position but are very different from the rocks in the Santa Lucía Formation; Smits (1965) introduced the name Caldas Formation for these rocks. The Santa Lucía Formation is best exposed south of Mallo, east and south of Abelgas, just north of Lumajo and just SW of Cacabelos.

Age. – Top Emsian (Brouwer, 1967) to Middle Couvinian (Comte, 1959).

Lithology. – The Santa Lucía is of a rather constant development in the region mapped. Being very well developed and well exposed, section 38 will be used as a reference section. Macroscopically the Santa Lucía can be subdivided into five members.

In its lower part member A consists mainly of grey coarse-grained, detrital, crinoidal limestone and some thin greenish marl beds. Next come dark grey, cherty, internal finely laminated, limestones with locally undulating bedding contacts. Thin-bedded, dark grey limestones with argillaceous partings constitute the upper part of member A.

Member B consists of massive or very thick bedded grey limestone with white calcite veins. Bird's-eye structures were found in the lower part. In the upper part large coral reef debris occur, in the lower part only a few crinoid remains.

Member C begins with an alternation of grey limestones and nodular limestones, overlain by 2 metres of crinoidal limestone. Next comes a breccia bed one metre in thickness with reef debris and limestone boulders, overlain by 2 metres of marls with lenses which are completely composed of fossils. The following limestones contain many corals and crinoids which have been chertified in the upper 7.5 metres. An unexposed part of the section is followed by very fossiliferous limestones and argillaceous limestones (20%). Some of the beds consist entirely of corals, brachiopods, bryozoans and crinoids.

The lower part of member D consists of limestones of medium bed thickness, partly argillaceous, with crinoids, brachiopods and bryozoans. Next comes a coarse-grained limestone, overlain by red and green argillaceous limestones. The next limestones are generally thick bedded and only in the lower part do they contain a small number of fossils. They are overlain by thin-bedded limestones with grey, red and green shale partings. At 4 metres below the top a one metre thick red marl bed with crinoids occurs.

Member E consists largely of red, grey and greenish coarse-grained crinoidal limestones, and red, grey and green marls and calcareous shales. The limestones are locally arenaceous or contain thin, hard ferruginous partings; cross-bedding is rather common. About 16 metres from the base a burrowed, cross-bedded sandstone, 2 metres in thickness, was found. Crinoids, bryozoans, brachiopods and corals are abundant in this member.

In section 38 the Santa Lucía-Huergas contact is sharp. The sandstone bed was also found in the western limb of the Palomas syncline; northwards it thickens rapidly on both flanks (section 39). North of Lumajo (just on the map limit) markasite concretions occur on the Santa Lucía-Huergas contact (fig. 35). In the Luna unit the Santa Lucía-Huergas transition is gradual; many brachiopods and complete calyxes of crinoids were found in the transitional beds there. Member E is very much thinner in the Luna unit than in the Babia Alta unit; the other members are generally constant in thickness and lithology. The Santa Lucía in the southern flank of the Abelgas syncline is much thinner (about 150 metres) than elsewhere. Large pyrite crystals were found there is isolated outcrops. Near Mallo limestone structures were observed in the shape of a reef.

Thickness. – East of Abelgas 248 metres, south of Mallo approximately 150 metres, north of Lumajo 290 metres, west of Cacabelos 351 metres.

Fig. 35. Markasite nodules at the Santa Lucía-Huergas contact, north of Lumajo.

Fossils. – Corals, brachiopods, crinoids, bryozoans, stromatoporoids.

Depositional environment. – The lower part of member A was deposited under the same conditions as member C of the La Vid. The cherty, laminated limestones were deposited in a quiet, possibly restricted, environment.
Bird's-eye structures in member B indicate sub-aerial solution. Coarse reef debris indicate deposition on a reef flank. The same may be concluded for the breccia bed of member C. During the deposition of member C a rich fauna could develop in a generally quiet environment. Member D was deposited under increasingly turbulent conditions which culminated during the deposition of member E, where cross-bedded detrital limestones and even sandstone beds were formed.

The palaeogeography will be dealt with after description of the Caldas Formation.

Caldas Formation

Smits (1965) proposed the name Caldas Formation and selected the type section east of Caldas de Luna. This is the only publication on this subject. The best exposures exist NNW of Caldas and along the road to Puerto de la Cubilla.

Age. – Upper Emsian to Lower Couvinian (Smits, 1965).

Lithology. – A detailed description has been given by Smits (sections 40 and 41), and therefore only some general aspects will be listed here. The Caldas Formation consists of grey, black, pink and red limestones, some of them nodular, and marls, shales and sandstones of the same colours. Red nodular limestones can be used as marker beds. The percentage of terrigenous material is much higher than in the Santa Lucía. Fossils are not so abundant as in the Santa Lucía. Filled-up cavities. bird's-eye structures and stromatolites were found in the limestones. The sandstones and detrital limestones often show cross-bedding. The sections in the Robledo and Villasecino anticlines could be correlated, but the section along the road to Puerto de la Cubilla could not be correlated to the other sections.

Due to the pre-Ermita uplifts, the overlying formations and the upper part of the Caldas have been eroded; during this erosion a karst landscape was formed so that cavities, filled with red ferruginous sandstone of the Ermita, occur, even at 50 metres below the unconformity.

Thickness. – The upper part of the formation has been eroded everywhere. North of Caldas 224 metres remained, east of the Arroyo de las Rozas 228 metres and near Puerto de la Cubilla 382 metres. Further to the north, in the Sobia unit, more than 600 metres of limestones, marls and shales have been measured between the La Vid and the Huergas (Mr. A. de Hoop, pers. comm.), a considerable part of the formation in the present region must therefore have been eroded.

Fossils. – Brachiopods, crinoids, corals, bryozoans, gastropods, stromatoporoids, algae, ostracods, lamellibranchs and spicula of spongae.

Depositional environment. – The intercalation of more terrigenous material than in the Santa Lucía indicates that the Caldas Formation was deposited closer to the shore than the Santa Lucía Formation. The percentage of terrigenous material increases in a NNE direction (Smits, 1965). For several levels, the presence of stromatolites indicates an intertidal environment. Bird's-eye structures were frequently found; they may indicate sub-aerial solution; the filled up cavities also suggest this.

The presence in the Sobia unit, where the Ermita overlies the Portilla, of a complete Caldas Formation, with the La Vid below and the Huergas above it, indicates that the Caldas Formation lies at the same stratigraphic level as the Santa Lucía, and that they probably intertongue. Mr. J. de Coo (pers. comm.), who compared both formations, is of the same opinion. We may therefore conclude that during the deposition of the Caldas and Santa Lucía Formations the land was situated in the NNE, on the Asturian Geanticline. Parallel to the coast a reef facies was present (Santa Lucía), while between the coast and the reefs the Caldas Formation was deposited. Shifting of these facies areas, however, frequently occurred.

Huergas Formation

The name 'Grès et Schistes de Huergas' was introduced by Comte (1959), who selected a type section near the village of Huergas in the Bernesga valley.

Most of the previous work was carried out by Comte (1959).

Complete sections only occur in the Babia Alta unit. The best-exposed sections occur north of Lumajo and in Quejo.

Age. – Upper Couvinian and Lower Givetian (Comte, 1959).

Lithology. – The Huergas Formation consists mainly of black micaceous and carbonaceous silty shales and drab-coloured, decalcified, argillaceous sandstones with burrows.

No complete sections occur in the Abelgas syncline. After a gradual transition from the Santa Lucía, mainly shales with some clay ironstone concretions follow; several sandstone beds and, north of the Sierra Blanca, three limestone beds, are intercalated. The transition from silty shale to argillaceous sandstone and *vice versa* is gradual.

Better-exposed sections were found in the Babia Alta unit. The Huergas in the SW limb of the Palomas syncline is very thin (section 43). A layer of black calcareous silty shales, 4.60 metres in thickness, which, with a rather sharp contact, overlies the marls and detrital limestones of the Santa Lucía, is followed by 28 metres of greenish siltstones and arenaceous siltstones, containing in the lower part clay ironstone and calcareous nodules and a thin calcareous sandstone bed with fossil moulds. Next come 32 metres of greenish and brownish arenaceous siltstones and sandstones with bed thicknesses of between 2 and 7 cm. Bed thicknesses and grain sizes diminish downwards. The upper 36 metres are not exposed, but elsewhere mainly black micaceous silty shales were found at this level.

The section in Quejo is much thicker and a higher percentage of sandstones occurs there. The basal 10 metres are not exposed, then follow 110 metres of greenish grey, decalcified, micaceous, argillaceous sandstones (subgreywackes; fig. 36) with many burrows and small fossil moulds; the bed thickness is about 5–20 cm.

The following 49 metres are not exposed. They are followed by 8 metres of sandstones of the same type as those in the lower part, but fine grained and with more silty intercalations. About 20% of the sandstones are purer and show a lamination with micaceous partings.

The following 46.5 metres are mainly composed of black, micaceous and slightly calcareous silty shales with an increasing percentage of thin burrowed argillaceous sandstones (up to 20%) in the upper part.

The following 43 metres are the most interesting. The lower 18 metres consist of sandstones, parallel laminated (or with a very flat cross-bedding) with micaceous partings and decalcified, often burrowed, micaceous subgreywackes with a low iron percentage. About 5 metres from the base a very ferruginous, micaceous coquinoid limestone bed 40 cm in thickness

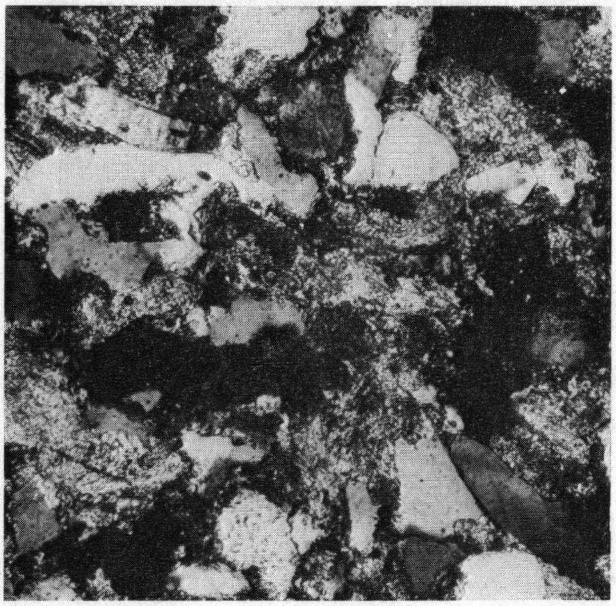

Fig. 36. Argillaceous sandstone from the Huergas Formation near Quejo (100x).

crops out. The upper 25 metres consist of micaceous subgreywackes and some greenish quartzites, with in the lower 12 metres 50% of black micaceous silty shales. All rocks are decalcified. Burrows and cross-bedding are present. Most of the rocks are slightly ferruginous and at two metres from the base a ferruginous, micaceous, decalcified sandstone, full of fossil moulds, crops out.

The upper 56.5 metres consist of nearly pure, fine, black micaceous silty shales with only a few sandstone lenses. The subgreywackes and silty shales often contain a rather high percentage of carbonaceous matter, often macroscopically visible.

A gradual Santa Lucía-Huergas transition exists NE of the Lagunas de las Verdes (just north of the map) where an alternation of impure sandstones, shales, marls and detrital limestones 50 metres in thickness was found between the last pure detrital limestones and marls of the Santa Lucía and the first black silty shales of the Huergas. This is possibly a more terrigenous development of member E of the Santa Lucía.

Thickness. – At least 150 metres in the Abelgas syncline; 100.6 metres north of Lumajo; 314 metres between La Vega de los Viejos and Cacabelos; 323.50 metres in Quejo.

Fossils. – Crinoids, brachiopods, bryozoans, cephalopods and corals.

Depositional environment. – The dark colour of the sediments and the presence of often rather large amounts of carbonaceous matter indicate that large parts of the formation were deposited in a more or less reducing environment. Burrowing animals were, however, still abundant. At several levels red coquinoid limestones, rich in hematite, and sandstones occur, too, indicating oxidizing conditions near the coast (fig. 32).

The often high percentage of matrix in the sandstones indicates that they were buried rapidly and that no winnowing could take place. The laminated, clean sandstones with micaceous partings and locally a very flat crossbedding, indicate that these sandstones were possibly deposited on a beach where winnowing could take place (p. 151). Clay ironstone and calcareous nodules were found only in the silty shales in the Abelgas syncline and the western limb of the Palomas syncline. From the publications of Comte (1959), Evers (1967) and Rupke (1965) it could be deduced that in more easterly regions, too, nodules only occur in the southern sections. The clay ironstone nodules indicate deposition under slightly reducing conditions at some distance from the coast (fig. 32).

The largest thicknesses were measured in the central part of the Babia Alta unit (more than 300 metres); towards the SW (western limb of the Palomas syncline) and the NW (Sobia unit) the thickness decreases rapidly (about 100 metres); the sediments grow coarser from the western limb of the Palomas syncline to the Saliencia syncline.

It may be concluded that the Huergas Formation was deposited in a rather rapidly subsiding region with the source area in the NE (Asturian Geanticline), generally under slightly reducing conditions. In the areas further to the NE oxidizing conditions could also prevail, these areas being closer to the shore; beach deposits were also found there. The sediment association of the Huergas is typical of an unstable shelf (Krumbein & Sloss, 1963, p. 505).

Portilla Formation

Comte (1959) introduced the name 'Calcaires de la Portilla' and selected a type locality in the valley of the Arroyo de la Portilla, west of Matallana-Estación in the Torío valley.

Previous work was carried out by Comte (1959).

The Portilla is best exposed in the places where the sections have been measured (sections 45–49).

Age. – Upper Givetian and Lower Frasnian (Comte, 1959).

Lithology. – The Portilla Formation was subdivided into four members. Member A consists mainly of cross-bedded, coarse-grained, arenaceous, detrital limestone beds with a brown weathering colour and in the lower part some thin black micaceous silty shale and greenish marl beds (together 20% of the rock volume). The limestones contain generally hard ferruginous streaks, which accentuate the sedimentary structures. In the detritus remains occur of brachiopods, corals and crinoids. In the extreme north-easterly section (section 48) many impure sandstone and calcareous sandstones are intercalated. Still further to the NE, in the NE flank of the Saliencia syncline (north of the region mapped),

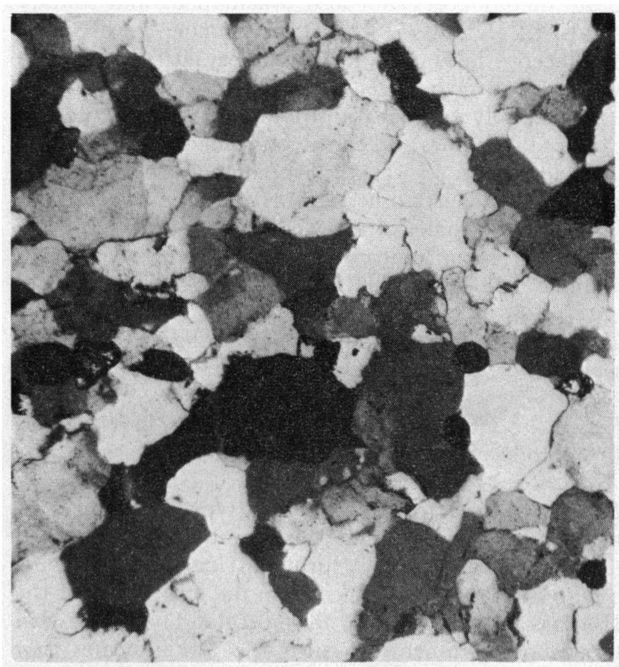

Fig. 37. Pure quartz sandstone with heavy mineral streak, from member C of the Portilla Formation near Quintanilla de Babia (40x).

this member consists mainly of burrowed sandstones and siltstones with many bryozoans and with a smaller quantity of calcareous rocks.

Member B consists mainly of light grey, well-bedded or massive limestones. In sections 46 and 47 an oolite occurs in the lower part of this member. All sections in the Palomas syncline contain a biostromal limestone above this oolite, composed of solitary corals and crinoids. Above this level massive limestones locally occur in section 46, whilst coarse reef debris were encountered in sections 47 and 48. Section 49 is generally thick-bedded; in section 45, however, thin-bedded limestones and some marls were found. Brachiopods, solitary corals, crinoids and bryozoans are common. The top of member B (below the sandstone-limestone contact) in section 46 has been slightly dolomitized.

Member C overlies member B always with a sharp contact, and, in the SW sections, consists nearly entirely of pure, parallel-bedded quartz sandstone with heavy mineral streaks (fig. 37) and with locally a very flat cross-bedding. In section 47 these sandstones grade into thamnoporal limestones, the thamnopores often constituting nearly the entire rock. In section 48 only thin-bedded limestones and some marls and shales occur which become sandy in the upper part of member C. The upper 10 metres there are cherty and contain many thamnopores. This member contains much sandstone in the section in the NE flank of the Saliencia syncline: first 30–40 metres of burrowed, cross-bedded sandstones and mudstones (30%), next 3–4 metres of thamnoporal limestone, followed by 12–15 metres of burrowed calcareous sandstones with thamnopores.

Member D consists mainly of locally silicified, thick-bedded, grey limestones often with many fossils, especially thamnopores, but also brachiopods and solitary corals. In section 45, where this member is very thin, argillaceous layers occur. In member D in the Quejo syncline very coarse reef debris occur.

A generally gradual transition takes place to the calcareous sandstones and calcareous shales of the Nocedo. Only in section 45 was a sharp contact found between thin-bedded cherty limestones of the Portilla and very fine, greenish shales of the Nocedo.

In the Sobia unit, bird's-eye structures are very abundant in the Portilla Formation.

Thickness. – The thicknesses can be derived from the sections.

Fossils. – Solitary and reef-building corals, brachiopods, stromatoporoids, bryozoans, crinoids.

Depositional environment. – The type of sediment of member A indicates that this member was deposited under slightly oxidizing, turbulent marine conditions; a regression thus took place from the upper part of the Huergas to member A of the Portilla. The increase in terrigenous sediments towards the NE indicates that the source area was still situated there.

Oolites in the lower part of member B (in section 46 and 47) show that this part of the member was deposited there under very turbulent shallow conditions, in the zone of breaking waves. The reef debris found locally indicate that reefs were present during the deposition of the upper part of member B. An irregular, slightly dolomitized member B-member C contact in section 46 might indicate a short period of non-deposition.

The pure, parallel-bedded quartz sandstones of member C, with heavy mineral streaks, are an indication of beach deposits (p. 151). In section 47 and the section in the NE flank of the Saliencia syncline these sandstones grade into thamnoporal sandstones and limestones.

During the deposition of member D reefs were formed.

It was concluded that during the deposition of member A the source area was situated in the NE. The presence of argillaceous layers and the absence of reef debris and thick-bedded or massive limestones in member B in section 45 might be an indication of deposition at a greater distance from the reefs, which were then present in more north-easterly sections. During the deposition of member C intertonguing beach deposits and thamnoporal biocoenoses were formed. During the deposition of member D reefs were present, especially in the NE. The rapidly diminishing thickness, the absence of reef debris, and the increase in argillaceous sediments in a south-westerly direction indicate that member D in the SW was deposited at some distance from the reefs. The abundant bird's-eye structures in the Portilla of the Sobia unit indicate subaerial erosion in a region still closer to the Asturian Geanticline.

Nocedo Formation

Comte introduced the names 'Grès de Nocedo' and 'Schistes de Fueyo'. Because the Fueyo shales have only been found very locally, they have been included in the Nocedo Formation as the Fueyo Shale Member (Evers, 1967). The type locality for the Nocedo Formation lies in the Bernesga valley, near the village of Nocedo; that of the Fueyo Member in the Arroyo del Fueyo, a tributary of the Bernesga, north of Puente de Alba.

Previous work was carried out by Comte (1959) and Mr. P. E. Pieters (internal report).

Only in the SW limb of the Palomas syncline does a complete Nocedo crop out. In all other places this formation has been partially or completely eroded. The base of the formation is exposed in the westernmost outcrop between Mallo and Abelgas and east of Quejo (section 51); the upper part is very well exposed along the road between Villaseca and Puente de las Palomas (section 52).

Age. – Lower and Middle Nocedo: Middle and Upper Frasnian and Lower Famennian; Fueyo Member: Middle Famennian (Comte, 1959).

Lithology. – A gradual transition takes place from the

Portilla to the Nocedo. The upper beds of the Portilla are generally arenaceous or argillaceous. The lower part of the Nocedo consists of yellowish, brown or drab-coloured, thin-bedded, argillaceous limestones, calcareous sandstones, marls and shales (section 51). The often very fossiliferous sandstones have frequently been decalcified, so that only moulds of crinoids, brachiopods and especially bryozoans remained. North of Lumajo (section 45) more than 40 metres of pure fine shales form the base of the Nocedo. More upwards, bed thicknesses and grain sizes increase and the rocks become less calcareous. This succession ends in the flank of the Palomas syncline with at least 43 metres of cross-bedded, coarse-grained, white quartzites.

These quartzites are overlain, with a sharp contact, by dark grey shales, with a few calcareous nodules at their base, in which, after a few metres, an increasing number of thin, originally slightly calcareous sandstone beds are intercalated (section 52). Their rock properties and their stratigraphic position indicate that these shales might be correlated to the Fueyo shales (Comte, 1959). After 25 metres, sandstones constitute the major component of the rock. In the following number of thin, originally slightly calcareous sand-bed thicknesses increase graduallyy and slump levels occur. The Nocedo-Ermita boundary was drawn below the first ferruginous sandstone bed.

Thickness. – Approximately 300 metres in the SW flank of the Palomas syncline.

Fossils. – Brachiopods, crinoids, bryozoans.

Depositional environment. – During the deposition of the lower part of the Nocedo, the finest sediments accumulated in the SW flank of the Palomas syncline. Mr. P. E. Pieters (internal report) showed that the sediment transport took place from north to south in the Alba syncline and the eastern part of the Abelgas syncline. Hence it may be concluded that the Asturian Geanticline was again the source area. The gradual transition from thin-bedded, fine-grained sediments in the lower part of the formation to coarse-grained, thick-bedded, cross-bedded quartzites in the middle part, indicates that deposition took place under increasingly turbulent conditions with a supply of increasingly coarse material; therefore this is clearly a regressive sequence. The increase in the amount of coarser detritus may indicate stronger erosion in the source region, possibly due to epeirogenic uplifts there. The Fueyo shales, which overlie these quartzites with a sharp contact, indicate that these movements culminated during the Lower Famennian.

A comparable regressive development occurred from the base of the Fueyo Member to the Ermita Formation. The deposition of black shales with some calcareous nodules could take place in a very quiet environment, but an increasing amount of coarse material in a more and more turbulent and unstable environment (slumps) gave rise to the deposition of the cross-bedded quartzites of the Ermita.

It may be concluded that the Asturian Geanticline was the source area, where the erosion increased during the Upper Frasnian possibly due to epeirogenic uplifts; these uplifts culminated during the Lower Famennian. After a short period of quiet sedimentation new uplifts began, culminating during the Upper Famennian. This picture is in accordance with the conclusion of van Adrichem Boogaert (1967), who also stated that important uplifts already took place during the uppermost Frasnian and Lower Famennian.

Palaeogeography and concluding remarks (fig. 38)

After the uplifts in the Asturian Geanticline which gave rise to the erosion of large parts of the San Pedro, the calcareous facies of the La Vid could spread out over the entire region. The upper detrital limestones are much thicker in the NNE, so that the Asturian Geanticline was possibly again the source area. During the deposition of the intertonguing Santa Lucía and Caldas Formations, a thick sequence of limestones and terrigenous sediments accumulated near the Asturian Geanticline, while further seawards more calcareous sediments were formed, partly in a reef facies. Occasional emersion took place, especially near the Asturian Geanticline.

The general palaeogeographic picture did not change very much during the deposition of the Huergas. Near the Asturian Geanticline coarse-grained sediments were deposited, sometimes near to or on the shore, whereas in the more southerly and south-westerly areas fine-grained sediments were deposited under slightly reducing conditions.

A regression took place during the deposition of the upper part of the Huergas and the lower part of the Portilla. The increasing amount of terrigenous detritus towards the NE in member A of the Portilla indicates that the Asturian Geanticline was still the source area. After deposition of member B, partly in a reef facies, the regression culminated in the deposition of intertonguing beach sands and thamnoporal biocoenoses in member C. Reefs were built during the deposition of member D. The reefs of members B and D did not extend further to the SW than as far as the present axis of the Palomas syncline.

During the deposition of the Nocedo Formation increasing uplifts took place at the Asturian Geanticline. culminating during the Lower Famennian. After a brief quiet period a second uplift took place during the Upper Famennian.

LOWER CARBONIFEROUS

Ermita Formation

The type locality of this formation is situated near the 'Ermita de Buen Suceso' on the eastern bank of the Río Bernesga, near the village of Huergas. Comte (1959) introduced the name 'Grès de l'Ermitage'.

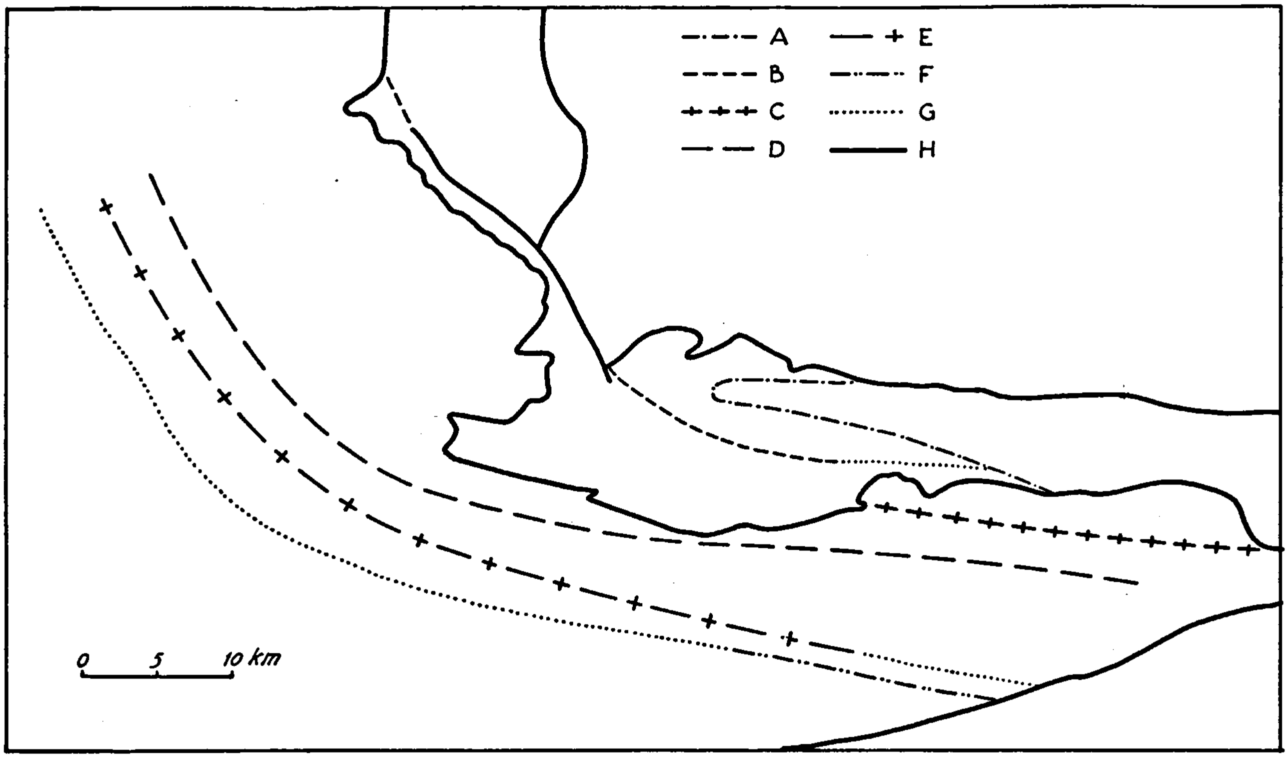

Fig. 38. Facies boundaries during the Devonian in the Luna-Sil region and adjacent regions.
A. Rocks younger than member B of the La Vid not exposed NNE of this line.
B. Boundary between Caldas with less and more than 50% of terrigenous material.
C. Boundary between Caldas and Santa Lucía facies (Mr. J. de Coo, pers. comm.).
D. 1) Boundary between Huergas with many clay ironstone nodules and Huergas with coarse sediments containing much hematite.
2) Boundary between member A of the Portilla Formation and member E of the Santa Lucía Formation containing much terrigenous material and these same members containing little terrigenous material.
E. Boundary between fine-grained and coarse-grained Nocedo.
F. SSW boundary of the occurrence of the Portilla Formation (van Staalduinen, 1969).
G. Boundary uncertain.
H. Fault of Hercynian age.

Van Adrichem Boogaert (1967), Comte (1959) and P. E. Pieters (internal report) carried out the most important previous work.

The sections best exposed and most characteristic of the differently developed Ermita in the present region were found west of the Puente de las Palomas, NE of Villafeliz and south of Puerto de la Cubilla.

Age. – Upper Famennian and Strunian (Comte, 1959); Upper Famennian to lowermost Tournaisian (van Adrichem Boogaert, 1967); Upper Famennian to Lower, locally even Upper, Tournaisian (Higgins *et al.*, 1963).

Lithology. – In the SW part of the Palomas syncline the Ermita apparently overlies the top of the Nocedo conformably. More to the NE progressively older rocks are unconformably overlain by this formation (fig. 43). The angular unconformity, however, can not be observed in a single outcrop. The unconformity surface is flat; deep sink holes, however, were found where this surface lies upon limestones. These sink holes, found in places at even 50 metres below the unconformity itself (section 41), were filled with very ferruginous sandstones. In the eastern part of the Babia Baja unit, where the formation is generally less than 6 metres thick, it is composed of red ferruginous quartz sandstones and micro-conglomerates; only a thin iron crust was found where the formation is absent. More to the west, in the western part of the Villasecino anticline and east of Torrebarrio, 60–85 metres of quartz sandstones were deposited; brownish particles containing iron give the rock a speckled appearence. These particles are often concentrated in thin layers, forming a clearly visible lamination; a flat cross-bedding is rather common. About 20–25 metres below the top of these sandstones 2–3 metres of dark grey shales, weathering yellowish, are exposed. Only in this area do the sandstones grade into a very coarse detrital limestone of a pink or greyish colour and with a constant thickness of 6–7 metres. Coarse detrital limestone fragments, sandstone and ferruginous sandstone particles and fossil fragments are the most important components. Very thin, undulating, red clay partings are frequent.

In the Quejo syncline and in the NE part of the Palomas syncline the Ermita unconformably overlies the lower part of the Nocedo; the unconformity is marked by several ferruginous soils (section 51). The Nocedo-Ermita boundary was drawn at the base of the lowest red soil. The basal part of the Ermita here consists of greyish, cross-bedded quartz sandstones slightly ferruginous in places, with fossil moulds and ferruginous soils. The upper part is composed of white, cross-bedded quartzitic sandstones and quartzites, medium to coarse grained, with some ferruginous silty layers not more than 25–30 cm thick. The quartz grains are well rounded; the sorting is good. The Ermita-Alba contact was nowhere exposed.

In the SW limb of the Palomas syncline the Ermita apparently conformably overlies the Nocedo (section 50). A gradual coarsening upwards occurs from the base of the Fueyo Member to the upper part of the Ermita; the Fueyo-Ermita boundary was drawn below the first ferruginous sandstone. The lower part of the Ermita Formation consists here of ferruginous sandstone beds and thin shale beds; further upwards the sandstones become less ferruginous. Small white quartz pebbles are distributed irregularly over these ferruginous sandstones. Three levels with sand flow rolls (Thomas, 1968; fig. 39) 0.7–1 metre in thickness, whose upper part is eroded, indicate that the sediment there underwent flowing shortly after deposition.

The upper part of the formation consist of coarse-grained, white and pink quartzites with major cross-bedding (fig. 40), ripples and channels. Just as in the lower part of the formation some irregularly distributed, well-rounded, quartz pebbles were found between the quartz grains of sand size. Locally thin conglomerates were found at the base of the channels. About 40–45 metres below the top a very coarse quartzite boulder bed was found 5 metres in thickness, with well-rounded boulders up to 1 metre in diameter (fig. 41). The top of the Ermita Formation there consists of a coarse quartzite breccia 3–4 metres in thickness (fig. 42). The contact with the Alba Formation was nowhere exposed.

In the Abelgas syncline the Ermita, with an erosional contact, overlies the Nocedo with some ferruginous beds at its base. These beds are very fossiliferous. The formation is composed here of pale pink to yellowish white quartzites; the upper part is not exposed.

Thickness. – 194 metres of sandstones and 4 metres of quartzite breccia in the SW limb of the Palomas syncline; 210 metres of sandstone east of Meroy; approximately 90 metres sandstone in the SE flank of the

Fig. 39. Sand flow rolls in the Ermita Formation near Puente de las Palomas.

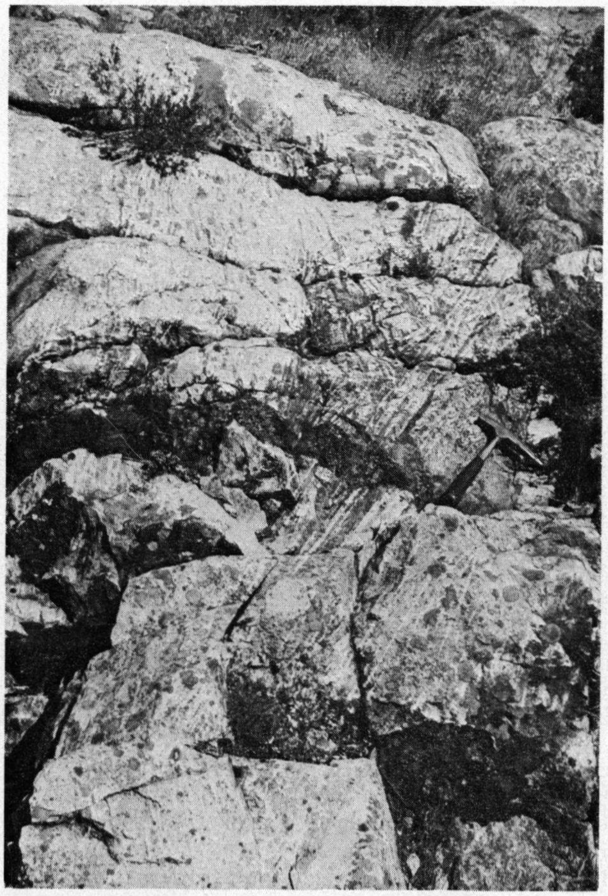

Fig. 40. Major cross-bedding in the Ermita Formation near Puente de las Palomas.

Fig. 41. Well-rounded boulder in the Ermita Formation near Puente de las Palomas.

Fig. 42. Quartzite breccia on the top of the Ermita Formation near Puente de las Palomas.

Quejo syncline; 85 metres of sandstone and 6 metres of coarse detrital limestone south of San Emiliano; 65 metres of sandstone and 7 metres of coarse detrital limestone NE of Villafeliz; 6 metres of sandstone and 1 metre of coarse detrital limestone south of Puerto de la Cubilla. In all other portions of the eastern part of the Babia Baja unit: less than 6 metres of ferruginous sandstone.

Fossils. – Brachiopods, crinoids, bryozoans, corals and conodonts.

Depositional environment. – The smooth unconformity surface indicates an advanced stage of peneplanation after the strong epeirogenic uplifts during the preceding period (fig. 43). The Ermita could easily spread out over this peneplain and cover the entire Cantabrian Zone; this is confirmed by the absence of a basal conglomerate.

The coarse, generally well-sorted sediments in the Babia Baja unit indicate an high energy environment, probably caused by wave action, hence a littoral environment. This environment spread out transgressively over the largest part of the region, probably from SSW to NNE, continuous sedimentation taking place in the SSW and the major uplifts in the NNE. We may compare this with a blanket sand which originates when an abundance of detritus is fed to a stable area undergoing very slow subsidence at a steady rate, so that beach deposits are distributed as a nearly continuous succession of parallel beaches. This is an indication of extreme tectonic stability (Krumbein & Sloss, 1963, p. 550).

In the SW limb of the Palomas syncline, where no erosion took place before the deposition of the Ermita, sand beds underwent flowing shortly after their deposition. This may have occurred during rapid sedimentation in an area with some relief. The irregularly distributed pebbles in the coarse sandstone beds cause a bimodal grain-size distribution. This is mainly found in river deposits (Krumbein & Sloss, 1963, p. 162).

The well-rounded, large quartzite boulders in the upper part of the formation have been transported and rounded by extremely strong currents; the steep cross-bedding and the channels also indicate highly turbulent conditions. The boulder conglomerate was deposited in a very turbulent stream or along a shore with cliffs in the zone of the breaking waves. The iron of the ferruginous sandstones may have derived from the ferruginous soils on the land (section 51). The cross-bedding and the channels indicate transport from NNE to SSW. It may be concluded that the sediments of the Ermita here were deposited near the shore, possibly partly in a fluvial environment or in a delta.

In the uppermost part of the Ermita, a very coarse detrital limestone was found in the Babia Baja unit. These deposits were only found in places having at least 6 metres of sandstones below them. In the region between the León line and the Rozo thrust and its eastern continuation sandstone thicknesses of more than 6 metres and detrital limestones were only found in the eastern part of the Forcada unit (Evers, 1967; fig. 44) and in the region mapped.

Alba Formation

Comte (1959) introduced the name 'Griotte de Puente de Alba' for the red, Lower Carboniferous marker bed.

The type locality lies near Puente de Alba in the Bernesga valley. Winkler Prins (1968) subdivided it into three members: the Gete Limestone Member, the Valdehuesa Siltstone Member and the La Venta Limestone Member which grades into the black platy limestone of the Caliza de Montaña.

Important previous work was carried out by van Adrichem Boogaert (1967), Higgins *et al.* (1963), Sjerp (1967) and Winkler Prins (1968).

Only at a few places in the Babia Baja unit are complete sections present from Ermita to Alba. The best of these sections were found along the road south of San Emiliano, NE of Villafeliz, east of Robledo de Caldas and north of Puerto de la Cubilla.

Being a very thin formation, its thickness had to be exaggerated on the map.

Age. – Lower Viséan to uppermost Viséan (van Adrichem Boogaert *et al.,* 1963; Budinger & Kullmann, 1964; Kullmann, 1961, 1963); Lower Viséan to Lower Namurian A or Middle Namurian A (Higgins, 1962; Kullmann, 1966; Wagner-Gentis, 1963; Winkler Prins, 1968).

Lithology. – In most places a hiatus between Ermita and Alba was demonstrated palaeontologically (van Adrichem Boogaert, 1967; Brouwer & van Ginkel, 1964; Budinger & Kullmann, 1964; Higgins *et al.,* 1963), this was also found at many places in the field where the griotte, with a sharp contact, overlies the Ermita. In the western part of the Villasecino anticline. however, apparently gradual transitions were found (sections 52 and 53). In section 53 the coarse pink detrital limestone grades into the red nodular limestone of the Gete Member and in section 52 the Ermita limestones grade into a grey sandstone 1.20 metres in thickness, containing some brownish weathering grains; this sandstone grades into the Gete Member, through a transition zone, a few cm in thickness. The subdivision into three members does not agree everywhere: in section 52 the Valdehuesa Member is only a few centimetres thick or absent, but in section 53 and north of Puerto de la Cubilla all three members are well developed. At the base of the Gete Member there, a pink to greyish fine-grained, slightly nodular limestone was found, followed by red nodular limestones with cephalopods. The latter grades into the shales and silts of the Valdehuesa Member. This member consists of red siltstones and shales with bedded red chert. The thickness of the chert beds, which often contain white radiolarians, is 2–2.5 cm. The separation between the chert and the shale beds is generally very sharp. The La Venta Member consists in section 53 of 3.5 metres of grey, slightly nodular limestones with four pink nodular limestone beds, each 25 cm in thickness; below the third bed 3 cm of red shales were found. A gradual transition takes place from the La Venta Member into the Caliza de Montaña Formation.

On the Alto de los Grajos only a few metres of greyish and yellowish limestones were found between the Caldas Formation and the Caliza de Montaña. In the NE limb of the Palomas syncline the Alba Formation is very thin and generally not exposed.

Thickness. – The thicknesses measured south of San Emiliano are: La Venta Member 3.50 metres; Valdehuesa Member 4 metres; Gete Member 6.20 metres. North of Puerto de la Cubilla: La Venta Member 3.50 metres; Valdehuesa Member 2.25 metres; Gete Member 8.50 metres.

Fossils. – Cephalopods, small solitary corals, crinoids, radiolarians.

Depositional environment. – The gradual transitions between Ermita an Alba mentioned above, are probably constituted by reworked transitional beds, because at least the Upper Tournaisian and lowermost Viséan are not represented (Budinger & Kullmann, 1964; Higgins *et al.,* 1963; Kullmann, 1963). During this period a condensed sequence of bituminous black shales with phosphate and some pyrite nodules, chert and limestone beds of the Vegamián Formation were deposited in a transgressive sea which rapidly spread out over the Cantabrian Zone: the wide distribution of this very thin formation (now cropping out between the rivers Bernesga and Pisuerga) suggests a transgression over a very flat surface.

During its deposition (Upper Tournaisian), and especially afterwards (lowermost Viséan), slight erosion took place (Higgins *et al.,* 1963; fig. 45).

Rock types of the Vegamián and Alba Formations can occur in one rock assemblage. McKelvey *et al.* (1959) and McKelvey (1967) found the lateral sequence of rocks on a shoaling bottom on which shoreward moving, relatively cold water, is progressively warmed, to be:

1. dark carbonaceous shale, phosphatic shale, phosphorite, and dolomite
2. chert or diatomite
3. several facies of carbonate rock
4. saline deposits, red or light coloured sandstone or shale

These rocks were deposited synchronously and intertongue. As these environments shift with epeirogenic movements, the rocks are found in the same vertical order or the reverse. Incomplete sequences can, of course, frequently occur. The first environment can be compared with the Vegamián Formation, the second with the diatomite chert of the Valdehuesa Member and the chert of the Vegamián, the third with the limestones of the Gete and La Venta Members, and the fourth with the red siltstones and shales of the Valdehuesa Member. From these facts it may be concluded that generally a gradual shallowing of the sea took place from the Lower Tournaisian to the Upper Viséan.

Davis (1918) concluded that bedded chert containing radiolarians originated, partly organically and partly inorganically, from gelatinous chert at the bottom of the sea.

The scarcity of current and wave marks and clastic sediments in both the Alba and the Vegamián Formations can be explained by the presence of a very flat coast which developed during the deposition of the Ermita on the already flat peneplain below it, where waves lose their power (Keulegan & Krumbein, in Winkler Prins, 1968). During Vegamián times the presence of large surfaces of seaweed may also have prevented wave action and caused stagnant water and oxygen deficiency (Krumbein & Sloss, 1963, p. 228). According to Bitterli (1963), Higgins *et al.* (1963) and Winkler Prins (1968) the type of sediment encountered in the Vegamián Formation was deposited in a shallow sea in a warm and moist climate. Bitterli (1963) stated that black bituminous shales, which rapidly spread out over a peneplanated region, mark major changes in the geological history of a region.

The nodular limestones of the Alba Formation may have originated as a lime mud in an oxidizing environment in a sea containing many cephalopods and radiolarians. After burial part of the limestone was dissolved (p. 147) giving the rock its nodular appearance.

It may be concluded that during Tournaisian, Viséan and lowermost Namurian times, very quiet tectonic conditions reigned over a large flat region. Only during the lowermost Viséan did some faint epeirogenic movements take place, giving rise to the erosion of the Vegamián in the present region (fig. 45). Epeirogenic movements also displaced the different sedimentary environments; the vertical succession of these environments indicates a slow regression from the base of the Vegamián to the Valdehuesa Member of the Alba.

Palaeogeography and concluding remarks

The uplift of the Asturian Geanticline attained its maximum in the centre of the San Isidro region (de Sitter, 1966), attaining a second maximum along the León line (fig. 43). In the western part of the region where the unconformity plane in a large area overlies the Portilla, the angle of unconformity is smaller than in the eastern part where the unconformity plane rapidly cuts through progressively older formations. After the peneplanation of the uplifted areas, the Ermita sandstones could spread out transgressively from SSW to NNE under quiet tectonic conditions. The thickest deposits were found in the SSW and SW parts of the region (fig. 44). The Ermita sediments spread out as a blanket over the karst landscape locally present so that the very thin Vegamián black bituminous shales, phosphorites, and chert could, under very quiet tectonic conditions, rapidly spread out over the very flat surface. Deposition took place in an euxinic environment. During the lowermost Viséan only slight local eustatic uplifts were needed for the erosion of this thin formation in the western part of the Cantabrian Zone, SE of the Porma fault and along the Sabero-Gordón line (Higgins *et al.*, 1963; fig. 45).

After this faint uplift the thin red limestones, shales and bedded chert of the Alba Formation were deposited over the entire, very flat, region in a still shallower marine environment than the Vegamián, under oxidizing, very quiet conditions.

NAMURIAN AND WESTFALIAN

Caliza de Montaña Formation

This is an old Spanish name (Boschma & van Staalduinen, 1968). Winkler Prins (1968), who calls it

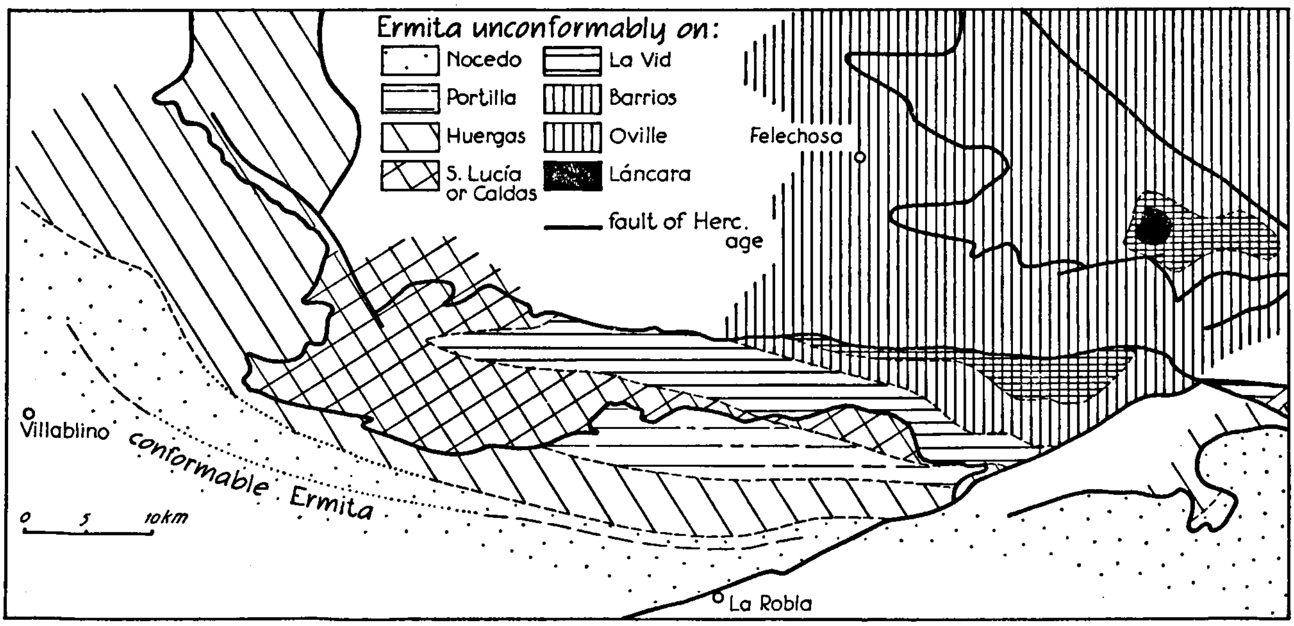

Fig. 43. Map showing the distribution of the various formations, which lie directly below the Ermita in the Luna-Sil region and adjacent regions.

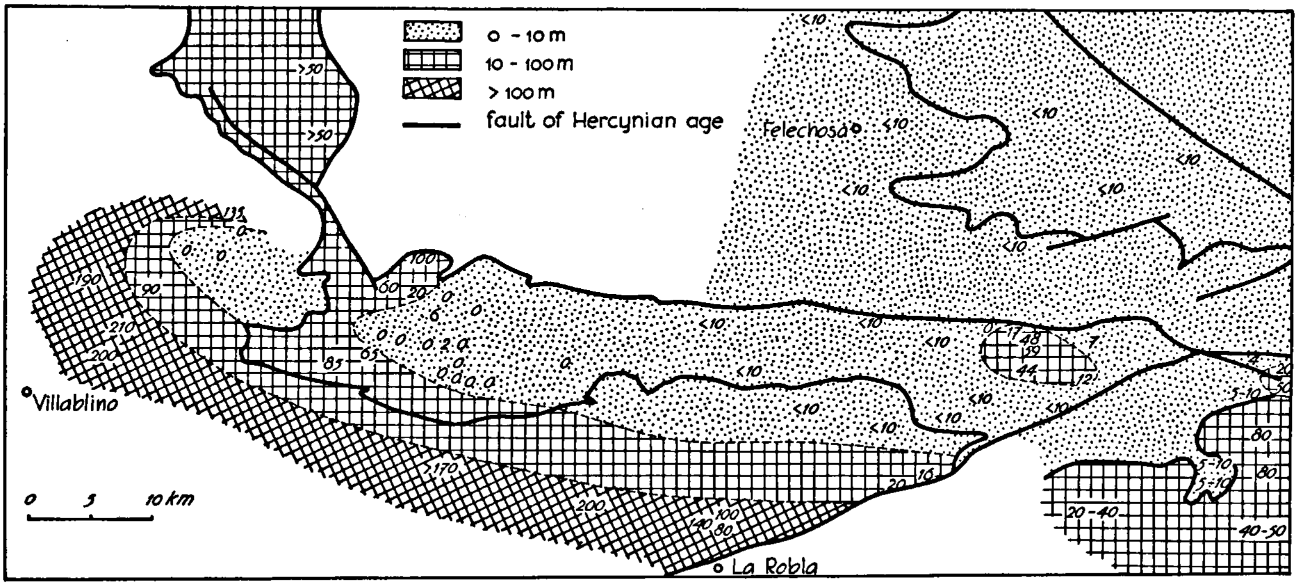

Fig. 44. Thickness distribution of the Ermita Formation in the Luna-Sil region and adjacent regions.

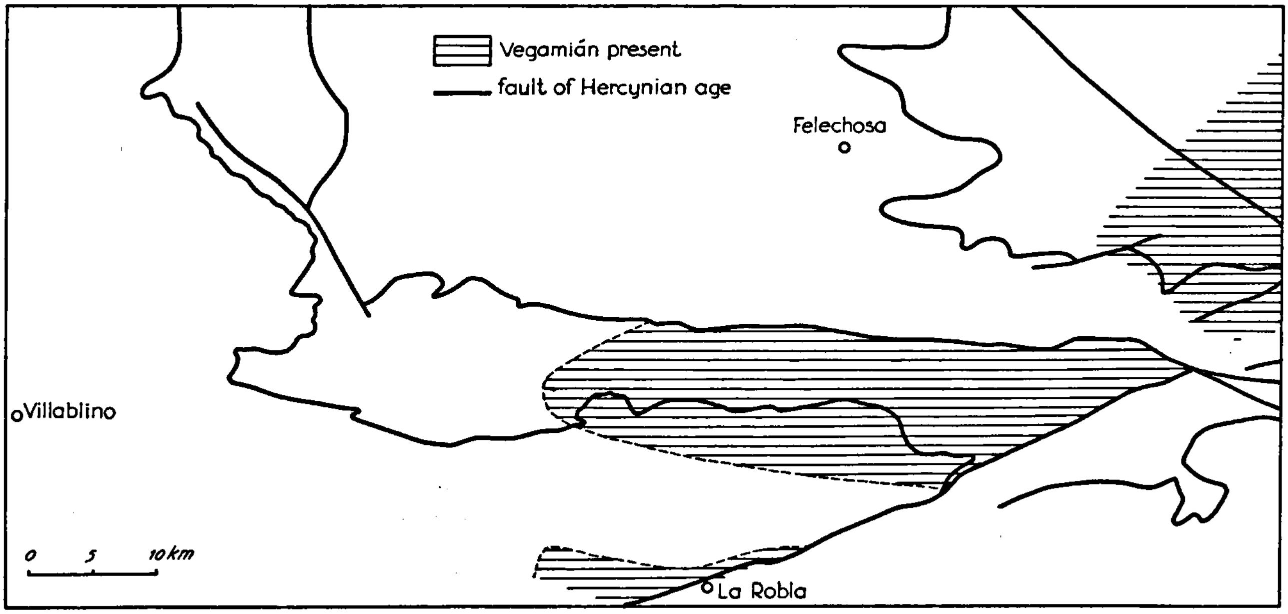

Fig. 45. Distribution of the Vegamián Formation in the SW part of the Cantabrian Mountains.

the Escapa Formation, subdivided it into a lower Vegacervera Micrite Member and an upper Valdeteja Biosparite Member. The type section for the Vegacervera Member lies in a road section along the Río Torío in the Hoces de Vegacervera; the type section for the Valdeteja Member lies east of Valdeteja, along the road to the Río Curueño. Sjerp (1967) and Winkler Prins (1968) carried out the most important previous work.

The Vegacervera Member is well exposed NW of Puente de las Palomas, complete sections of the entire formation being found NE of Vega de Robledo and along the Arroyo del Valle de Retuerto (NW of Puerto de la Cubilla).

Age. – Lower Namurian A to Upper Namurian B (Brouwer & van Ginkel, 1964); Middle to Upper Namurian A (Higgins *et al.*, 1963; Wagner-Gentis, 1963) to Namurian B (Kullmann, 1962) or Upper Namurian B (Winkler Prins, 1968).

Lithology. – The non-fossiliferous Vegacervera Micrite Member consists of black micritic limestone with white calcite ceins. It is a well-bedded, platy limestone with bed thicknesses of between 5 and 20 cm (fig. 94 and 95); the beds, which, in a lateral direction, are constant in thickness, are laminated in many places. At some places grading was observed. When struck with a hammer the rock emits a powerful fetid odour, for which the substance forming the laminae may be responsible. Very locally chert and pyrite crystals were formed. Dolomitization, not related to the bedding-plane and thus postdepositional, and weathering often give these limestones a very massive character.

The Valdeteja Biosparite Member consists mainly of a light to dark grey, thick-bedded limestone with shale and marl beds or laminae, especially in the upper part. The limestones are chiefly coarse to medium grained and well sorted. The thickness of the beds is not constant in a lateral direction. An intraformational breccia was once found near the base, in the SW flank of the Palomas syncline. Due to dolomitization this member locally has a very massive appearance and yellowish to orange-brown weathering colours. At some places authigenic quartz crystals, cinnabar or fluorite were found. Especially the upper part of this member contains many fossils.

The upper part of the Valdeteja Member intertongues with the San Emiliano Formation: intercalations of sandstone, marl or shale are very common there. This intertonguing is particularly clear in the southern part of the Babia Baja unit.

Thickness. – 506 metres NW of Puente de las Palomas; 325 metres NE of Vega de Robledo.

Fossils. – Absent in the Vegacervera Member; crinoids, brachiopods, corals, gastropods, calcareous algae and foraminifers in the Valdeteja Member.

Depositional environment. – The Vegacervera Micrite Member was deposited under inoxic conditions as a fine lime mud by rapid, probably chemical or biochemical precipitation (Teichert, 1965), and covers a very wide flat region (fig. 51). The fine lamination indicates that the accumulation of lime mud and the probably organic matter (Teichert, 1965), which emits the fetid odour, proceeded below the turbulent zone while the supply of terrigenous material was very small. The euxinic conditions suggest that the region was sheltered to some extent from the open sea. Teichert (1965) concluded that a comparable fetid dolomite in Arizona was deposited in a large embayment 30–50 metres in depth.

As corals and blue-green algae were living there at the time, the sea must have been somewhat shallower, well aerated and well lighted during the deposition of the Valdeteja Member (Rácz, 1964). The intercalation of sandstones and shales in the upper part of this member indicates less stable conditions and a transition to the type of sediments encountered in the San Emiliano Formation.

San Emiliano Formation

The name San Emiliano Formation was introduced by Brouwer & van Ginkel (1964); the type locality lies east of San Emiliano, in the region mapped, between the villages of Pinos and Vilargusán.

The most important previous work was carried out by Rácz (1964) and Winkler Prins (1968).

This formation is best exposed in the type section (section 54), along the road between San Emiliano and Candemuela, between Robledo de Caldas and Sena (section 55), and between Vega de los Viejos and Puente de las Palomas. Because the lower part of the San Emiliano Formation intertongues with the Caliza de Montaña, no exact boundary could be drawn in most places.

Age. – Upper Namurian B to base Westfalian A (Brouwer & van Ginkel, 1964); Upper Namurian B to Westfalian A (Winkler Prins, 1968); Upper Namurian B to Lower Westfalian A (Rácz, 1964); Top: Westfalian A (van Ginkel, 1965), Uppermost Namurian C or lowermost Westfalian A (Wagner, 1959).

Lithology. – The transition from the Caliza de Montaña to the San Emiliano is chiefly gradual; a thin limestone conglomerate or breccia was, however, locally found in the NE part of the Babia Baja unit.

In the Babia Baja unit the lower part of this formation consists mainly of poorly stratified grey silty shales and mudstones, greenish or reddish in places, with plant remains and some brownish subgreywacke beds, thin argillaceous limestone bands weathering yellow, and reefs. Near Pinos a quartzite bed was found. The basal part varies in thickness between some tens of metres and hundreds of metres.

In the middle part more pure, light grey limestones occur between the mudstones together with argillaceous limestones and subgreywackes (fig. 46); in the subgreywacke beds slumps were found and ripples were locally formed on top of them. Graded bedding was often found in the mudstones; the individual beds have a medium thickness of 10–15 cm. At some places angular limestone boulders occur in poorly stratified mudstones (fig. 47). The subgreywackes contain about 15% of rock fragments, the quartz grains are subrounded to rounded and sorting is poor.

A preferred direction of sediment transport was not found.

Locally the limestone beds vary in thickness over a short distance, especially in the lower part, where algae and corals built reefs (fig. 48); most of the other rock types change little in bed thickness and lithology in a lateral direction.

Some limestones have a brownish weathering colour, due to the large amount of clay in the matrix, and are fine grained; the individual beds are generally not more than 5–10 cm thick and show some local grading.

Fig. 46. Middle part of the San Emiliano Formation north of San Emiliano (looking north-eastwards).

Other limestones are pure and have a light grey weathering colour; they are also fine grained, but the individual beds are thicker than in the argillaceous limestones. Oolites were found twice at the base of such a pure limestone. These limestones are richer in fossils than the argillaceous ones; one pure limestone is completely composed of algae. Often grading was found: first a coquinoid limestone, followed by a fossil fragment limestone which grades into a fine lime mud. These sequences closely resemble the allodapic limestones of Meischner (1964). The upper part of the formation in the Babia Baja unit contains more greywackes and less limestones; the shales generally have a dark colour and are often carbonaceous. Most of the coal beds are poor. In the lower part some graded bedding is still present, and east of Torrebarrio a layer with convolute bedding is exposed. East of Riolago, just in front of the Rozo thrust fault, gypsum occurs in a carbonaceous shale. More to the east, in an abandoned mine south of Truébano, coal balls were found. These dolomite concretions usually have a crust of pyrite and a diameter of 3–20 cm. They contain a large quantity of organic material such as plants, corals, crinoids and brachiopods (Gómez de Llarena & Rodríguez Arango, 1946). Silica concretions were found in coal beds near Genestosa and La Majua.

Gómez de Llarena & Rodríguez Arango (1948) estimated a total coal reserve of 24,000,000 tons for the entire Babia Baja unit.

Fig. 47. Angular limestone boulder in mudstone in the lower part of the San Emiliano Formation, east of Pinos.

Fig. 48. Reefs in the middle part of the San Emiliano Formation, southwest of Robledo de Caldas (looking westwards).

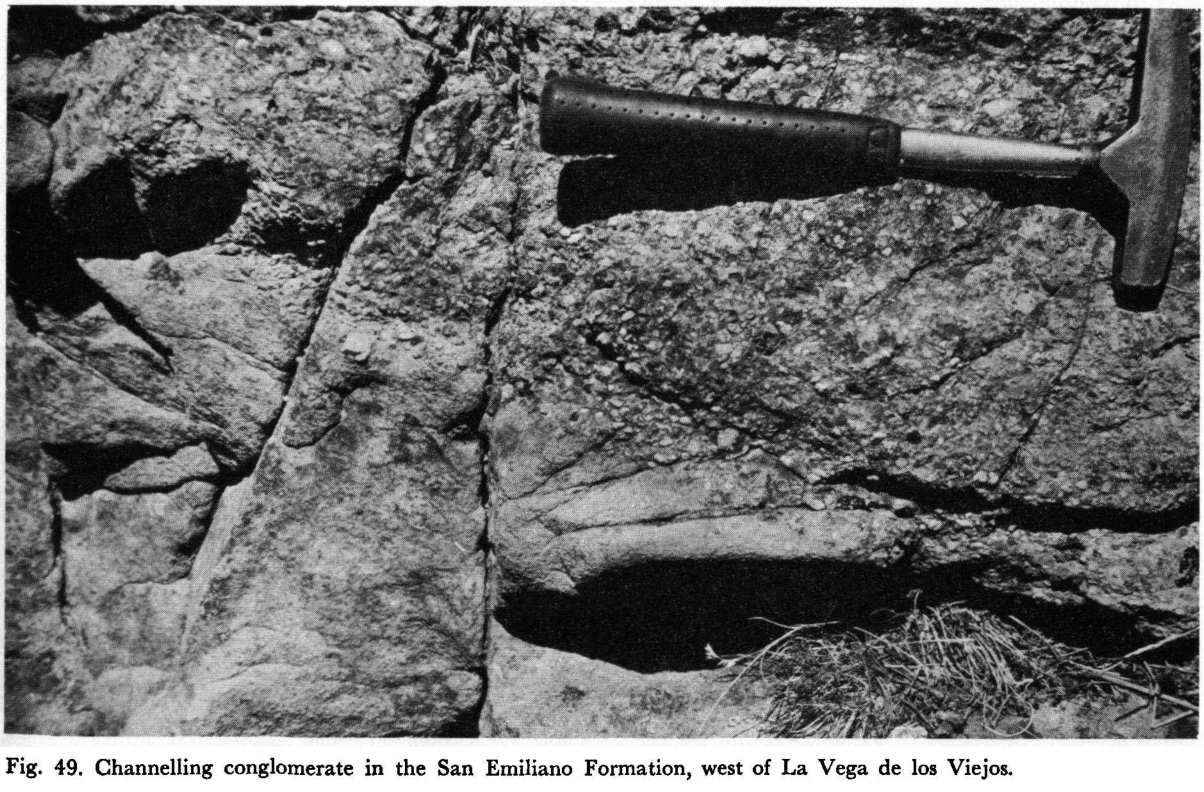

Fig. 49. Channelling conglomerate in the San Emiliano Formation, west of La Vega de los Viejos.

Fig. 50. Greywacke from the San Emiliano Formation, west of La Vega de los Viejos (100x).

The sediments in the Palomas syncline are clearly much coarser than in the Babia Baja unit. Shales with red-brown clay ironstones, only occur to a small extent in the lower part of the formation, and NE of Puente de las Palomas only one limestone band was found. Channelling conglomerates with a clear erosive base (fig. 49) grade upwards into coarse lithic greywackes with plant remains. The diameter of the pebbles varies between 0.5 and 3 cm. The poorly sorted greywackes (fig. 50) contain up to approximately 50% of rock fragments; the quartz grains are subangular to subrounded, the rock fragments are rounded. The channels run from SSW to NNE.

Thickness. – More than 2400 metres in the type section.

Fossils. – Calcareous algae, foraminifers, corals, bryozoans, brachiopods, gastropods, lamellibranchs, ostracods, trilobites, crinoids, plants.

Depositional environment. – The appearance of the first greywackes and shales of the San Emiliano indicates that in the source area the rising increased. The contemporary subsidence of the basin was very rapid, more than 2400 metres of sediment being deposited between Middle Namurian B and Lower Westfalian A. Due to this rapid subsidence no time was left for winnowing of the sediments by currents or wave action so that only very badly sorted sediments with rather badly rounded grains were formed. Periods of relatively slow sedimentation with fine-grained sediments were followed by a sudden inflow of coarse detritus, probably the result of tectonic activity. The presence of many rock fragments in the sediment indicate strong erosion of a not too distant land. The great variety of fossil assemblages also indicates numerous irregular and rapid depth changes (van Ginkel, 1964; Rácz, 1964; Winkler Prins, 1968). Slumps, graded bedding, convolute bedding, angular limestone boulders embedded in clay, and rapid facies changes are also indications of very unstable conditions which may have rendered possible the development of turbidity corrents.

Most of the sediments were deposited in rather shallow water, generally between 10 and 40 metres deep (Rácz, 1965; Winkler Prins, 1968). According to Rácz (1964), the pure limestones were deposited in clear water with a moderate to strong wave action, in which the sunlight could penetrate, whereas the argillaceous limestones were deposited in water of about the same depth, in a calm undisturbed sea. The greywackes and shales were deposited in very shallow water.

The sediments of the upper part of the San Emiliano Formation in the Babia Baja unit were deposited in a still shallower environment; even coal beds could often form.

The presence of conglomerates and coarse greywackes, with subangular to subrounded grains, in the Palomas syncline, of shales, limestones and subgreywackes with subrounded to rounded grains in the Babia Baja unit, and of mudstones in the Central Asturian Coal Basin (Sjerp, 1967) during the period between Upper Namurian B and Lower Westfalian A, shows that the main source area was situated in the SSW; this is also confirmed by the SSW-NNE direction of the channels in the Palomas syncline.

Lena Formation

Barrios (1882) introduced the name 'Grupo de Lena'. Van Ginkel (1965) introduced the stratigraphic term 'Lena Formation' and described a type section 5 km NE of Canseco in the Vegarada valley.

This formation covers only a very small portion of the region mapped and therefore only some general aspects will be dealt with.

Van Ginkel (1965), Rácz (1964) and Sjerp (1967) gave descriptions of the Lena Formation.

Age. – Namurian C to Westfalian D (Sjerp, 1967)

Lithology. – Mainly shales, mudstones, subgreywackes, conglomerates, marine limestones and coal.

Thickness. – Nearly 4000 metres (Sjerp, 1967).

Depositional environment. – During the deposition of the San Emiliano Formation in the Leonides, much thinner and finer sediments were deposited in the Central Asturian Coal Basin in a condensed sequence (Ricacabiello Formation and the lowest part of the Lena Formation). During the Westfalian A more rapid sedimentation began with sediments of the type described above (Sjerp, 1967).

Palaeogeography and concluding remarks

After the deposition of the Alba Formation under very quiet tectonic conditions, the black platy limestones of the Vegacervera Member of the Caliza de Montaña Formation were, during the Namurian A deposited over a large part of the Cantabrian Zone. This again took place under very quiet tectonic conditions. Only south of the Sabero-Gordón line is the Caliza de Montaña absent (fig. 51); there the Cuevas Formation, with sediments of the type encountered in the San Emiliano and Lena Formations, was deposited (Boschma & van Staalduinen, 1968).

The Valdeteja Member of the Caliza de Montaña Formation was deposited in a more turbulent environment during the Lower Namurian B, and the upper part intertongues with the San Emiliano Formation. This indicates that during the Upper Namurian B the greywacke facies spread out over the entire Leonides, except the Forcada and Armada units (Evers, 1967, p. 112; fig. 51). The deposition of the greywackes and other sediments of the San Emiliano took place under very unstable conditions; meanwhile mudstones and shales of the Ricacabiello and lowest part of the Lena Formation were deposited in the Central Asturian Coal Basin, the San Isidro region and the Forcada and Armada units. In these regions deposition of the coarser material began during the Westfalian A, while sedimentation in the other regions ceased. The deposition of greywackes thus started in the SSW during the Namurian A and shifted towards the NNE during the Namurian and Westfalian. The source area was situated in the SSW and migrated towards the NNE, together with the line of maximum sedimentation.

STEPHANIAN

Prado Formation

The type locality of this formation lies in the Cea and Sabero basins (Boschma & van Staalduinen, 1968). The name 'Prado Member' was introduced by Helmig (1965); this member constituted a part of the Cea Formation. The Prado Formation is named after the Hulleras de Prado, S. A. (Helmig, 1965). Important previous work (mainly palaeontological) was carried out by Alvarez Ramis (1965), de la Concha & Jorissen (1961), de la Vega Rollán (1964) and Wagner (1965).

Good exposures exist north of Villayuste, NW of La Urz and near Manzanedo in the southern basins, and along the Río de San Miguel in the Villablino basin.

Age. – The sediments in the Villablino basin are of a

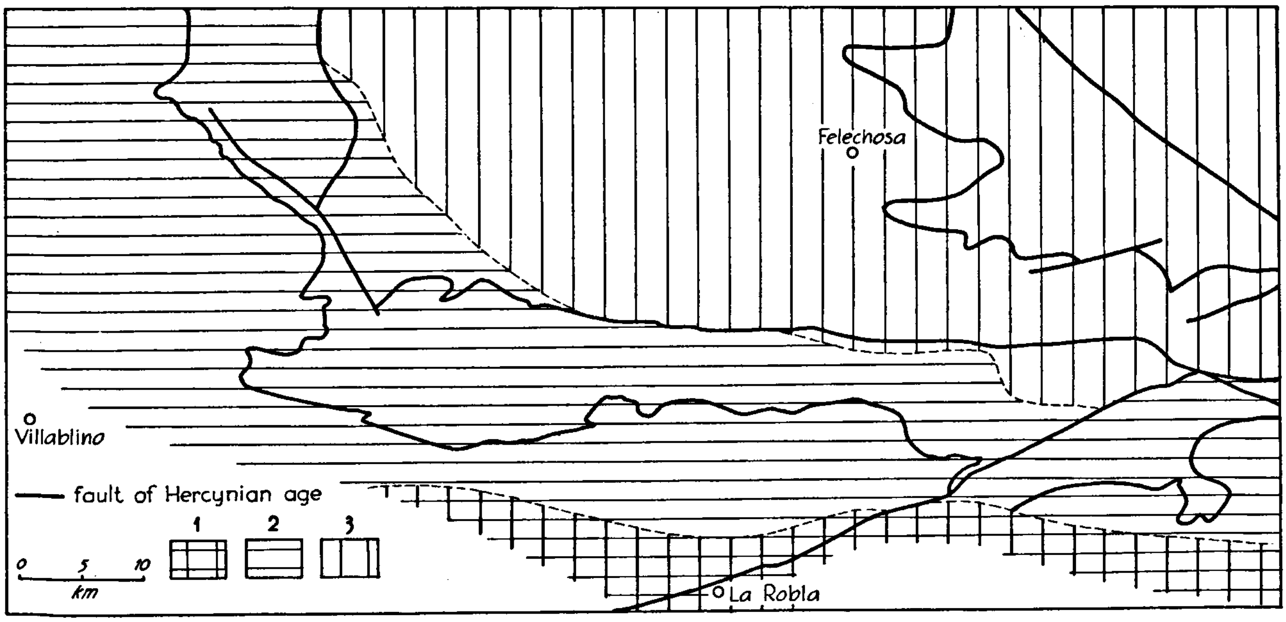

Fig. 51. Distribution of the greywacke facies in the Luna-Sil region and adjacent regions during the Upper Carboniferous.
1. Distribution of the greywacke facies during the Namurian A-Middle Namurian B; Caliza de Montaña in regions 2 and 3.
2. During the Middle Namurian B-Middle Westfalian A; condensed sequence of fine-grained sediments in region 3.
3. During the Middle Westfalian A-Upper Westfalian D; youngest sediments in regions 1 and 2 are of a Westfalian A Age.

Stephanian B–C age (flora determinations by Alvarez Ramis, 1965 and Wagner, 1965).
P. & A. Hernández Sampelayo (1947) determined limnic, brackish and marine faunas and proposed a Middle Namurian to Westfalian B age, but as all other authors arrived at a Stephanian B–C age, this determination might be erroneous. Most age determinations in the La Magdalena basin indicate a Stephanian B age (Almela, 1951; Alvarez Ramis, 1965; Wagner, 1965); the sediments in the subbasin of Sosas del Cumbral are also of a Stephanian age (Hernández Sampelayo & Almela, 1942).

Lithology in the Villablino basin. – The Prado Formation always unconformably overlies the older rocks. A clear, sharp, unconformable contact was found between Puente de las Palomas and Villaseca (fig. 52). The basal beds vary in lithology from place to place, depending on the kind of rock immediately below them: quartzite breccias, mixed breccias, sandstones or shales. A thick sequence of very coarse to fine-grained greywackes, sandstones, argillaceous sandstones, mudstones and coal beds was deposited in the asymmetrical Villablino basin; the thickest sequence

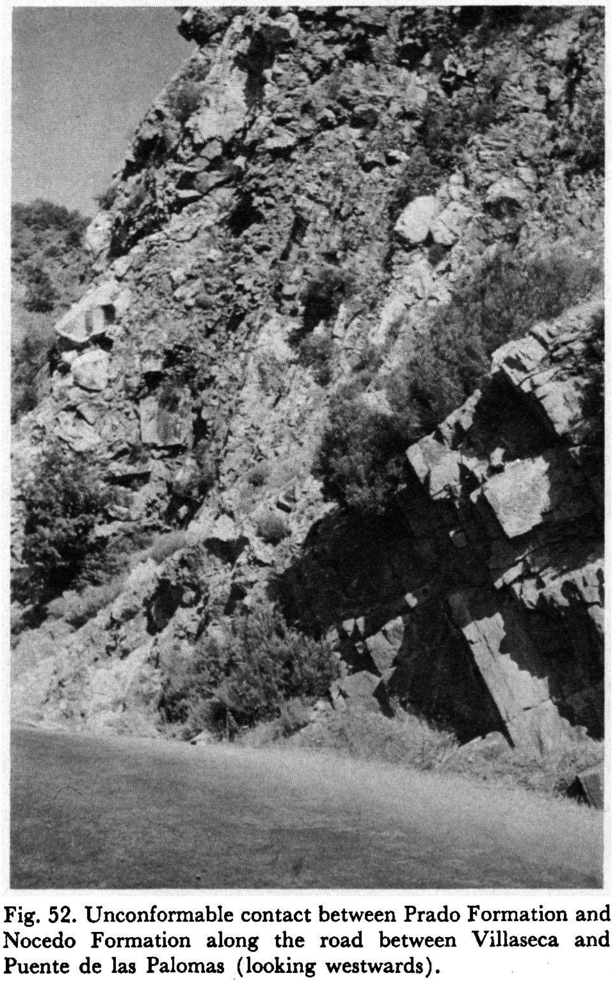

Fig. 52. Unconformable contact between Prado Formation and Nocedo Formation along the road between Villaseca and Puente de las Palomas (looking westwards).

accumulated in the southern part of the basin. Limestones are absent and conglomerates were only encountered in the western part of the basin, outside the region mapped, except some thin microconglomerates with well-rounded pebbles of very limited lateral extent. The only marker beds are the granite-porphyry sills which could locally be traced over some distance. Washouts, coarse and irregular cross-bedding and ripple marks were frequently encountered. Graded bedding is rare, but cyclothems are a common phenomenon. Sedimentation in these cyclothems began, in the ideal case, with grey or yellowish channelling greywackes or sandstones, and continued with sandy shales or argillaceous sandstones of a greenish colour. These are overlain by dark-coloured shales which towards the top become carbonaceous and pass into coal. In two places a laminated calcareous shale with lamellibranchs was found upon this coal bed. These cyclothems have been found in all stages of development, previously deposited sediments having been eroded during the emergent stages before the basal sandstone was deposited.

Due to thermometamorphism during the intrusion of the granite-porphyry sills, the surrounding rocks were altered. The alteration zone is, however, generally thin, but at some places the coal was transformed into anthracite or even into natural coke as far as several tens of metres from the sills (de Alvarado, 1951; Vidal Box, 1943). Sulphur was locally found near the sills.

De la Concha & Jorissen (1961) found 15 workable coal beds, together 7.5 metres in thickness. They reported an annual production in the entire basin of 155,000 tons of anthracite and 783,000 tons of coal, and a total reserve of 10 million tons of anthracite and 269 million tons of coal.

Thickness in the Villablino basin. – A total thickness of more than 2450 metres was measured in the valley of the Río de San Miguel.

Fossils in the Villablino basin. – Abundant plant fossils; two outcrops with lamellibranchs. P. & A. Hernández Sampelayo (1947) described marine and brackish water fossils: *Lingula, Posidoniella* and a nautiloid.

Depositional environment in the Villablino basin. – Because the rocks of the Prado Formation lie unconformably on all other rocks, they must have been deposited after or during the last stages of the main deformation phases. The abundance of continental fossils, the coal beds and the type of sedimentary structures suggest a mainly continental environment. The breccias found in places at the base of the sequence often are scree material. As the sediments have their greatest thickness near the southern boundary fault, this fault might have determined the place of this intramontane basin during sedimentation (fig. 109). Although no grain sizes were measured, it may be assumed that the basin is a clastic wedge (Krumbein & Sloss, 1963, fig. 13.13). The cyclothems in this clastic wedge may not have been caused by the rhythmic movements along the boundary fault which gave rise to a rhythmic supply of coarse sediments (Helmig, 1965), but it is quite possible that the different grain sizes in the cyclothems represent successive stages in a complex system of stream channels, flood plains and swamps, as elements of a shifting river system, possibly with once or twice a short marine transgression.

The sills intruded some time after the sedimentation, contact zones having been found on both sides of the sills.

Lithology in the southern basins. – Most of the deposits lie between two WNW trending normal faults, which form a graben (p. 215). Where the basins are not bounded by a fault, they unconformably overlie the Precambrian Mora Formation. The upper 15–20 metres of this formation below the Stephanian sediments has a red weathering colour and is often brecciated by the weathering. The part of the La Magdalena basin covered by this map is mainly filled with grey, well-cemented, very coarse quartzite conglomerates with highly rounded cobbles and boulders of a high sphericity (fig. 53). At several places at the base of these deposits channels, filled with Stephanian conglomerates, were found in the Mora. Only west of La Urz and east of the Arroyo de Carrocedo were shale, sandstone and coal beds found.

Finer sediments were also encountered in the subbasins near Sosas del Cumbral, but conglomerates are also present there; the northernmost subbasin consists entirely of conglomerates.

Fig. 53. Boulders in the Prado Formation south of Salce.

The subbasins along the southern boundary fault between Manzanedo and Villayuste were filled with red conglomerates with pebbles and cobbles consisting of red slates, derived from the underlying weathering zone of the Mora. They are less sorted than the quartzite conglomerates, and the percentage of cobbles and pebbles in the rock is much lower (about 90% for the quartzite conglomerates and about 50% for the slate conglomerates). Almela (1951) erroneously considered these conglomerates to be a coarse development of the San Pedro.

In the two small Prado outcrops near Soto, in the extreme SE of the map, sandstones and coal are exposed.

Thickness in the southern basins. – 550 metres, west of La Urz.

Fossils in the southern basins. – Only plant fossils.

Depositional environment in the southern basins. – Because the sediments here are of about the same age as the sediments in the Villablino basin, they must also have been deposited during or after the last stages of the main deformation phases. Coal beds and plant fossils indicate a continental environment. First sandstones, shales and coal were deposited in large parts of the southern basins under quiet conditions, possibly comparable to the picture of the Villablino basin, followed by the deposition of coarse conglomerates. Channels in the underlying Mora, the high rounding and sphericity, and the large size of the quartzite pebbles and cobbles indicate a very turbulent environment and probably a rather long transport by a river. Less turbulent conditions prevailed in the southern part of the graben; this may be concluded from the lower degree of sorting and the presence of easily erodable slate pebbles and cobbles which indicate a rather rapid deposition. Probably for most of the time the river channel lay in the northern part of the graben, where now the axis of the syncline is found, so that the most turbulent conditions prevailed there.

The Villabandín basin and its subbasins. – The subbasins and the SW part of the Villabandín basin itself were filled with the same quartzite conglomerates as found in parts of the La Magdalena basin. An equal environment may be decided upon. The NE part of the basin was filled with sandstones, sandy shales and thin coal beds, presumably deposited under the same conditions as the sediments in the Villablino basin.

TERTIARY

Riacos Formation

Mabesoone (1959) subdivided the Tertiary deposits in the Duero basin into different facies. The Vega de

Riacos facies of Miocene age closely resembles the red bed deposits NE of Villayuste. Van Staalduinen (1969) gave comparable deposits the name Riacos Formation and this name will also be used in the present region.

The outcrops of post-Hercynian sediments, without fossils, in the Riello basin and north of Robles de Laciana closely resemble the Miocene deposits further to the south (Miss E. van de Wilk, pers. comm.). They also resemble the Upper Cretaceous Voznuevo Formation (Evers, 1967; van Staalduinen, 1969), but their flat-lying position on the ESE dipping peneplain or pediplain, similar to that of most of the Miocene deposits further to the south, suggests these deposits to be of a Miocene age. As the Riacos Formation is of a Miocene age, too, the sediments in the Riello basin and north of Robles de Laciana will be provisionally included in the Riacos Formation.

Lithology NE of Villayuste. – Almela (1951) already observed the large quartzite cobbles on the flat hill top NE of Villayuste and the red soil material on the slopes of the same hill, which seems to have been washed down from the hill top. These deposits lie at the same level as the subhorizontal identical red beds SW of La Magdalena (van Staalduinen, 1969).

Lithology in the Riello basin. – Chiefly fine and coarse unconsolidated micaceous sands were found here. They have striking white, yellow, orange and red colours, and often white clay beds and fine conglomerate lenses, with pebbles up to 5 cm, are intercalated (fig. 54); locally clay balls of more than 50 cm in diameter were found. Coarse cross-bedding and channels are common (fig. 54), as are hard residual ferruginous layers. Coarse quartzite cobbles were found south-west of Oterico, at the base of the formation; the uppermost deposits contain rather coarse pebbles deposited in channels.

Thickness in the Riello basin. – Approximately 200 metres.

Depositional environment in the Riello basin. – The oxidized deposits with micas and white clay indicate deposition in a warm and humid climate; the residual ferruginous layers may have formed in a soil. This indicates a terrestrial environment. The steep cross-bedding and the channels, in the often coarse sediments indicate transport from WNW to ESE by a fast running river.

The exposure north of Robles de Laciana. – A restricted occurrence of deposits of the same kind as found in the Riello basin crops north of Robles de Laciana, on the old Luna landscape (p. 186). Its position on this old surface and the analogy with the sediments in the Riello basin made it plausible to include these deposits in the Riacos Formation.

Fig. 54. Cross-bedded sands with white clay lenses in the Riacos Formation, west of Amío.

Younger deposits (p. 188) are all of a Quaternary age. Only lateritic soils on a high terrace (p. 188) may be of a Pliocene age, just as the peneplain or pediplain on the Precambrian (fig. 64), which was formed after the deposition of the Miocene (Miss E. van der Wilk, pers. comm.).

CONCLUSIONS

There are indications that the Asturian Geanticline already existed during the deposition of the Herrería and Láncara Formations (Lotze & Sdzuy, 1961). Its presence really manifested itself, however, during the deposition of member C of the Oville Formation. From that time on, the Asturian Geanticline played an important rôle during the Palaeozoic history of the Luna-Sil region. During the Lower Palaeozoic an eugeosyncline, in which about 12,000 metres of sediments accumulated, developed SW of this region; this development began during the Upper Middle Cambrian (during the deposition of member C of the Oville Formation). The Cantabrian Zone, of which the Luna-Sil region forms a part, was generally a stable shelf, except during the deposition of member C of the Oville, which took place under unstable conditions; only 2000 metres of sediments accumulated on this shelf during the Lower Palaeozoic. At the end of the Arenig, epeirogenic uplifts took place which first brought about the formation of a secondary rise in the Aralla zone and later caused the emergence of the entire Cantabrian Zone. Less important uplifts during the Gedinnian in the Asturian Geanticline gave rise to the erosion of nearly the entire San Pedro Formation in the NE part of the Luna-Sil region.

During the Devonian, mainly limestone deposition took place on a shallow stable shelf; only during the deposition of the Huergas did more unstable conditions prevail and deposition of sandstones and shales took place. It could be concluded from the facies distribution that the Asturian Geanticline was the source area. Due to the deep erosion in the regions SSW of the Luna-Sil region, no Devonian and Lower Carboniferous sediments crop out there, and therefore nothing can be remarked on the development of these rocks there. The facies in the SSW part of the region concerned show an increasing distance from the coast in a NNE to SSW direction and hence the presence of a possible rise in the place of the Narcea anticlinorium seems very unlikely.

Due to uplifts along fundamental faults during the Famennian, whose centres were located along the León line and in the San Isidro region, strong erosion could take place, so that in a SSW to NNE direction the erosion surface overlies successively older formations. After peneplanation of the uplifted region, the Ermita rapidly spread out over the entire Cantabrian Zone, and was overlain by the Vegamián black shales. Minor uplifts along the same fundamental faults gave rise to the erosion of the Vegamián Formation in the region concerned and in several other regions.

The Alba Formation and the Vegacervera Member of the Caliza de Montaña Formation were deposited under very quiet conditions.

During the deposition of the latter, greywacke sedimentation began in the SSW and later spread out over the entire Cantabrian Zone: first over the Leonides during Upper Namurian B to Lower Westfalian A, and later over the Central Asturian Coal Basin during Westfalian A to Westfalian D. Folding and thrusting of these sediments followed shortly after their deposition in directions more or less parallel to the former Asturian Geanticline.

Along normal faults, more or less parallel to the axis of the orogene, oblong basins formed in the Luna-Sil region, which were filled with a thick sequence of continental, coal bearing sediments of a Stephanian B and C age.

It is striking that restricted conditions prevailed after epeirogenic movements (Lower Formigoso, Upper San Pedro-Lower La Vid, Vegamián and Vegacervera Member of the Caliza de Montaña), whilst after orogenic movements continental sediments accumulated (Herrería, Prado).

The present relief has mainly been caused by Eocene, Miocene and Pliocene uplifts.

CHAPTER II

IGNEOUS ROCKS, ORES AND COAL

Igneous rocks

Several types of igneous rocks have been found in the Lower Palaeozoic rocks. In the basal conglomerate of the Herrería rhyolite pebbles occur, deriving from the Precambrian. The Láncara griotte south of Riolago contains green rocks composed of idiomorphic muscovite, idiomorphic quartz crystals and clino-zoisite, in a ground-mass of sericite and micro-crystalline chlorite (fig. 55). It could not be ascertained which kind of rock this is (Dr. P. Floor, pers. comm.). In the middle and upper part of the Oville and the lower part of the Barrios many dolerites (fig. 56) and pyroclastic rocks have been encountered.

In the San Pedro, especially in the Babia Baja unit, many palagonite particles are embedded. Thus some of the sandstones are tuffaceous (fig. 30) but real

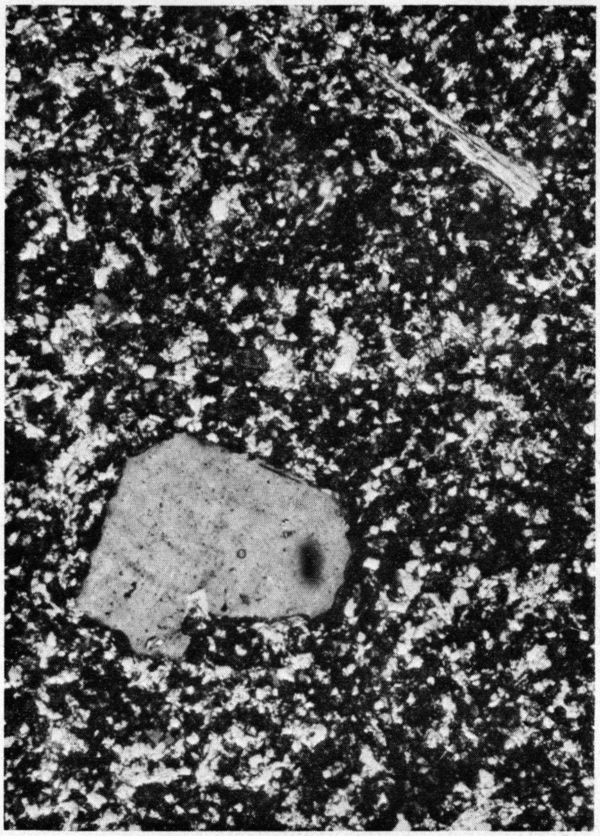

Fig. 55. Not determined rock from the Láncara Formation south of Riolago (40x).

Fig. 57. Muscovite granite from the Prado Formation near Sosas de Laciana (40x).

Fig. 56. Dolerite from the Oville Formation in Riolago (40x).

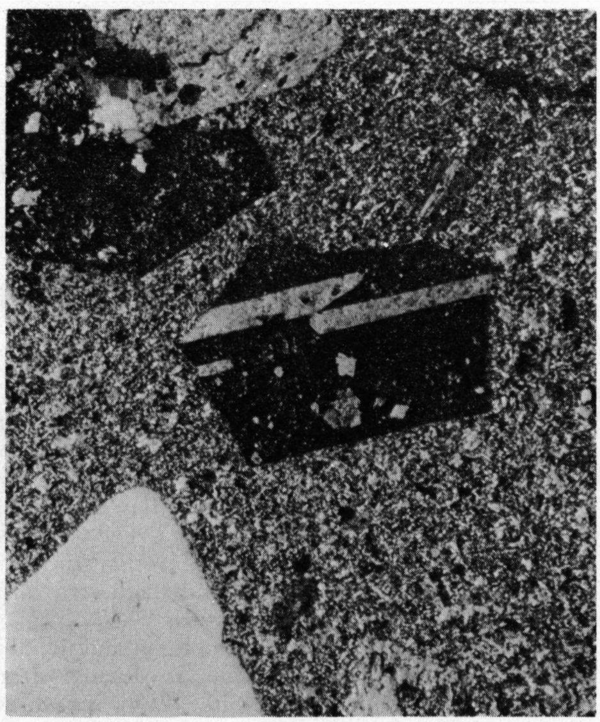

Fig. 58. Granite porphyry from the Prado Formation north of Sosas de Laciana (25x).

thick tuffites have also been encountered (section 26, p. 160).

The youngest igneous rocks intruded into the Stephanian sediments of the Villablino basin. Muscovite granites occur north of Sosas de Laciana (fig. 57); all other igneous rocks are granite porphyries (fig. 58), which often form extensive sills. These rocks contain quartz and alkali feldspath and some of them also plagioclase or pyrite phenocrysts in a ground-mass with plagioclase, sericite and quartz (Dr. P. Floor, pers. comm.). Rocks of the same composition intruded into the Mora NNW of Sosas de Laciana and along the Grajos fault, east of the Villablino basin.

Ores

The following ores were found in small, usually abandoned mines:

1. antimonite in the Mora near Sosas del Cumbral
2. malachite near Curueña
3. barite in the Láncara near the fault contact or at the limestone-griotte transition; east of Riolago, SE of Torre, south of Rabanal
4. galenite and pyromorphite SW and east of Riolago and between Riolago and Truébano along the Arroyo de Villasecino (still in operation)
5. sfalerite SE of Torre
6. azurite, malachite, siderite and barite east of Riolago
7. hematite in the San Pedro north of Caldas and in the Ermita north of Meroy and NNE of Lumajo
8. cinnabar at three places in the Caliza de Montaña in the eastern part of the Babia Baja unit
9. kaolinite in the Riello basin
10. laterite on the Carboniferous limestones east of Riolago and south of Robledo de Babia (fig. 59). These soils consist of very white levels, composed entirely of muscovite, and of pink or red coloured, more lateritic material. This originated on an old terrace possibly of Pliocene age (p. 188). The lateritic material has the following composition: SiO_2 = 59.69%, Al_2O_3 = 24.38%, Fe_2O_3 = 2.70%, FeO = 0.15%, CaO = 1.25%, TiO_2 = 1.34%, H_2O + CO_2 = 11.42%, remainder = 0.20%.

Coal

The coal in the Babia Baja unit occurs only in the upper part of the San Emiliano Formation. Gómez de Llarena & Rodríguez Arango (1948) estimated a total reserve of 24,000,000 tons for the whole Babia Baja unit. The same authors (1946) analyzed coal from a mine south of Truébano, which proved to contain 36–68% of carbon.

Coal has also been mined north of Villayuste and west of La Urz in the La Magdalena basin, and in the Villabandín basin. Coal is only being mined on a large scale in the Villablino basin. De la Concha & Jorissen (1961) found 15 important coal beds with a total thickness of 7.5 metres. They estimated a total production of 155,000 tons of anthracite and 783,000 tons of coal, and a total reserve of 10 million tons of anthracite and 269 million tons of coal.

CHAPTER III

GEOMORPHOLOGY

The Cantabrian Mountains are the result of Tertiary uplifts during the Eocene, late Miocene and Pliocene. They have a medium relief caused by fluvial, glacial and karst agents.

River system

There are four drainage areas in the region mapped: those of the Luna, the Sil, the Omañas and the Huerna rivers (fig. 59). The Huerna is a river with a very large gradient, flowing directly to the Bay of Biscay, and is capturing an affluent of the Luna. The Omañas is cutting downwards into an old erosion surface, forming steep V-shaped valleys, and discharges into the Luna, south of the region mapped.

The most interesting rivers are the Luna and the Sil. The headwater area of the Luna is situated near Quintanilla. In its upper course the Luna meanders through a broad W–E trending valley (fig. 61); its gradient between Quintanilla and Villasecino is 0.63%. In some locations the affluents of the Luna deposited extensive alluvial fauns in the main valley; locally these brooks lose all their water there, which disappears in the coarse material of the Luna valley. This occurs near Piedrafita, Peñalba, Riolago and Torre. NE of Villafeliz, in the valley of the Arroyo de Río del Puerto, beautiful river terraces have been preserved (fig. 60); just SW of Huergas, too, remains of terraces were found in the continuation of the valley between Mena and Huergas, which is probably the former rivercourse (fig. 59). The Sil is rapidly cutting downwards, having a gradient of 3.1% between Quejo and Rioscuro. All affluents of the Sil in this region were formerly affluents of the Luna, which is proved by the barbed junctions near Puente de las Palomas (fig. 59), and especially by the presence of an old landscape in the continuation of the old landscape near Piedrafita, cut by the young incisions of the Sil and its affluents (fig. 61 and 62). It is thus very clear

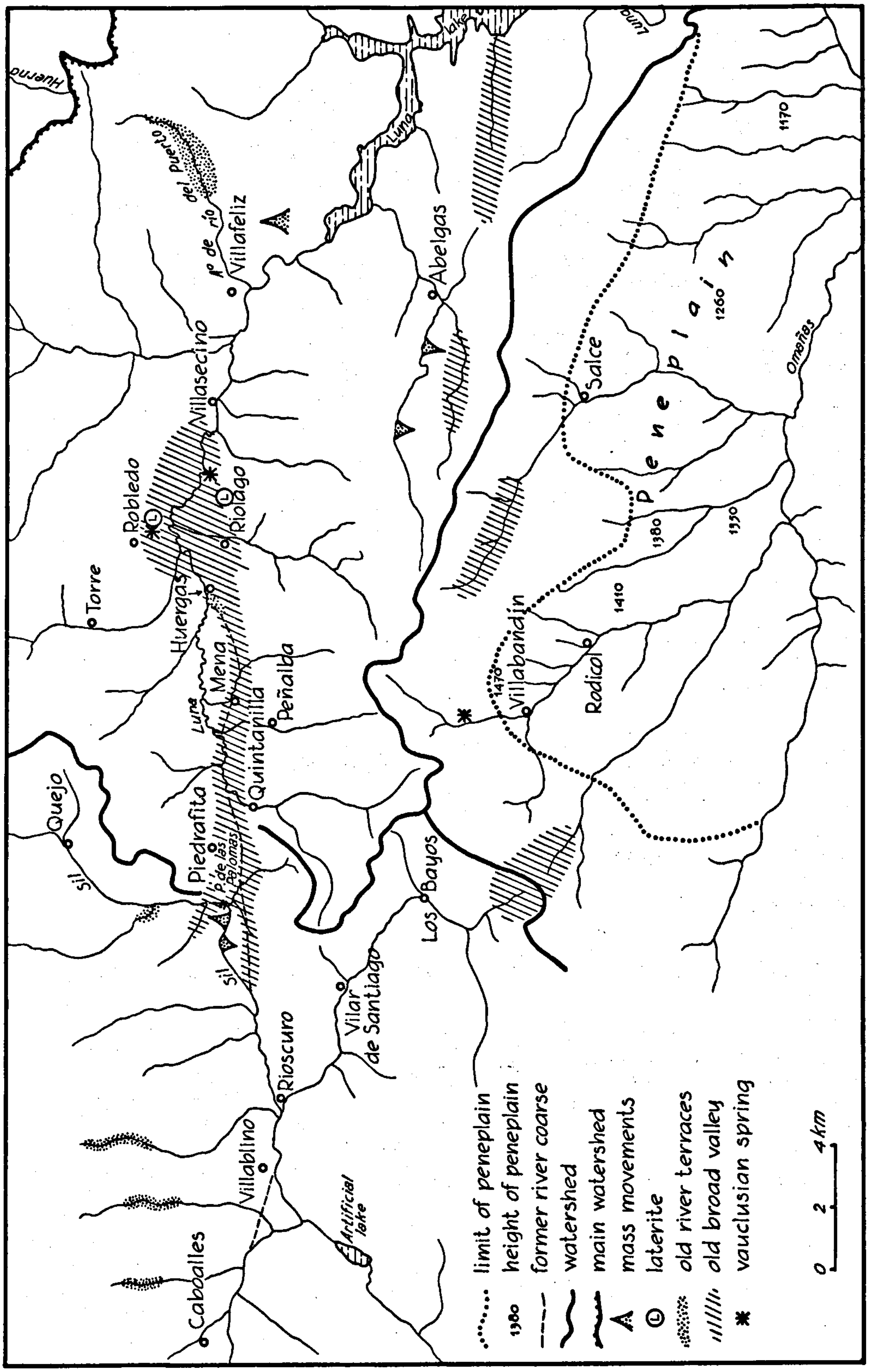

Fig. 59. Geomorphological phenomena in the Luna-Sil region.

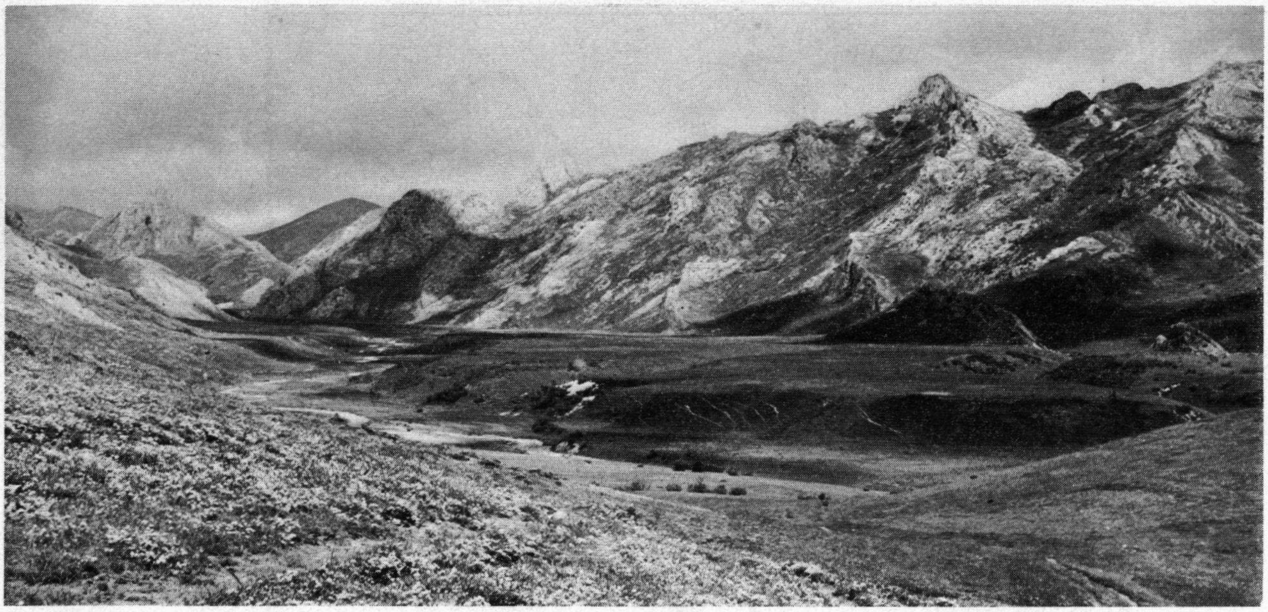

Fig. 60. River terraces in the valley of Arroyo del Río del Puerto (NE of Villafeliz), looking eastwards.

that this river is capturing the meandering Luna. In its upper course the Sil locally eroded very steep canyons in the old broad and flat valleys; that of Puente de las Palomas (fig. 62) is the deepest (82 metres). Near Los Bayos and Vilar de Santiago tributaries of the Sil are capturing tributaries of the Luna. Possibly the Luna once flowed eastwards from Caboalles, while the Sil ran to the south and had its headwater region south of Villablino. Due to late-Miocene sinking in the Bierzo basin, 70 km south of Villablino (Sluiter, 1964), it developed a larger gradient and captured the headwaters of the Luna near Villablino and later on the part between Villablino and Puente de las Palomas as well (Vidal Box, 1943).

Mass movements

Landslides here are due to undercutting of a rock mass by a river. Due to the underlying soft La Vid shales being cut away by the river a large mass of Santa Lucía limestone slumped down in two locations west of Abelgas. Along the Sil, just west of Puente de las Palomas, thick layers of Stephanian sandstones have been undercut by the river, and have slumped down along the bedding-plane.

Further more, mechanisms such as creep, rockslides, debris falls, etc. have also been active.

Karst

Especially in the Caliza de Montaña, but also at many places in the Santa Lucía and Láncara, karstification was encountered; sandpipes, dolines and uvalas are all present (fig. 63). Vauclusian springs are found east of Riolago, south of Robledo and north of Villabandín (fig. 59).

Lateritization

In the broad Luna valley, near Robledo and Riolago, about 30–40 metres higher than the present river course, an old terrace, which, however, lost the greater part of its terrace cover, bears some lateritic material. White material, consisting completely of muscovite flakes, and red laterite show that tropical weathering must have taken place. The height of the terraces suggests that these deposits were possibly formed during the Pliocene (Pannekoek, 1968).

Glacial Forms

At least during the Riss and Würm glaciations, large glaciers filled the valleys of the Cantabrian Mountains; the lowest glacial deposits have been found at 550 metres (Würm) and 950 metres (Riss); the snow limit lay at a level of 1350 metres (Lotze, 1962; Nussbaum & Gigax, 1953).

Many moraines have been preserved in the Babian mountains (see geological map). Those near Piedrafita (fig. 61a) and especially those between Torre and Huergas, both at a level of about 1250 metres, are impressive. The youngest moraines, occompanied by cirque lakes in several locations, are found NNE of Torre, NW of Salce, south of Mena, and in several other places at a level of 1650–1700 metres. Deposits which might be fluvio-glacial fans (Torre and Piedrafita) and glacial valleys are rather common; *roches moutonnées* have been preserved in the village of Torre and between Quintanilla and Piedrafita.

Relief

The relief is mainly old; in Oviedo and also in parts of

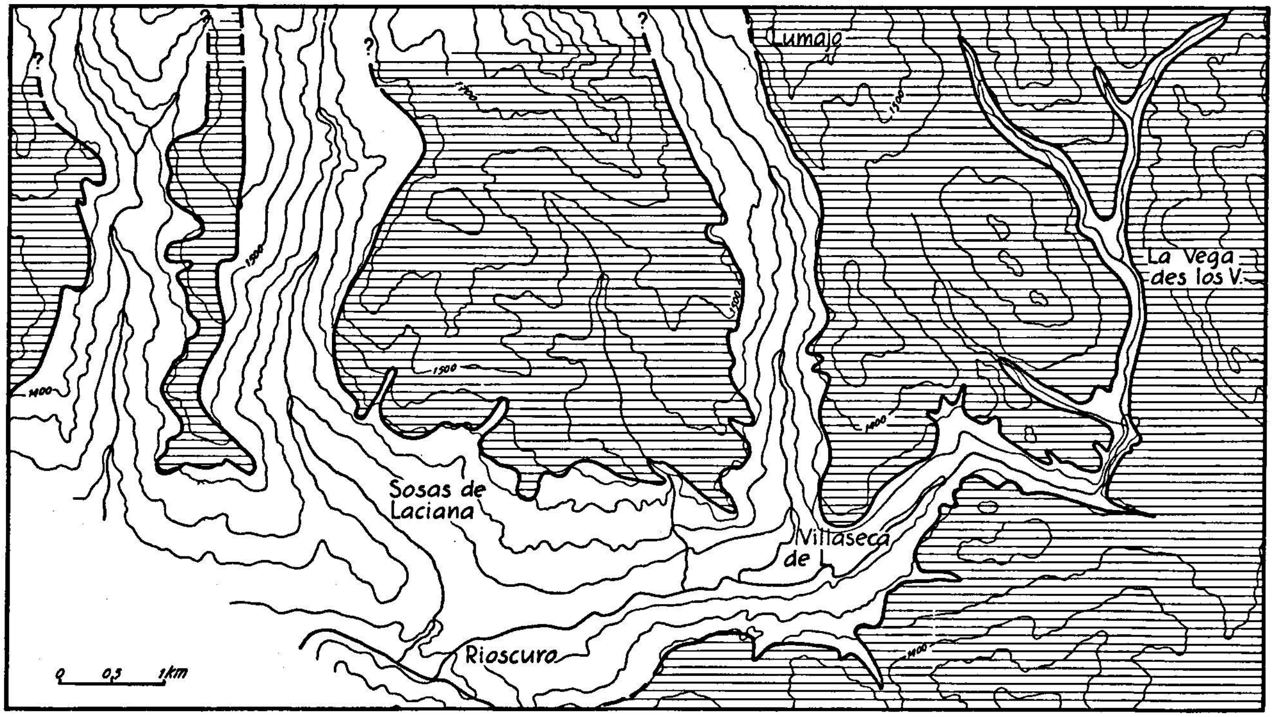

Fig. 61a. Flat old valley near Piedrafita; a moraine in foreground (looking westwards). b. Young valleys of the Sil and its tributaries in the old landscape of the former Luna drainage area.

Laciana it is mature. In the southern part of the region, on the Mora schists, lies an extensive erosion surface, either a pediplain, or a peneplain, incised by V-shaped valleys (fig. 64). Its altitude varies from 1470 metres near Villabandín to 1200 metres in the east (fig. 59). Analogously to comparable erosion surfaces more southwards, this one may be of late Miocene age (p. 183). An infuence of the geology on the relief can be ascertained: the resistant Herrería, Barrios, San Pedro and Santa Lucía form the highest parts, but the courses of the most important rivers and brooks are generally not influenced by geology.

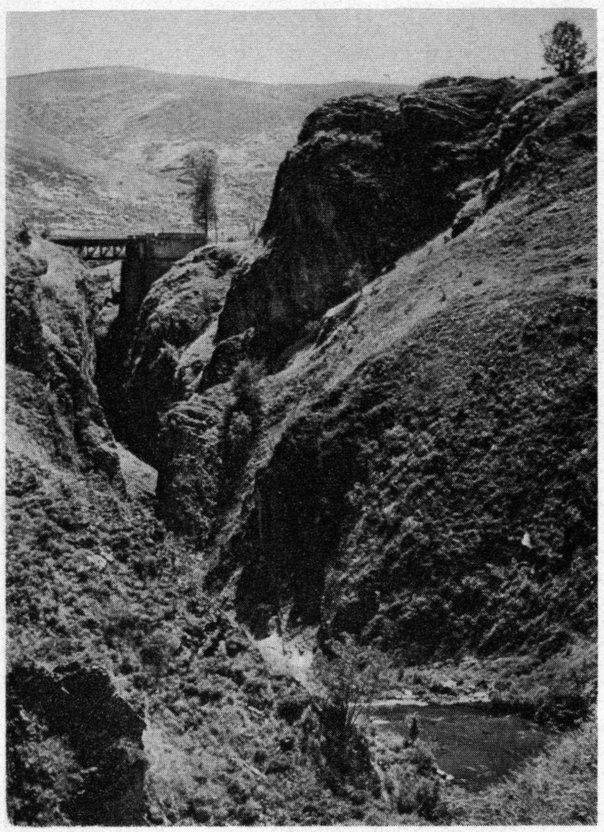

Fig. 62. Canyon of the Río Sil (82 metres deep); in the background the Puente de las Palomas and the Villablino Basin (looking southwards).

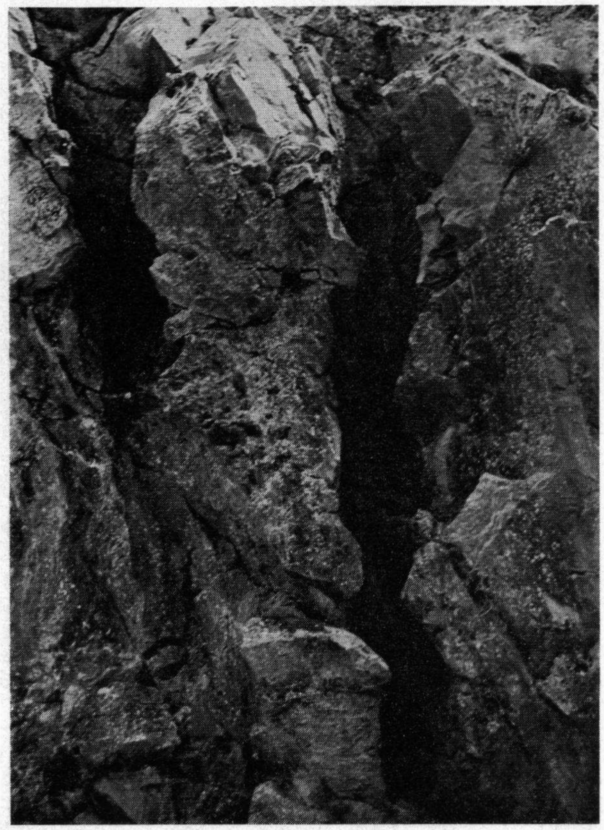

Fig. 63. Sandpipes in the Caliza de Montaña Formation, along the road east of Villasecino.

Fig. 64. Peneplain or pediplain on the Precambrian near Sosas del Cumbral (looking southwards).

CHAPTER IV

STRUCTURES

INTRODUCTION

The stratigraphic succession of the Precambrian to Upper Carboniferous rocks in the Luna-Sil region is typified by many contrasts in lithology. In order to obtain a better insight into the tectonic problems, the structural properties of the different rock units will be discussed first and with these rock properties in mind, we shall attempt to analyse the various kinds of deformation.

Several levels of incompetent rocks cause disharmonic structures which are characteristic of the region. The deformation of these incompetent beds is clearly governed by the structures in the competent ones.

Generally the incompetent rocks show a tendency to direct faults parallel to their bedding-plane, enabling the overlying competent rocks to move over those beneath for sometimes great distances. The more competent strata, on the other hand, are mostly intersected diagonally by faults. In accordance with the structural properties of the beds, which were folded during the Sudetic and Asturian phases, the stratigraphic succession may be subdivided into the following 'structural rock units'.

San Emiliano	Shales cause slight intraformational disharmonies; In limestones parasitic and cascade folds.
Caliza de Montaña	Platy limestones with numerous parasitic folds.
Alba Ermita Nocedo Portilla	Competent succession; concentric folding of the Upper Palaeozoic beds in synclines usually ends at the level of the Alba.
Huergas	Rather incompetent; at some places allowing folds to develop disharmonically; locally small-scale folding.
Santa Lucía Crinoid Limestone Member of La Vid	Competent succession; box-folds may develop in the lower part.
Calcareous Shale Member of La Vid	Highly mobile calcareous shales; cause important disharmony between Lower and Upper Palaeozoic rocks; crumpling, bedding-plane shearing, small-scale folding.
Limestone Member of La Vid San Pedro Upper Formigoso	Competent succession; at many places large-scale cascade folding.
Lower Formigoso	Mobile black shales; at some places 'intruding' into fissures of the top of the Barrios; small-scale folding; locally causing disharmony between San Pedro and Barrios.
Barrios Upper Oville	Very competent Barrios quartzites control the shape and width of most folds; many cross-faults; locally formation of imbricate structure behind the main thrust fault.
Lower Oville Láncara	Thin base of Láncara (black shales and dolomites) was most effective layer for bedding-plane shearing; all large thrust faults located at this level; Láncara and Lower Oville parallel to the main thrust faults.
Herrería Mora	Fault tectonics

Table B. Structural rock units.

This vertical accumulation of disharmonic levels with different tectonic styles is called a tectonic 'Stockwerk' (Cloos, 1964; Metz, 1967, p. 137). These properties and many bedding-plane measurements provided the basic data required for the construction of the structural sections. As far as possible, use was made of the arc-construction method, most of the folds being concentric.

The region has been subdivided into the following structural zones, shown on the structural index map (fig. 65):

Central Asturian Coal Basin } Cantabrian Zone
Leonides }
West-Asturian-Leonese Zone
Stephanian basins
Miocene deposits

The Leonides have been subdivided into:

Luna unit
  Abelgas syncline
  Aralla zone
Babia Alta unit
  Torre zone
  Saliencia zone
  Meroy zone
  Corralines zone
  Muxivén zone
Babia Baja unit; the NE part is called the Ubiña zone

The portion of the West-Asturian-Leonese Zone south of the Ocedo fault is called the Salce unit, forming a part of the Narcea anticlinorium, mainly consisting of rocks of Precambrian age (Julivert & Pello, 1967; Matte, 1968b).

First the most important structures in these units well be described geometrically in the order of sequence given above, and later on, in the last chapter, the dynamics and kinematics will be analyzed.

LUNA UNIT

In the region mapped the Luna unit constitutes the southern part of the Leonides. The Leonides have here a roughly WNW strike and are situated between the Ocedo fault and the thrust faults in the Ubiña zone. They are characterized by the presence of thrust faults at the base of the Láncara Formation. The structures with a WNW trend are interrupted by a zone with a WSW trend (fig. 65). The Luna unit, subdivided into the Abelgas syncline and the Aralla zone, is the western continuation of the Bernesga unit (de Sitter, 1962b). As shown in fig. 65, it is bounded, in the region concerned, by the Rozo thrust fault in the north, the Ocedo fault in the south and the Grajos fault in the NW.

Abelgas syncline

The Abelgas syncline trends mainly WNW and is bounded in the north by the Abelgas thrust fault and in the south by the Ocedo fault, named after the Alto de Ocedo, NNW of Villabandín. SE of Abelgas, and again near the nose of the syncline, this WNW trend is deflected into a WSW dirction. The Abelgas syncline forms the western continuation of the Alba syncline, the Mirantes anticline, and the Pedroso syncline (de Sitter, 1962b); the Sabero-Gordón line ends in the eastern part of the syncline. Two synclines and an anticline can be recognized in the Abelgas syncline, which thus properly constitutes a synclinorium. This synclinorium is asymmetric, having an overturned northern limb.

Sediment thicknesses mostly grow thinner towards the south.

A striking feature is the discontinuity horizon in the La Vid calcareous shales: important structures below these shales do not penetrate into the strata above them.

The rocks in the Abelgas syncline form a tectonic 'Stockwerk': three competent levels, each with its characteristic structures (Láncara-Barrios, Upper Formigoso-La Vid limestones, and the Santa Lucía and younger formations), are separated by the mobile Formigoso black shales and La Vid calcareous shales. The silty shales and sandstones of the Huergas also separate structures above and below them, but less distinctly than the former two.

Structures below the upper part of the Formigoso. – The Abelgas syncline is thrusted upon the Aralla zone along a thrust fault in the black shales at the base of the Láncara; where exposed, the fault contact is extremely thin. Near the Cáscaros the thrust fault leaves this level and starts cutting upwards through younger strata up to the level of the San Pedro, whereas the beds in front of the thrust are the same as those more westerly; thus the amount of thrusting diminished considerably (sections 1 and 2). Due to continued folding the thrust fault stands steep or even overturned at many places.

In analogy with the Cospedal thrusts and their associated folds and faults in the Torre zone, which very much resemble the structures in the NE part of the Abelgas syncline, it seems probable that the Abelgas thrust fault remains at the base of the Láncara and merges into the Rozo thrust fault at depth (sections 1 and 2); the Abelgas thrust, just as the Cospedal thrusts (fig. 65), does not pass into beds younger than the La Vid calcareous shales; this shows clearly the disharmonic character of these shales, and their property to direct faults parallel to the bedding-plane.

Many structures have been found which show the great difference in competency between the Formigoso black shales and the Barrios quartzites. The highly competent Barrios quartzites are dissected by many small cross-faults. In the northern flank of the syncline this faulting is restricted to the Barrios and the top of the Oville, as most faults get lost downwards in the incompetent Oville siltstones and upwards in the even more incompetent Formigoso black shales; in the

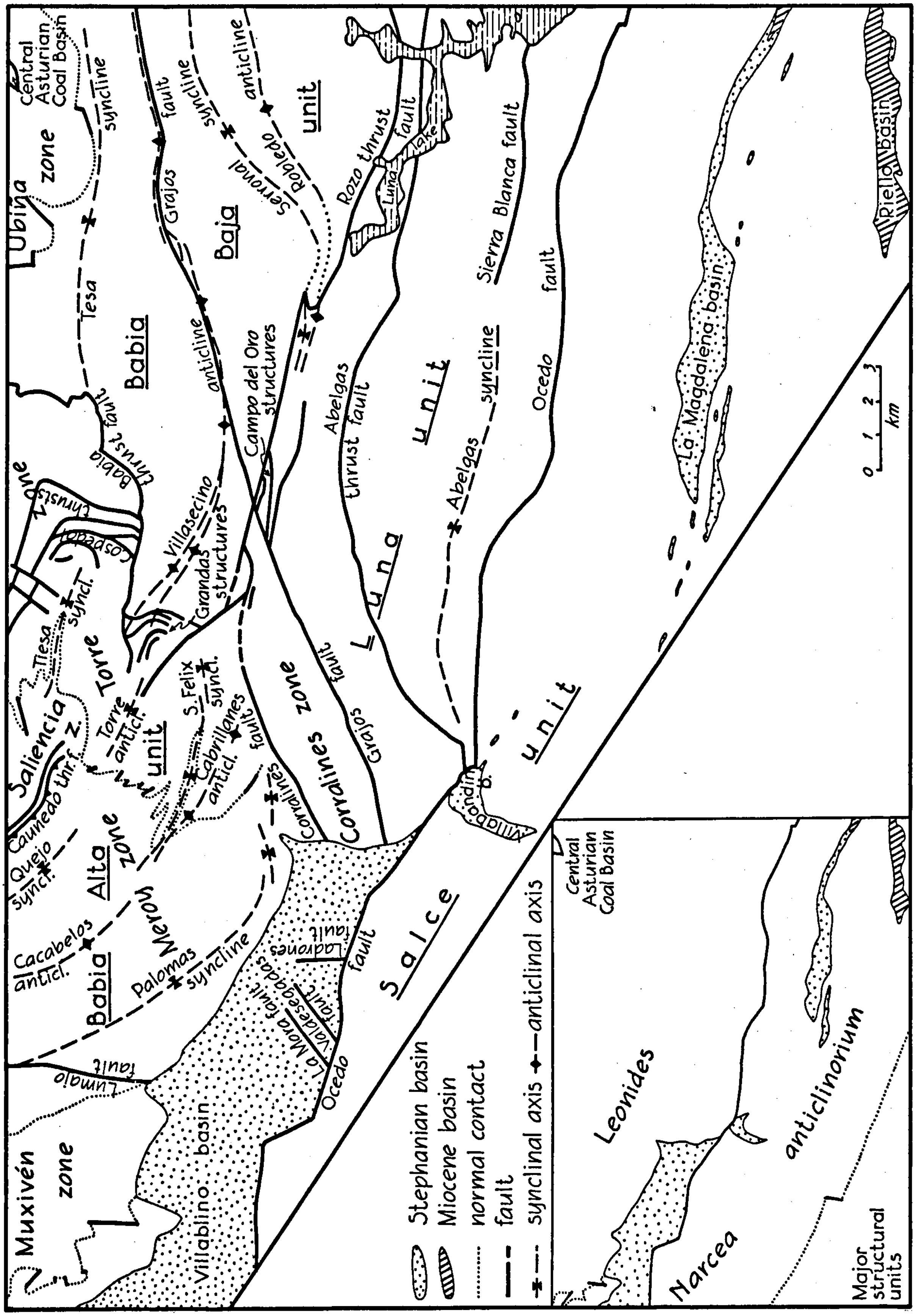

Fig. 65. Structural index map.

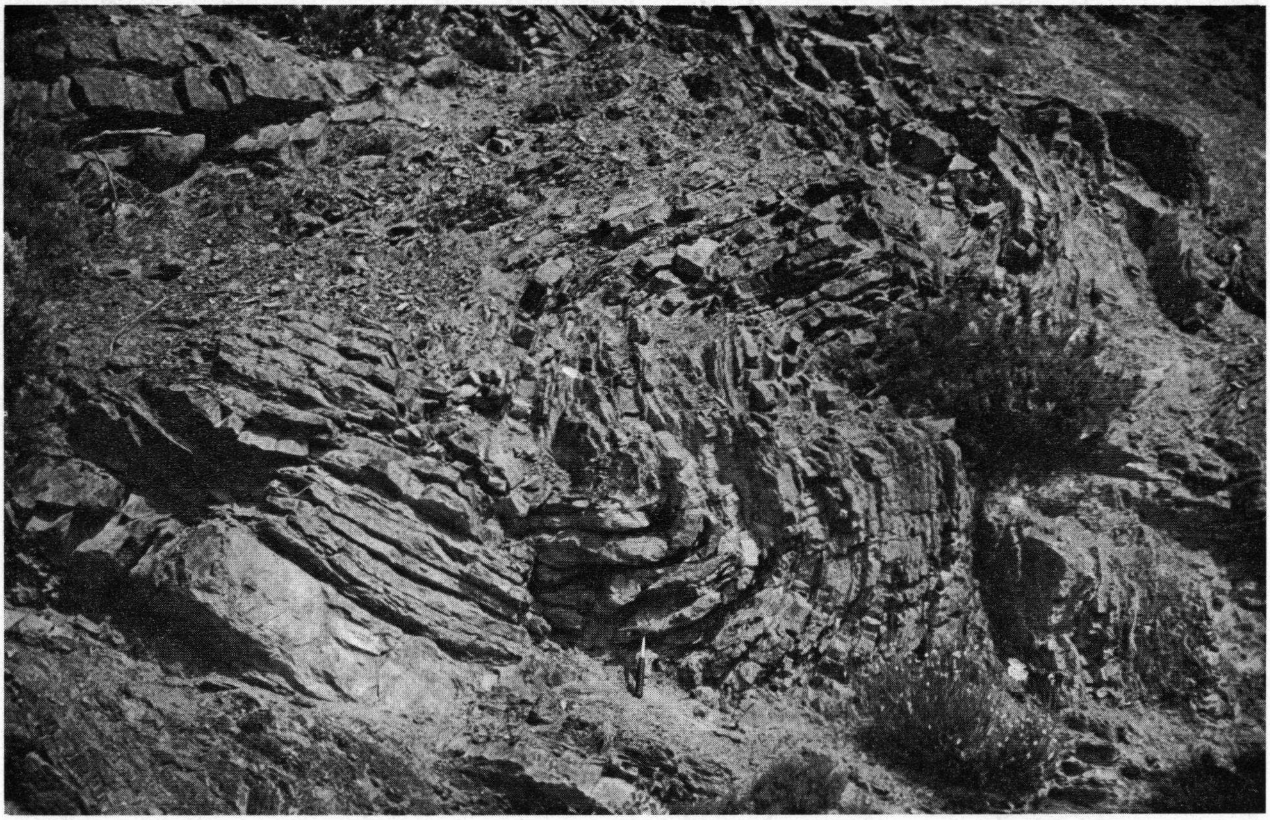

Fig. 66. Minor folds in the Formigoso Formation north of Abelgas (looking westwards).

southern flank, however, several faults continue into the San Pedro. Almost all these small faults are perisynclinal wrench-faults which originated from the stretching of the developing syncline; these faults have a tendency to lengthen the syncline (de Sitter, 1964, p. 193).

The mobility of the Formigoso black shales is demonstrated north of the Churros where the mobile black shales were pressed into a fissure in the very competent Barrios quartzites. Minor folds which are encountered very often in the Formigoso shales, are very well exposed north of Abelgas (fig. 66) and demonstrate again the low competency of the Formigoso shales. Another, less frequently encountered, phenemenon in the Formigoso black shales is a slightly developed fracture cleavage, in general more or less parallel to the bedding and thus difficult to detect.

The structures near the Cáscaros show that the Abelgas thrust fault and probably all other thrust faults in the mapped region, originated from breaking through of an anticline (section 3). The northern limb of this anticline lies very flat and forms an impressive dip slope, with local remnants of Formigoso at its foot; this explains the great apparent thickness on the map. The concave nature of the thrust fault is explained by the passing downwards of the diagonal fault in the Barrios into the zone of bedding-plane shearing at the base of the Láncara. Small extra thrusts were formed which pass downwards into the initial bedding-plane fault where the latter merges into the diagonal shear fault in the Barrios. This took place just there, because the greatest frictional resistance during the movements along the fault plane existed at that location (section 4). The strata in this imbricate structure stand 50° overturned; this can be explained partly by the concave nature of the main fault which caused a rotation of the strata behind it during the further thrusting and partly by a subsequent further folding. An anticlinal bend in the Barrios just south of Santa Eulalia indicates that one of the faults there did not reach the uppermost layers of this formation (section 3). A variably plunging anticlinal axis gives rise to a separate culmination with Barrios in the core SE of Láncara de Luna; this anticline has the same asymmetry at the level of the Barrios as it has more westerly (sections 2 and 3).

East of the Villabandín basin lies an intensively disturbed, badly exposed zone. In the western part of this zone the Láncara is repeated by a small upthrust (section 9). This zone is bounded in the north by a W-E trending, probably vertical fault which cuts off the Abelgas thrust fault and thus originated later than the thrusting.

Structures between the Formigoso black shales and the La Vid calcareous shales. – Behind the Abelgas thrust fault, near the Luna lake, beds are no longer parallel to this fault, but folds were formed, mainly at the level of the Upper Formigoso-San Pedro-Lower La Vid, of the same shape as in the Torre zone (compare sections 1 and 7). Note the disharmonic folding between the Barrios and San Pedro by the incompetent Formigoso shales (section 1).

Gravity folds are very common in the La Vid Limestone Member and the San Pedro. This can be explained by the presence of the mobile La Vid calcareous shales on one side and the Formigoso black shales on the other side of this sequence, which did not provide sufficient support when the unconsolidated beds were set steep, so that sliding could start. Large cascade folds have been encountered in the La Vid limestones and San Pedro near the nose of the Abelgas syncline. Dip slopes developed because the strata there were brought into a very overturned position; this makes the structures look more complicated on the map than they are in reality (fig. 67). The most western part (section 7) shows a broken cascade fold; just eastwards of this section two small horizontal faults are found in the overturned limb (fig. 68); they probably originated during the development of the cascade fold (fig. 69). All these structures are cut off in the east by a large cross-fault. On the other side of this fault an unbroken

Fig. 67. Cascade fold west of the Penouta (looking eastwards), after photograph.

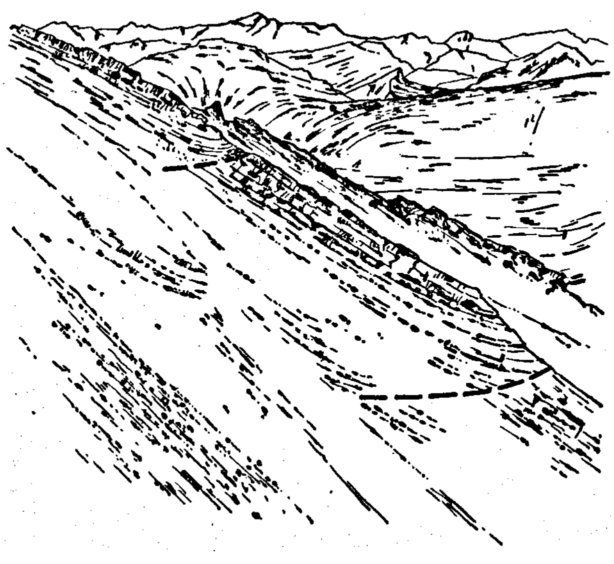

Fig. 68. Small thrust faults in a cascade fold east of the Cañada (looking westwards), after photograph.

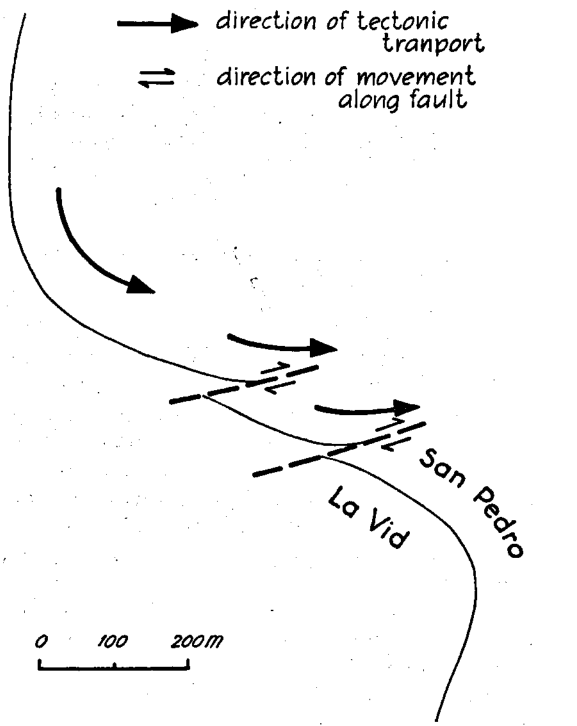

Fig. 69. Explanation of the origin of the thrust faults of fig. 68.

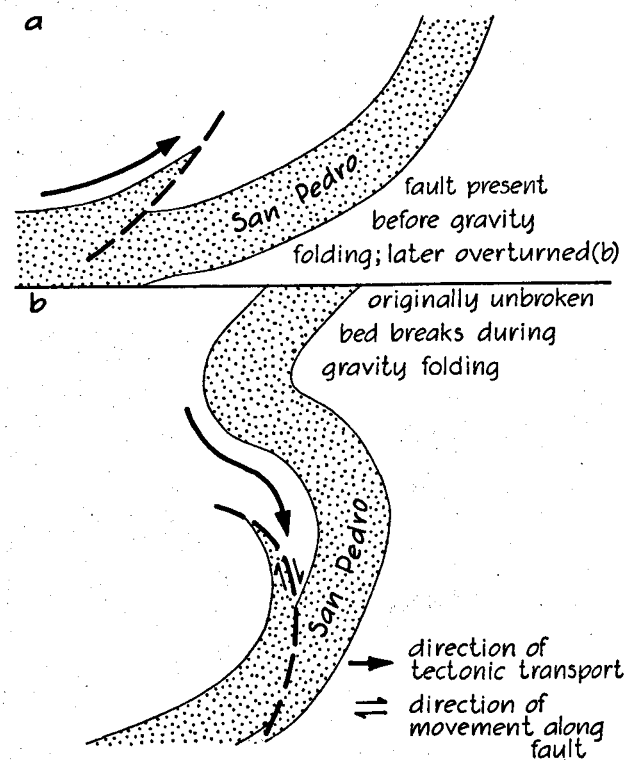

Fig. 70. Two possible explanations of the origin of the structures SE of the Penouta.

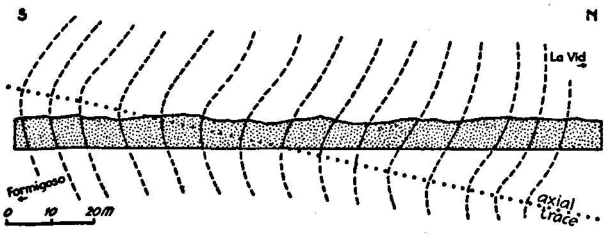

Fig. 71. Sketch of a cascade fold in the San Pedro Formation between Los Barrios and Mallo.

with one large and two very small upthrusts south of this fold, which upthrusts may have originated before the thrusting and were overturned during the development of the cascade fold. Another possibility is that they originated during the development of the cascade fold (fig. 70). Dip measurement east of this fold indicated more (smaller) folds, but as they can nowhere be seen in profile, it is not sure whether they are cascade folds or not.

Another cascade fold in the La Vid limestones and San Pedro was found on the southern flank of the Abelgas syncline between Los Barrios and Mallo (fig. 71). Smaller folds of the same type occur there in the middle part of the Formigoso. Cascade folds have also been encountered on both sides of the Abelgas thrust fault (section 1).

Structures above the La Vid calcareous shales. – As stated above, the complicated eastern part of the core of the Abelgas syncline forms the continuation of the Pedroso syncline, Mirantes anticline, and Alba syncline (fig. 72). The syncline with Ermita in the core can be connected with the Alba syncline (fig. 73). A longitudinal fault, of which the fault breccia is exposed, locally brings Portilla into contact with Santa Lucía.

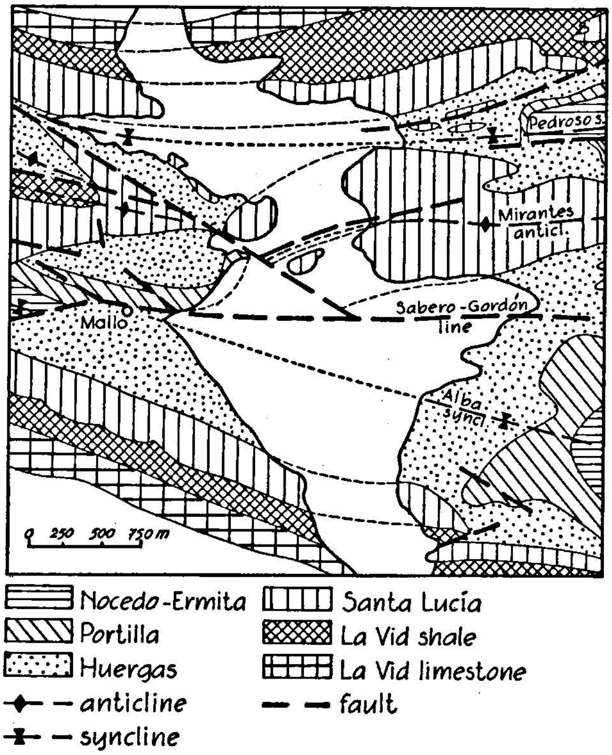

Fig. 72. Presumable relationships between the structures on the eastern and western side of the Luna lake.

The two broken anticlines with La Vid shales in the core (fig. 74) form the prolongation of the Mirantes anticline. The tectonic position of the isolated outcrops of Santa Lucía west and east of these anticlines could not always be ascertained (fig. 72). A longitudinal fault in the core of the continuation of the Pedroso

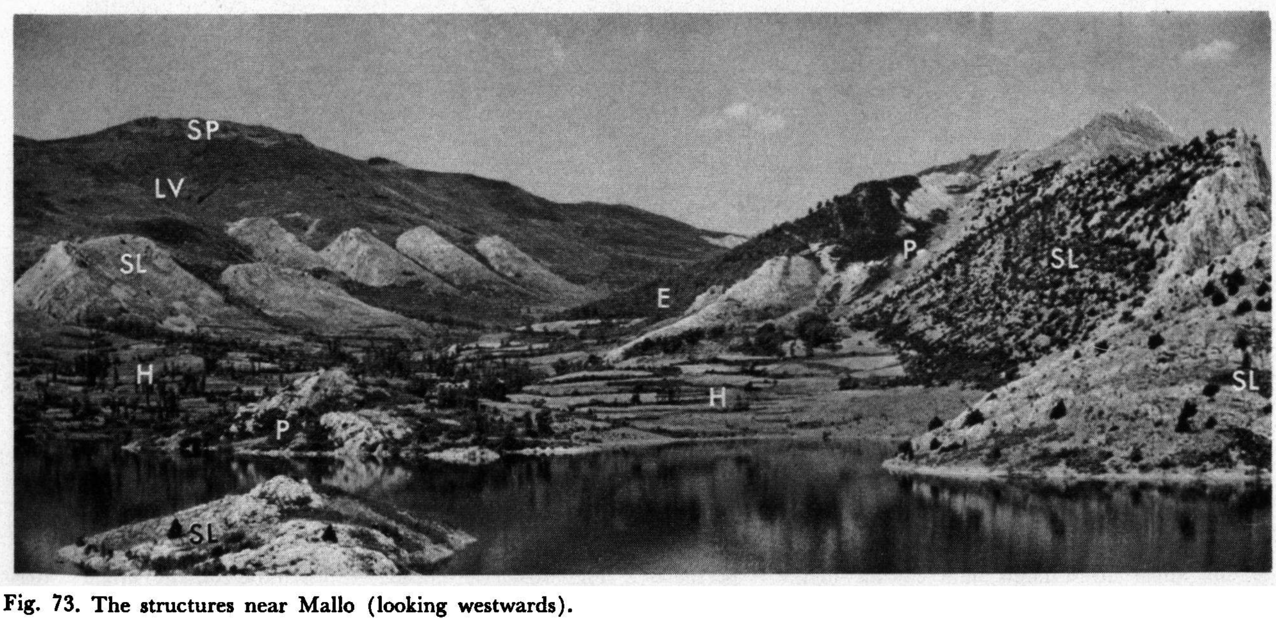

Fig. 73. The structures near Mallo (looking westwards).

syncline brings Santa Lucía of the southern flank into contact with the same level of the northern flank. Due to the incompetent character and the poor exposure of the Huergas shales it could not be ascertained whether the three longitudinal faults north of the Portilla continue to the west.

The Sabero-Gordón line (Rupke, 1965) more eastwards is a most important fundamental line which manifests itself on the map as faults. This zone which has been described in the Stephanian basins of the Cea region and south of the Esla structures (Helmig, 1965; Rupke, 1965), continues north of the Alba syncline (Evers, 1967; van Staalduinen, 1969) and towards the west can be traced into the core of the Abelgas syncline as far as Abelgas. In the areas east of the Luna lake this zone was a facies division during the Upper Devonian and Lower Carboniferous. This suggests vertical movements and a vertical position of the faults (Evers, 1967; van Staalduinen, 1969). The present author is of the same opinion, although in the Abelgas syncline insufficient evidence can be found to support these views.

Overturning accompanied by cascade folding, locally even to an upside-down position, occurred in the most western part of the area where Portilla, Nocedo and Ermita crop out (fig. 75). This folding must have occurred after the Huergas had been thrusted upon the younger formations, because the beds at both sides of the thrust (fig. 76) participated in the cascade folding (section 3).

Although locally badly exposed, the area south of the Sierra Blanca fault shows a synclinal form, cut off in the west by a cross-fault. Two strike-faults on the southern flank of the syncline repeat the Santa Lucía-La Vid contact twice; they probably represent upthrusts along which movement took place towards the south.

Fig. 75. Overturning to an upside-down position of the Portilla Formation between Mallo and Abelgas (looking eastwards).

Fig. 76. Overturned core of a syncline in the Huergas Formation south of the structure of fig. 75 (looking northeastwards).

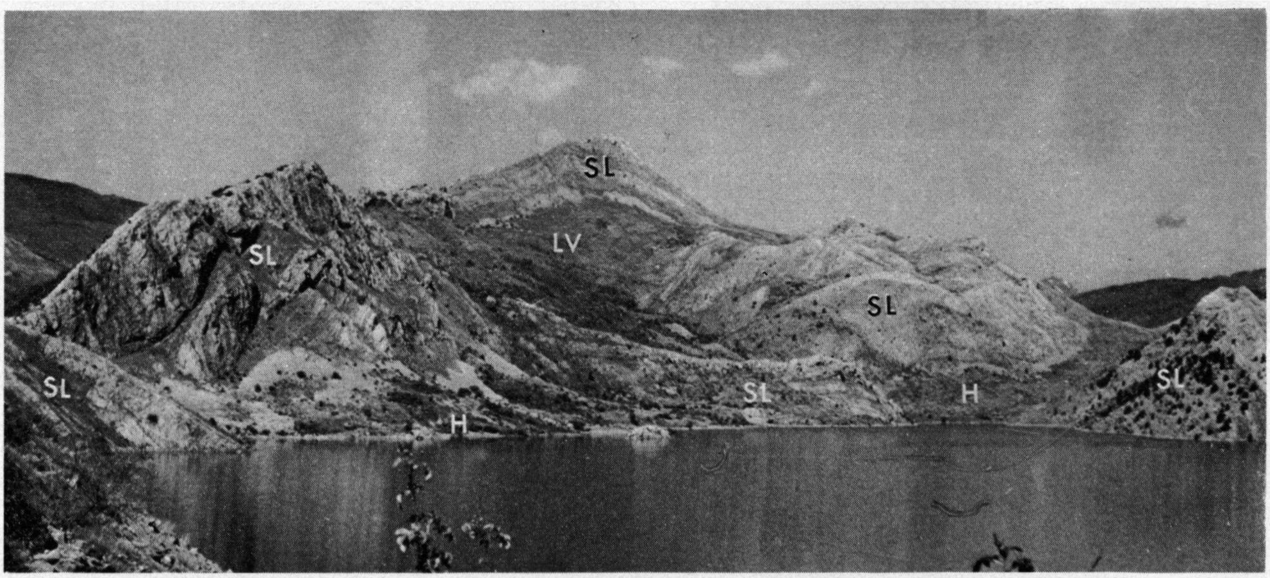

Fig. 74. The structures north of Mallo (looking westwards).

Similarly to the thrusts north of these, which are directed to the north, they are of local importance, but have a direct relation to the folding in the core of the Abelgas syncline. The thrusts do not continue deeper than the La Vid calcareous shales.

Other structures also demonstrate the already mentioned very mobile character of the La Vid calcareous shales. The most striking is the great disharmony between the beds above and below these shales, especially in the western part of the syncline, where 3 minor synclines and 2 anticlines are found (fig. 77) in the younger rocks, whereas the older rocks were folded to only one broken syncline. Moreover, only one fault and no fold can be traced through the La Vid shales, while the construction of the anticlines in the Santa Lucía clearly indicated a detachment-plane in these shales (sections 2 and 5).

Locally, especially in the nose of the syncline, a slight fracture cleavage was found in the calcareous shales and also in the black shales in the upper part of the La Vid limestones.

In the western part of the core of the Abelgas syncline (fig. 77) three minor synclines and two anticlines are found, just as in the eastern part of the syncline. Both structures seem related, but, however, are separated by an area with only uniform, poorly exposed Huergas. The southernmost syncline is bounded in the north by an upthrust, of which the dip varies from 30°–35° to the SE in the west to vertical in the east. The nose of this syncline is repeated several times by normal cross-faults, due to the stretching of the developing syncline (p. 194). The overturned nose of the central syncline (fig. 78) is repeated by a similar normal cross-fault.

The northern flank of the northernmost syncline is cut by four diagonal faults in the Santa Lucía lime-

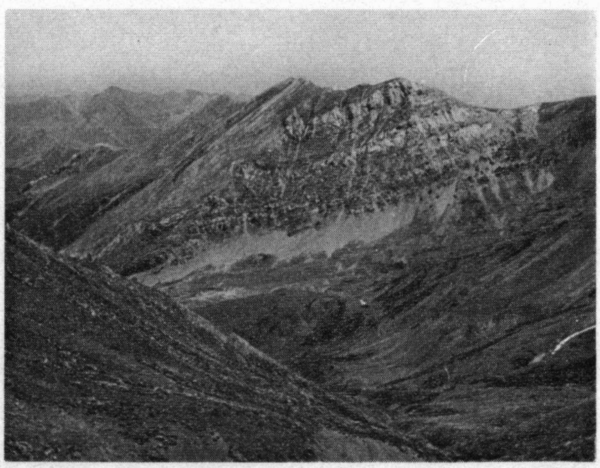

Fig. 78. Overturned nose of the Abelgas syncline (Santa Lucía Formation), looking eastwards.

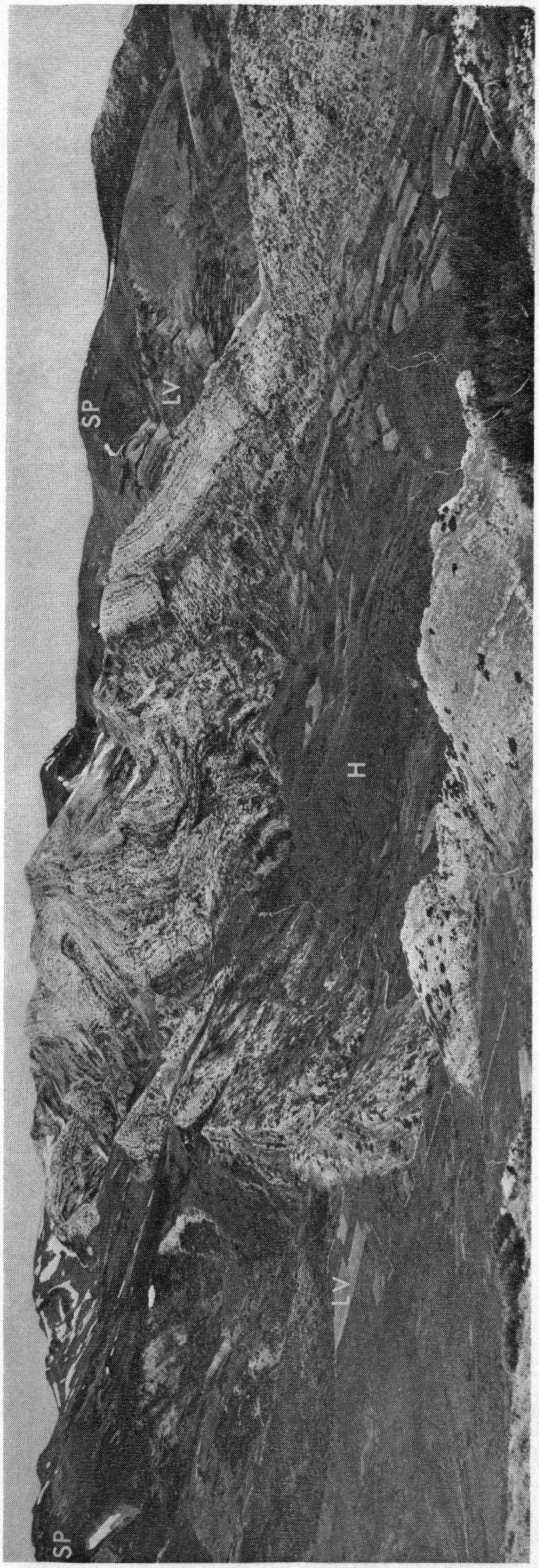

Fig. 77. Western part of the Abelgas syncline (looking westwards).

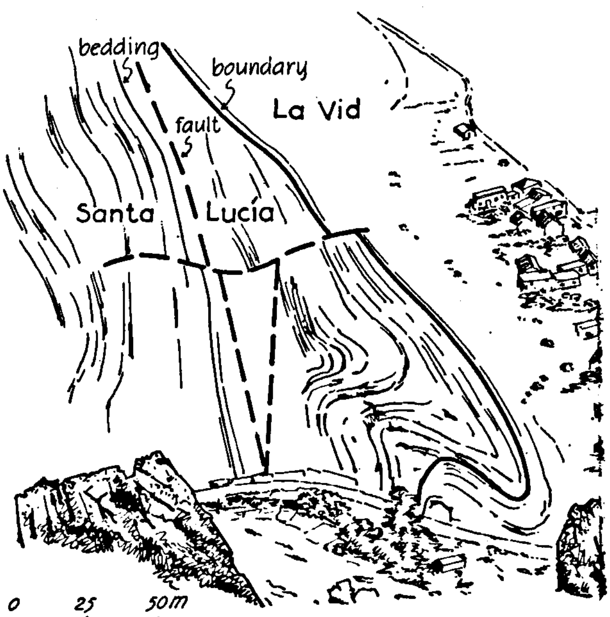

Fig. 79. Cascade fold at the end of a diagonal fault at the La Vid-Santa Lucía contact, south of Abelgas (looking eastwards), after photograph.

stone, which show identical structures where the faults pass into the La Vid shales. Such a structure is well exposed just south of Abelgas where a cascade fold developed at the end of one of these faults (fig. 79). At the fault more to the west a segment of Santa Lucía with La Vid shales on both sides slid into a 35° dip from its original vertical position (fig. 80) where the faults pass from Santa Lucía into La Vid. The faults successively further to the west show similar features.

Aralla zone

The Aralla zone was originally a syncline which was thrusted upon the Babia Baja unit along the Rozo thrust fault. It has been overthrusted by the Abelgas syncline, so that in the eastern part only the northern limb crops out. The Aralla zone can be traced from the Río Bernesga and continues westwards through the Leonides until it ends at the Ocedo fault, which

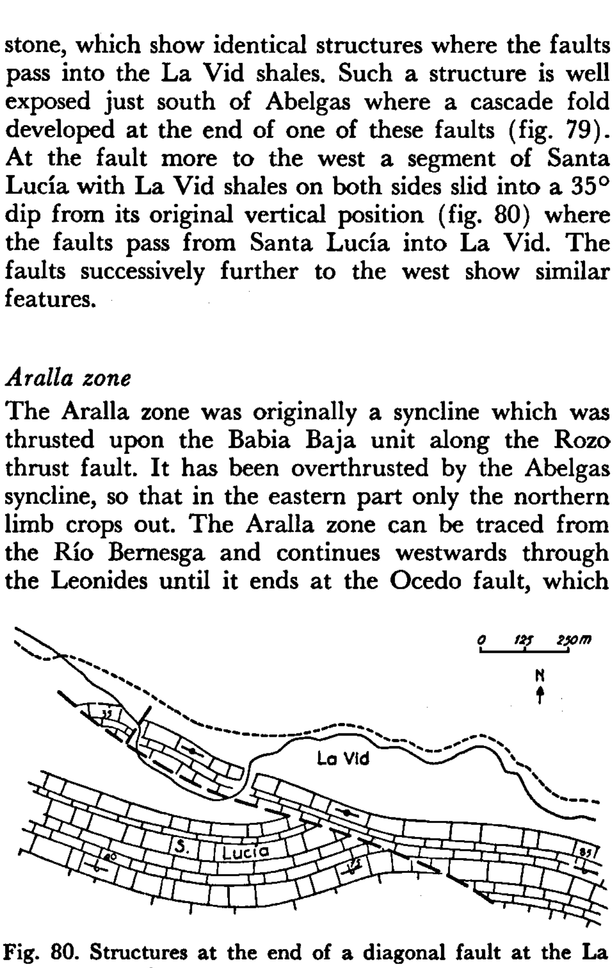

Fig. 80. Structures at the end of a diagonal fault at the La Vid-Santa Lucía contact, west of Abelgas.

cuts off the Leonides in the south and southwest.

As the structures have developed more harmonically than in the Abelgas syncline, no 'tectonic rock unit' subdivisions were made.

The WNW trending eastern part of the Aralla zone is a rather simple structure (sections 1–4), whereas in the WSW trending western part several folds and imbricate structures were found. The youngest strata to crop out are the lowermost La Vid shales.

A good example of the relation between the thickness of the strata involved in the folding and the size and depth of the folds is provided by the central and eastern part of the Aralla zone: folds in the thinner formations of the eastern part have a much smaller amplitude than the folds in the thicker formation in the central part (compare sections 2 and 5); towards the west the sediment thicknesses and fold sizes again diminish (section 9).

At the front of the Aralla zone the Rozo thrust fault brings Láncara upon San Emiliano limestones and greywackes of the Babia Baja unit. Generally the fault contact is extremely thin. It seems probable, in analogy with the Cospedal thrusts, that the Abelgas thrust fault merges into the Rozo thrust fault. Assuming that the sediments involved in the thrusting have an average thickness of 4000 metres (3500 metres in the Babia Baja unit and 5000 metres in the Luna unit), and that the thrust fault had an average initial dip of 30° (the average angle of a diagonal fault with the stress field), the minimum net slip amounts up to at least 8 kilometres.

During continued folding after the thrusting, the front of the Aralla zone was folded into an anticline and a syncline near Rabanal (section 4). Both folds lie exactly in the continuation of the Serronal syncline and the Robledo anticline and we believe that this is not coincidental: just as the Babia thrust fault underwent subsequent folding in the continuation of the Villasecino anticline, the Rozo thrust fault here might have undergone folding which was related to the final folding of the Serronal syncline and the Robledo anticline.

The Grajos fault, named after the Alto de los Grajos (NE of Villafeliz), is basically a wrench-fault, but its wrench character is obscured by thrusting in a NW direction along that part of the fault which forms the NW boundary of the Aralla zone. The deflection of the beds of the WNW trending Luna and Babia Alta units to the WSW shows that shortening mainly normal to the faults has taken place. The magnitude of the strike-slip movements is thus small (Garfunkel, 1966).

Complicated structures developed in the area where the WSW trending Grajos fault and the WNW trending Rozo fault meet; these structures have been called the Campo del Oro structures after the low hills east of Riolago (fig. 81). The Láncara has been repeated three times in the eastern part of these structures and all these Láncara wedges reflect the interference of the two local stress fields associated with the Grajos and Rozo faults in these strongly faulted Láncara exposures. In the two northern Lán-

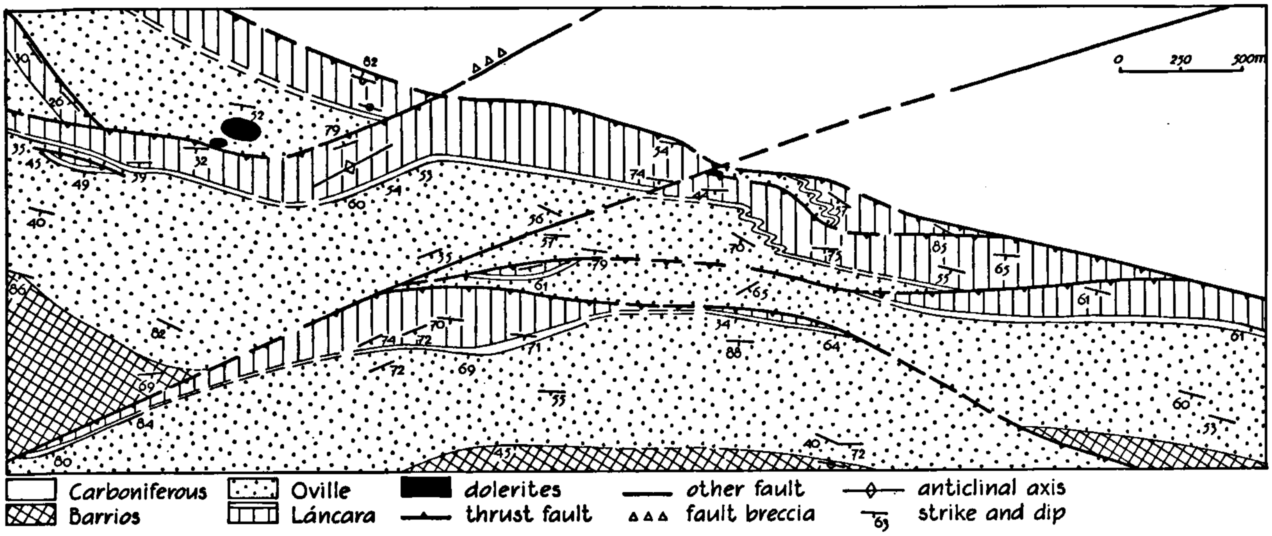

Fig. 81. Detailed map of the Campo del Oro structures.

cara wedges, which have WNW strikes, small folds have developed with WSW trending axes. The strike of the other two Láncara wedges curves sharply from WNW to WSW. Where the Grajos and Rozo faults intersect, neither fault seems to displace the other. An explanation for this unexpected phenomenon will be given in the last chapter.

In the western part of the Aralla zone, just behind the Grajos fault, an imbricate structure was found which closely resembles the structures near the Cáscaros (fig. 82; section 8 and 9) and is also of the same origin. Here two minor thrusts developed on the main thrust fault. Westward their throw increases rapidly until the base of the Barrios is in contact with the Formigoso. Just as at the Cáscaros structure, this imbrication occurs where the main thrust fault leaves the Láncara level and cuts upwards through the Barrios. Contrary to the Cáscaros structures the Barrios does not stand vertical where it is cut off by the main thrust fault but dips gently westwards. The whole imbricate structure is cut by a N-S trending cross-fault with a rotary movement; where it cuts the Barrios-Formigoso contact, black shales of the Formigoso have been pressed between both fault surfaces, just as in the exposure near the Churros.

Another example of a thrustfaulted anticline with Barrios cropping out in the core can be seen near the Alto de Salgadina (section 6). The Barrios-Formigoso contact is repeated twice by two minor strike-faults. The map has been simplified on this point, the faults being too small to be represented at the present scale.

A cascade fold was found in the most eastern part of the Aralla zone. Here again, La Vid limestones and San Pedro sandstones are involved (section 1). West of this fold three clearly WSW trending fault cut the Barrios and San Pedro diagonally, just in the continuation of the important WSW trending fault in the Babia Baja unit. The relationship between these and other structures in the Luna unit and the important fault in the Babia Baja unit will be discussed in the last chapter.

Beyond doubt, the most complicated part of the Luna unit is the SW part of the Aralla zone (fig. 83). A slice with Oville, Láncara and even Herrería (A) has been transported upwards in front of the Abelgas thrust fault; its SW part is an anticline with an overturned northern limb. This is regarded as an autochtonous anticline, on the back limb of which the Láncara started thrusting (p. 217). The front of this slice is formed by white quartzite 20 metres thick, which closely resembles the Barrios, and is bounded at both sides by a strike-fault. North of this slice a gentle anticline (B), plunging to the NE and with Láncara in the core crops out, bounded by two WSW trending faults (section 9). North of this anticline another anticline (C) and a small syncline (D) were found, both plunging to the NE and cut off in the north by a thrust fault along which a north to south movement took place. This may be the result of 'underthrusting' due to lack of space in the core of the anticline (fig. 112).

BABIA ALTA UNIT

The Babia Alta unit, named after the high western part of Babia, is bounded in the southeast and east by the Babia thrust fault, in the south by the Grajos fault, and in the west by the Mora/Herrería unconformity and by the unconformable Stephanian Villablino basin. The Babia Alta unit continues many kilometres to the NW with generally WNW strikes which turn northwards 20 kilometres north of the region mapped (García Fuente, 1959). The Babia

* Billings, 1947, p. 158.

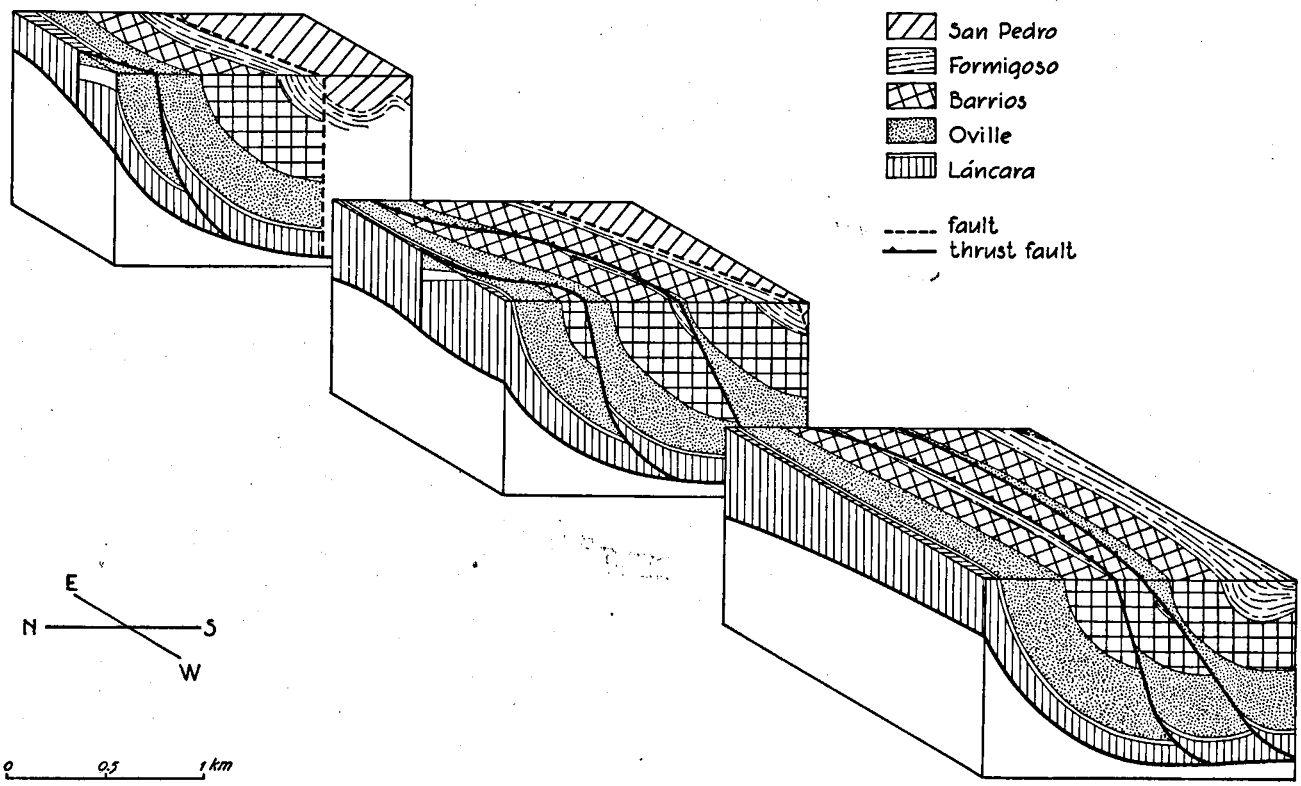

Fig. 82. Sketch of the imbricate structures in the Aralla zone, behind the Grajos fault.

Alta unit has been subdivided into the Torre zone, named after the village of Torre de Babia; the Saliencia zone, named after the lakes of Saliencia; the Meroy zone, named after the village of Meroy; the Muxivén zone, named after the mountain top of Muxivén, and the Corralines zone, named after the Corralines hill, SE of Mena.

The Torre zone is separated from the Meroy and Saliencia zones by the La Vid calcareous shales. Above and below these shales folding is extremely disharmonic (fig. 84). As in the Abelgas syncline, the structures below these shales can not be correlated with structures above them. Bedding-plane shearing, crumpling and

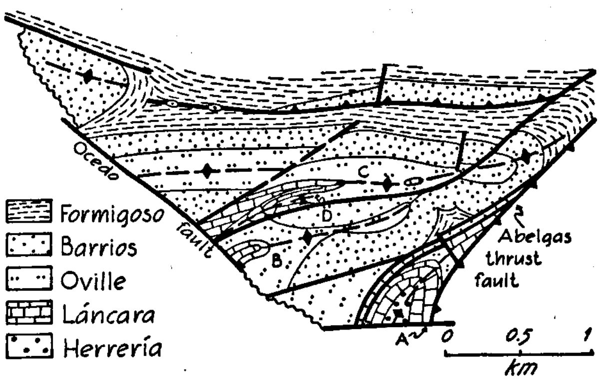

Fig. 83. Index map of the SW part of the Aralla zone.

Fig. 84. A large box-fold in the Santa Lucía Formation of the Saliencia zone, NE of the Lagunas de las Verdes (just north of the map). This picture also very well shows the disharmonic folding on either side of the La Vid calcareous shales (after photograph).

Fig. 85. Very well exposed, extremely sharp, thrust fault along which Láncara was brought upon San Emiliano, between Torre de Babia and Robledo de Babia; the fault breccia is only 20 cm thick.

flowage were the most important mechanisms for the deformation of the La Vid shales themselves. This complete disharmony between the competent beds on either side of the La Vid calcareous shales could only occur because the thickness of these shales was greater than the fold wave lengths in the Santa Lucía (Ghosh, 1968). The La Vid limestones may be compared with a stiff basement and the overlying beds with a detached sedimentary cover. It appears that the incompetent beds nearest to the La Vid limestones remain parallel to these limestones in large flat segments with some narrow, sharp-crested anticlines. In the upper part of the La Vid shales and the base of the Santa Lucía, these flat parts constitute the flat crests of the box-folds.

Torre zone

This zone is limited in the west and NW by the calcareous shales of the La Vid, in the south by the Corralines fault, in the east and SE by the Babia fault, and in the north by a W-E trending fault west of Genestosa which is only partly covered by this map.

The axes of the WNW trending folds plunge about 45° to the WNW, thus affording us a much better view of the structures than in the Luna unit, where the fold axes generally plunge to a lesser extent.

The Babia thrust fault merges southwards into the Rozo thrust and continues many kilometres north of this map. Complications of the same type as the Grandas structures (fig. 65) were found west of Genestosa (de Sitter & van den Bosch, 1968) of which the most southern part appears on the present map. An excellent exposure of the Láncara-San Emiliano fault contact was found west of Robledo: an extremely sharp fault, with about 20 centimetres of fault breccia. This fault runs exactly parallel to the beds, both in the Láncara and in the San Emiliano, and is exposed over a length of about 300 metres (fig. 85). This parallelism of the Láncara and the San Emiliano could be followed up to Cospedal, whereas more to the east this parallelism was likewise found, though not so perfect as between Huergas and Cospedal.

Subsequent folding of the Babia thrust fault, as found in several places in the above-mentioned excellent exposure west of Robledo (section 7), shows that the Villasecino anticline, and hence also the Torre anticline (fig. 65), just as probably the Robledo anticline and the Serronal syncline (p. 209), were formed, at least partially, after the thrusting. A less distinct relationship exists between the Tesa syncline and the Tiesa syncline, named after the mountain top La Tiesa, north of Robledo (fig. 65). The fold axis of the latter, however, can not be directly connected with the fold axis of the Tiesa syncline. This is due to the large mass of very competent Barrios quartzites of the

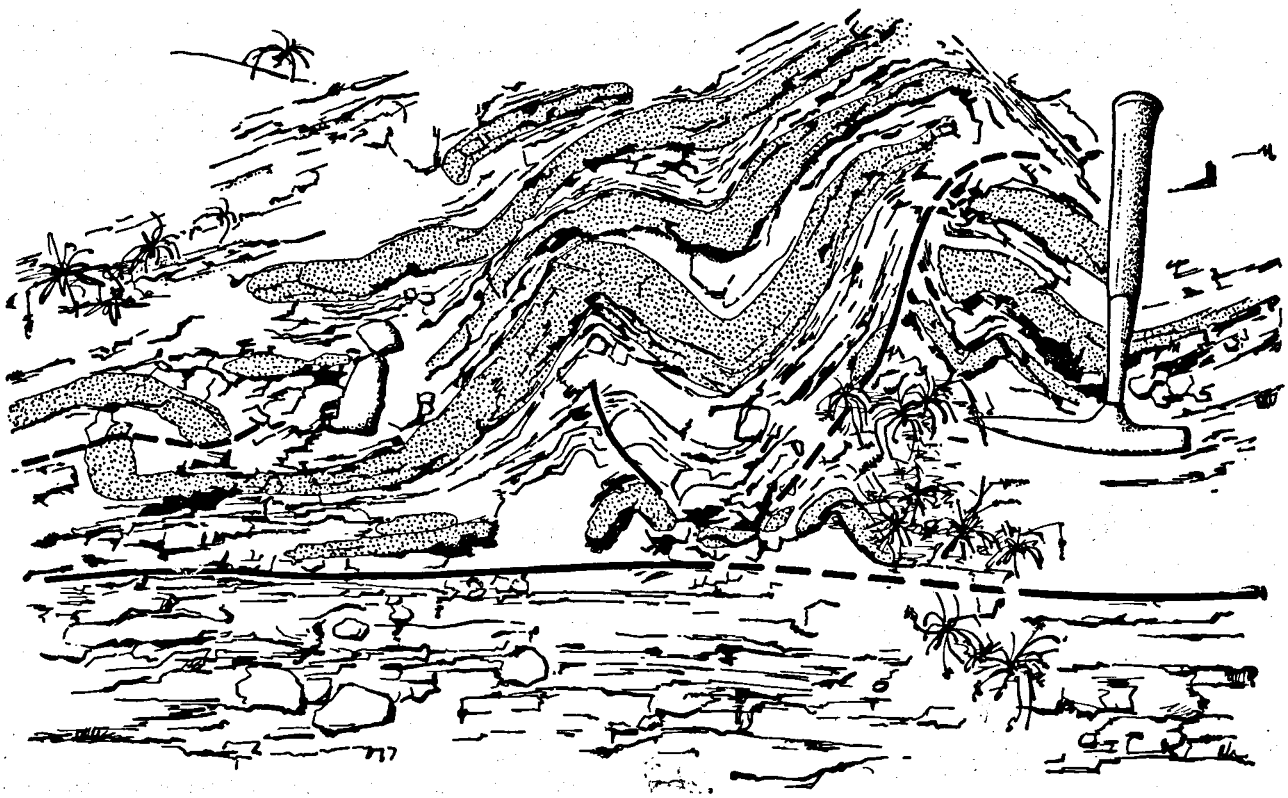

Fig. 86. Disharmonic folding in the Formigoso Formation, NNE of Torre de Babia. This broken fold very well demonstrates the origin of a thrust fault (after photograph).

Cospedal thrusts which determined the place of the axis of the Tiesa syncline: the fold axis did not originate in this competent mass but close to it.

The Torre anticline and its continuation, the Villasecino anticline, in the Babia Alta unit, show a small syncline at their crest (section 7). The Láncara in the core of the Torre anticline, the Grandas structures, underwent more intensive deformation than the younger beds above it: another example of disharmonic folding (section 9).

The imbricate structure between Robledo and Huergas developed first as mainly WNW trending small thrusts. All thrusts ended in the Formigoso shales or even below that level. This imbricate structure was later folded into two anticlines and a syncline which are also present, but less pronounced, in the younger beds of the Torre anticline. Note that the small thrust fault which passes through Huergas, just as several thrusts in the Luna unit, ends in a broken anticline, in front of which a slice crops out.

The incompetent character of the Formigoso is demonstrated at several locations in the Torre zone. It was found to contain interesting minor folds just NNE of Torre, such as the disharmonic folds represented in figure 86. Their axes stand steep and are parallel to bedding-planes in the neighbouring San Pedro and Barrios. Possibly these folds were set steep during a later stage of the folding. One of these folds reveals much about the origin of a thrust fault (fig. 86). Furthermore, tight parasitic folds were encountered in the black shales in the lower part of the formation with axes parallel to the bedding-plane. Other arguments in favour of the refolding are the cross-folds with their axes parallel to the direction of refolding (fig. 87). Refolding of a parasitic fold by one of these folds was found only once.

The Cospedal thrusts, named after the village of Cospedal, merge into the main thrust fault at the base of the Láncara, which has been crushed intensively. At this base their throw is very small, but going upwards it increases until Barrios lies upon La Vid limestones. Especially in the western part of the structures, anticlinal bends were found in the top of the Barrios, just behind the thrusts. At the higher San Pedro level complete folds exist (section 7) showing the association of thrusts and folds. The thrusts originated from the breaking through of an anticline. Note that the Cospedal thrusts with their associated folds are very similar to the structures behind the eastern part of the Abelgas thrust and presumably have a comparable history. Two small NNE trending faults in the Barrios of the most eastern thrust, and two larger ones cutting both thrusts, must have originated later than the thrusting, because they displace the thrusts.

On the northern flank of the San Félix syncline, named after the village of San Félix de Arce, several small thrusts developed, parallel to its axial trend; they are of the same type as those in the older strata near

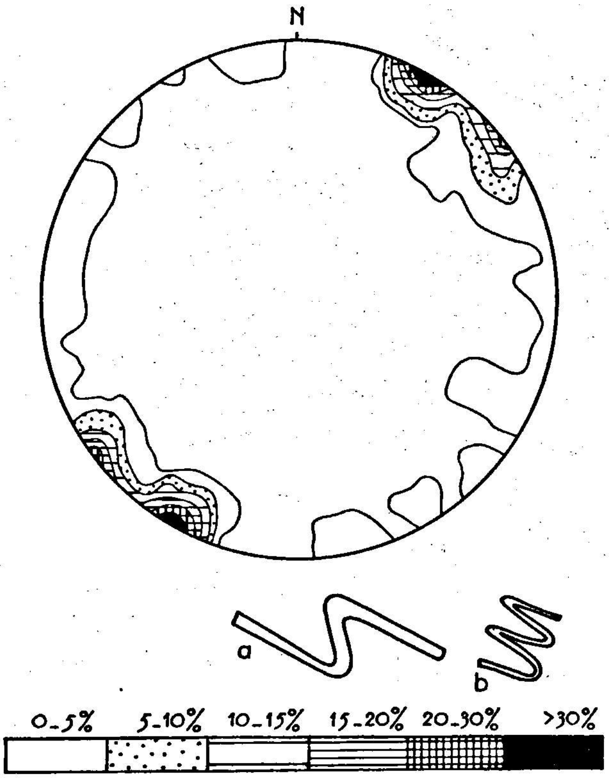

Fig. 87. Poles of axial planes, measured in the Formigoso Formation, NNE of Torre de Babia. The folds of type *b* are parasitic folds; the folds of type *a* appear to be of later origin than those of type *b*.

Torre. The WNW trend of the synclinal axis, and also the parallel axis of the Cabrillanes anticline, named after the village of Cabrillanes, curve eastwards at their SE end, in the neighbourhood of the Corralines fault and the associated parallel faults. The Corralines fault cuts the Cabrillanes anticline longitudinally and the fold curves there to an even WSW trend. A regular pattern of WSW and NW trending faults has been mapped west of Huergas where the thrusts interfere with the wrench movement.

Saliencia zone

The Saliencia zone is a synclinal structure with La Vid shales at its base. In the SW it is bounded by the Caunedo thrust fault. In its SW part the syncline shows some smaller additional folds (de Sitter & van den Bosch, 1968); only the latter small folds are covered by this map (section 10).

The structures in the Santa Lucía, situated between the mobile calcareous shales of the La Vid and the incompetent Huergas shales are of great interest. Box-folds were found at two locations south of the Yegüero, and also in the Meroy zone (fig. 88). These box-folds were probably caused by mobile La Vid shales, which locally migrated upwards under high pressure during the folding (fig. 111). An enormous box-fold in the Santa Lucía above the unfolded La Vid basal limestones just west of long. 2°26′ and immediately north of this map, clearly demonstrates the disharmony of folding above and below the La Vid calcareous shales (fig. 84).

The throw of the Caunedo thrust fault of which only the SE part is covered by this map, increases towards the NW until Láncara was thrusted on to Huergas. On the present map, where throws are small, the thrust is

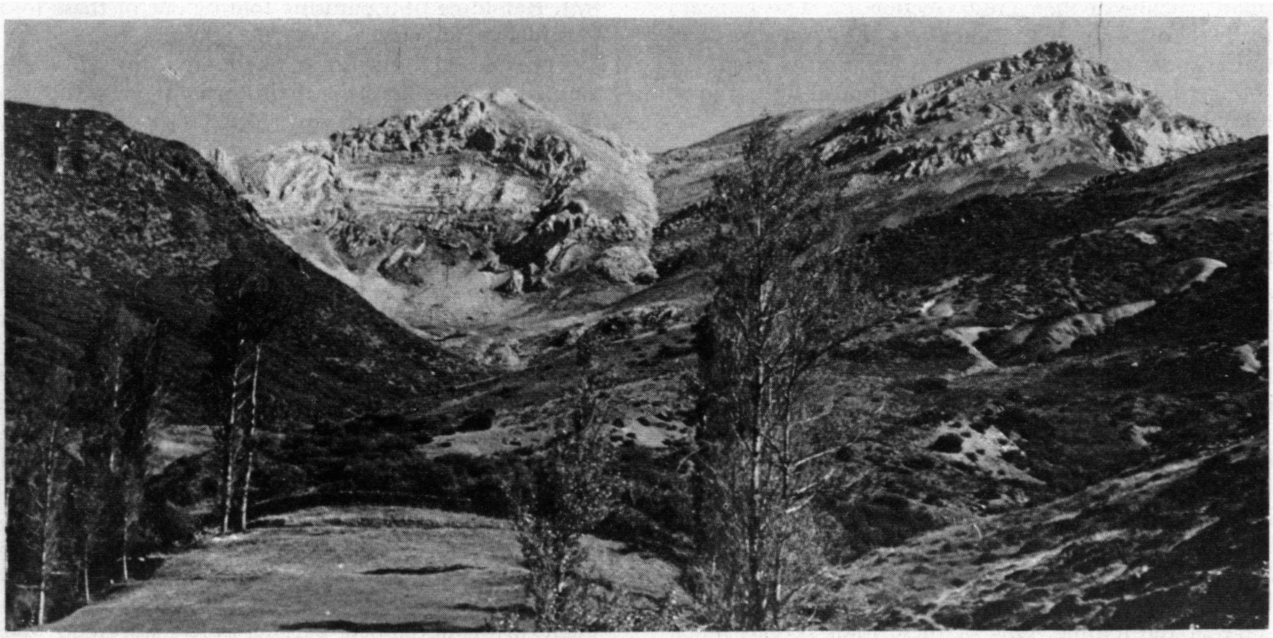

Fig. 88. Box-folds in the Santa Lucía Formation, NNW of La Riera (photograph by Mr. E. S. Jharap).

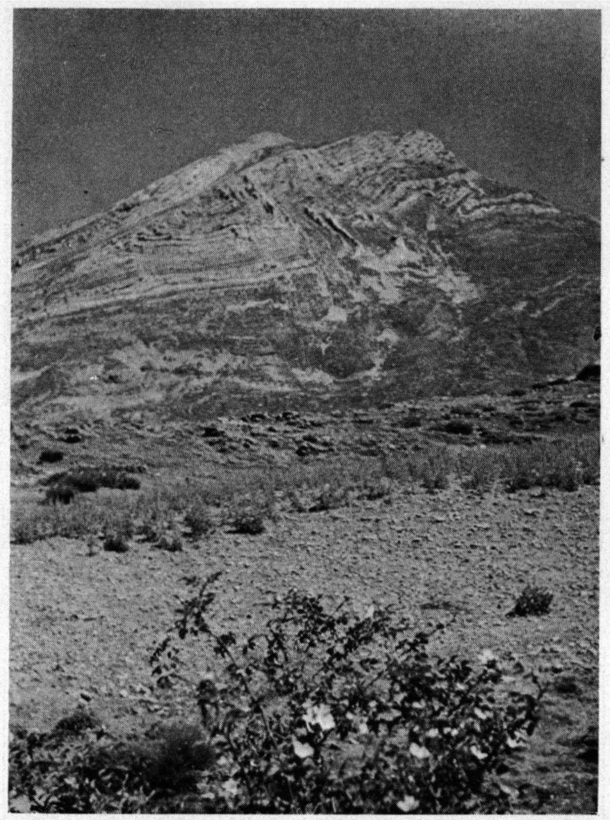

Fig. 89. Box-fold in the Santa Lucía, north of Las Murias.

a double fault. The Santa Lucía between the two faults is a rootless anticlinal crest which can be seen very well along the path from La Cueta north-eastwards, just north of the region mapped (section 10). This section also shows that this fault zone originated from the breaking through of an anticline along two faults; the syncline behind this broken anticline seems to be of the same age.

Meroy zone

The Meroy zone is bounded in the NE by the Caunedo thrust fault and furthermore by the calcareous shales of the La Vid, except in the SE where the unconformable Villablino basin overlies this contact. It can be subdivided into the Palomas syncline, named after the Puente de las Palomas, the Cacabelos anticline, named after the hamlet of Cacabelos, and the Quejo syncline, named after the hamlet of Quejo.

As stated above, the Quejo syncline is contemporaneous with the thrusting. The relations of the Palomas syncline and the Cacabelos anticline, either with respect to the thrusting or to the refolding, are difficult to establish. From sections 10 and 11 it seems probable, however, that these three folds and the Caunedo thrust zone are contemporaneous. Subsequent steepening occurred due to continued folding.

The Quejo syncline is an open fold, its axial plane inclined to the NE and its axis plunging gently to the NW. A normal cross-fault lengthens the nose of the

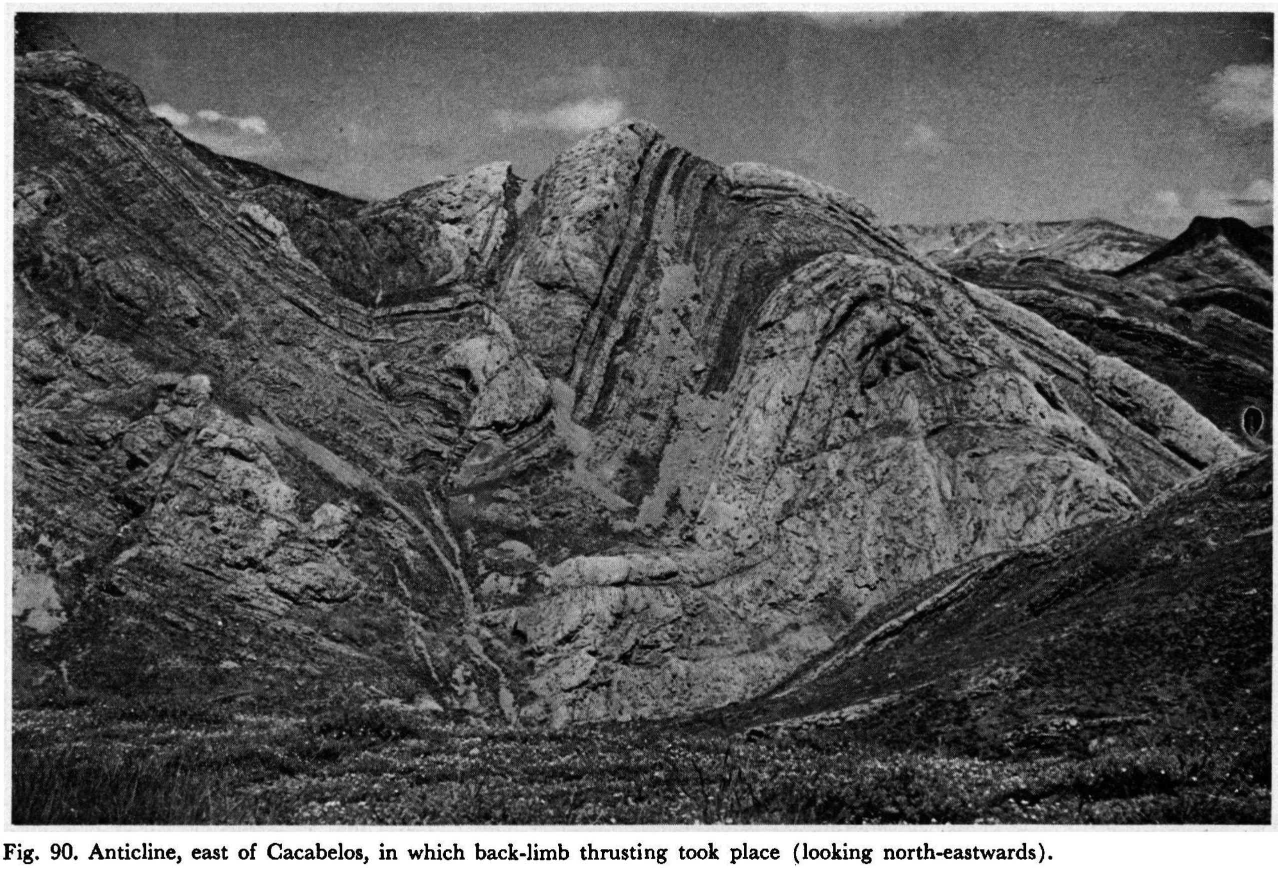

Fig. 90. Anticline, east of Cacabelos, in which back-limb thrusting took place (looking north-eastwards).

syncline. The syncline has a simple core of Portilla up to Caliza de Montaña. The Santa Lucía at the base of the structure shows a subsidiary small anticline. In between these two formations, the Huergas facilitates this disharmonic folding. Again, the base of the Santa Lucía has been distorted intensively, but on so small a scale that not all these structures could be represented on the map. Box-folds were encountered at the same level (fig. 89), just as in the Saliencia unit.

The Cacabelos anticline is an open fold of Santa Lucía limestone with La Vid calcareous shales in the core; its axial plane dips to the NE and its axis plunges to the NW. Four minor folds, two synclines and two anticlines, were found on its NE flank (section 11). The two anticlines have their detachment-planes in the La Vid shales, which again shows the disharmonic character of these shales. The easternmost of these minor folds (fig. 90) is an illustration of 'back limb thrusting', which does not develop from breaking through at the frontal limb of an anticline but from a fault parallel to the bedding-plane in the back limb (fig. 91). In the opposite limb of this anticline several minor faults were encountered, parallel to the back limb faults; these faults, however, are not parallel to the bedding-plane, but cut it obliquely (fig. 92).

The NW-SE trending Palomas syncline is an open fold, its axial plane dipping to the NE. Near the

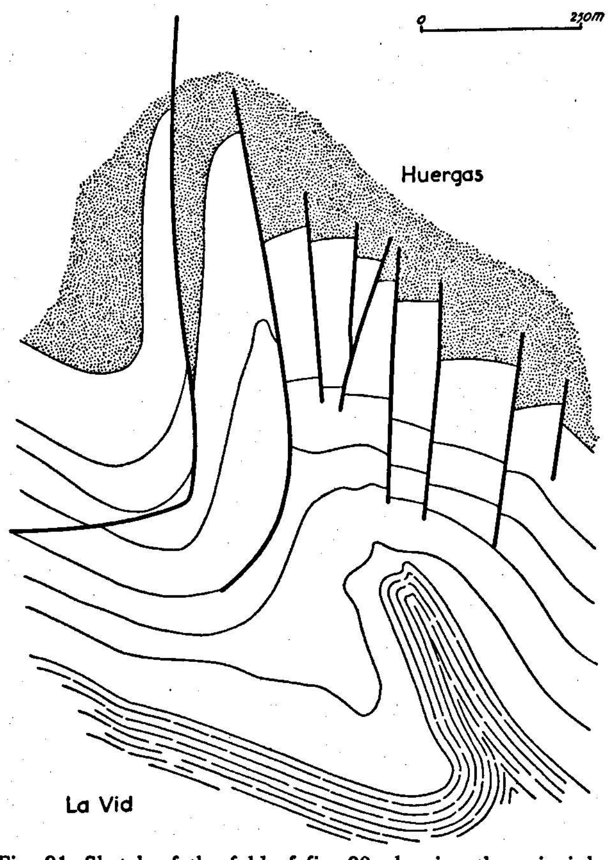

Fig. 91. Sketch of the fold of fig. 90, showing the principle of back-limb thrusting.

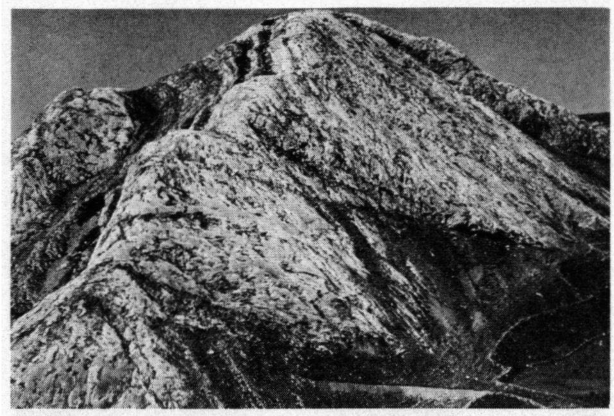

Fig. 92. Cross-faults on the frontal limb of the fold of fig. 90 (looking north-westwards), photograph by Mr. E. S. Jharap.

Lumajo fault, named after the village of Lumajo, and the Corralines fault, the fault axis respectively curves slightly to the north in the northern part and to the ENE in the southern part. The important Lumajo

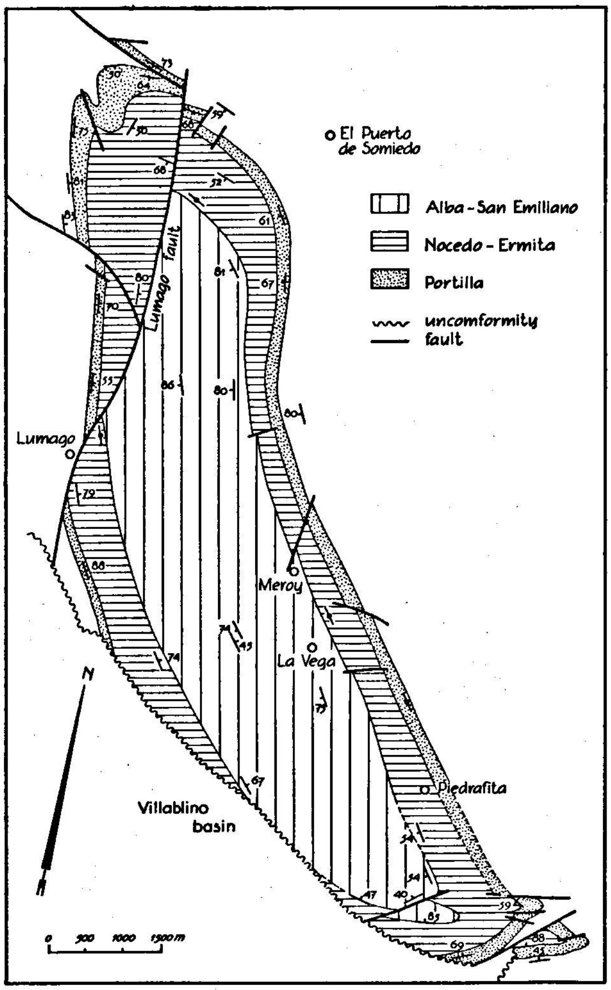

Fig. 93. The entire Palomas syncline and Lumajo fault. This figure shows that a deviation of the strikes occurs near the Lumajo fault and also that strike-slip movement took place along the Lumajo fault.

Fig. 94. Parasitic folds in the Caliza de Montaña of the eastern flank of the Palomas syncline, north of La Vega de los Viejos (looking south-eastwards).

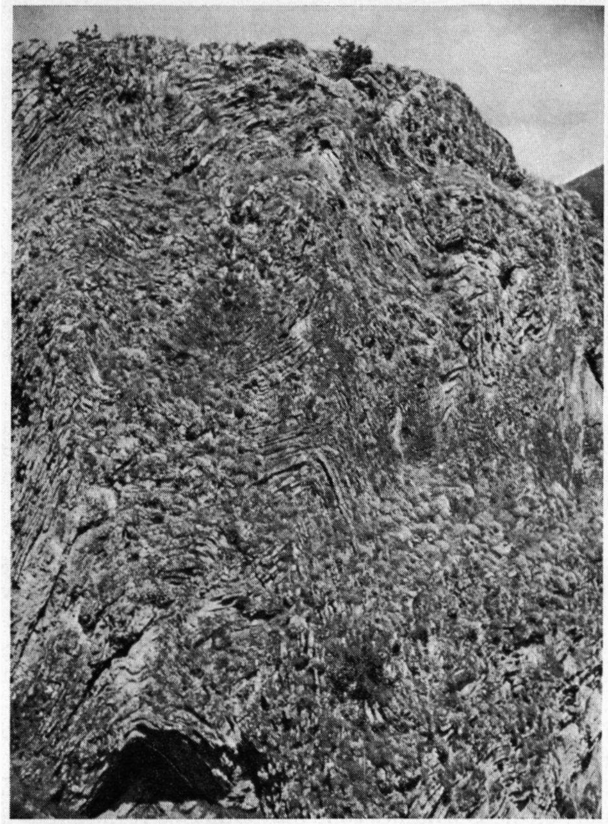

Fig. 95. Parasitic folds in the Caliza de Montaña near the Puente de las Palomas, looking northwards.

fault has a vertical throw of about 700 metres (section 12) and a horizontal displacement of at least 250 metres which could be established north of this map (fig. 93). It cuts off a WSW trending cross-cutting fault which can be followed as running into the Precambrian (Mr. P. den Hengst, pers. comm.). The nose of the Palomas syncline is cut by many perisynclinal faults, two of them being normal cross-faults. A thickening is caused, especially in the core of the syncline, by numerous parasitic folds which are found on both flanks (fig. 94) as well as in the hinge zone (fig. 95) in the platy limestones of the Caliza de Montaña Formation.

Corralines zone

The Corralines zone is limited by the Grajos fault in the SE, the Corralines fault in the NW, the Babia thrust fault in the NE, and the unconformable Villablino basin in the west. As the Corralines fault runs parallel to the Grajos wrench-fault, it is regarded to have originated as a wrench-fault. Subsequent vertical movements obscured the original character of the fault. Structures in the Luna unit are separated from those in the Corralines zone. The Barrios in the latter clearly constitutes the southern flank of the Cabrillanes anticline, the Grajos faults thus being the obvious separation between the Babia Alta and the Luna unit.

The Corralines zone shows the interference of the WSW trending structures associated with the wrench movement and the NW-SE trending structures associated with the thrusting. In this zone the incompetent Formigoso shales caused disharmony between structures in the Barrios and structures in the San Pedro; these structures will be dealt with separately.

The front of the Corralines zone is formed by the western part of the Campo del Oro structures. East of Riolago the strike of the Láncara suddenly changes from WNW to WSW under influence of a WSW trending fault which cuts the San Emiliano of the Babia Baja unit and the Láncara of the Corralines zone. At the same place the thrust plane leaves the base of the Láncara and cuts upwards through a small anticlinal fold (section 6). The Barrios also reflects the interference of the two directions very well: several NW trending faults and folds cross the WSW trending Cabrillanes anticline.

The beds above the Formigoso black shales show a somewhat different pattern: several approximately W-E trending folds, arranged *en échelon,* are cut off by SW-NE trending faults, forming an angle of 10–15° with the main wrench direction, and by a NNE trending thrust fault, forming an angle of approximately 45° with that direction. The faults in the Corralines

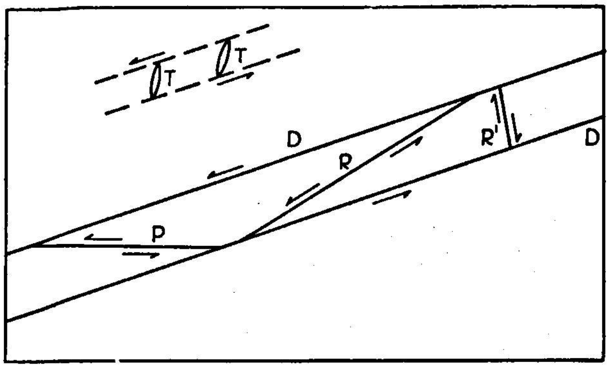

Fig. 96. Riedel shears (R) and their conjugate set R′, tension fractures (T) and shear fractures or thrust shears (P) can originate between two surfaces along which horizontal slip takes place (Grajos fault and Corralines fault), Skempton (1966) and Ashgirei (1963, p. 59).

zone are of the R and R′ type (fig. 96). Experiments showed that Riedel shears appear in the sedimentary cover before the main faults are visible there. Small-scale overthrusting along the R′ fault in the Corralines zone is due to further compression. The arrangement of the folds *en échelon* also indicates that the Corralines zone is bounded by two wrench-faults, oblique to the main stress (Hubbert, 1928).

The long narrow zone, limited by faults, in the SE part of the Corralines zone may either be regarded as a thrust slice, or as a shear lens, i.e. a lens-shaped mass between a Riedel shear and the main slip surface.

Muxivén zone

The Muxivén zone is limited in the east by the calcareous shales of the La Vid, in the west by the Mora-Herrería unconformity, and in the south by the Villablino basin. It continues northwards until being cut off by a WNW trending fault, parallel to the Caunedo thrust (de Sitter & van den Bosch, 1968). The block-faulted Muxivén zone forms the transition between the Leonides and the West-Asturian-Leonese Zone.

The interference of the NNW trending general strike and the WNW strikes of the cross-faults resulted in a rather complicated pattern, in which folds appear, whose axes are parallel to these cross-faults. This shows that, although the general stress was directed WSW-ENE, local deviations exist near the WNW trending faults. Several other faults are parallel to the general strike. One fault with a NNE strike determined the pre-Stephanian topography and thus was active before the deposition of the Stephanian. As in many other places, cascade folds are common here in the La Vid limestones and the San Pedro (section 12).

BABIA BAJA UNIT

The Babia Baja unit is named after the eastern, lower part of Babia. It is bounded by the Rozo thrust fault in the south, the Babia thrust fault in the west and thrust faults in the continuation of the León line, which thrust faults bound the Asturian Coal Basin, in the NE; to the east it merges into the Bodón unit (de Sitter, 1962). The Babia unit is characterized by mainly ENE-WSW trending, rather simple folds with their vergence to the south, and is cut by the fundamental Grajos fault, named after the Alto de los Grajos NE of Villafeliz, with the same trend.

In the Babia Baja unit pre-Carboniferous sediments are much thinner and the style of deformation is much simpler than in the Luna and Babia Alta units. As only slightly disharmonic structures are present, mainly

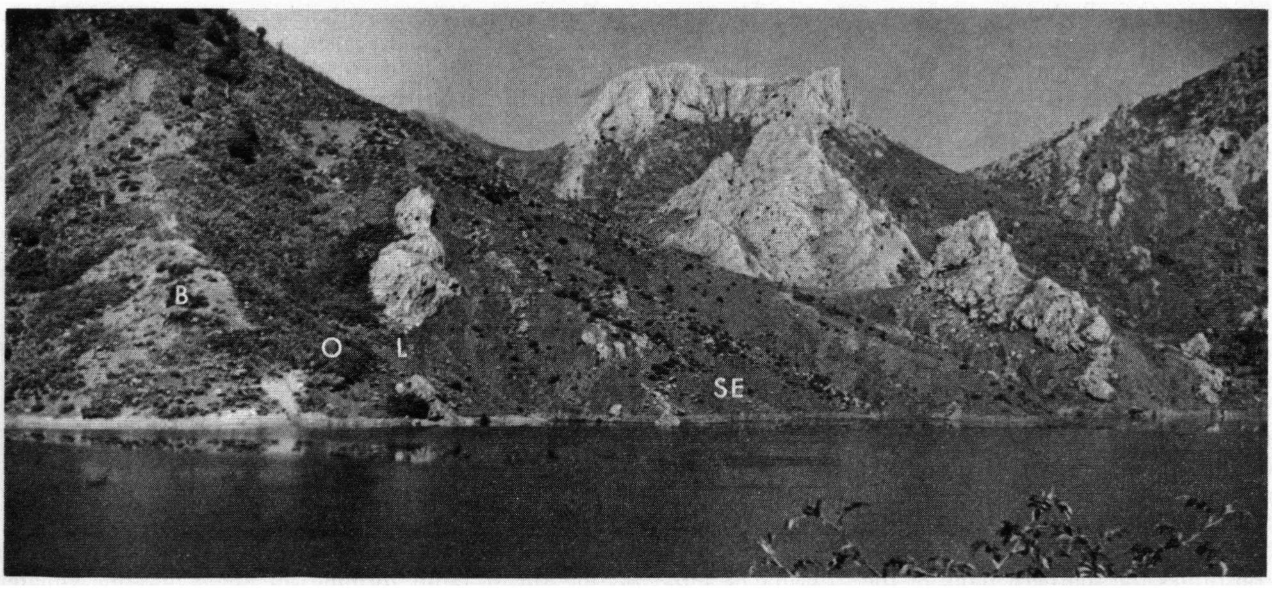

Fig. 97. Small syncline in the San Emiliano, just north of the Rozo thrust fault (looking westwards).

Fig. 98. The Robledo anticline, northwest of Caldas de Luna (looking westwards).

in the hinge zones of the anticlines and due to the incompetence of the Formigoso shales, no subdivision into 'structural rock units' has been made.

The only indications of a larger syncline south of the Robledo anticline are the two U-shaped limestones east and west of Láncara de Luna (fig. 97). It was difficult to assess whether these forms are intertonguing limestones and shales, folds caused by drag along the thrust fault, or indeed the core of a large syncline; the first hypothesis seems by far the least likely.

The WSW trending fault in the Carboniferous east of these structures does not seem very important on this map, but more to the east it appears to be highly important, a wedge of Devonian having been thrusted up along this fault (van Staalduinen, 1969). The rôle of this fault will be discussed in the last chapter.

The Robledo anticline, named after the village of Robledo de Caldas, is a symmetrical upright open fold, plunging to the west (fig. 98), with its axial plane dipping about 85° to the north; only in its western part complications are found in the hinge zone. East of Robledo de Caldas a set of small rectangular flexures is exposed in the top layers of the Caliza de Montaña, whereas west of this village a gentle minor syncline was encountered on the anticlinal crest. More to the west the exact trace of the plunging fold axis could not be found in the thick, badly exposed San Emiliano shales and greywackes.

The axial plane of the simple open Serronal syncline (van Staalduinen, 1969) dips about 75° to the north. Just as the Robledo anticline, the anticlinal axis in the thick San Emiliano shales could not be traced further than as far as Sena. Near this village both axes approach each other, and the amplitudes of the folds diminish (p. 199).

Vertical and lateral facies changes occur frequently in the San Emiliano Formation of the Serronal syncline; the northern limb consists mainly of thick limestone sequences, the eastern limb consists of an alternation of mainly shales and limestones. The synclinal axis often lies in places where thin limestone beds pass laterally into one massive limestone bed.

The most important feature of the Babia Baja unit is doubtlessly the Villasecino anticline, named after the village of Villasecino. This anticline is cut longitudinally by the almost straight Grajos wrench-fault. The axial plane is vertical or dips subvertically to the north. In the east, where the core of the anticline crops out, it is a tight fold; towards the west it becomes more and more an open fold (sections 1–6 and fig. 99). Between Robledo and Villasecino a minor syncline was found at the crest of the anticline (fig. 99). West of Villasecino the axis of the structure is no longer parallel to the Grajos fault but turns to the WNW. Through the very poorly exposed area near Robledo the anticlinal structure continues into the overthrust Babia Alta unit.

Dolomitization of the Caliza de Montaña and the

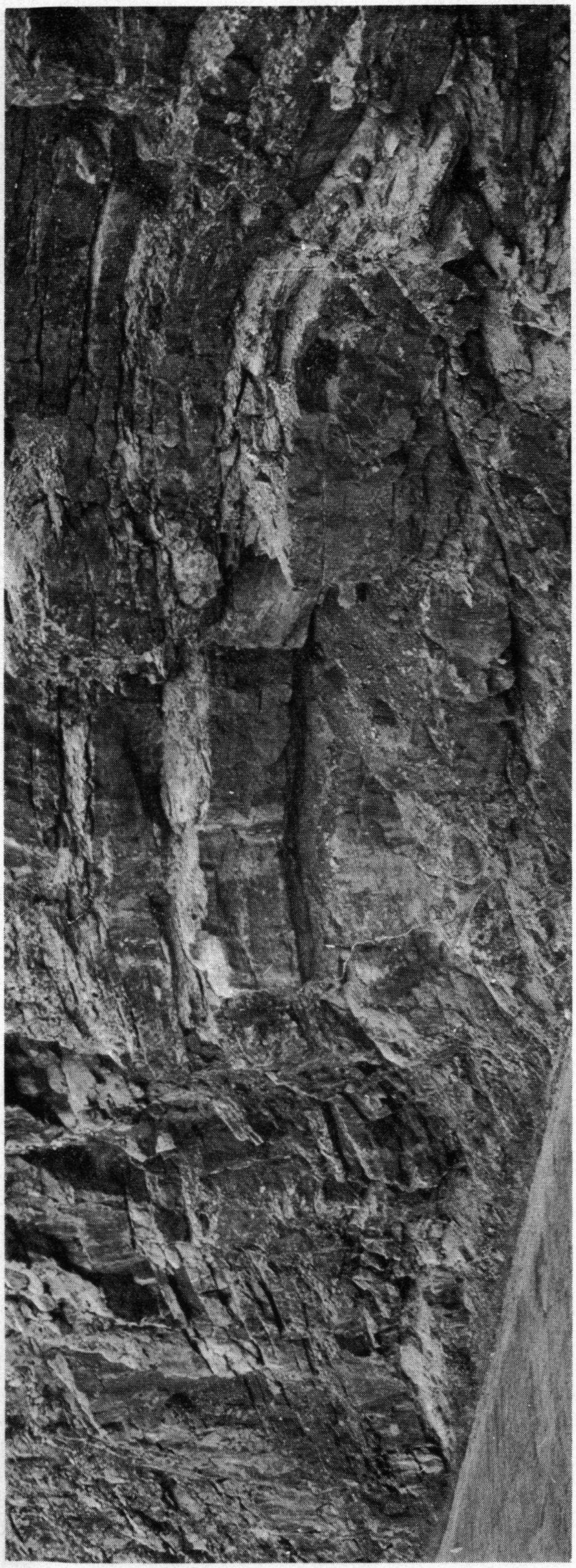

Fig. 99. Minor syncline at the crest of the Villasecino anticline west of Villasecino (looking north-eastwards).

San Emiliano limestones in the very poorly exposed area south of Robledo and Cospedal rendered mapping difficult. Moreover, the various San Emiliano limestone beds SE of Robledo constitute a structureless, dolomitized mass. The WSW trending San Emiliano beds are cut off by the WNW trending limestones, probably Caliza de Montaña. Similarly to the Rozo thrust near Rabanal, the Babia thrust fault near Robledo was folded during the final folding stage of the Babia Baja unit (p. 218).

The Grajos fault is essentially a WSW trending left-lateral wrench-fault cutting through the Leonides (p. 220). The interference of the Grajos wrench-fault with W-E trending faults, NE of Villafeliz, resulted in a complicated structure (fig. 100). These faults diverge from the Grajos fault at an angle of 20°, and are more or less vertical. Left-lateral shearing and also upthrusting took place along these faults. They are called splay faults (Anderson, 1942; Chinnery, 1966) or thrust-shears (Skempton, 1966) and were generated by secondary stresses caused by the action of the main fault. Probably the western splay fault continues into the faults north of Villasecino and south of Robledo. If this is true, we are confronted with an arcuate rotary fault or propellor fault. A minor fault north of Villafeliz, which also diverges from the main fault at an angle of about 20°, but in a WSW direction, may be a second-order shear, also generated by secondary stresses along the main fault (Chinnery, 1966).

The strikingly straight contact between the San Emiliano shales and the Caliza de Montaña near the Puerto de la Cubilla is not parallel to the structures in the Alba griotte and older strata in the Villasecino anticline. The probable fault character of this contact could not be established in the field.

In the core of the Tesa syncline (van Staalduinen, 1969) several parasitic, accordion and cascade folds are exposed (fig. 101 and 102). The form of the syncline itself is very tight in the eastern part, having a vertical southern limb and an overturned northern limb, but further to the west it becomes an open syncline with moderately inclined limbs.

A NNW trending normal fault 20 kilometres in length, its western block thrown downwards, ends somewhere in the shales east of Torrebarrio (fig. 103). Final movement along this fault took place during and after the Stephanian, this fault limiting a Stephanian basin more to the north.

The WSW trending faults and their associated folds in the NE part of the syncline lie in the continuation of a wrench-fault zone in the Ubiña zone, have the same movement and hence are probably associated with this fault zone. The eastern part of the Ubiña zone, named after the more than 2400 metres high Peña Ubiña (fig. 103), shows a steep WSW trending thrust fault along which the Babia Baja unit was thrusted upon the Asturian Coal Basin. This fault was later cut off by a steep N-S trending thrust fault along which the western part of the Ubiña zone was thrusted upon the eastern part. A slice, with San Pedro and La Vid cropping out, was caught along it. Simultane-

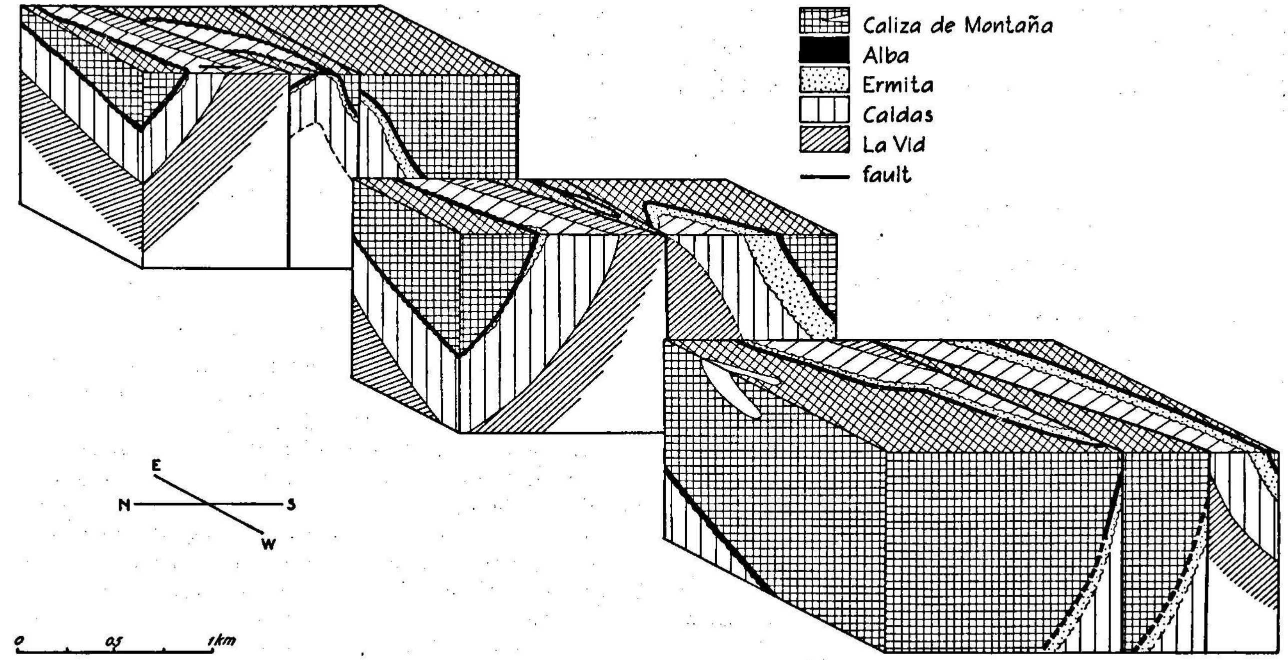

Fig. 100. Sketch of the structures SW of the Alto de los Grajos.

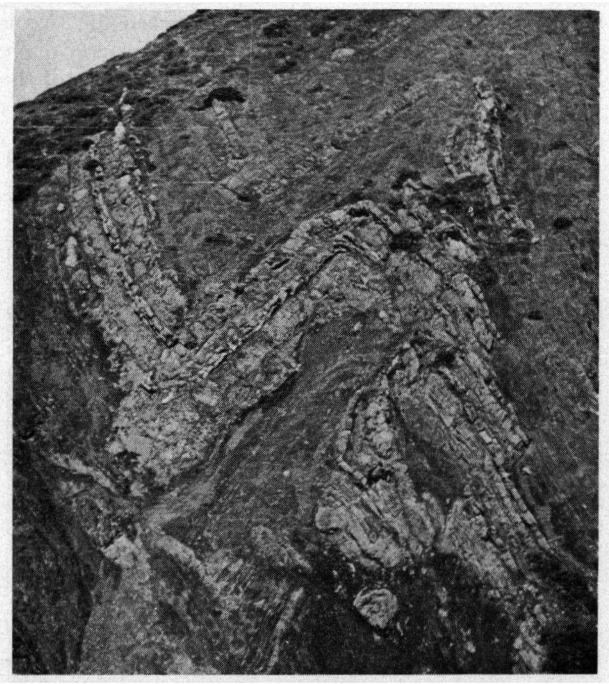

Fig. 101. Broken parasitic folds in the San Emiliano Formation on the northern flank of the Villasecino anticline, between Pinos and Puerto de la Cubilla (looking south-westwards).

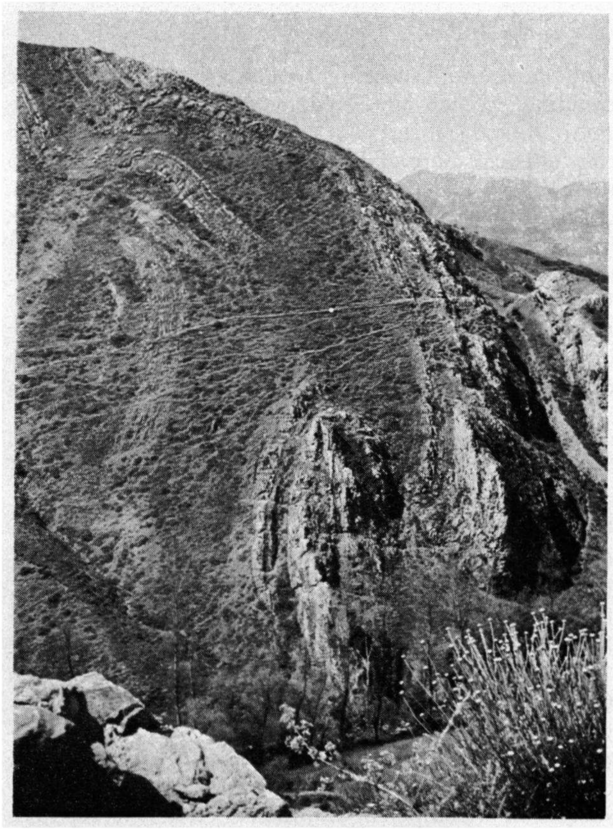

Fig. 102. Cascade fold in the San Emiliano Formation on the northern flank of the Villasecino anticline, just east of Pinos (looking south-westwards).

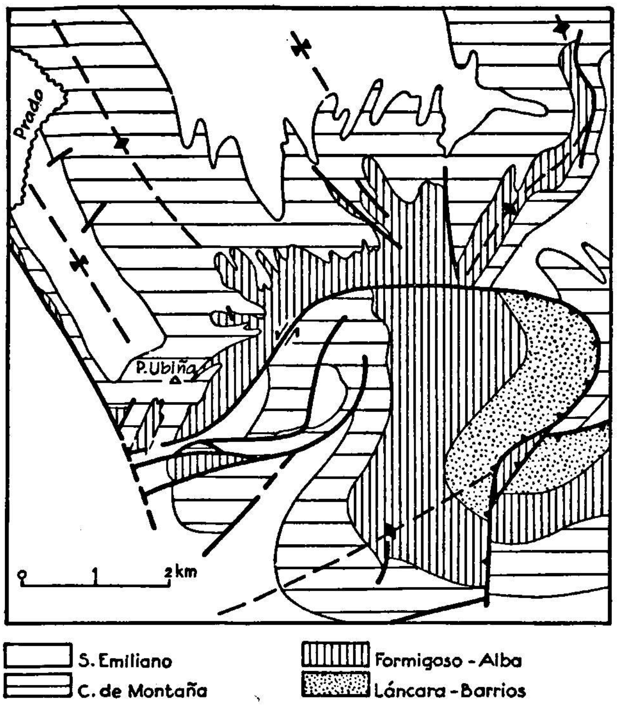

Fig. 103. The entire Ubiña zone. This figure shows the combination of thrusting and strike-slip movement along one fault.

ously horizontal shearing took place along a WSW trending wrench-fault system, causing a clock-wise rotation of the thrust block (fig. 103). The strata and the fold axes here are rather steep but flatten out to the SW. A wedgelike block of Devonian east of Villargusán has been pushed upwards by vertical movements in the wrench-fault system.

Comparison of the northern and southern part of the Ubiña zone shows that the N-S trending thrust fault in the southern part must have originated from the breaking through of a fold which was probably the continuation of one of the NNW trending folds north of it.

CENTRAL ASTURIAN COAL BASIN

Only a very small portion of this basin is covered by the present map. The rocks here are mainly of Westfalian age. The Leonides were thrusted upon these rocks along a thrust fault at the base of the Láncara.

Maps of the southern part of the Central Asturian Coal Basin (Sjerp, 1967; van Staalduinen, 1969) show that the fold axes have interfering WSW and WNW to NW strikes and generally plunge to the west; the axial planes generally dip to the north. Due to the succession of competent and incompetent beds the folds are disharmonic; parasitic and gravity folds occur frequently. A slaty cleavage, more or less parallel to the bedding was found locally in the incompetent argillaceous beds.

NARCEA ANTICLINORIUM

The Narcea anticlinorium (Julivert & Pello, 1967; Matte, 1968b) is a narrow anticlinorium with Precambrian in the core, along the north-eastern and eastern boundary of the West-Asturian-Leonese Zone. Lotze (1945, 1966), Lotze & Sdzuy (1961) and Matte (1968b) stated that the West-Asturian-Leonese Zone is the metamorphic internal zone of the Iberian orogene, and is bounded in the SW and west by the Galician-Castilian Zone (the axial zone) and in the NE and east by the Cantabrian Zone (the external zone) of which the Leonides form a part (fig. 4).

The Narcea anticlinorium is bounded in the north and NE by the Mora-Herrería unconformity or by faults, and in the SW and west by the same unconformable contact (Matte, 1968b). In the east the anticlinorium disappears below Tertiary sediments (Pastor Gómez, 1965). It is cut by the E-W trending Ocedo fault; in the region concerned the area south of this fault is called the Salce unit, named after the village of Salce. The Precambrian-Cambrian unconformity on the northern flank of the anticlinorium demonstrates a Precambrian orogeny (fig. 104). The anticlinorium itself was formed during the Hercynian orogeny. The rocks of the Mora, which are usually coloured a typical green, display at least two cleavages, and are slightly metamorphic. The presence of knick zones distinguishes the Mora schists in this region from the shales in the lower part of the Herrería, which shales are nearly identical in places. The Hercynian orogenesis caused fault-structures, very different from the structures in the Leonides. In the south these two kinds of structures are clearly separated by the Ocedo fault, but in the west a transition between the two is gradual, formed by the Muxivén zone. Because in the Leonides the thrusting took place along the base of the

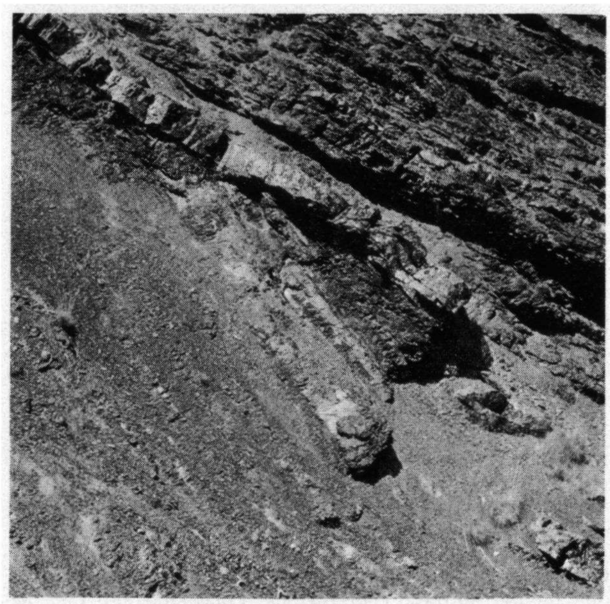

Fig. 104. Mora-Herrería unconformity near the cemetary of Irede (looking north-westwards).

Láncara, all rocks below it, Herrería and Mora, may be considered as basement. Thus the structures encountered in the Narcea anticlinorium may provide us with some information regarding the kind of structures below the thrust faults in the Leonides. First the Hercynian structures will be dealt with, and after this the Precambrian.

Structures in the Palaeozoic rocks

As Palaeozoic rocks only crop out in the Salce unit, only structures in this unit will be dealt with. The northern boundary of the Salce unit is the Ocedo fault. This fault dips 66° to the south in a mine west of Villablino, 47° to the SW in a mine NW of the Alto de Ocedo, whereas more to the east this fault stands vertically or subvertically. Along a splay of the Ocedo fault, NE of Villabandín, a slice was caught, parallel to the fault, dipping 54° overturned to the SW. This shows that the Ocedo stands steep in the deeper parts where Mora or Herrería was brought into contact with Lower Palaeozoic rocks, but in places where it brings Mora into contact with Stephanian sediments it dips only about 50° to the south. Hence it may be concluded that the Ocedo fault steepens downwards. These facts, as also the overturned slice and the NNW trending wrench-faults which displaces the Ocedo fault east of Rioscuro, suggest that the Ocedo fault is a fold-thrust (fig. 105 and 106). Berg (1962) and Sales (1968) gave field examples of this kind of faults, and Sales (1968) and Sanford (1959) created them in experiments.

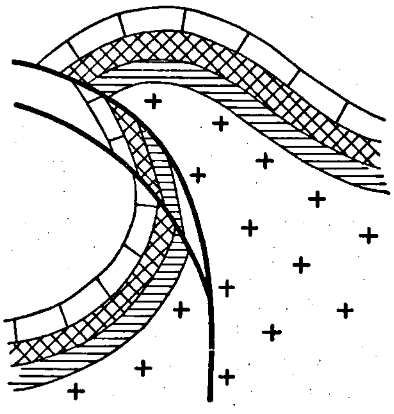

Fig. 105. Fold-thrust (after Sales, 1968).

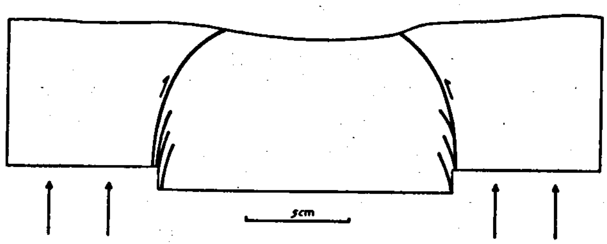

Fig. 106. Upthrusts which flatten upwards, in an experiment with beach sand (after Sanford, 1959).

A N-S trending fault cuts off the Ocedo fault near Los Barrios de Luna where the latter has little or no throw. East of this cross-fault a different structure was mapped. Contrary to the northwards directed thrusts often found, we find there, on the southern flank of the Abelgas syncline, small thrusts directed southwards along which the lowest part of the Láncara or the uppermost part of the Herrería were brought into contact with the Láncara griotte or the basal part of the Oville. Similarly to comparable structures in the Santa Lucía and La Vid in the central part of the southern flank of the Abelgas syncline, these structures may have been caused by the outwards directed movements during the development of the Abelgas syncline.

As the Ocedo fault cuts off structures in the southern flank of the Abelgas syncline and the Aralla zone, it is clear that movement along this fault must have taken place after the main folding phases; its position suggests that it played an important rôle during, and also after, the formation of the Villablino basin, cutting off as it does all structures in this basin. The faults limiting the La Magdalena basin, and probably some of the WNW trending faults in the Herrería north of Sosas del Cumbral, are presumably of the same age as the Ocedo fault (p. 215).

In its western part, where the Ocedo was displaced by the WNW trending wrench-fault, a small outcrop of Herrería was found; due to poor exposure the tectonic relations of this isolated outcrop could not be ascertained. The angle between the two NE-SW trending faults north of Rodicol and the main fault there, suggests that primary left-lateral horizontal shearing took place along these faults (fig. 114).

The remaining faults in the Salce unit can be subdivided into two groups: NNE trending faults the western block of which is generally thrown down and WNW trending faults the SW block of which is generally thrown down. Along the NNE trending faults between Villabandín and Salce, Herrería with strikes parallel to these faults crops out. These strikes are perpendicular to the generally WNW strikes encountered elsewhere in this part of the region mapped. It is most likely that these faults are normal faults which originated after the main folding phases, together with the N-S and WNW trending faults which limit the Stephanian basins.

The WNW trending, southerly directed, thrusts are of the same origin as those in the Láncara near Los Barrios de Luna and in the Santa Lucía and La Vid in the central portion of the southern flank of the Abelgas syncline, which were associated with the outwards directed movement during the development of the syncline.

Structures in the Precambrian rocks

As no detailed study was made of the Precambrian Mora Formation, only some general aspects will be dealt with here.

The succession of greenish quartzites, greywackes

Fig. 107. Mora slates with well-developed cleavage, south of Los Barrios de Luna.

and slates of the Mora Formation several thousands of metres in thickness, is only slightly metamorphic in the mapped portion of the Narcea anticlinorium. In a south-easterly direction metamorphism increases, so that locally even a mesozonal metamorphism was found at both sides of the unconformable Precambrian-Cambrian contact on the SW flank of the Narcea anticlinorium (Matte, 1968b). Just NW of the region mapped, the Precambrian is non-metamorphic and has a dark grey colour. These facts show that the isogrades cut through both Precambrian and Palaeozoic and hence the metamorphism must, at least partly, be of a Hercynian age.

The age of the cleavages and the folds is less easy to establish. The steep dips of the Cambrian Herrería Formation on both flanks of the Narcea anticlinorium show that this anticlinorium itself acquired its present form during the Hercynian folding phases. The smaller structures, however, seem to be partly of a Precambrian, partly of a Hercynian age.

There is an angular unconformity between the Mora and Herrería, the angle of which is generally rather constant over great distances (fig. 104); in the centre of the anticlinorium parallel bedded rocks of a monotonous green colour are found, hundreds of metres in thickness. This shows that the Precambrian folds have a large radius. It was possible to map the core of one of these folds in the upper course of the Río de San Miguel.

One schistosity, with mainly a WSW strike (fig. 107) in the eastern part of the Salce unit, there cuts both the Mora and the basal beds of the Herrería. This can be observed NE of Curueña, south of Irede, and along the main road south of Los Barrios de Luna. Hence this cleavage is probably Hercynian and may be related to the locally observed fracture cleavage in the younger Palaeozoic rocks (p. 194 and 198).

At other places, however, (e.g. NE of Rodicol) relatively small folds with a well-developed axial plane cleavage are cut off by the unconformity plane. In the same area the structure was difficult to analyze, owing to the lack of sufficient measurements and the interference of minor folds (often accompanied by axial plane cleavage) with axes and axial planes in all attitudes between vertical and horizontal and generally WNW and often WSW strikes. All these facts indicate, however, that the tectonic picture here may not be quite as simple as described by Matte (1968 a and b), who stated that all cleavages and minor folds were formed during the Hercynian orogenesis and are parallel to the general strikes of the Hercynian orogene.

A late phase caused WNW trending, more or less horizontal knick zones, which are characteristic of the Mora Formation in the region mapped.

STEPHANIAN BASINS

Unconformable Stephanian basins, filled with non-marine sediments, overlie the older rocks unconformably.

La Magdalena basin and its subbasins

Only the western part of the La Magdalena basin is covered by the present map. This part of the basin originated in a graben bounded by two WNW-ESE trending normal faults; steeply dragged beds near the faults show them to be more or less vertical. A narrow zone of quartzite- and schist-conglomerates, sandstones and shales is exposed east of La Urz in the deeper, northern part of the basin, whereas in the southern part only a few remains of red-schist conglomerates were preserved along the southern border fault. The red-schist cobbles (p. 182) derived from the red weathered zone of the Mora, directly beneath the lowest Prado beds.

A reversed basin asymmetry was found west of La Urz; great masses of white quartzite-conglomerates accumulated there in the deeper, southern part of the graben; they probably derived from the Herrería and Barrios quartzites. A synclinal structure could be mapped here, whereas in the eastern part of the basin a monocline developed the beds of which all dip to the north.

Small faults bordering the small subbasins, and faults parallel to these in the Herrería more to the north, all with the same strike as the two faults bordering the graben, show that the graben ended there. The two small outcrops of Prado beds below and near the church of Soto lie against the continuation of the southern border fault of the La Magdalena basin. Geophysical investigation showed that this is a vertical fault with a throw of 1.7 kilometres (Mr. P. E. Pieters, written communication).

Villabandín basin and its subbasins

The NNE trending Villabandín basin, named after the village of Villabandín, is mainly filled with coarse quartzite conglomerates. Only in the NE part have sandstones, shales and coal been found.

The basin is bounded in the NE by the Ocedo fault; an unconformable contact is exposed in the SE and east at several places. The western boundary, however, is not so clear: the beds stand overturned and dip about 60° to the west, and the Mora-Prado contact could not be found. The NNE trend of the basin can perhaps be explained by the presence of a NNE trending fault, parallel to faults more eastwards in the Herrería, which fault forms the western boundary of the basin.

The two subbasins more eastwards, developed along a splay fault of the Ocedo fault; the Prado beds there stand vertical due to drag along this fault.

Villablino basin

Only the eastern half of this basin is covered by the present map. It is bounded in the south by the Ocedo fault which dips 47° to the south in a mine at the SE end of the basin and 66° to the south in a mine west of this map. A N-S trending fault partially limits the basin in the east, and the northern boundary is an unconformity, except in two places: south of Quintanilla and north of Sosas de Laciana, where the basin is limited by faults.

Detailed structural mapping of the Prado deposits was very difficult owing to a lack of marking horizons;

Fig. 108. Anticline in the Villablino basin, west of Villaseca. In the background the old landscape of the former drainage area of the Luna (looking north-westwards).

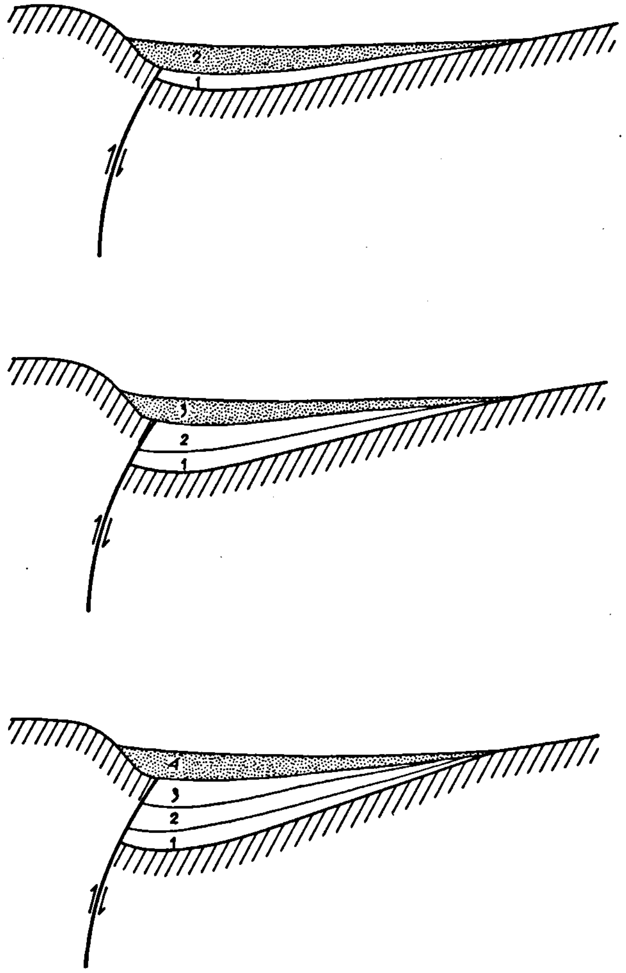

Fig. 109. Three stages in the development of the Villablino basin.

no conglomerates were even found. The absence of marking horizons is mainly caused by the rapid lateral facies changes which are typical of these sediments. The poor exposure in the areas with a mature relief (p. 186 and fig. 108) also rendered mapping difficult.

A view on the map shows that the Villablino basin is, structurally as well as sedimentologically, asymmetric. This must be the result of a gradual growth of a WNW trending intramontane depression along the Ocedo fault and the other bordering faults (fig. 109). This depression was rapidly filled with the erosion products of the uplifted areas, which were deposited in cyclothems.

The faults bordering the basin originated after the folding of the older Palaeozoic rocks during the main folding phases. These rocks broke into rigid blocks which started moving in a mainly vertical direction. Material deriving from erosion of the upthrown blocks was deposited in the intramontane basin, on the downthrown blocks. In this manner a 'basement'-sedimentary cover relationship was obtained. Although probably some of the deformation occurred during deposition, the main part occurred later and was apparently controlled by the same faults which governed the origin and growth of the basin (Helmig, 1965). These increased movements along the 'basement' faults, possibly accompanied by rotational block movements, caused a lack of space and subsequent deformation of the sediments deposited in the Villablino basin. The already existing, slightly synclinal form of the basin (fig. 109) was accentuated by the movements along the Ocedo fault, giving rise to dips ranging between 35° to 75° in the relatively undisturbed northern flank of the basin. As the Ocedo fault flattens upwards, shortening there will be greatest when movements take place along it. This explains why the upper beds of the Villablino basin have been disturbed more intensively than the lower beds.

West of Sosas de Laciana only the northern limb of the basin is exposed over a long distance in front of the Ocedo fault; east of Carrasconte only one syncline is found, the southern limb cut off by the Ocedo fault. Between Sosas de Laciana and Carrasconte, however, more folds developed (sections 10, 11 and 12; fig. 108); the strikes of these folds were clearly influenced by the movement along the NNW trending wrench-fault which displaces the Ocedo fault between Robles de Laciana and Rioscuro. In the axial portion of these folds secondary minor faults and wrinkles were found, generally parallel to the axes of the main folds. They are, however, difficult to trace in the field and are so small in scale that they have largely been omitted.

The main folds are cut by several cross-faults which can be subdivided into two groups: N-S and NE-SW trending faults. Subsurface data (kindly provided by the director of the mines) showed that at least the La Mora fault, the Valdesegadas fault and the Ladrones fault (fig. 65) are wrench-faults; the Valdesegadas fault has a horizontal displacement of even 500 metres. All these faults form angles of about 60° with the Ocedo fault, and it may therefore be concluded that they all belong to a primary wrench-fault system (Moody & Hill, 1956; fig. 114). These faults end before reaching the lower beds of the basin, which are undisturbed.

The presence of a fault along the NE boundary, as mapped by Vidal Box (1943, 1959), could not be established.

There are indications that the structures encountered at the surface are manifestations of the activity of large faults in the 'basement' in the incompetent Prado cover, just as in the Stephanian basins in the eastern part of the Cantabrian Mountains (Helmig, 1965), some boundary faults of the basin and also other faults in the 'basement' possessing the same strike and/or lying in the continuation of faults in the Villablino basin. Details of this relationship are, however, lacking.

Granite porphyry dykes intruded as feeders of a Prado deposits (p. 186). They are presumably of the series of sills, mainly restricted to the lower part of the same age as granites in the western part of the West-Asturian-Leonese Zone, namely Late-Stephanian or Early-Permian (absolute dating by Matte, 1966). Dykes of this composition were also found along the

Grajos fault, just east of the Villablino basin, and in the Mora north of Sosas de Laciana. All these dykes possibly intruded during the late-tectonic intrusive phase (de Sitter, 1964, p. 404).

MIOCENE DEPOSITS

Only a few remnants of Miocene have been preserved in the region. The outcrop nort of Robles de Laciana lies on the old landscape described on p. 186.

The Riello basin is an asymmetric former river valley, its beds generally dipping 20°–35° to the south (p. 183).

The Miocene NE of Villayuste lies at the same level as the extensive Miocene deposits on the Meseta and has the same composition; thus both deposits may be connected; the beds lie more or less horizontal.

CHAPTER V

CONCLUSIONS

Development of the structures

A very low degree of distortion was found along the thrust faults at the base of the Láncara Formation. Hubbert & Rubey (1959) explained this by the presence of abnormally high water pressures in the layer along which movement was due to take place, which movement could originate after an increased rate of sedimentation occurred in the region in which the thrusts were to be formed. A rate of sedimentation of 500 metres/1 million years is more than sufficient to attain this (Bredehoeft & Hanshaw, 1968). The same order of magnitude was reached in the Luna-Sil region during the deposition of the San Emiliano Formation, of more than 2400 metres in thickness, between Middle Namurian B and Lower Westfalian A. The Láncara was a convenient layer for high water pressures, and lay at a depth where the critical relationships between the pressure of the overburden and the lateral stresses were exceeded. The selection of the Láncara depended more on the contrast between the competencies than on absolute values (de Sitter, 1964, p. 172).

The structures mapped show clearly that the thrusts developed from breaking through of anticlines. Two minor folds (fig. 86 and 90) illustrate the development of a thrust fault very well. When the fold has formed, the maximum principal strain always occurs in its core. This strain increases very rapidly as the curvature of te fold increases, and can result in the development of conjugate shear planes in the core of the structure (Ramsay, 1967, p. 397–403; fig. 110a). The most competent beds, in this case the Barrios quartzites, break first. When the tangential longitudinal strain increases further, the strain will be relieved by movements along the shear planes. With further increments of the tangential longitudinal strain, the shear planes become thrusts, one of which passes downwards into a zone on the back-limb which is suitable for bedding-plane shearing (fig. 110b and 110c). In our case the vergence of the thrusting is towards the NNE, to that mainly those thrusts which were originally dipping to the south became of importance. The bedding-plane shearing in this growing fold, which would otherwise be distributed over many bedding-planes, was now concentrated along this single bedding-plane: the black shales at the base of the Láncara Formation (fig. 112). Only in the western part of the Aralla zone is the conjugate set present (fig. 112; section 9). As the thrust fault cuts the Barrios obliquely, or even at right angles, and is parallel to the bedding-plane at the base of the Láncara, it must be a concave fault in its lower part (fig. 110c). Models (Hafner, 1951; Hubbert, 1951) and data from oil wells in comparable structures in the Foothills Belt of Alberta in Canada (Douglas, 1950; Fox, 1959; Link, 1949), which have many properties in common with the structures in the region mapped, also show that the dips of the thrust faults decrease with the depth.

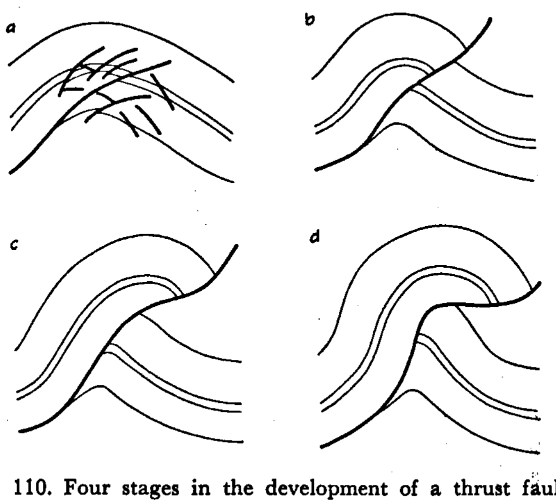

Fig. 110. Four stages in the development of a thrust fault.

The width of the syncline growing behind the thrust-faulted anticline is controlled by the very competent Barrios quartzites. Further compression results in the fault migrating upwards. Fig. 86 shows a later stage of this kind of thrust-faulting. This thrust-faulted fold also shows very clearly the relatively low amount of deformation in the lower parts of the structure; an anticline of the type encountered in front of the thrust fault in the lower part of the structure, is exposed in the western part of the Aralla zone, just in front of the Abelgas thrust fault (fig. 112; section 9).

Structures which also very well show the development of the thrusts are the Torre anticline (section 9), the structures near the Cáscaros (sections 3 and 4) and the structures on the Alto de Salgadina (section 6).

The greatest differential movement between competent and incompetent strata is found where the major thrust fault passes upwards from a bedding-plane fault at the base of the Láncara to a shear fault which passes diagonally or at right angles through the Barrios quartzites. Frictional resistance is greatest there, so that small secondary back-limb thrusts formed in order to diminish this resistance; they pass downwards and merge into the initial bedding-plane fault (fig. 112). This is illustrated by structures in the western part of the Aralla zone behind the Grajos fault, near the Cáscaros and in the Cospedal thrusts. The structures east of and around the Cáscaros and the Cospedal thrusts are very much alike. Therefore it was concluded that they must have the same, above mentioned, origin. Thus the Abelgas thrust fault merges downwards into the Rozo thrust fault (sections 1 and 2).

During a later stage in the folding the thrust faults south and west of the Babia Baja unit were involved in the folding. This folding of a thrust fault can be very well observed west of Robledo and also, though less clearly, in the northern part of the Torre zone and near Rabanal. This folding took place because the rocks involved in the thrusting abutted upon the next folds or because the thrust faults grew too steep to permit of further thrusting, and still further compression took place (de Sitter, 1964, p. 209). The folds in the Babia Baja unit could now penetrate into the lower levels of the rocks which were thrusted upon them (fig. 1d). Near Rabanal, only the Láncara and Oville underwent this subsequent folding in the continuation of the Robledo anticline and the Serronal syncline, and near Torre all rocks below the La Vid calcareous shales were refolded in the continuation of the Villasecino anticline and the Tesa syncline. As the folds in the Babia Baja unit are related to the WSW trending previously existing faults, they generally have WSW strikes, but in the west the Villasecino anticline curves to the WNW. Subsequent folding in a WSW direction is also known from the Porma and San Isidro regions (de Sitter, 1963; Evers, 1967; Sjerp, 1967).

The youngest folding has a vergence to the south, whereas the older folding and thrusting have a vergence to the north. An explanation for this has been given by van den Bosch (1969). The two NNE trending faults in the northern part of the Torre thrusts cut the folded thrust faults and are therefore younger than the youngest folding. Faults of the same relative age and with the same direction were found in the Torío area (Evers, 1967).

After the main folding phases, the deformed and consolidated rocks were locally divided into rigid blocks which started moving along their generally steep boundary faults in a mainly vertical direction. On the downthrown blocks a number of elongated basins developed in which a thick sequence of continental sediments of a Stephanian B and C age accumulated. At least the Ocedo fault, which limits the Villablino basin in the south, flattens upwards. Continued movement along the boundary faults and folding of the Stephanian sediments occurred during the Lower Permian, because Permian and Triassic sediments unconformably overlie the Stephanian sediments in the eastern part of the Cantabrian Mountains (Llopis Lladó, 1954; de Sitter & Boschma, 1966). In the Villablino basin movement along the boundary faults was accompanied by the intrusion of granites and granite porphyries.

Tectonic 'Stockwerk'

As stated above, the stratigraphic succession in the Luna-Sil region can be subdivided, according to the structural properties of the beds, into 'structural rock units'. Together they form an accumulation of disharmonic levels with different tectonic styles, a tectonic 'Stockwerk' (Cloos, 1964; Metz, 1967, p. 137). Due to the less intensive style of folding in the Babia Baja unit, this tectonic 'Stockwerk' is less pronouncedly developed there.

As the Mora and Herrería Formations are not involved in the thrusting, they may be considered as basement. Where these formations crop out, they generally show fault tectonics which for a large part originated during a later stage of the deformation (p. 213). During the thrusting the basement below the Babia Alta and Luna units was only locally disturbed. These structures will be dealt with in the following section.

The black shales at the base of the Láncara were best suited for bedding-plane shearing: all large thrust faults are situated at this level. Only a very narrow disturbance zone was found along the thrust faults (fig. 85), although 5000 to 6000 metres of sediments were involved in the thrusting. The Láncara and the lower part of the Oville were folded parallel to the thrust faults.

The upper part of the Oville and the Barrios formed a very competent level. The thrust faults generally cut the Barrios quartzites at a steep angle or even perpendicularly, so that a great shearing stress was produced there (sections 1–4). In order to release this stress imbricate structures were formed behind the thrusts (sections 4 and 6; fig. 112). Due to the very competent character of the Barrios quartzites, these rocks determined the shape and width of most folds in the region mapped. Because they are difficult to fold, these quartzites have been intensively broken by numerous small cross-faults. Parts of the crests of the broken anticlines from which the thrust faults developed, were locally preserved along these faults (sections 3, 4, 6, 9).

The black shales in the lower of the Formigoso form an incompetent layer between two competent levels each with its own tectonic style. The Formigoso has generally been intensively folded so that its original thickness could not be established. These minor folds are not always simple parasitic folds; in the upper part, where thin sand layers are intercalated, other minor folds represent subsequent refolding (fig. 87). Many of the cross-faults in the competent beds on both sides of the black shales end in the black shales, just as some of the smaller thrusts which were directed parallel to the bedding-plane, so that movement could be distributed along the shale flakes (sections 4 and 9). Moreover, the larger thrust faults were locally directed parallel to the bedding-plane over some distance (NW of La Riera). Especially in the hinge zones of folds a clear detachment took place between the Upper Formigoso and the top of the Barrios (sections 1, 2, 7, 9). Locally a slight fracture cleavage was found.

The upper part of the Formigoso, the San Pedro and the basal limestones of the La Vid constitute the next competent succession. This was the best suited level for gravity folds (sections 1, 7, 11; fig. 67). At this level the beds are no longer more or less parallel to the thrust faults but were folded into an anticline and a syncline behind the thrusts (sections 1, 2, 7, 9).

The highly mobile calcareous shales of the La Vid constitute the level which caused the most important disharmonic structures in the beds above the Láncara. Structures in which all younger rocks were involved, were formed practically independently of structures involving all rocks between the Láncara and the La Vid shales. The shales themselves were generally intensively deformed by bedding-plane shearing, crumpling and small-scale folding, the latter two often being restricted to the upper part and the former to the lower part of these shales. At several places fracture cleavage has been found. The map and the sections show that nearly all folds and faults encountered in the competent beds at both sides of the shales end in these shales; the Torre thrusts and probably the Abelgas thrust as well end in them, no rocks younger than the lowermost La Vid shales being found in front of the latter. Similarly to the thrusts ending in the black shales of the Formigoso, it appears that these thrusts, too, were directed parallel to the bedding-plane of the shales, and that the shearing movement was distributed along the shale flakes. Up to a level of approximately 30–50 metres above the base of these shales, the shales are generally parallel to the La Vid limestones (fig. 111a); above this level they have been intensively disturbed. Thus it appears that the detachment-plane between the competent beds above and below the calcareous shales lies at that level. The box-folds in the upper part of the La Vid and the lower part of the Santa Lucía provide a good illustration of the behaviour of the La Vid shales (fig. 84 and 88). According to a theory developed by Ghosh (1968) the box-folds were formed during an early stage of folding. During their formation in the lower part of the Santa Lucía, the shape of the beds in the upper part of the La Vid had to be accommodated to the shape of the growing fold. These beds were folded to a complex mass of small crumples and contortions in the core of the main fold. The adjacent competent La Vid limestones did not shorten by the same amount, which caused the development of a discontinuity horizon

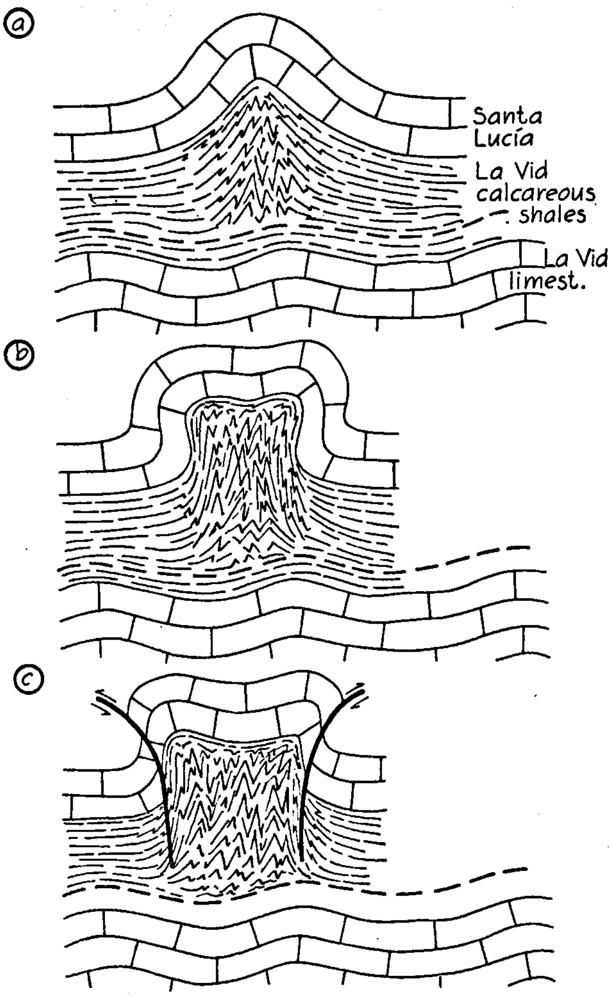

Fig. 111. Three stages in the development of a box-fold.

in the mobile La Vid calcareous shales at approximately 30–50 metres from the base.

The crumpled mass in the inner part of the fold grew more and more resistant during the continuing folding, and restricted the development of the fold (Ghosh, 1968). Due to these circumstances the shape of the folds was modified, the hinge zone of the fold having to move upwards (fig. 111b). At places where the limbs of the resulting box-fold broke, this mass in the hinge zone could move even further upwards and respond to the lack of space in the core of the fold (Ramsay, 1967, p. 415–421; fig. 111c).

The Santa Lucía limestones and the red crinoidal limestones of the La Vid at their base, were folded almost independently of the competent beds below the La Vid calcareous shales. In the Babia Alta unit box-folds are characteristic of this level (fig. 88 and 89). Other small folds, though somewhat larger than the box-folds, and small thrusts are common. One of these small folds (fig. 90) illustrates the mechanism of 'back-limb thrusting' very well.

The Huergas silty shales and sandstones also permitted the development of disharmonic structures (Quejo syncline and eastern part of the Abelgas

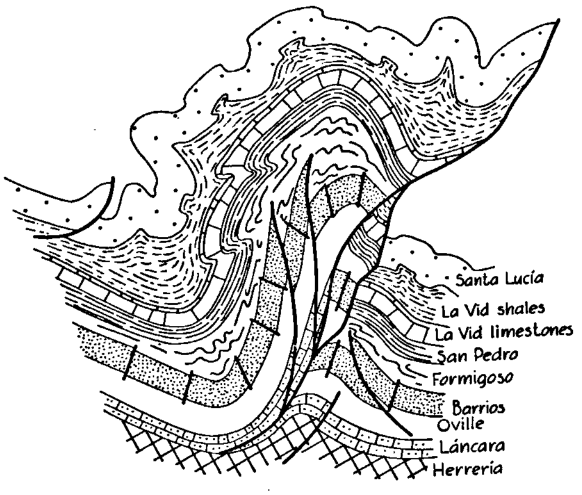

Fig. 112. Sketch of a thrust-faulted anticline in which the different properties of the 'tectonic rock units' have been combined.

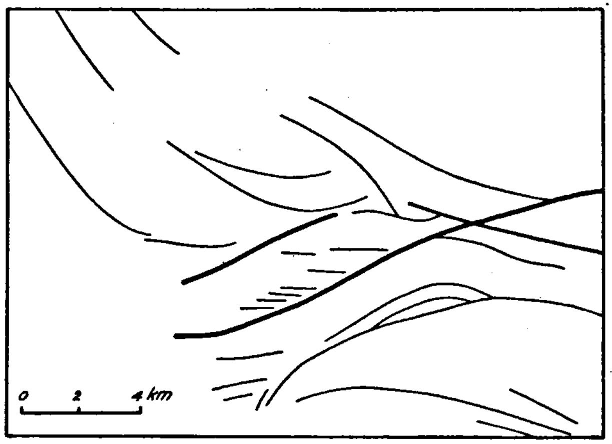

Fig. 113. Sketch of the deviation in the strikes of fold axes and thrust faults near the Grajos fault and Corralines fault.

syncline). Locally a fracture cleavage was developed (fig. 76).

The relatively few outcrops of the competent succession of the Portilla, Nocedo, Ermita and Alba Formations do not show characteristic structures. Some of the structures end in the relatively incompetent Huergas. Concentric folds developed up to a level just above the Alba. Above this level complications occur.

The well-bedded platy limestones of the Caliza de Montaña were very well adapted for the development of parasitic folds on the flanks and crumpling in the core of some of the synclines (fig. 94 and 95).

In the relatively thin competent limestone and sandstone beds in the incompetent San Emiliano Formation of the Babia Baja unit, parastic and gravity folds were found on the flanks and crumpling in the core of the folds (fig. 101 and 102). Disharmonic structures between these competent beds, caused by the incompetent shales between them, are frequent. The axis of the Serronal syncline lies where the competent limestone beds possess their smallest thickness (p. 209).

Fig. 112 a combination of all properties of the different 'tectonic rock units' up to the Huergas.

Interference of WSW trending structures with the general strikes

The WSW trending structures crossing the general WNW fold direction are most impressive phenomena in the Luna-Sil region (fig. 113). Comparable structures in the Jura Mountains were interpreted by Lloyd (1964) as resulting from an already existing fault in the hidden basement, which was reactivated and acted as a wrench-fault during the folding of the sedimentary cover. Other comparable structures have been mapped in the Pisuerga region in the SE part of the Cantabrian Mountains (de Sitter, 1957) and on the South Island of New Zealand (Schofield, 1960). The map shows that the Grajos fault (a left-lateral wrench-fault, with a net horizontal slip of 800 metres near Villasecino, along which fault considerable vertical movements also took place locally) determined the directions of the adjacent structures. During the folding it separated the region into two blocks in which, more or less independently, thrusts were formed.

Wrench-faults normally form an angle of 30° with the main stress direction (Moody & Hill, 1956), so that there are two possible wrench directions (fig. 114). On comparing this picture with the geological map, it is evident that the N-S trending Lumajo wrench-fault is the complementary right-lateral wrench-fault of the WSW trending system. The primary-fold direction is also exactly in accordance with the theoretical picture. Anderson (1942) estimated theoretically how the stress field in a thrusting mass around an existing fault is deviated and causes a deflection of the thrust

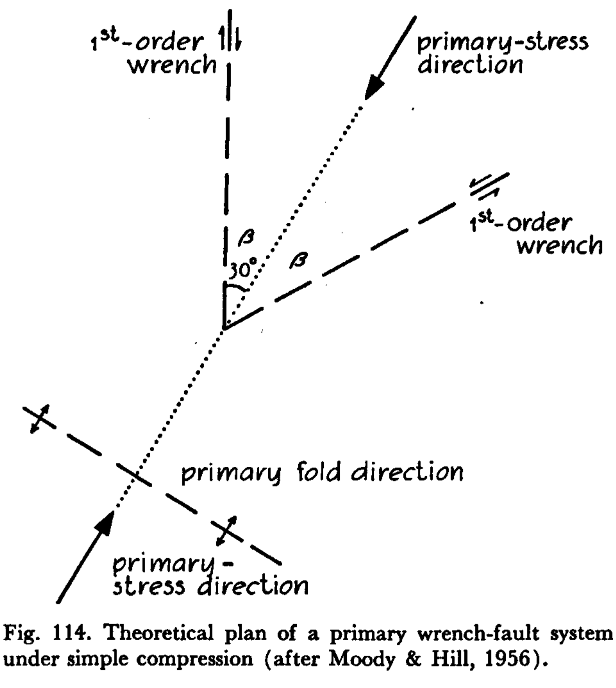

Fig. 114. Theoretical plan of a primary wrench-fault system under simple compression (after Moody & Hill, 1956).

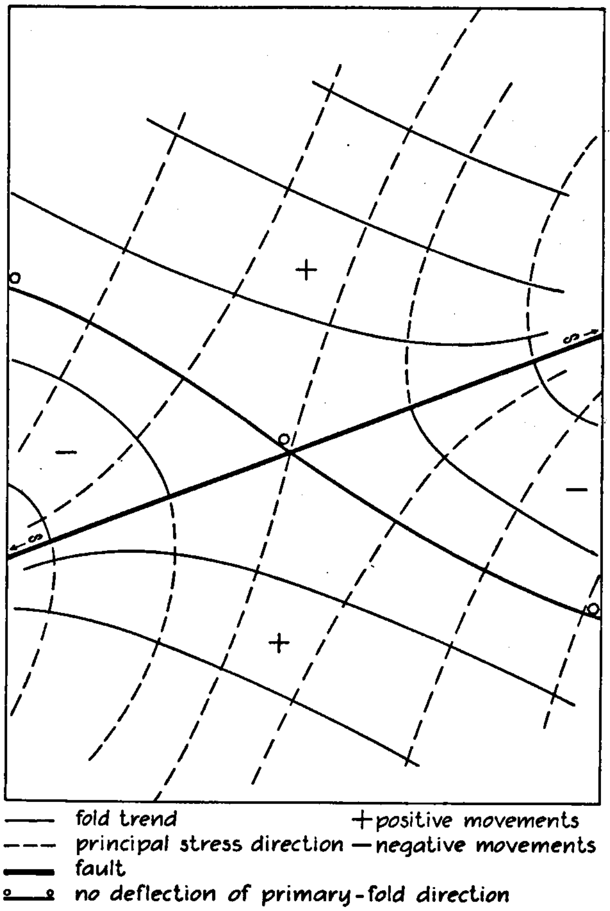

Fig. 115. Principal directions of pressure and fold trends in a detached part of the crust around a fracture developed under shearing stress (after Anderson, 1952 and Chinnery, 1961).

planes along it (fig. 115). The maximum stress exists at both ends of the fault, whereas no stress, neither perpendicular nor parallel to the fault, exists near its central part. Chinnery (1961, 1966) arrived at the same conclusion. He also found that the northern and southern quadrants move upwards and the southern and eastern quadrants downwards (fig. 115).

Comparing this theoretical case with the structures in the Luna-Sil region, we find that in the middle part of the Grajos fault the Rozo thrust fault was neither displaced nor deflected along the Grajos fault, but cuts it obliquely (zero line in fig. 115). Towards both ends of the Grajos fault, however, the thrusts and other structures run exactly parallel to it. The strike of the Villasecino anticline near the zero line, too, is no longer determined by the Grajos fault, but curves to the WNW. It has been observed that the southern and northern quadrants of fig. 113 (Luna unit and Torre zone) moved relatively upwards, whilst the western and eastern quadrants (SW part of the Babia Baja unit and the Meroy zone) moved relatively downwards.

Riedel shears and the arrangement of folds *en échelon* in the zone between the two main slip surfaces in the Babia Alta unit (Corralines zone), indicate that this zone is bounded by two wrench-faults (the Grajos fault and the Corralines fault) to which the main stress was applied obliquely (p. 208). Thus it may be concluded that the interference of WSW and WNW trending structures may be explained in the manner proposed by Anderson and Chinnery.

North of the Alto de los Grajos the generally WSW trending Grajos fault runs over some distance in a W-E direction, parallel to nearby splay faults. This may be explained by the fact that splay faults (and also second-order shears) are formed at the ends of wrench-faults, where the greatest shearing stresses accumulated (Chinnery, 1966). Thus the Grajos fault was possibly first restricted to a level below the Barrios and later extended itself upwards (fig. 116). The small fault in the core of the Robledo anticline did not penetrate to levels higher than the Oville (section 1).

Besides the Grajos fault, other WSW trending faults affecting the deformation in the Babia Alta and Luna units were also mapped in the Babia Baja unit: one east of Riolago and one in front of the eastern part of the Rozo thrust fault. The influence of the latter on the deformation of the Luna unit is of special interest. Further to the east this is a very important fault along which a wedge of Devonian rocks was pushed upwards (van Staalduinen, 1969). The Rozo thrust fault there curves to a WSW strike, and remains parallel to this fault over a long distance. Furthermore, several WSW trending structures were mapped in the Luna unit in the continuation of the WSW trending fault. Several WSW trending faults cut the Barrios and San Pedro of the Aralla zone, the strikes in the Abelgas syncline curve from WNW to WSW and folds with WSW trending axes were found on the southern flank of the Abelgas syncline. Possibly the WSW strike of the Láncara WSW of these structures is related to it as well. All these features indicate an influence of the WSW trending fault in the Babia Baja unit on the structures in the Luna unit in the same manner as the Grajos fault, but much less intensively.

Along the Lumajo fault a slight deviation of strikes was also observed (fig. 93). The strikes of the Precambrian basement there are parallel to the Lumajo fault,

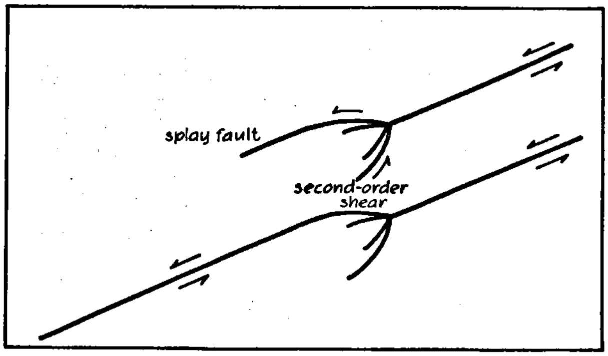

Fig. 116. Two stages in the development of a wrench-fault (after Chinnery, 1966).

and this might have stimulated the shearing in that direction.

It may be concluded that a stress perpendicular to the basin axis has formed the folds and thrust faults with WNW to NNW strikes, while simultaneously local stress fields developed perpendicularly to previously existing WSW and N-S trending faults, especially the WSW trending Grajos fault, which gave rise to the WSW and N-S trending folds and thrust faults.

REFERENCES

Adrichem Boogaert, H. A. van, 1967. Devonian and Lower Carboniferous conodonts of the Cantabrian Mountains (Spain) and their stratigraphic application. Leidse Geol. Med., 39, p. 129–192.

Adrichem Boogaert, H. A. van, Breimer, A., Krans, Th. F. & Sjerp, N., 1963. A new stratigraphic interpretation of Palaeozoic sections in the region between San Isidro Pass and Tarna Pass (province of León, Spain). Notas Comun. Inst. Geol. Min. España, 70, p. 131–135.

Almela, A., 1951. Delimitación del Carbonífero de la zona La Robla – Vegarienza (León). Bol. Inst. Geol. Min. España, 63, p. 221–256.

Almela, A., García-Fuente, S. & Ríos, J. M., 1956. Explicación de la hoja no. 52 Proaza (Asturias). Mapa geol. España escala 1:50.000. Inst. Geol. Min. España, núm. 264.

Alvarado, A. de, 1951. Límites estratigráficos del Carbonífero del N-W de León. 3e Cong. Strat. Géol. Carbonifère, Heerlen, 1951, p. 5–12.

Alvarez-Ramis, C., 1965. Primera aportación al estudio de los 'Sphenopteris' del Carbonífero español. Est. geol., 21, p. 1–141.

Anderson, E. M., 1942. The dynamics of faulting and dyke deformation with applications to Britain. Oliver and Boyd, Edinburgh and London, 191 p.

Ashgirei, G. D., 1963. Strukturgeologie. Veb. Deutscher Verlag Wiss., Berlin, 572 p.

Bäcker, H., 1959. Die Furada-Zone in Zentral-Asturien und Nord-León (Eine feinstratigraphische Untersuchung der Grenzschichten Gotland-Devon). Unpublished thesis, geol. dep. univ. Münster.

Barrois, C., 1882. Recherches sur les terrains anciens des Asturies et de la Galice. Mém. Soc. Géol. du Nord, 2 (1), 630 p.

Berg, G., 1944. Vergleichende Petrographie oolitischer Eisenerze. Arch. Lagerstättenforschung, 76, p. 1–128.

Berg, R. R., 1962. Mountain flank thrusting in Rocky Mountain foreland, Wyoming and Colorado. Am. Assoc. Petroleum Geologists Bull., 46, p. 2019–2032.

Billings, M. P., 1947. Structural Geology. Prentice-Hall, Inc., New York, 473 p.

Bitterli, P., 1963. Aspects of the genesis of bituminous rock sequences. Geol. & Mijnb., 42, (6), p. 183–201.

Borchert, H., 1953. Die Bildungsbedingungen mariner Eisenerzlagerstätten. Chemie der Erde, 16, p. 49–74.

Bosch, W. J. van den, 1969. The relationship between orogenesis and sedimentation in the SW part of the Cantabrian Mountains (NW Spain). Leidse Geol. Med., 44, in press.

Boschma, D. & Staalduinen, C. J. van, 1968. Mappable units of the Carboniferous in the southern Cantabrian Mountains. Leidse Geol. Med., 43, p. 221–232.

Braun, H., 1964. Zur Entstehung der marin-sedimentären Eisenerze. Claustaler Hefte zur Lagerstättenkunde und Geochemie der min. Rohstoffe, 2, p. 1–133.

Bredehoeft, J. D. & Hanshaw, B. B., 1968. On the maintainance of anomalous fluid pressures. Geol. Soc. America Bull., 79, p. 1097–1122.

Brouwer, A., 1964. Deux facies dans le Devonien des Montagnes cantabriques méridionales. Breviora Geol. Astúrica, 8 (1–4), p. 3–10.

—, 1967. Le Dévonien inférieur des Montagnes cantabriques (Espagne du Nord-Ouest). Mém. Bureau Rech. Géol. Minière, 33, p. 197–203.

Brouwer, A. & Ginkel, A. C. van, 1964. La succession carbonifère dans la partie méridionale des Montagnes cantabriques (Espagne du Nord-Ouest). C. R. 5ème Cong. Int. Strat. Géol. Carbonifère, T.1, p. 307–319.

Budinger, P. & Kullmann, J., 1964. Zur Frage von Sedimentationsunterbrechungen im Goniatiten- und Conodonten-führenden Oberdevon und Karbon des Kantabrischen Gebirges (Nordspanien). N. Jahrb. Geol. Paläont., Mh., 1964, p. 414–429.

Capdevila, R., 1965. Sur la géologie du Précambrian et du Paléozoique dans la region de Lugo et la question des plissements assyntiques et sardes en Espagne. Notas Comun. Inst. Geol. Min. España, 80, p. 157–174.

—, 1967. Extension du métamorphisme régional hercynien dans le Nord-Ouest de l'Espagne (Galice orientale, Asturies, León). C. R. Somm. Seances Soc. Géol. France, 1967, 7, p. 277–279.

Chinnery, M. A., 1961. The deformation of the ground around surface faults. Seismol. Soc. America Bull., 51, p. 355–372.

—, 1966. Secondary faulting. Can. Jour. Earth Sci., 3, p. 163–190.

Cloos, E., 1964. Appalachenprofil 1964. Geol. Rundsch., 54, p. 812–834.

Comte, P., 1959. Recherches sur les terrains anciens de la Cordillère cantabrique. Mem. Inst. Geol. Min. España, 60, 440 p.

Concha, S. de la & Jorissen, R. R., 1961. Nota explicativa del mapa de los yacimientos carboníferos de España, escala 1:2.500.000. Notas Comun. Inst. Geol. Min. Espanña, 64, p. 87–92.

Davis, E. F., 1918. The radiolarian cherts of the Franciscan group. Univ. California Publ. Dept., Geol. Sci. Bull., 11, p. 235–432.

Delépine, G., 1943. Les faunes marines du Carbonifère des Asturies (Espagne). Mém. Acad. Sci. Inst. France, 66, 3, p. 1–122.

Douglas, R. J. W., 1950. Callum Creek, Langford Creek and Gap map-areas, Alberta. Geol. Surv. Canada, mem. 255.

Drot, J. & Matte, P., 1967. Sobre la presencia de capas del Devoniano en el límite de Galicia y León (N.W. de España). Notas Comun. Inst. Geol. Min. España, 93, p. 87–92.

Evers, H. J., 1967. Geology of the Leonides between the Bernesga and Porma rivers, Cantabrian Mountains, NW Spain. Leidse Geol. Med., 41, p. 83–151.

Färber, A. & Jaritz, W., 1964. Die Geologie des westasturischen Küstengebirges zwischen San Estaban de Pravia und Ribadeo (NW-Spanien). Geol. Jahrb., 81, p. 679–738.

Fairbridge, R. W., 1967. Phases of diagenesis and authigenesis.

In: Diagenesis in sediments, G. Larsen & G. V. Chilingar, ed., 1967, Elseviers, Amsterdam, p. 19–89.

Fox, F. G., 1959. Structure and accumulation of hydrocarbons in southern Foothills, Alberta, Canada. Am. Assoc. Petroleum Geologists Bull., 43, p. 992–1025.

García-Fuente, S., 1959. Explicación de la hoja no. 77 La Plaza (Teverga) (Asturias). Mapa geol. de España, escala 1:50.000. Inst. Geol. Min. España, núm. 305 H.

Garfunkel, Z., 1966. Problems of wrench faults. Tectonophysics, 3 (5), p. 457–473.

Ghosh, S. K., 1968. Experiments of buckling of multilayers which permit interlayer gliding. Tectonophysics, 6 (3), p. 207–249.

Ginkel, A. C. van, 1965. Carboniferous fusulinids from the Cantabrian Mountains (Spain). Part II. Spanish Carboniferous Fusulinids and their significance for correlation purposes. Leidse Geol. Med., 34, p. 173–225.

Gómez de Llarena, J. & Rodríguez Arango, C., 1946. Las tacañas fosilíferas (coal balls) de la mina Rosario, de Truébano (León). Notas Comun. Inst. Geol. Min. España, 16, p. 215–236.

—, 1948. Datos para el estudio geológico de la Babia Baja (León). Bol. Inst. Geol. Min. España, 61, p. 79–206.

Hafner, W., 1951. Stress distribution and faulting. Geol. Soc. America Bull., 62, p. 373–398.

Helmig, H. M., 1965. The geology of the Valderrueda, Tejerina, Ocejo and Sabero Coal Basins (Cantabrian Mountains, Spain). Leidse Geol. Med., 32, p. 75–149.

Hernández Sampelayo, P. & Almela, A., 1942. Mancha estefaniense de Sosas del Cumbral (Murias de Paredes-León). Notas Comun. Inst. Geol. Min. España, 9, p. 28.

Hernández Sampelayo, P. & A., 1947. Fauna carbonífera de Villablino (León). Notas Comun. Inst. Geol. Min. España, 17, p. 1–24.

Higgins, A. C., 1962. Conodonts from the 'griotte' limestone of North-West Spain. Notas Comun. Inst. Geol. Min. España, 65, p. 5–22.

Higgins, A. C., Wagner-Gentis, C. H. T. & Wagner, R. H., 1963. Basal Carboniferous Strata in Part of Northern León, NW Spain. Stratigraphy, Conodont and Goniatite Faunas. Bull. Soc. belge géol. paléont. hydrol., 72, (2), p. 205–260.

Hollmann, R., 1962. Ueber Subsolution und die 'Knollenkalke' des Calcare Ammonitico Rosso Superiore im Monte Baldo (Malm; Norditalien). N. Jahrb. Geol. Paläont., Mh, 1962, p. 163–179.

Houten, F. B. van, 1948. Origin of red-banded early Cenozoic deposits in Rocky Mountain region. Am. Assoc. Petroleum Geologists Bull., 32, p. 2083–2126.

Hubbert, M. K., 1928. The direction of the stresses producing given geologic strains. Jour. Geology, 36, p. 75–84.

—, 1951. Mechanical basis for certain familiar geological structures. Geol. Soc. America Bull., 62, p. 355–372.

Hubbert, M. K. & Rubey, W. W., 1959. Role of fluid pressure in mechanics of overthrust faulting. Geol. Soc. America Bull., 70, p. 115–206.

Internationale Kommission für Geochronologie, 1968. Vergleichende Tabelle der neuesten veröffentlichten geochronologischen Skalen für die Zeiten des Phanerozoikums. Geologie, 17 (2), p. 210–215.

Julivert, M. & Martínez García, E., 1967. Sobre el contacto entre el Cámbrico y el Precámbrico en la parte meridional de la Cordillera Cantábrica y el papel del Precámbrico en la orogénesis herciniana. Acta Geol. Hisp., 2 (5), p. 107–110.

Julivert, M. & Pello, J., 1967. Las dos etapas principales de deformación herciniana en la Cordillera Cantábrica y el trazado de sus estructuras. Acta Geol. Hisp., 2 (4), p. 77–81.

Kegel, W., 1929. Das Gotlandium in den kantabrischen Ketten Nordspaniens. Zeitschr. deutsch. geol. Ges., 81, p. 35–62.

Koopmans, B. N., 1962. The sedimentary and structural history of the Valsurvio Dome (Cantabrian Mountains, Spain). Leidse Geol. Med., 26, p. 121–232.

Krumbein, W. C. & Sloss, L. L., 1963. Stratigraphy and Sedimentation, 2nd ed. W. H. Freeman and comp., San Francisco and London, 660 p.

Kullmann, J., 1961. Die Goniatiten des Unterkarbons im Kantabrischen Gebirge (Nordspanien). N. Jahrb. Geol. Paläont., Abh., 113, p. 219–326.

—, 1962. Die Goniatiten der Namur-Stufe (Oberkarbon) im Kantabrischen Gebirge, Nordspanien. Akad. Wiss. Lit., Abh. math.-naturw. Kl., 6, p. 1–119.

—, 1963. Die Goniatiten des Unterkarbons im Kantabrischen Gebirge (Nordspanien) II Paläontologie. N. Jahrb. Geol. Paläont., Abh., 116, p. 269–324.

—, 1966. Goniatiten – Korallen – Vergesellschaftungen im Karbon des Kantabrischen Gebirges (Nordspanien). N. Jahrb. Geol. Paläont., Abh., 125, p. 443–466.

Link, T. A., 1949. Interpretations of Foothills structures, Alberta, Canada. Am. Assoc. Petroleum Geologists Bull., 33, p. 1475–1501.

Llopis Lladó, N., 1954. Sobre la tectónica de la cuenca carbonífera de Asturias. Est. Geol., 21, p. 79–101.

—, 1955. Estudio geológico del reborde meridional de la cuenca carbonífera de Asturias. Monografías geol., 3. Inst. geología aplicada, Oviedo, 177 p.

Lloyd, A. J., 1964. Cover Folding in the Sonmartel Chain (Jura Neuchatelois). Geol. Rundsch., 53, p. 551–580.

Lotze, F., 1945. Zur Gliederung der Varisziden der Iberischen Meseta. Geotekt. Forsch., 6, p. 78–92.

—, 1960. El Precambrico de España. Notas Comun. Inst. Geol. Min. España, 60, p. 227–240.

—, 1962. Ueber pleistozäne Vergletscherungen in der Valnera Gruppe (östliches Kantabrisches Gebirge). N. Jahrb. Geol. Paläont., Mh., 1962, p. 377–381.

—, 1966. Die variszischen Gebirgszusammenhänge im westlichen Europa. Zeitschr. deutsch. geol. Ges., 115 (2–3), (1963), p. 919.

—, 1968. Schlusswort: Der gesammtektonische Rahmen. Geotekt. Forsch., 27, p. 147–154.

Lotze, F. & Sdzuy, K., 1961. Das Kambrium Spaniens, Teil 1, Stratigraphie. Akad. Wiss. Lit., Abh. math.-naturw. Kl., 6, p. 283–498.

Mabesoone, J. M., 1959. Tertiary and Quarternary sedimentation in a part of the Duero basin (Prov. Palencia, Spain). Leidse Geol. Med., 24, p. 31–180.

Matte, P., 1964. Remarques préliminaires sur l'allure des plis hercyniens en Galice orientale. C. R. Acad. Sci. France, 259, 12, p. 1981–1984.

—, 1966. La schistosité primaire dans l'arc hercynien de Galice; variation de son pendage parallèlement et perpendiculairement aux structures et role des phases de déformation ultérieures. Etages tect., colloque de Neuchatel, 18–21 avril 1966, p. 243–251.

—, 1968a. Précisions sur le Précambrien supérieur schisto-gréseux de l'Ouest des Asturies. Comparaison avec les autres affleurements précambriens du Nord-Ouest de l'Espagne. Rev. Géog. Phys. Géol. Dyn. (2), 10 (3), p. 205–211.

—, 1968b. La structure de la virgation hercynienne de Galice (Espagne). Thesis, Extr. Trav. lab. géol. fac. sci. Grenoble, t. 44.

McKelvey, V. E., 1967. Phosphate deposits. U.S. Geol. Surv. Bull. 1252-D.

McKelvey, V. E., William, J. S., Sheldon, R. P., Cressman, E. R., Cheney, T. M., Swanson, R. W., 1959. The Phosphoria, Park City, and Shedhorn formation in the western phosphate field. U.S. Geol. Surv. Prof. Paper, 313-A, p. 1–47.

Meer Mohr, C. G. van der, 1969. The stratigraphy of the Cantabrian Láncara Formation between the Luna river and the Esla river in the Cantabrian Mountains, Spain. Leidse Geol. Med., 43, p. 233–316 (in press).

Meer Mohr, C. G. van der & Okulitch, V. J., 1967. On the occurrence of Scyphomedusa in the Cambrian of the Cantabrian Mountains (NW Spain). Geol. & Mijnb., 46, p. 361–362.

Meer Mohr, C. G. van der, & Schreuder, G. A. N. H., 1967. On the petrography of the Láncara Formation from the Sierra de la Filera (Spain). Leidse Geol. Med., 38, p. 185–189.

Meischner, K. D., 1964. Allodapische Kalke, Turbidite in Riffnahen Sedimentationsbecken. In: Bouma, A. H. & Brouwer, A.: Turbidites-Developments in Sedimentology, 3, Elseviers, Amsterdam, p. 156–191.

Metz, K., 1967. Lehrbuch der tektonischen Geologie, 2nd ed. Ferdinand Enke Verlag, Stuttgart, 357 p.

Moody, J. D. & Hill, M. J., 1956. Wrench-fault tectonics. Geol. Soc. America Bull., 67, p. 1207–1246.

Nollau, G., 1966. El desarrollo estratigráfico del Paleozoico en el oeste de la provincia de León (España). Notas Comun. Inst. Geol. Min. España, 88, p. 31–48.

Nussbaum, F. & Gigax, F., 1953. La glaciación cuaternaria en la Cordillera Cantábrica. Est. Geog., 7 (nr. 51), p. 261–270.

Oder, C. R. L. & Bumgarner, J. G., 1961. Stromatolitic Bioherms in the Maynardville (Upper Cambrian) Limestone, Tennessee. Geol. Soc. America Bull., 72, p. 1021–1028.

Oele, E., 1964. Sedimentological Aspects of four Lower-Paleozoic Formations in the northern part of the province of León (Spain). Leidse Geol. Med., 30, p. 1–99.

Pannekoek, A. J., 1968. Geomorphology of the Iberian Peninsula. Syllabus, dep. geol. univ. Leiden.

Pastor Gómez, V., 1962. Probable area precambriana al NO de León. Notas Comun. Inst. Geol. Min. España, 67, p. 71–79.

—, 1963. Explicación de la hoja núm. 129 La Robla (León). Mapa geol. España, escala 1:50.000. Inst. Geol. Min. España núm. 348H.

—, 1965. El momento actual de los estudios geológicos en la provincia de León. Bol. Inst. Geol. Min. España, 76, p. 3–21.

Peacock, M. A., 1930. On chlorophacite and palagonite. Geol. Mag., 67, p. 170–178.

Poll, K., 1963. Zur Stratigraphie des Altpaläozoikums von Belmonte (Asturien/Nordspanien). N. Jahrb. Geol. Paläont., Abh., 117, p. 235–250.

Prado, C. de, 1850. Sur les terrains de Sabero et ses environs (León). Bull. Soc. Géol. France, 2ème sér., 7, p. 137–155.

Prado Enriquez, Antonio de, 1781. Relación de los minerales reconicidos en las immediaciones del Principado de Asturias en la provincia del Bierzo, Valle de Valdeorras y Galicia.

Rácz, L., 1964. Carboniferous calcareous algae and their associations in the San Emiliano and Lois-Ciguera Formations (Prov. León, NW Spain). Leidse Geol. Med., 31, p. 1–112.

Radig, F., 1962. Ordovizium/Silurium und die Frage prävariszischer Faltungen in Nordspanien. Geol. Rundsch., 52, p. 346–357.

Ramsay, J. G., 1967. Folding and fracturing of rocks. McGraw-Hill Book Company, New York, 1967, 568 p.

Riemer, W., 1963. Entwicklung des Paläozoikums in der südlichen Provinz Lugo (Spanien). N. Jahrb. Geol. Paläont,. Abh., 117, p. 273–285.

Rupke, J., 1965. The Esla Nappe, Cantabrian Mountains (Spain). Leidse Geol. Med., 32, p. 1–74.

Rusnak, G. A., 1960. Some observations of recent oolites. Jour. Sed. Petrology, 30, p. 471–480.

Sales, J. K., 1968. Crustal mechanics of Cordelleran foreland deformation: a regional and scale-model approach. Am. Assoc. Petroleum Geologists Bull., 52, p. 2016–2044.

Sanford, A. R., 1959. Analytical and experimental study of simple geologic structures. Geol. Soc. America Bull., 70, p. 19–52.

Schofield, J. C., 1960. Some theoretical structures associated with transcurrent faulting applied to the Alpine Fault. New Zealand Jour. Geol. Geophys., 3, p. 461–466.

Seilacher, A., 1967. Bathymetry of trace fossils. Marine Geology, 5 (5–6), p. 413–428.

Sitter, L. U. de, 1957. The structural history of the SE corner of the Palaeozoic core of the Asturian Mountains. N. Jahrb. Geol. Paläont., Abh., 105, p. 272–284.

—, 1962a. El Precambriano de la cadena cantabrica. Notas Comun. Inst. Geol. Min. España, 67, p. 145–146.

—, 1962b. The structure of the southern slope of the Cantabrian Mountains; explanation of a geological map with sections scale 1:100.000. Leidse Geol. Med., 26, p. 255–264.

—, 1964. Structural Geology, 2nd ed. McGraw-Hill Book Company, New York, San Francisco, Toronto, London, 551 p.

—, 1966. De relatie tussen orogenese en epirogenese in het Cantabrische gebergte (voordracht). Verslag: Kon. Ned. Acad. Wet., 75 (7), p. 114–120.

Sitter, L. U. de, & Bosch, W. J. van den, 1968. The structure of the SW part of the Cantabrian Mountains. Leidse Geol. Med., 43, p. 213–216.

Sitter, L. U. de & Boschma, D., 1966. Explanation geological map of the Palaeozoic of the Southern Cantabrian Mountains 1:50.000, sheet 1, Pisuerga. Leidse Geol. Med., 31, p. 191–238.

Sjerp, N., 1967. The geology of the San Isidro-Porma Area (Cantabrian Mountains, Spain). Leidse Geol. Med., 39, p. 55–128.

Skempton, A. W., 1966. Some observations on tectonic shear zones. Proc. 1st cong. Int. Soc. Rock Mechanics, Lisbon, 1966, p. 329–335.

Sluiter, W. J., 1964. Sedimentologie en Geomorfologie van El Bierzo (N.W. Spanje). Unpublished thesis, dep. geol. univ. Leiden.

Smits, B. J., 1965. The Caldas formation, a new Devonian unit in León (Spain). Leidse Geol. Med., 31, p. 179–187.

Staalduinen, C. J. van, 1969. The geology of the area between the Luna and Torío rivers, southern Cantabrian Mountains, NW Spain. Leidse Geol. Med., in press.

Teichert, C., 1965. Devonian rocks and paleogeography of Central Arizona. U.S. Geol. Surv. Prof. Paper, 464, p. 1-181.

Teijón, E., 1946. El valle del Río Luna. Est. Geog., 7, p. 419–477.

Thomas, W. A., 1968. Contemporaneous normal faults on flanks of Binningham anticlinorium, Central Alabama. Am. Assoc. Petroleum Geologists Bull., 52, p. 2123–2136.

Vega Rollán, C. de la, 1964. Contribución al estudio de la flora carbonífera del occidente de Asturias. Breviora Geol. Astúrica, 8 (1–4), p. 37–51.

Verneuil, E. de, 1850. Note sur les fossiles dévoniens de Sabero. Bull. Soc. Géol. France, 2ème sér., 7, p. 155–186.

Vidal Box, C., 1943. Notas previas a un estudio morfológico y geológico de la alta cuenca del río Sil, cuencas de

Laciana y Babia Alta (prov. de León). Rev. Real. Acad. Cienc. Madrid, 37, p. 95–117.
—, 1959. La estructura geológica del borde de la depresión hullera de Laciana con la Babia Alta (prov. de León). Est. Geol., 15 (nr. 41–44).
Wagner, R. H., 1959. Flora fósil y estratigrafía del Carbonífero de España NO y Portugal. Est. Geol., 15, p. 393–420.
—, 1965. Palaeobotanical Dating of Upper Carboniferous Folding Phases in NW Spain. Mem. Inst. Geol. Min. España, 66.
Wagner-Gentis, C. H. T., 1963. Lower Namurian Goniatites from the griotte limestone of the Cantabric Mountain Chain. Notas Comun. Inst. Geol. Min. España, 69, p. 5–42.
Walker, T. R., 1967. Formation of red beds in modern and ancient deserts. Geol. Soc. America Bull., 78, p. 353–368.
Walter, R., 1965. Die Unterschiedliche Entwicklung des Alt-Paläozoikums östlich und westlich der Kristallins von Vivero-Lugo (Nordwest Spanien). N. Jahrb. Geol. Paläont., Mh., 1965, p. 740–753.
Winkler Prins, C. F., 1968. Carboniferous Productidina and Chonetidina of the Cantabrian Mountains (NW Spain): systematics, stratigraphy and palaeoecology. Leidse Geol. Med., 43, p. 41–126.