LEIDSE GEOLOGISCHE MEDEDELINGEN, Deel 50, Aflevering 2, pp. 75–114, 1–4–1980

GEOLOGICAL MAPS OF THE SOUTHERN CANTABRIAN MOUNTAINS (SPAIN)

BY

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ABSTRACT

This compilation of stratigraphic and structural data accompanying the (re)issue of the 1:50 000 sheets completes the project initiated by Prof. L. U. de Sitter in 1950. The total area mapped comprises about 400 km² in a strip more than 150 km from east to west.

This part of the Hercynian tectogene is characterized by a very consistent sequence of Palaeozoic shelf sediments only interrupted by syn- to late-orogenetic flysch-molasse development. Neither of these sequences lend themselves to a simple geosynclinal model.

Only the suprastructures of the orogene are exposed here; essentially decollement thrusting and folding. Fold and thrust vergences vary through 180° giving the centripetal pattern of the well-known Knee of Asturias.

Very minor amounts of igneous rock have been mapped although activity in some form has been registered throughout most of the systems represented. The degree of metamorphism is so slight to have been negligible for the mapping.

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CHAPTER 1

INTRODUCTION

The present series of maps of the southern slope of the Cantabrian Mountains is the result of 25 years of field work, that started in 1950 under the direction of Prof. Dr. L. U. de Sitter.

De Sitter's field trips served a double purpose: instruction for students, and investigation of hitherto unknown areas; and hence the students had the advantage to be trained in new areas. As 'which lies behind next hill' was new, both to teacher and pupil, each kept high interest in their findings. In general de Sitter started obtaining a general map of the area by running it over with a group of young students, the area to be mapped more accurately as a 'doctoraal' thesis, and finally to be concluded by publication of the coloured map as part of a doctor's thesis.

In the Cantabrian Mountains de Sitter started in the Barruelo area, with one student. These first years, progress of course was very slow. In his 1950 internal report de Sitter writes about that area that the coals are of Westphalian age. In 1951 the adjoining Visean and Stephanian were established. In the area around Cervera de Pisuerga, where another student started, it was concluded that Dinantian limestones rest unconformably upon older quartzites, and non-fossiliferous shales are on top of these limestones (Culm). When in 1952 the 'complete' Westphalian C-D in the Sierra Corisa was seen to be represented, the link between Cervera and Barruelo was completed. That year the Dinantian of Polentinos and Ventanilla was also discovered (the presence of Silurian has only recently been demonstrated).

In the following years, with an increasing number of students introduced in the area, the knowledge grew, as did the problems!

In 1955 the Valsurvio area between Cervera and Guardo was tackled, so together with the Polentinos area, field work was in progress in the Palaeozoic east of the Carrion River. Up till that time rock-stratigraphic units, especially in the Carboniferous were not very distinctive from the point of view of mapping purpose. This changed when de Sitter moved to the Esla area with its marvelous lower Palaeozoic stratigraphy. The sequence, from Cambrian sandstones through Silurian, to Lower Carboniferous had been excellently described from this area and several more westerly sections by Pierre Comte who unfortunately due to wartime circumstances interrupted his career as a geologist: of his – at that time unpublished – thesis, the Leiden group took full advantage (see Table I). Working with large groups of students – e. g. in 1963 de Sitter had some 50 students, through all grades, at work in the Cantabrian Mountains – the general map of the area grew rapidly. In 1962 a geological map on a 1:100 000 scale was compiled and published with the primary aim to settle the state of affairs. With so many people working at the same project some information is likely to get lost. The 1:100 000 map, of course, was far from complete, but distributed amongst all co-operators it provoked them to criticism and contribution of more correct information. At the same time de Sitter took advantage of the 1962-meeting in Oviedo to present the work established till that date. The Río Luna area up to the sources of the Río Sil was the last to be mapped. Beyond this area a vast area of Cambrian and Precambrian lies, which did not interest de Sitter at that moment, as he preferred to consolidate the present area. Neither did he wish to penetrate into Asturias, where the Spanish colleagues were at work. Thus the Leiden contribution to the geology of the Cantabrian Mountains is restricted to the Palaeozoic of the southern slope. The maps produced, based on a mixture of rock-unit stratigraphy and time stratigraphy are by no means considered to be complete, but should serve as a base for further investigation by specialists in fields as sedimentology, palaeontology and structural geology, thus enlarging the knowledge of the area and its development, and improving the mapping.

The main purpose of the present publication is to make the complete 'oeuvre' of maps of the area available in one volume. A very short description of the geology involved should suffice. Those readers interested in more detail, of course, are referred to the previous descriptions of the maps.

In that context I like to repeat what de Sitter wrote in 1962: "From the beginning I want to emphasize the fact that the present outcrop of Palaeozoic rocks in the Cantabric-Asturian mountain chain represents only a more or less fortuitous portion of a much larger Hercynian block, the present outcrop being determined largely by an Alpine deformation."

There are two main longitudinal units, a northern one dominated by Cambro-Ordovician and Carboniferous rocks, with very little Devonian, and a southern one, in particular in the western half of the map where rocks range from Precambrian to Upper Carboniferous. The two units have been named the Asturides and Leonides, and are separated by the León Line.

The Leonides, south of this line, consist of a western Luna Zone, a central Bernesga-Esla Zone and the eastern Ruesga Zone. On the southern border of the Leonides we find discordant Upper Carboniferous intramontane coal basins. The Luna Zone is essentially a complicated synclinorium resting on Precambrian, at its southern flank with Lower Cambrian; in its northern flank the Cambrian has been thrust over the Namurian of the frontal part of the Leonides. Along the León Line Lower Palaeozoic with Lower Cambrian at its base again shows up. Just before reaching the Bernesga River this Luna Synclinorium splits up into many thrust sheets, which end against the Pardomino Fault along the Porma River. Further east the Esla Thrust Sheet has taken over the thrusting movement. Together they form the Luna-Esla Zone. Still further east follows first the discordant Valderrueda or Cea Basin and then the Ruesga Unit with the Valsurvio Dome. These units of the Leonides have in common the almost complete absence of Westphalian rocks, except in the Tejerina Syncline, and the gradual disappearance of Devonian rocks going from south to north. At the León Line the Famennian or the Lower Carboniferous rests directly on Ordovician or even Cambrian, and in the south of the Leonides the Devonian sequence is complete. Between these we find a gradual development of this hiatus.

In the Asturides we can distinguish the Beleño area folds and thrust structures, mostly in Cambro-Ordovician and Carboniferous west of Riaño, and an eastern zone of more variable structure, in which the Lower Westphalian Curavacas Conglomerate plays a dominant role (the Yuso area), and a zone of Devonian rocks in the upper Carrión River (the Carrión area), and finally the Pisuerga and Barruelo Basins with their Westphalian-Stephanian sequence of coal-bearing sediments.

In the western end of the Leonides the structures known from the eastern part of the Luna Unit do not continue but are replaced by one syncline and a number of thrusts. Major block faulting of the basement determined the curved structure of the 'Asturian Knee', and also the diverse contemporaneous directions of thrusting and folding (de Sitter & van den Bosch, 1968).

ACKNOWLEDGEMENTS

It is not possible to mention by name everyone who (and how they) contributed to the maps presented. The major contributors will be found in the reliability diagrams given with most sheets. There have been many others from undergraduate students to foreign guests in various capacities. It is quite clear that this work could never have been done by one man but without that man none of it would have been accomplished. It has been a privilege for us all to be involved in this work.

Technical assistance through the years has also been credited in the individual publications but we cannot omit grateful acknowledgement of the superb work of Miss C. P. J. Roest whose draughting style elevates the appearance of most of the maps of our series.

One group does not figure in our account except perhaps in the place names of the towns and villages: the people of Old Castille who have welcomed and accepted the generations of geologists with the rightly famous Castillian hospitality. Those names must recall for all who have worked there the many true and faithful friends who enriched our experiences during the long field seasons.

CHAPTER 2

STRATIGRAPHY

The availability of Comte's work of course was an invaluable advantage to our mapping as it provided an excellent set of mappable horizons. Our stratigraphical implications - hence - should be regarded from the one major point of view: does it contribute towards the mapping process. In the same sense, sedimentology, a rapidly developing specialism, during de Sitter's time, meant another useful tool to our handcraft.

For the lithostratigraphic subdivision of the Lower Palaeozoic deposits of the present area, Comte (1959) chose a type section along the Bernesga River between Busdongo and La Robla. In the type section, Comte defined the Palaeozoic formations, but the names of the pre-Devonian formations and those of the Devonian Portilla Formation were derived from reference sections elsewhere. Later publications either complete the lithostratigraphic subdivision of this area, or detail the pre-existing units (see Table I and Fig. 1). The presence of Precambrian in this area remained unknown until de Sitter (1961, 1962a) described an angular unconformity in the Herrería Formation. He suggested the rocks underlying the unconformity to be of Precambrian age and called them Mora Slates and Greywackes, now Mora Formation.

The Department of Structural Geology of the University of Leiden has produced a number of map sheets contributing to the regional geology of the Southern Cantabrian Mountains. Those either reissued with this volume* or used in the compilation of new sheets are listed in Table II. Unpublished work is credited on the individual sheets.

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

Table I. Comte's (1959) stratigraphic subdivisions and their correlation with the formations used here (modified after van Staalduinen, 1973).

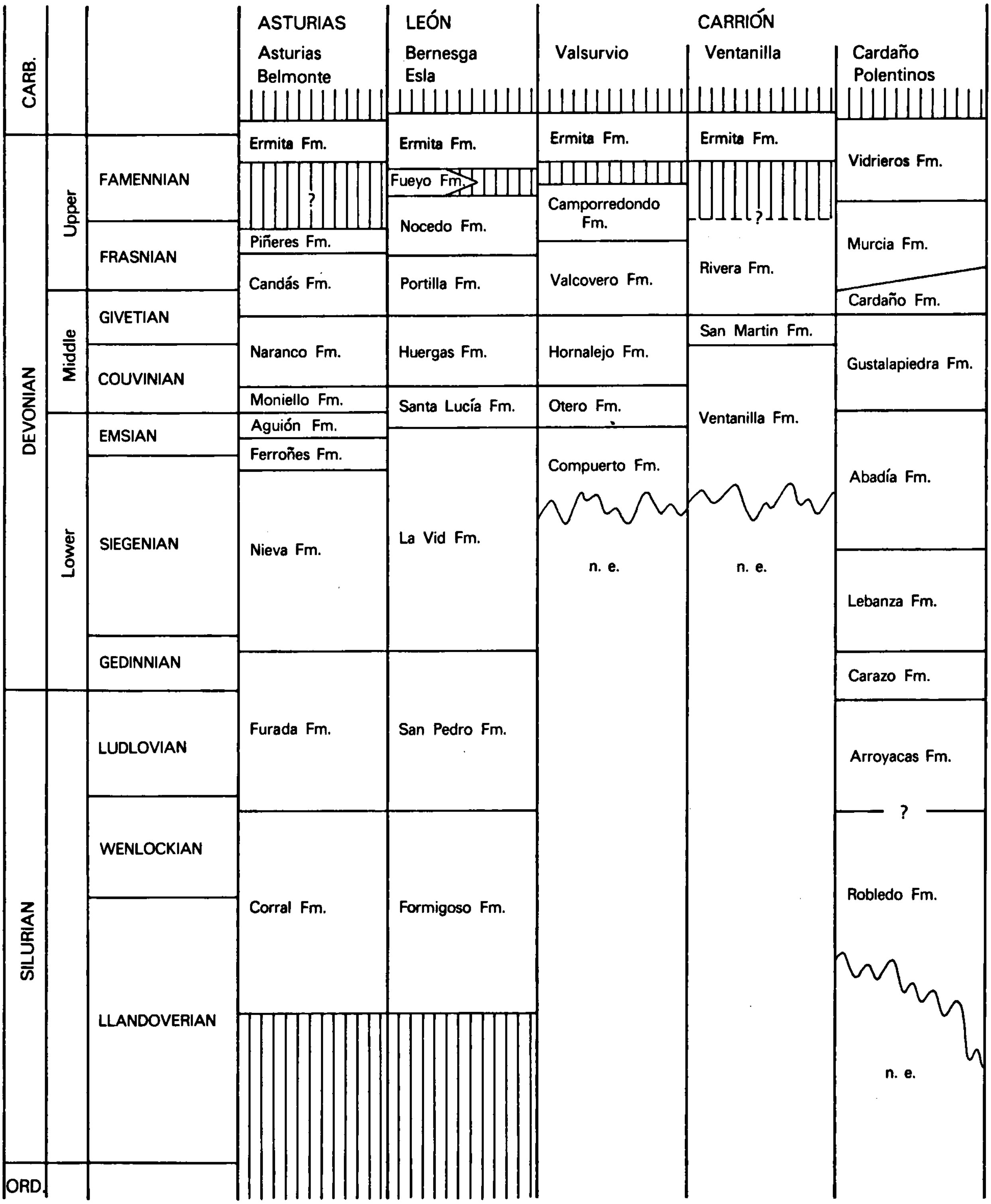

Fig. 1. The regional and local correlation of Upper Palaeozoic formations throughout the Cantabrian Mountains (modified after Brouwer, 1968).

The following account has been largely extracted from the articles accompanying the Map Sheets listed in Table II. Some of the later research in the Carboniferous flysch and molasse has, however, also been incorporated.

The bibliography presented here has been restricted to articles cited in our text. Exhaustive lists of both Spanish and foreign literature can be found in the detailed map descriptions as well as in some of the stratigraphic and palaeontological monographs also cited.

Sheet	Author	Date
Luna-Sil*	van den Bosch, W. J.	1969
Provisional Geological Map of the Valdéon-Liébana-Polaciones area	Boschma, D.	1968
Bernesga-Torío-Curueño-Porma*	Evers, H. J.	1967
Southern Pisuerga and Santibañez de Resoba	Frets, D. C.	1965
Cea-Esla-Porma*	Helmig, H. M.	1965
Sierra del Brezo	Kanis, J.	1956
Valsurvio Dome	Koopmans, B. N.	1962
Nansa-Deva*	Maas, K.	1974
Pisuerga	Nederlof, M. H.	1959
Cea-Esla-Porma*	Rupke, J.	1965
Yuso Basin	Savage, J. F.	1967
Yuso*	Savage, J. F.	in press
Carrión*	Savage, J. F.	in press
Provisional Geological Map of the Southern Slope of the Cantabrian Mountains	de Sitter, L. U.	1962
SW part of the Southern Cantabrian Mountains	de Sitter, L. U. & van den Bosch, W. J.	1968
Pisuerga*	de Sitter, L. U. & Boschma, D.	1966
San Isidro-Porma*	Sjerp, N.	1967
Luna-Bernesga-Torío*	van Staalduinen, C. J.	1973
Cardaño de Arriba	van Veen, J.	1965

Table II. List of Map Sheets published by members of the Department of Structural Geology, Leiden (*indicates reissued with this volume).

PRECAMBRIAN

Mora Formation (van den Bosch, 1969)
The Mora Formation crops out in a long arc, the Narcea Anticlinorium, between the Asturian coast and La Magdalena where Tertiary deposits overlie it.

The name Mora Formation has been introduced by de Sitter (1962b); the type locality lies 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 (1968b), Pastor Gómez (1962) and de Sitter (1962a).

The best-exposed sections occur north of Villabandin and in the N–S striking mountain chain west of the Río de San Miguel.

Age. – No fossils were encountered, but as the Herrería Formation of lowermost Cambrian age, overlies the Mora Formation unconformably, it must be of a Precambrian age.

Lithology. – The Mora Formation consists mainly of very slightly metamorphic slates, graywackes and quartzites, all possessing a green colour. Several sets of cleavages and knick zones were found. 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. Sedimentary structures are rather abundant. Slumps, 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 1 cm in diameter was encountered, and rhyolite tuffs occur, especially in the western part of the region. Due to weathering before the deposition of the unconformable Herrería Formation, the upper 5 to 20 m of the Mora below the unconformity have a characteristic red colour. This unconformity plane is very smooth.

CAMBRO-ORDOVICIAN

Herrería Formation (van den Bosch, 1969)
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 base of the formation is exposed 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 m of the formation which are characteristic of Lower Cambrian, hence the hundreds of metres of rocks below it may be lower Lower Cambrian, too.

Lithology – The Herrería Formation overlies the Mora Formation unconformably. This unconformity is accentuated by a red weathering zone 10–25 m thick on the Mora, and by a conglomerate at the base of the Herrería, generally 0.5–3 m thick. The conglomerate does not always rest directly upon the weathering zone but somewhat higher in places, though never more than 8 m 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. The conglomerate 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 m consists of red and green silty shales and white and pink quartzites. Some slumps were found in the quartzites.

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

Yellow-weathering dolomite beds, alternating with shales and sandstones, come next. This dolomite horizon could be traced only in the west of the area.

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 m) occur; drab-coloured sandstones, and red and green silty shale layers are rather irregularly distributed over this succession with bed thicknesses 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 feldspar grains may be mixed between the quartz grains.

In the upper 50 m, 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. The base of the Lancara was drawn at a level where more than 50% of dolomite occurs.

Thickness. – Approximately 850 m south of Los Barrios de Luna; approximately 550 m NW of Salce; approximately 1100 m along the Río de Sosas. In other areas the base is absent.

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

Depositional environment. – The lithostratigraphic continuity of the bedding and the rock characteristics such as cross-bedding, ripple marks and all sorts of flute and groove-like casts in the poorly sorted sediments of the Herrería lead to the opinion of near-shore deposition (deltaic?) in the aftermath of the tectonic movements, which produced the unconformity of the Herrería on the folded Precambrian schists known to the west.

Observations on cross-bedding and sole markings in the Herrería beds of Pardomino point to a generally southward transport direction (Rupke, 1965).

Lancara Formation (van den Bosch, 1969)
Comte (1959) introduced the names 'Dolomies', 'Calcaires' and 'Griotte de Lancara', named after the village of Lancara de Luna. The type section is inundated now.

The formation is subdivided into the Dolomite Member, the Limestone Member and the Griotte Member. Accounts have been published by Comte (1959), Lotze & Sdzuy (1961), Oele (1964), and by van der Meer Mohr & Schreuder (1967), and van der Meer Mohr (1969).

From a structural point of view the Lancara 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, and west of Salce.

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 Lancara is of a rather uniform lithological composition, description of the very well exposed section west of Los Barrios de Luna suffices.

The gradual transition from the Herrería to the Lancara 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. This alternation of dolomites and black shales is followed by a grey limestone, 3.60 m thick, completely composed of well-developed stromatolites, overlain by a laminated limestone 80 cm thick, with cross-bedding. On the following cross-bedded dolomite lies a laminated argillaceous dolomite, overlain by dolomite beds with ripples, cross-bedding and intra-formational breccias. Next comes an alternation of laminated black shales and dolomites with ripples and cross-bedding. These beds are overlain by 20 m of thick-bedded dolomites, weathering to a typical orange-brown.

The dolomitization of the limestone (member) is partly of secondary origin, because it is not restricted to the bedding plane: small tongues of yellow-weathered dolomite penetrate into the limestone layers perpendicular to the bedding (Sjerp, 1967; Rupke, 1965; van Staalduinen, 1973). Sandstone dykes also have been observed.

Well-sorted oolites, small angular limestone fragments and subrounded sandgrains indicate a shallow neritic depositional environment (Evers, 1967). Van der Meer Mohr & Schreuder (1967) concluded intertidal conditions from the laminated algal-growth structures (stromatolites).

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. Rupke (1965) found glauconite. The upper 4 m are of a pink colour, and are rich in fossil debris. Sjerp (1967) reports a layer which is highly fossiliferous, the main component being parts of brachiopods and trilobites.

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. At the contact with the Limestone Member many irregular contacts were found. In the northern part of the region the griotte in places loses its red colour. There we find grey-green nodular limestones and yellow nodules in grey shales. The transition to the Oville Shales is well exposed near the church of Lancara: between the last thick griotte bed and the olive green Oville Siltstone lie 2 m of greenish, red and black shales with green limestone concretions, weathering to an orange colour, and a few griotte beds were also found.

Thickness. – Thicknesses vary over the area, up to 150 m in the Luna area.

Fossils. – Stromatolites, trilobites, brachiopods, carpoids. A concentration of broken trilobite fragments is very common in the griotte.

Oville Formation (van den Bosch, 1969)
Comte introduced the name 'Grès et Schistes d'Oville' and selected a type locality just south of the village of Oville, in the Porma Valley.

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

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.

There is a gradual transition between the Griotte of the Lancara Formation and the argillaceous siltstone of member A of the Oville Formation. The 25 to 50 m thick member consists of finely laminated, green or bluish argillaceous siltstones 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. 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, and glauconite grains of the same colour. These glauconite grains are typical of the entire formation.

After the basal 25 or 50 m, 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 m; 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 m 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. Micas are concentrated on the bedding planes and sharp contacts between the generally parallel-bedded sandstones and siltstones. This parallel bedding is typical of member C; coarse to fine gradations also occur, as does internal lamination. 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, and parallel lamination was also found. Especially in the upper part, the channels have a clearly erosional basis. Flaser and linsen structures, ripples, clay pebbles and cross-bedding are frequent. Loading was found in the upper part. The presence of slump structures indicates the unstable character of these deposits (Evers, 1967). The top of the Oville is drawn below the first thick sequence of thick-bedded, cross-bedded quartzites.

Tuffite beds and dolerite sills, parallel to the bedding, and without contact metamorphism, were found locally in the various parts of the formation.

Eastward, the formation can be subdivided into two parts only, a lower shale-siltstone member and an upper sandstone member (van Staalduinen, 1973).

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

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

Thickness. – The thickness varies: in the Abelgas Syncline: 430 m, west of La Majua 192 m. The thinnest development of the formation occurs in the Aralla Zone (some 140 m), where the upper member was deposited under conditions which allow erosion, as is shown by several eroded bedding planes. In the Esla area the Oville is about 220 m.

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

Barrios Formation (van den Bosch, 1969)
Comte (1959), who introduced the name 'Quartzites de Barrios', selected the type locality near Los Barrios de Luna. Previous work was done by Comte (1959) and Oele (1964).

In the formation five members could be recognized: 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) on the southflank of the Narcea Anticlinorium, some 50 km SW 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. The mica flakes have been determined as phengite by Oele (1964). 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 m 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. In the Esla area, Rupke (1965) reports igneous rocks in the midst of the formation.

The following 21 m are not exposed in the type section, but in other places in the Abelgas Syncline, 8.5 m 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 km to the NNW; these levels can therefore be traced over a great distance.

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. 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. 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 m.

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 m 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. At the base, intensive loading took place. Large, steep cross-bedding and deep erosive channels occur frequently in these beds, but beds with parallel lamination or planar cross-bedding also occur. Bed thicknesses of more than 10 m are not uncommon. The upper metres of member E near Torre are thin bedded and very micaceous. This member was not found in the eastern part of the Aralla Zone.

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.

Thickness. – 301 m in the type section; further to the east 100 m. Small thicknesses have also been encountered in the SW part of the Aralla Zone. 350 m in the Esla area.

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

Luarca Formation
Between the Barrios Quartzites and the Formigoso Shales, van den Bosch (1969) reports 'transitional beds' in the west of the Cantabrian area. Their stratigraphic position suggests they should be correlated with the Luarca Formation (Poll, 1963). The pure bluish dark shales, however, which constitute more than 90% of the Luarca Formation, are rare in these transitional beds. Especially, impure, often burrowed ferruginous sandstones and siltstones, containing a large amount of glauconite, are common (van den Bosch, 1969). The thickness

rapidly decreases from a 170 m in the west of the area; east of the Luna the beds are not found.

According to Evers (1967) the sharp, undulating boundary between the Barrios and Formigoso Formations as exposed north of Felmin, represents a nonconformity which is marked by concentrations of well-rounded, white quartzite pebbles.

Comte (1959) also concluded the existence of an important hiatus after the deposition of the Barrios Formation, equalling the Luarca Beds in Asturias. His main argument is the occurrence of the same Upper Llandoverian graptolites at the base of the Formigoso and Corral Formations.

SILURIAN

Formigoso Formation (van den Bosch, 1969)
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).

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. These black shales are the most indicative sediments of the formation. 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. A weathering clay or an iron concentration is often found at the base of the Formigoso. Markasite and pyrite were also found at higher levels in the black shales.

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. In the upper ten metres, thick sandstone beds with cross-bedding occur. The shales have at first a black or grey colour, but in the upper tens of metres they become more and more green.

Thickness. – The variation of thickness in the Formigoso partly is due to intensive folding, especially of the finely laminated shales.

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

Robledo Formation (Ambrose, 1974)
The Robledo Formation consists of sandstones with shale partings, some silty and calcareous near the base, usually bioturbate. Beds vary from 10 cm to 1 m tending to thicken upwards, some of the thicker beds are finely laminated, fine- to medium-grained quartzites. Sorting also tends to improve upwards in the sequence. Shallow channelling can be detected, and ripple marks with mica coatings are common together with euhedral pyrite and haematite, as well as haematite ooliths. The interbedded shales are dark grey to black, probably recrystallized although no sign of cleavage has been found.

The best, practically only, section is exposed just north of Lebanza (Encl. VII: Carrión Sheet) where these rocks have been thrust over shales of the Abadia Formation. Upwards the Robledo is conformably overlain by the Arroyacas Formation. The exposed sequence is 160 m thick and although the lower limit is faulted this is probably near to the true thickness.

The age of these strata can only be deduced indirectly because of the absence of identifiable fossils. In contrast with Ambrose (1974) we correlate the Robledo with the Formigoso Formation of León on the basis of lithology, taking into consideration the correlation of the overlying Arroyacas with the lower Furada Beds of Asturias (Poll, 1969) and the San Pedro Formation of León (Comte, 1959). A Lower to Middle Silurian age is thus most probable (Fig. 1).

San Pedro Formation (van den Bosch, 1969)
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.

Previous work was done by Bäcker (1959) and Comte (1959).

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 m 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 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. Approximately 6–15 m from the base of member A, such a level, with coarse phosphorite pebbles up to 4 cm in diameter was encountered in several sections (Bäcker, 1959; van den Bosch, 1969).

Clay pebbles and flakes are common in the sandstones. Burrows were only found at the generally sharp sandstone-shale contact. Bed-thicknesses of up to 4 m 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 are channelling, cross-bedded sandstones. Ripples, however, 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 m for the red sandstones. Laterally the bed thickness does not change very much. Current ripples are very abundant, as are flaser and linsen structures and burrows. 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 cross-bedding or ripples which have been locally 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.

Thickness. – The thickness varies from 150 m in the west, some 200 m in the Luna area, a 100 m in the Bernesga area, and again a 180 m in the Esla area. In general the thickness of the San Pedro Formation as well as that of the underlying Formigoso Formation decreases towards the north.

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

EASTERN SHEETS (YUSO, CARRION AND PISUERGA)

The stratigraphy of the Valsurvio area (Koopmans, 1962) and the Pisuerga area (de Sitter & Boschma, 1966) resembles the stratigraphy of the western nappe area only in rough outlines. In these areas other formation names have been used. Van Veen (1965) correlates his Carazo Formation with the San Pedro of the Comte stratigraphy. The Carazo is not as ferruginous as the San Pedro but the shale-sandstone-quartzite sequence is very much alike.

In general, correlations of the formations with those of the nappe area are mainly based upon comparison of biostratigraphy (Fig. 1).

Arroyacas Formation (Ambrose, 1974)

The Arroyacas Formation is mainly silty, sandy mudstones often strongly bioturbate, interbedded with a few horizons of splintery, blue-black shales and even fewer layers of sandstone. The sandstones are more frequent and cleaner towards the top of the formation.

The type section north of Lebanza is not continuous and the total thickness of 350 m may be in error but not by more than 100 m. The thickness of 75 m plus formerly attributed to this sequence called Member a of the Carazo Formation by Binnekamp (1965) and van Veen (1965), is certainly wrong by an order of magnitude.

Fossils are relatively abundant and include brachiopods, crinoids, trilobites, pelecypods, bryozoans, corals, orthocones and graptolites.

The age of the uppermost beds of the Arroyacas Formation is fixed by the determination of a guide form graptolite found in them which belongs to the Lower Ludlow Stage of the Upper Silurian (Rickards in Ambrose, 1974). This dating together with the lithological characteristics are considered to justify correlation with the San Pedro and Furada Formations in León and Asturias, respectively (Fig. 1).

Carazo Formation (Ambrose, 1974; Binnekamp, 1965; van Veen, 1965)

The Carazo Formation as modified by Ambrose comprises the members b and c of the formation as originally defined by Binnekamp and van Veen. The lower member, the Carazo Quartzite, is typified by sandstone and ironstone deposits, the thicker beds often cemented to a true quartzite. They are medium to coarse grained, usually parallel laminated but sometimes cross-bedded. The ironstones with occasional hematite ooliths and phosphorite pellets are typical of thinner sequences where a higher proportion of interbedded shale is to be found.

The upper member (c of Binnekamp and van Veen) is predominantly shaly; silty in the lower part, becoming calcareous upwards. The brown-weathering, splintery shales are interbedded with darkgray, yellowishbrown weathering, well-stratified, thin to medium bedded fine-grained sandstones. The shales are noticeably less micaceous than those of the underlying Arroyacas Formation, and the sandstones typically decalcified with detailed casts of brachiopods. Fossils occur throughout but are especially abundant in the upper member in which brachiopods, lamellibranchs, ostracods, trilobites, pteropods and gasteropods have been found.

The thickness of the Carazo Formation, as well as its members and beds, varies rapidly through the relatively small area of its outcrop. From the type section at Peña Carazo where a total of more than 400 m of strata are present, the formation thins rapidly north and southeast to less than 100 m.

The faunas identified by Binnekamp and Brunton (in Ambrose, 1972) show that the redefined Carazo Formation is of Gedinnian age.

DEVONIAN

La Vid Formation (van den Bosch, 1969)
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.

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 gradual transition takes place from member C of the San Pedro to member A of the La Vid. At several places, however, an erosional contact has been found. In a gradual transition 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.

Dolomite beds and limestones up to 60 cm thick, alternate, interbedded with dark shales. In some dolomites, quartz grains, mica flakes (largely replaced by dolomite) and some pyrite grains have been observed. 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. In large parts of the Babia Baja and Luna Units, several metres of calcareous black shales, locally with detrital, very fossiliferous, limestone beds occur between the last thick limestones and the green shales of member B.

Member B consists mainly of very fine, often splintery, olive-green calcareous shales with several thin, very fossiliferous beds. Several detrital limestone lenses occur. The shales often contain small pyrite grains and iron silicate grains; the latter probably give these shales their green colour. Due to tectonic disturbance the thickness of this member is generally difficult to establish.

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.

Thickness. – The thickness of the members varies considerably: member A is reported some 380 m in the west of the area, member B 279 m and C 158 m. In the Esla area a total of 200 m are found in the nappe as well as the autochthonous. In general the middle member is well developed, members A and C often are diminished or absent.

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

Compuerto Formation (?La Vid, Koopmans, 1962)
This formation crops out in the centre of the Valsurvio Dome and can be subdivided into three members, from top to bottom:
c. Light-coloured slates and phyllites, with a chloritoid-sericite schist on top, and locally a soil of strongly weathered red-yellow sandy shale.
b. Sandy slates with sandstone and limestone lenses, grading upwards into black slates.
a. Coarse crystalline limestones.

Lebanza Formation (Binnekamp, 1965; van Veen, 1965)
The Lebanza Formation consists basically of limestones; detrital, well-sorted, calcarenites at the base. These basal strata are usually medium bedded with parallel lamination although cross-bedding and even a slumped horizon also occur. Fossils though common are usually broken or distorted. Upwards these limestones pass into more shaly limestones with poorly preserved brachiopods. A horizon of pure white, thin-bedded limestones with well-preserved brachiopods forms the transition to the most characteristic thick to massively bedded part of the formation. The fauna is less abundant and consists mainly of stromatoporoids and rugose corals. The uppermost part of the formation is of dark muddy limestones with an extremely rich, well-preserved fauna of brachiopods together with lamellibranchs, tentaculites and gasteropods.

Both the upper and lower contacts of this formation are conformable and, in fact gradational, with the Abadia and Carazo Formations, respectively. The thickness at the type section near Lebanza, of 160 m is probably the maximum although many other sections seem to have been tectonically truncated. The formation almost certainly thins to the north and northwest, and the more fossiliferous levels seem to thin out in those directions too.

The abundant faunas yield determinations of Upper Gedinnian to uppermost Siegenian ages.

Abadia Formation (van Veen, 1965)
The Abadia Formation is essentially a shaly interval although two levels of limestone are sufficiently well developed as to have been distinguished as members. The shales are remarkably siliceous, silty and even sandy despite the tendency to nodular limestone deposition at all levels. The lower, Requejada Limestone Member is a

dark, thin-bedded, silty limestone at the base which passes upwards into lighter-coloured, wavy-bedded to nodular, limestones. The passage to the overlying shales is gradual, and nodular and detrital limestones recur throughout the rest of the shaly interval.

The Polentinos Limestone Member grades conformably from the shales below becoming medium-bedded, detrital bands. Upwards pyrite becomes common and the bedding becomes wavy until nodular limestones interbedded with shale complete the sequence. Fossils are common throughout the limestones, goniatites and conodonts being particularly important. Trilobites (especially in the Polentinos Member) together with corals, brachiopods, ostracods, pteropods and lamellibranchs complete the faunas.

The extensive faunas have enabled numerous datings to be made, and the Abadia Formation is known to range in age from Upper Siegenian to (lower?) Eifelian (Kullmann, 1960; van Adrichem Boogaert, 1967).

Santa Lucía Formation (155–230 m) (Evers, 1967)
The type locality (Comte, 1959) is found at the village of Santa Lucía in the lower Bernesga Valley, in the Bernesga Sheet (van Staalduinen, 1973). The massive limestone banks of the Santa Lucía Formation are resistant to weathering, and thus form well-exposed ridges flanked by the less resistant argillaceous beds of the La Vid and Huergas Formations. Because the general lithologic aspects of the Santa Lucía Formation are quite uniform, a differentiation into four prevalent limestone members could be made.

The transition from the La Vid Formation is gradual: several reddish detrital limestone-lenses still intercalate the grey limestones at the base of the Santa Lucía Formation. The lower member consists of yellowish-grey, thinly bedded argillaceous limestones (30–40 m), which are irregularly dolomitized.

A 5 m thick transition zone of shale and marl beds precedes the second limestone member, which is composed of thickly bedded biostromal limestone banks intercalated by black bituminous shale lenses. The biostromal banks are built up of well-cemented fossil debris and colonies of stromatoporoids; rugose corals are scarce. The irregular dolomitization and silicification appears to be post-depositional, but authigenic chert nodules elongated parallel to the bedding plane also occur. The thickness varies considerably from 24 m near La Vid to 56 m N of Aviados, and even 90 m N of Voznuevo.

The second and third limestone members are separated by a reddish-brown detrital limestone bed (ca. 8 m), consisting of concentrations of brachiopods, lamellibranchs, gastropods, rugose corals and stromatoporoids. These wave-built biostromal blankets indicate greater wave action and shallower water depth. An Emsian age was determined by Comte.

The third limestone member is composed of light-grey biomicrites (ca. 40 m) overlain by dark-grey argillaceous biosparites (50–60 m).

The fourth limestone member consists of a reddish detrital lower part (ca. 15 m) and a thin-layered arenaceous upper part (ca. 10 m), separated by a bryozoal-limestone layer (ca. 2 m). Concentrations of large specimens (4 to 7 cm) of *Spirifer cultrijugatus* indicate the top of the Emsian or Couvinian. The total thickness of the Santa Lucía Formation varies between 155 and 230 m.

Van Veen (1965) correlates the Santa Lucía with a part of his Abadia Formation, according to his map. These correlations are based upon fossil content.

Otero Formation (?Santa Lucía Formation, Koopmans, 1962)
This Otero Formation consists principally of calcareous sediments, showing great lateral variations. In the valley of the Río Carrión, directly south of Otero de Guardo it consists of five distinct limestone layers separated by shales and calcareous shales. The Santa Lucía limestone of Comte (1959) is of the same age as the Otero Limestone. Both formations are rich in corals and stromatoporoids. An upper zone of the Santa Lucía Limestone, rich in brachiopods, and known as the zone *Spirifer cultrijugatus* in the province of León, has not been found in the Otero Limestone.

Caldas Formation (van den Bosch, 1969)
Smits (1965) proposed the name Caldas Formation and selected the type section east of Caldas de Luna. De Coo (1974) regards the Caldas Formation as another facies of the Santa Lucía Formation. 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 (1965). 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.

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 m below the unconformity.

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

Fossils. – Brachiopods, crinoids, corals, bryozoans, gasteropods, stromatoporoids, Algae, ostracods, lamellibranchs and spicula of Spongae.

Huergas Formation (van den Bosch, 1969)
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).

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.

In general the Huergas overlies the Santa Lucía with a sharp contact, but according to van den Bosch (1969) a gradual transition is not uncommon. After a gradual transition from the Santa Lucía, mainly shales with some clay-ironstone concretions follow; several sandstone beds and sometimes a few limestone beds are intercalated. The transition from silty shale to argillaceous sandstone and vice versa is gradual.

The Huergas Formation is a sandstone formation consisting of silty shales, siltstones, arenaceous siltstones, sandstones, and very occasional quartzite. In general the formation is slightly ferruginous, occasionally very iron-rich layers are found. Sandstones are decalcified and burrowed, a few fossil-rich horizons are typical.

Thickness. – From the upper Luna area more than 300 m are reported, from the Abelgas Syncline 150 m (van den Bosch, 1969). Van Staalduinen (1973) reports 200–300 m from the Bernesga area, and according to Evers (1967) only the lower 100–150 m are exposed in the Porma area. In the Esla Nappe, mapped by Rupke (1965), the thickness is up to 240 m, and up to 320 m in the autochthonous.

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

Piedrasecha Formation (van Staalduinen, 1973)
The Piedrasecha Formation was introduced by van Staalduinen for the clastic sequence of mainly Devonian rocks deposited on top of the Santa Lucía Formation along the southern limb of the Alba Syncline. This clastic sequence is present from Portilla de Luna to the Bernesga Valley. The western boundary of the Piedrasecha with the Nocedo and Ermita Formations is purely arbitrary. It is assumed that there exists a lateral facies change between these three formations. The section near Piedrasecha has been chosen as type locality. The thickness decreases to the E.

The type section starts with about 200 m of brown shales in which concretions of claystone and limestone are common. The limestone concretions are often fossiliferous, especially in the basal part of the sequence. Apart from the concretions, thin ferruginous beds occur in this lower part of the formation. Then follows a monotonous sequence of about 160 m of dark brown to black shales with thin ferruginous sandstone beds. The following sequence of about 100 m is composed of thin-layered sandstones alternating with shale beds. In this part of the sequence, animal tracks, load casts and slump structures could be recognized. The upper 20 m of this sequence also contain concretions. White quartzites and sandstones form the following 10 m of the section, and 30 m of badly exposed brown shales cover the quartzites and sandstones. The upper 50 m of the sequence consist of alternating brown sandstones and mudstones, with cross-bedding in the sandstones of the uppermost 10 m. Casts of brachiopods have been found in the mudstone beds.

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

Thickness up to 600 m.

Gustalapiedra Formation (van Veen, 1965)
The Gustalapiedra Formation conformably overlies the Abadia Formation. It consists of alternating dark-grey and black slates and argillaceous black limestones. Upwards there is a gradational contact with the overlying nodular limestones of the Cardaño Formation. The limestones are medium crystalline and pyritic, while the slates contain crystalloblasts of chloritoid and sericite, in the type section. The appearance of these rocks is also strongly influenced by the cleavage development in highly deformed sections. A few, thin, lenticular beds of sandstone are interbedded with the limestones and slates. The proportion of limestone is less in the eastern Carrión area but the total thickness remains between 50 and 70 m.

A specialized fauna of trilobites, goniatites, conodonts, lamellibranchs and gasteropods has been found which have yielded dates from later Eifelian to Upper Givetian (Kullmann in van Veen, 1965; van Adrichem Boogaert, 1967).

Portilla Formation (Evers, 1967)
The type location (Comte, 1959) is found on the right bank of the Portilla brook, W of Matallana-Estación (see

also Reijers, 1972). In the field, the Portilla Formation can easily be differentiated into two resistant limestone members separated by a less resistant reddish, detrital limestone bed. The limits of this formation are difficult to set, because a gradual transition to calcareous shales or sandstones is developed on both sides.

The following composite section was measured from base to top, near Matallana-Estación, SW of Aviados and NE of San Adrian:

a. Thinly layered, coarse-grained arenaceous limestones, ranging in colour from yellow to grey, but weathering red to brown. Concentrations of brachiopods, solitary corals and crinoids give this member a biostromal aspect. Cross-bedding occurs locally. Thickness: 5–15 m.

b. Well-layered, blue-grey limestones. The massive parts are rich in reef-building fossils. Thickness in the Torío-section: 15–30 m; NE of San Adrian and SW of Aviados: 50–60 m. In the latter sections, bioherms with flank deposits have developed. In contrast to the Santa Lucía Formation, compound rugose corals are abundant but stromatoporoids are rare.

c. Reddish detrital marls and wave-built biostromal limestones. The thickness is generally 10–25 m, but SW of Aviados, a 5 m thick coarse-grained calcareous sandstone lens and thicker biostromal lenses account for a total thickness of 40 m.

d. Massive, light-grey reef limestones and few shale lenses: 40–60 m. Locally, the biohermal dome-shaped structures interfinger with well-bedded biostromal limestones and calcareous shales. Partly silicified corals, bryozoans, brachiopods and a few stromatoporoids are found.

e. Yellow-weathering, calcareous shales and sandstones with lenses of reef debris. Thickness: 10–25 m.

The total thickness of the Portilla Formation is generally up to 200 m. The age of the Portilla Formation is determined as Upper Givetian and Lower Frasnian (Comte, 1959).

Lateral facies changes are common in the Portilla. Van den Bosch (1969) points to its argillaceous character in the Luna-Sil area. In the Alba Syncline, van Staalduinen (1973) notes an extraordinary thickness of the limestones. Koopmans (1962) also mentions great lateral facies changes in his Valcovero Formation equivalent in the Valsurvio area.

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

Cardaño Formation (van Veen, 1965)
The Cardaño Formation consists of beige to grey nodular limestones with darker shaly partings. Gradations exist in the form of typical peanut shales, specific to the Palentian facies in Cantabria and strikingly similar to the Hercynian facies of northern Europe (Brouwer, 1964). In the type locality the rocks have been tectonized and metamorphosed, giving rise to slaty cleavage and a preferred orientation of nodules, as well as chloritoid porphyroblasts. The limestones, however, have remained fine grained, and have yielded abundant conodonts, a few goniatites and some broken shell fragments.

The contacts with both the underlying and overlying formations are conformable, and while condensation of the sequence is suspected, no separate hiatus level can be picked out. The thicknesses observed vary from 10 to 40 m.

The rich conodont faunas, in particular, have allowed very precise dating of this formation as ranging from the lowermost to Upper Frasnian (van Adrichem Boogaert, 1967).

Nocedo Formation (van den Bosch, 1969)
Comte introduced the names 'Grès de Nocedo' and 'Schistes de Fueyo'. Because the Fueyo Shales have only been found 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). In the west only in the SW limb of the Palomas Syncline does a 300 m Nocedo crop out. The most complete Upper Devonian section was described by Comte near the village of Nocedo in the lower Bernesga Valley, where approximately 500 m of calcareous sandstones, quartzites and shales were measured. The Nocedo Formation is reduced to 350 m in the Torio section near Matallana-Estación, to 250 m NE of La Valcueva, and about 175 m near San Adrian. From the Bernesga and Esla Basins towards the Pardomino Ridge, the sand grains tend to become coarser and increase at the cost of the shales and limestones (Evers, 1967).

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

Lithology. – The general aspect of the Nocedo is that of a calcareous sandstone sequence with diminishing grain size towards the top of the formation. 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. The often very fossiliferous sandstones have frequently been decalcified, so that only moulds of crinoids, brachiopods and especially bryozoans remained.

In the Esla Nappe the formation consists of clastic beds, the base is formed by soft coarse-grained yellow decalcified sandstones. These are topped by medium-grained white limestones (Cremenes Limestones). On top of this complex follows again a porous decalcified sandstone sequence with a thickness of 40–50 m.

Thickness. – About 500 m near the village of Nocedo. In the upper Luna area some 300 m. From the Esla Nappe, Kupke (1965) reports 250 m, however usually not recognizable.

Fossils. – Brachiopods, crinoids, bryozoans.

Murcia Formation (van Veen, 1965)
The Murcia Quartzite Formation consists of dark-coloured, well-bedded, quartzitic sandstones to quartzites alternating with thinner, dark-coloured shales. The formation grades conformably out of the uppermost beds of the Cardaño Formation and above passes into the overlying Vidrieros Formation. The basal part is usually very thin-bedded alternations of fine-grained, quartzitic sandstones with dark-grey shales. The sandstones are often cross-bedded and load casted whereas some are graded. These shales, often slates, often yield small lamellibranchs of the genus *Buchiola* which are most frequent inside rounded concretions. The major part of the formation is fine to medium grained, often a real quartzite and medium to very thickly bedded. Lighter coloured, well-rounded quartzites are typical of the thinner sections.

The thickness increases from about 130 m in the type section to more than 200 m further east where the grainsize is also noticeably coarser. Much thinner sequences of about 60 m are assumed along the Cardaño Line.

The identification of the lamellibranchs and the determinations of the conformable over- and under-lying sequences make it likely that this formation encompasses the whole of the Frasnian stage.

Ermita Formation (van Staalduinen, 1973)
The base of the formation is often typified by microconglomerates which grade upwards into coarse decalcified sandstones and ferruginous quartz sandstones. This part of the formation often has a reddish colour, mainly caused by the high iron content of the sandstones.

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

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

Age. – Comte (1959) restricts the Ermita Formation to the Upper Famennian but conodont determinations by Higgins et al. (1964) and van Adrichem Boogaert (1967, p. 160) indicate a continuation into the Lower Tournaisian.

Thickness. – At the type locality near the village of Nocedo, the Ermita Formation is about 140 m thick; north of the Sabero-Gordón Line, the Ermita Formation is usually less than 10 m thick (Evers, 1967).

Vidrieros Formation (van Veen, 1965)
The Vidrieros Formation is strikingly similar to the Cardaño Formation. It consists of thin- to medium-bedded limestone lenses or thin-bedded nodular limestone beds both interbedded with shales. The shales have often been converted to slates with strong cleavage development. The typical peanut shales in which the nodules have been tectonically oriented are also found. Trilobites, goniatites, conodonts and lamellibranchs make up the faunas found, but identification is often hampered by tectonic deformation. These rocks average 20 m in thickness ranging from 10 to 100 m in the extremes.

The ages found range from Lower Famennian to lowermost Tournaisian (van Adrichem Boogaert, 1967).

LOWER CARBONIFEROUS

Two units of Lower Carboniferous age are shown on our maps which have been generally recognized and are reasonably easy to correlate on lithostratigraphic as well as biostratigraphic grounds. That is excluding the earliest Carboniferous deposits which in fact belong to the Ermita and Vidrieros Formations already described. Despite the instability during the active transgression of these times, the Devonian-Carboniferous boundary has been recorded by continuing, if restricted sedimentation.

Vegamián Formation
The Vegamián Formation typically consists of black cherts and shales with phosphatic nodules, although sandy, even microconglomeratic beds also occur. Thicknesses up to 50 m have been reported but usually the sections are less than 30 m. This formation is not often fossiliferous but Radiolaria are to be found in the cherts: lamellibranchs, ostracods, brachiopods, cephalopods and trilobites in the shales in rare localities. Conodonts have been obtained from some sandy beds and, though many are probably reworked, they afford perhaps the best indication of the age of these rocks. A late Tournaisian age has been deduced with the possibility of some variation of the base in view of the suspicion of an unconformity there. The upper contact with the Alba Formation is also sharp in most places although in others a transition has been described. The nature of the overlying rocks and the short time-interval between their deposition do not give rise to any suspicion of extensive erosion at the contact.

Since the original description by Comte (1959) and redefinition by Evers (1967), there have been a number of investigations of this unit; Higgins et al. (1964), van Adrichem Boogaert (1967), Winkler Prins (1968) and Wagner et al. (1971), over and above the descriptions accompanying each map sheet.

Although widespread, the outcrops often seem to be discontinuous, and the unit is certainly not as persistent as implied by most of the 1:50 000 maps.

Alba Formation

The Alba Formation has a striking appearance – typically a red, sometimes green, 'griotte' (nodular) limestone often bearing a rich goniatite fauna – which has been noted from earliest times (e.g. de Verneuil & Collomb, 1853). The very fine micritic porcellaneous texture of the limestone as well as the colour set this unit off from the underlying and overlying rocks so that it has been useful as a marker horizon for mapping. Red and even green cherts and fine shales are often present and may even be traced over considerable distances (Winkler Prins, 1968). Again this formation is quite thin, usually not thicker than 30 m. In addition to the goniatites, conodonts, solitary corals and crinoid ossicles are commonly to be found as well as Radiolaria in the cherts and shales.

The Alba frequently overlaps the Vegamián but there is no lithologic evidence of erosion at the contact, and various descriptions speak of a gradual transition (Comte, 1959; Evers, 1967; Wagner & Varker, 1971). The absence of detrital deposits at localities where the Alba rests directly on Lower Palaeozoic rocks indicates that the overstep took place upon a considerably older erosion surface. However, it would seem unlikely that the break in sedimentation at this contact in any other part of the region was more important than other hiatuses to be expected in such a condensed sequence. Where clearly visible, the upper contact with the overlying Caliza de Montaña is conformable and in general gradual.

The age ranges from Lower Viséan to Lower Namurian (Kullmann, 1962; Wagner-Gentis, 1963) but, as mentioned above, breaks in this sequence have to be anticipated. There is not enough evidence to derive directions of transgression etc. as attempted by some authors.

The Alba Formation is remarkably persistent troughout southern Cantabria, especially considering the thin sequence it is (see Fig. 2). However, the outcrop on the 1:50 000 maps has had to be exaggerated for cartographic reasons, and this generalization has been extended laterally so that again the truly visible extent of the formation is considerably less than most maps imply.

UPPER CARBONIFEROUS

Generally well-differentiated, lithostratigraphic units so typical of the earlier Phanerozoic sequence are not developed in the Upper Carboniferous rocks of southern Cantabria. The majority of the sequences have been regarded as a flysch-molasse assemblage even though the Lower Carboniferous lacks the ophiolites and the 'becken' facies regarded as typical of the pre-flysch in a normal geosynclinal development (Reading, 1970). In addition the relation between the flysch and molasse facies is complex so that a simple geosynclinal model is not appropriate (Reading, 1970; Savage, 1979). Mapping

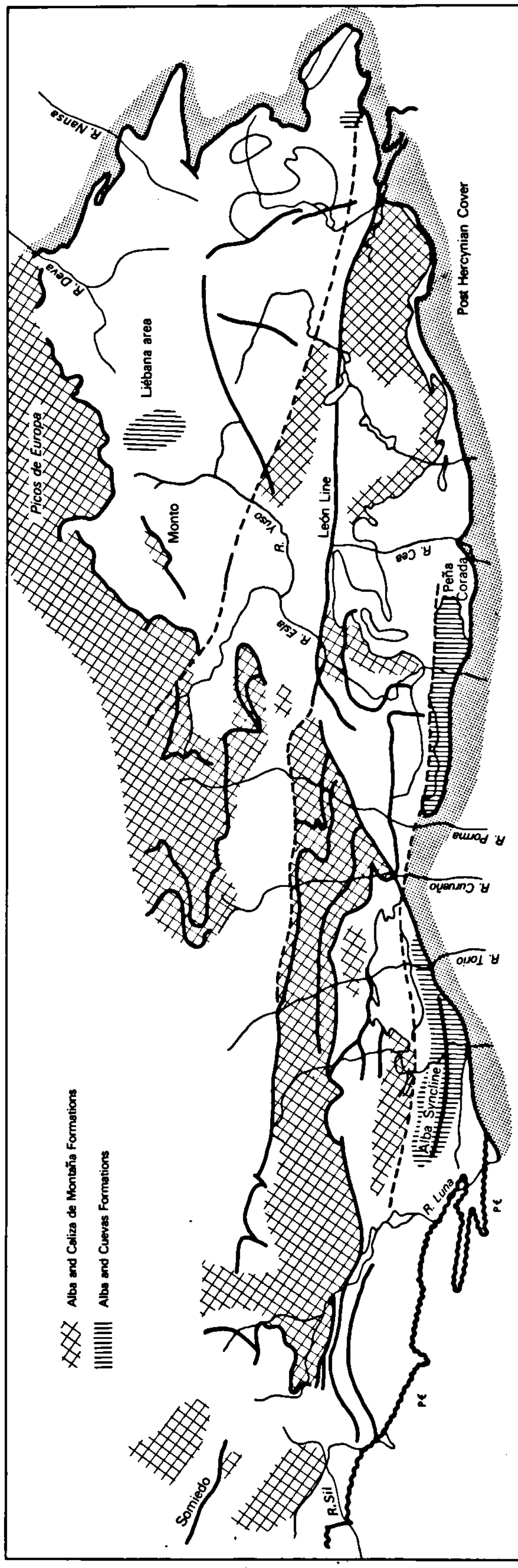

Fig. 2. Present distribution of Visean and Namurian carbonates (Alba and Caliza de Montaña Formations) and flysch (Cuevas Formation) (modified after Boschma & van Staalduinen, 1968).

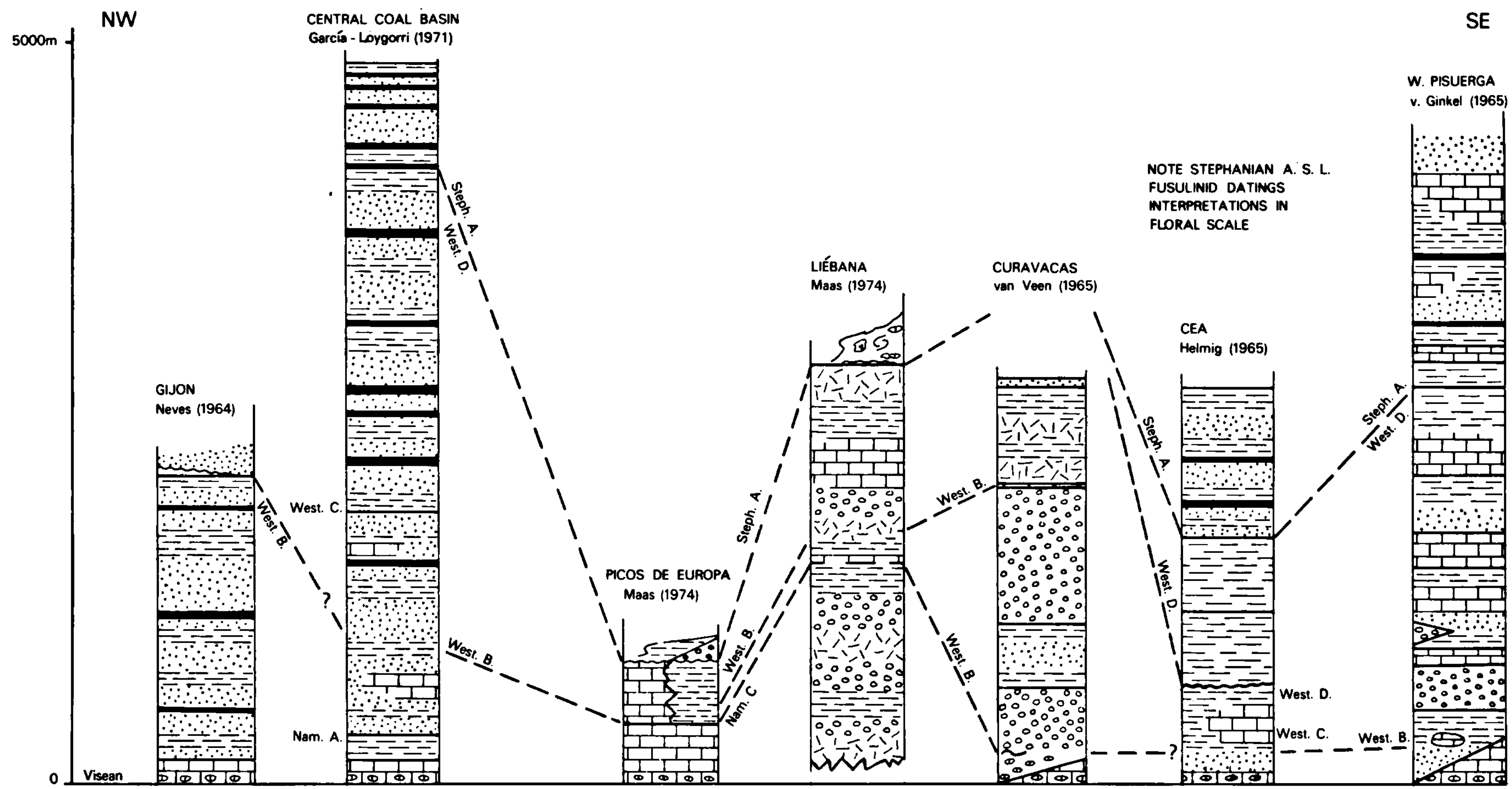

Fig. 3. Stratigraphic columns showing the variation in thickness and lithology of the Carboniferous rocks throughout Cantabria.

of these types of sediments is notoriously difficult and it is significant that, while a large number of units have been named, almost none of them have been actually mapped. In fact only very loosely defined lithosomes or individual beds can be traced in the field and even then gradual variations often force recourse to arbitrary boundaries. An impression of the variations in thickness and lithology of the Upper Carboniferous sequences to be found in Cantabria can be obtained from Fig. 3.

Basically our maps show a biostratigraphic subdivision of the sequence, roughly corresponding, as well as the data permit, to the Namurian, Westphalian and Stephanian stages. The groups, Ruesga, Yuso and Cea were formally proposed by Koopmans (1962) in the belief that mappable formations would be discovered within them. Only the Alba and Vegamián of the Ruesga Group have lived up to this expectation, and the failure to substantiate the rest of the succession with mappable formations effectively negates the proposal, although the names have been perpetuated on all the maps for the sake of uniformity. Within the units so defined, individual lenses and beds of sufficiently distinctive lithology are distinguished by colour or symbol overprinting. Not all the lithologic variations known can be expressed in this way, and there are differences between the individual styles of mapping, particularly in respect of dealing with unexposed terrain.

Admitting the primary role of biostratigraphy in classifying the sequence, reference has to be made to the source of this information. The data, neither complete nor unambiguous, are largely derived from marine Foraminifera and Algae (van Ginkel, 1959, 1965, 1971; Rácz, 1964, 1966) together with terrestrial floras (Wagner, 1971, with bibliography of many other articles; Stockmans & Willière, 1966). The majority of the samples are from spot localities but structural relationships are often complex, so that the relevance to the map picture may be dubious. Another major uncertainty is the true correlation between the differing reference sequences used – the Foraminifera of Russia, the marine Algae of N America and the terrestrial floras of France and Germany. Much work has been done since the mapping was completed (approximately 1974), and the continuing research will undoubtedly result in considerable revision of even the rough indications of our maps.

In general terms the same lithologies are to be found throughout the Upper Carboniferous although there is quite a marked tendency for terrestrial influence to increase through to the younger parts of the sequence (i.e. a progression from flysch to molasse). Local variations are also quite strong, appropriately for syn- and post-tectonic sediments, but interfingering of the major facies types indicates a complex history. The environments would seem to have corresponded to either the interior or marginal basins of Krumbein and Sloss (1963) with some shelf and unstable shelf zones developed upon continental crust (Reading, 1970; Savage, 1979).

The most striking contrasts are between the carbonates and the siliciclastics (flysch/molasse) both of which are developed into sequences more than a thousand metres thick in a number of places. The sketch maps (Figs. 2, 4 and 5) show the present distribution of these two types of the Carboniferous for various stages. These include the areas of dominant limestone development in the north (Picos de Europa), the extreme west (Somiedo), in the thrustsheets of the Río Torío (Bernesga area), as well as the southern flank of the Valsurvio Dome. In most of the other areas they interfinger with the siliciclastics. In contrast along the southern boundary, south of the Sabero-Gordón Line, the Liébana Basin south of the Picos de Europa, and even east of Mudá, the Visean Alba Griotte is overlain by siliciclastics with only minor intercalations of limestone beds. Examples of some of the sequences are shown in Fig. 2, but it must be appreciated that these are largely composite and may not represent true thickness at a single spot.

Most of the Upper Carboniferous limestones in the Cantabrian Mountains can be adequately described in terms of three main facies groups; following van de Graaff (1971b) these are: "1) the massive, mostly micritic biogenetic bank facies, which may occasionally form distinct bioherms instead of the normal biostromes; 2) the well-bedded micritic 'lagoonal' facies, which is normally unfossiliferous; 3) the clastic, fossiliferous limestones which may be differentiated as follows: 3a) the oosparitic-pelsparitic grainstone facies which originated in agitated waters and which probably indicates littoral to sub-littoral conditions, and 3b) the clastic wackestone-packstone facies, with occasional mudstone and grainstone patches. This facies often contains numerous corals, Foraminifera, calcareous Algae, brachiopods, crinoids, etc., and is intermediate between the other three, but more closely associated with 3a."

A general tendency may be noted that fossiliferous types become more common in the younger parts of the sequence although interfingering and repetition is usual, as must be expected from such closely related sediments.

The Carboniferous siliciclastics are considerably varied, and have been described under a large number of facies types of which the genesis of only a few can be said to be understood. These are the coarsening-upward (regressive), fining-upward (fluviatile), fining-upward (plant growth) and fining-upward (transgressive) 'predictable' sequences of Reading (1970). The remainder include many turbidites; Nederlof (1959), Savage (1967), Lobato (1975), van de Graaff (1971a), Maas (1974), Reading (1970). However, Rupke (1977) and Maas (1974) have shown the presence of other, mainly deep basin, facies types.

The rocks themselves range in grain-sizes from fine black shales through mudstones, siltstones, sandstones and conglomerates to the coarsest boulderbeds and olistostromes with mappable olistoliths. Generally limestone beds are associated with the finer grained shales and greywackes but they can pass into conglomerates laterally. Most of the conglomerates are in thick to massive beds, but gradational contacts are common;

particularly passage, laterally or vertically into pebble mudstones. Sharp contacts indicate predepositional erosion or mass-emplacement of the conglomerate by gravitational sliding. It should be obvious that such variations of lithology can never be adequately represented upon 1:50 000 maps.

Namurian (Upper Ruesga Group)

Carbonates of this age are shown on the maps as Caliza de Montaña Formation. All three facies types of van de Graaff (1971b) are to be found within these units, although, as a rule the Alba Griotte is overlain by a sequence of dark micrites (facies type 2) with a unit of bioclastic limestone (facies types 1 and 3) above it. This subdivision is shown (but not named) on the Bernesga-Porma Sheet but not elsewhere because weathering and postdiagenetic alteration make the distinction in the field difficult.

In the extreme north, the Picos de Europa (Fig. 2) only carbonates were laid down during the Namurian (Maas, 1974). Elsewhere siliciclastic flysch overlies and interfingers with the carbonates, except in the Liébana and south of the Sabero-Gordón Line where, in the absence of the usual basal limestones the Alba Griotte is conformably succeeded by siliciclastics (Fig. 2). All three facies limestones may be interbedded as autochthonous sediments or in slump and slide, resedimentation deposits (van de Graaff, 1971b). Brecciation is commonly observed at the contact as well as manganiferous-nodule zones (Ricacabiello Formation, Sjerp, 1967). It seems likely that these phenomena signify little more than the change in sedimentation which may occur repeatedly. These may have been controlled by tectonic movements but these will have been small and restricted in extent, not a regional tectonic event.

Siliciclastics of Namurian age also comprise the whole range of types already referred to (p. 93), although coarser conglomerates are very rare. A very typical deposit is the lydite-bearing microconglomerate. It is hardly likely that this rock type could be completely exclusive to this stage, but experience shows that it is not far from being so. The black chert (lydite) has almost certainly been derived from the Vegamián Formation, and it is logical that, with erosion and the accretion of more and more Upper Carboniferous sediments, the sources of supply of material from the youngest parts of the 'basement' would be either exhausted or choked off. Mudstone with or without pebbles is the typical matrix for slump and slide deposits but other gravity flow – turbidity, grainflow, fluidized sediment – are also to be found (Rupke, 1977).

Westphalian (Yuso Group)

Thick carbonate sequences of Westphalian age are only to be found in the Picos de Europa and near the León Line at Riaño (Fig. 4). The former consist almost entirely of van de Graaff's facies type 3 (a & b), the latter of facies 1 with interbeds of type 3 (a & b). These limestones near Riaño interfinger westward into the flysch sequence of the Central Coal Basin, usually called

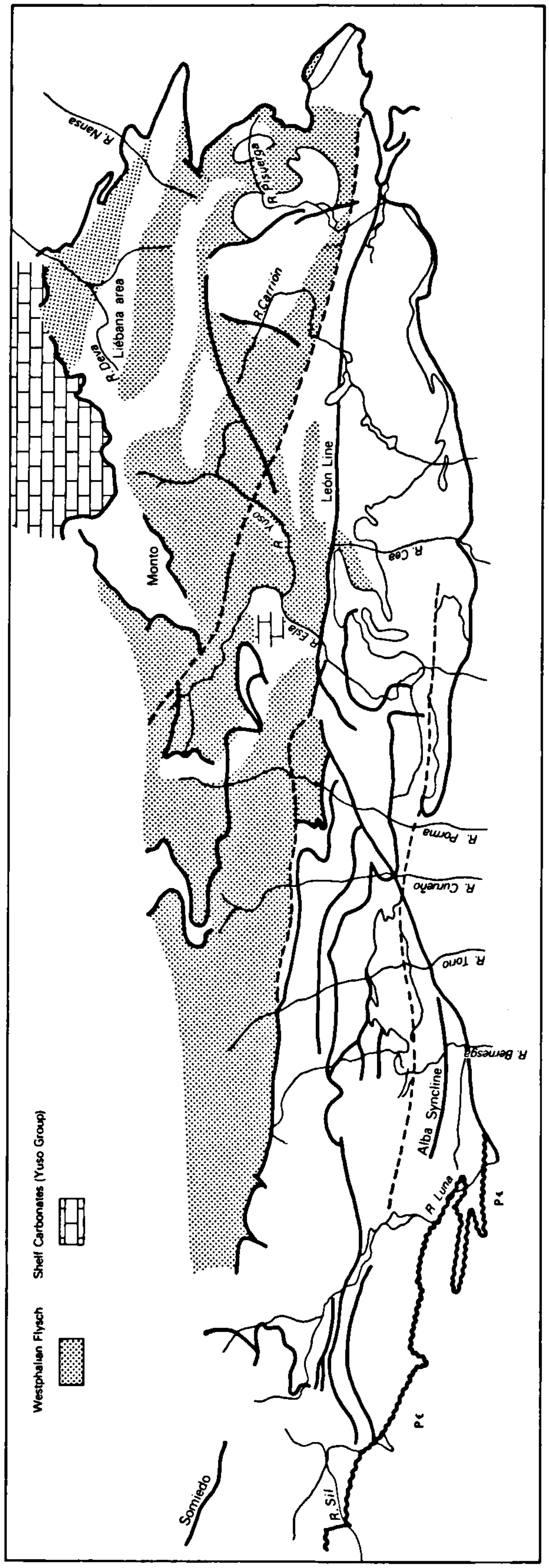

Fig. 4. Present distribution of Westphalian flysch and shelf carbonates (Yuso Group) (modified after Boschma & van Staalduinen, 1968).

the Lena Formation on our maps. In the Picos de Europa, terrestrial influence is largely restricted to thin shaly interbeds in the lower part of the Picos de Europa Formation; the siliciclastic Aliva Formation that Maas (1974) considers to be a lateral equivalent may in fact be much younger since the dates he quotes have all been derived from resedimented limestone bodies.

The siliciclastics of Westphalian age (Yuso Group) are lithologically indistinguishable from those of the older sequence, although they usually contain a higher proportion of coarser beds. Conglomerates are common, occasionally attaining enormous thicknesses as in Curavacas (Carrión Sheet) where an uninterrupted section of 800 m has been measured (van Veen, 1965). Turbidites in the classic and variant types are also common (van Veen, 1965; Savage, 1967; Maas, 1974; Reading, 1970; Lobato, 1976). Mudstone, often uncleaved as well as unbedded, is another typical sediment, and the occurrence of allochthonous blocks of limestones or other rocks within them demonstrate that many have been involved in mass or grain-flow movements (Maas, 1974).

In general, rocks of Westphalian age are not found in the Leonide thrust sheets (Fig. 4) although some have been identified south of the León Line around Prioro, Vegamián and near San Emiliano (Rupke, 1965; Evers, 1967; van den Bosch, 1969; van Ginkel, 1965). Considerable overlap exists between the ages of the rocks involved in the thrusting and those they cover so that this generalization cannot be extrapolated to detailed mapping (Evers, 1967, p. 116). In the absence of good mappable horizons no subdivisions have been made.

The sequence is mostly typically flysch although that in the Pisuerga area, to the east, has been interpreted as a prograding delta (van de Graaff, 1971a) and to the northeast as deep basin (Maas, 1974) in which the submarine delta of E Liébana developed (Rupke, 1977). Occasionally, seat-earths have been found but they are not widespread nor typical. They are most common along the northwestern and eastern edges of our mapping area, the Cuenca Central (Martínez Alvarez, 1962; García-Loygorri et al., 1971) and the Pisuerga Basin (de Sitter & Boschma, 1966; van de Graaff, 1971a), respectively.

The most striking and mappable lithosome is formed by the Curavacas Conglomerate mentioned above, although considerable difficulties in mapping are encountered due to pinchouts and gradational contacts with pebbly mudstones and limestone conglomerates (usually mapped separately). The mapped unit certainly included conglomerates of various genetic types and considerable range of age.

The frequent interbeds of autochthonous and allochthonous, marine limestones confirm the dominance of marine conditions and, since seat-earths and rootlet zones are rare, the majority of the land-plant material found in these rocks has probably been washed into the sea before deposition.

From the maximum thickness of the section at Curavacas the conglomerates thin very rapidly towards the east, beyond Los Cintos (Carrión Sheet). To the west the thinning is less rapid but perhaps more striking because the gradual pinching out is very clearly shown on the maps. This all yields a striking wedge-shaped form for the lithosome. The thicker end of the wedge rests unconformably upon Devonian between Curavacas and Cervera de Pisuerga, as well as near Peña Prieta (Carrión Sheet). Elsewhere the unconformity once considered to define the base of the Yuso Group (Koopmans, 1962; Boschma & van Staalduinen, 1968), is by no means clear (Sjerp, 1967; Maas, 1974). The presence of slide conglomerates in the Los Cintos area (Carrión Sheet; Reading, 1970) lessens the significance of the discordance with the Ruesga (Namurian) rocks to the south, so that it can by no means be taken as certain that this represents a single, regional tectonic event as proposed by Kanis (1956).

Our maps show the obvious lithosomes within this sequence such as the conglomerates, limestones and occasional, massive sandstones. These are not, however, constant facies indicators so that the relations to one another and to the sandstones and shales of the innumerable facies types they represent vary enormously. Understanding of the threedimensional shape of the lithosomes is very sketchy so that sections and palinspastic reconstructions must be very speculative.

Carboniferous Westphalian igneous rocks include basic and acid intrusives and extrusives. The latter are particularly common around Riaño (Yuso Sheet), where brecciated tuffs occasionally slumped, as well as vesicular lavas have been found. The form of the bodies has been obscured by intense deformation so that little can be said of relationships. While varying considerably, most of the extrusives can be classed as dacites (Dr. T. Perekalina, pers. comm.). One near Santibañez de Resoba (Carrión Sheet) has been described as spilitic (Koopmans, 1962)

The largest intrusive is that at Peña Prieta (Carrión Sheet), which appears larger on the map because the tabular body is exposed along a dipslope. It has been described as a granodiorite porphyrite (van Veen, 1965), although there seem to be considerable internal variations within the mass, and it may even be a composite intrusion. Contact metamorphism has formed chiastolite (andalusite), muscovite, reddish brown biotite and garnet (van Veen, 1965).

Chloritization, albitization, prehnitization and calcification affect many of the rocks rendering thorough analysis difficult. Pyrophyllite has also been identified in shales of the North Liébana accompanying the formation of the slaty cleavage there (Rupke, 1977). In the Murcia Anticline and the Valsurvio Dome, chloritoid porphyroblasts are common in the Devonian shales (Koopmans, 1962; van Veen, 1965). A general, if patchy, very low grade of lower greenschist facies metamorphism is deduced.

Stephanian (Cea Group)

Sediments of Stephanian age are scattered over the whole region, usually resting unconformably upon the older Palaeozoic and Precambrian rocks, as can even be seen in the sketchmap (Fig. 5), hence they have been mostly easy to map.

Helmig (1965) provided the first comprehensive study of these rocks in what has been considered as the type area along the Cea River. He called it a paralic sequence emphasizing the dominance of limnic deposits, although more recent workers interpret marine intercalations to a greater degree (Wagner et al., 1969; Reading, 1970; van Loon, 1972). Very large proportions of marine sediments have been described from outlying areas such as the Pisuerga Basin (Nederlof, 1959), the Valdeón (Boschma & van Staalduinen, 1968) and the Picos de Europa (Maas, 1974).

In general the rocks are "...fluviatile, piedmont fan and paralic deposits, the typical sediments of the molasse" (Reading, 1970), and hence reasonably distinctive on lithological grounds. Within the unit, however, changes are by no means distinctive or constant laterally so that, although some sequences have been given different formation names, these generally refer to geographically isolated patches of rocks of similar lithology and not to any distinctive sequences. Only Helmig (1965, p. 89) subdivided the succession in the Valderrueda Basin on lithologic and palaeobotanic grounds "... Thus, in fact, two cycles can be recognized in the Cea Formation: a Westphalian lower cycle with quartzite conglomerates at its base and no limestone conglomerates, and a Stephanian upper cycle with limestone conglomerates in its lower part and void of quartzite conglomerate." There are, of course, exceptions to the rule, for the occurrence of the conglomerates and more detailed knowledge of the floras may modify the biostratigraphic level quoted (e.g. Wagner et al., 1969). The scale of lateral changes in this sequence is illustrated on our maps by the variations in thickness and number of the conglomerate levels which are in the nature of the facies "...transitional between the predictable sequences of migrating rivers and the largely conglomeratic piedmont fans" (Reading, 1970, p. 29). In such conditions it seems hardly likely that any bedding plane will form a bio- and lithostratigraphically correlable horizon over the distances covered by our maps. The more recent studies in which many subdivisions have been recognized will not be commented on in this account but may be found in Wagner & Varker (1971), van Loon (1972), Knight (1971).

In one area – the first and most extensively studied – of the Pisuerga Sheet, no basal unconformity is evident, and interpretations of the sequence have differed; see van de Graaff (1971a) for discussion and bibliography. The rocks are largely marine sediments, contrasting the usual molasse, and consisting of alternations of shallow-, marginal-, and deep basin facies (Reading, 1970). The numerous coal seams here represent the extreme of the cycles in which they were formed and as opposed to the sequence of events in the earlier 'high-destructive' delta system of van de Graaff (1971a), later conditions allowed many coals to be preserved. The almost unique alternation of marine and terrestrial fossils has not yet led to the superior worldwide correlation that might be expected, but active research is still in progress (Wagner,

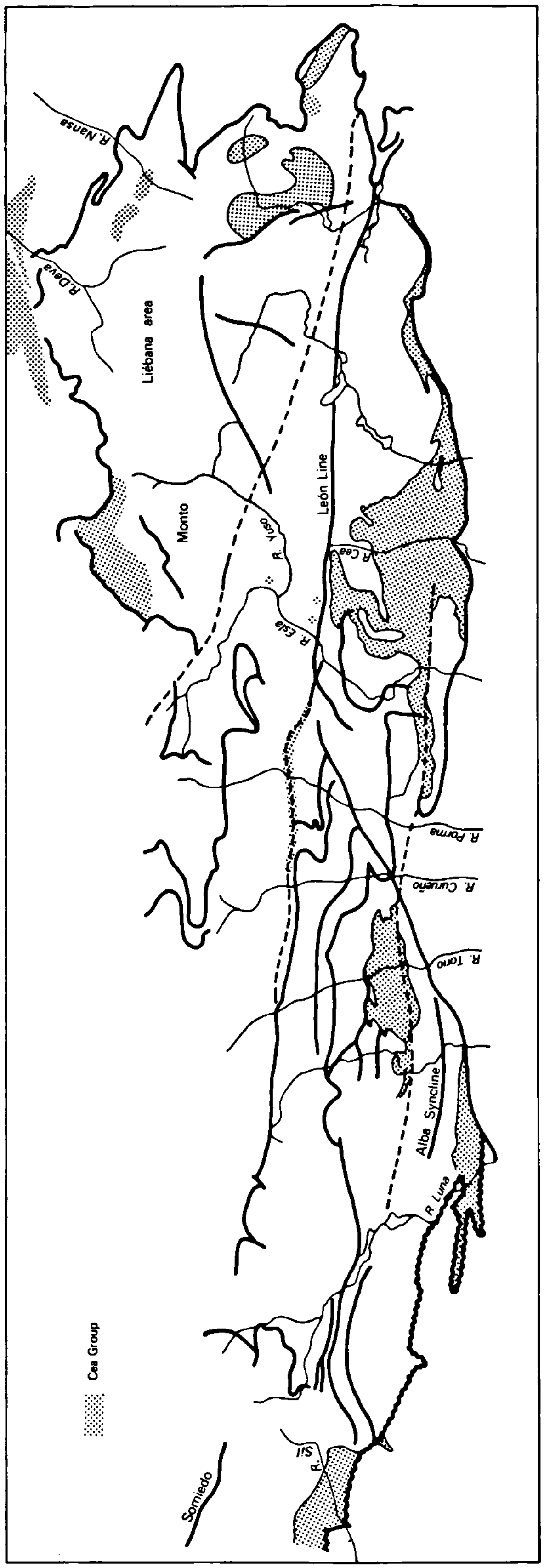

Fig. 5. Present distribution of Stephanian molasse (Cea Group) (modified after Boschma & van Staalduinen, 1968).

1971, and van Ginkel, 1971 with bibliographies). The truly continental deposits of Stephanian age in this area – the Peña Cilda Formation – also unconformably overlie the barely older deposits (Reading, 1970). The latter author postulates continuing tectonic activity to account for the gross, unpredictable changes in the sedimentary sequences. Accepting this does not, however, imply that such events were necessarily felt throughout the Cantabrian Mountains. Rather the nature of the model suggested by Reading (1975) with strike-slip faulting controlling the development of basins and their margins, would favour intermittent, migratory movements.

Another area with atypical sequences of Stephanian age is that in and around the Picos de Europa (Fig. 5). Maas (1974) describes them under different names in the different structural units although they are largely similar and show some similarities with the upper part of the Pisuerga succession. They are less well known and less well dated due to the absence of workable coals and floras. The limestones bearing marine faunas are often allochthonous slide masses so that their ages only set a maximum for their emplacement. Maas (1974, p. 406) even claims that units "belonging to the uppermost Westphalian or Lower Stephanian, have, as far as could be observed, an unconformable basal contact with older strata." The positive identification of unconformable relations is probably less common than might be deduced from the statement above, but the stratigraphic position of the strata remains unquestionable.

In the Valdeón area palaeontological data are very scarce, and Boschma and van Staalduinen (1968) correlated these rocks with the Yuso Group. As argued by Maas (1974, p. 405) it would seem more probable that they correspond to younger Stephanian units within and around the Picos de Europa (as shown on the Yuso and Deva Sheets). Conglomerates are particularly common in these sequences and some, at least, are allochthonous slide conglomerates (it should be noted that many beds and lenses are erroneously shown as limestone around Soto de Valdeón on the Yuso Sheet; they are in fact conglomerates).

PERMIAN

Labra Formation (Maas, 1974)
A sequence of volcanic conglomerates and breccias, tuffs, reworked tuffs and tuffaceous sandstones and shales, is exposed along the south slope of the Peña Labra. The lower boundary is a sharp angular unconformity with the Carboniferous rocks. At the top, the Labra is cut off unconformably under a low angle by the basal conglomerates or coarse sandstones of the Triassic Bunter facies. In the type section on the SW slope of the Peña Labra, a lahar member, a volcanic agglomerate member, a tuff member and red beds are distinguished.

The type section was chosen for its good exposure and accessibility, but the Labra Formation varies considerably throughout the known outcrops. The Labra Formation must be younger than the youngest Carboniferous known in this area. A restricted flora from the upper part of the tuff member NW of San Mames yielded a *Walchia piniformis* specimen, hence the suggested Permian age.

TRIASSIC AND JURASSIC

The folded Palaeozoic of the Cantabrian orogene is covered in the east unconformably by red continental deposits, which are supposed to be of Triassic or Permo-Triassic age. Along the southern border the Triassic forms a narrow band from Cillamayor to Ligüérzana, then it disappears, and from south of Cervera the Cretaceous in Wealden facies reposes directly on the Palaeozoic.

The Triassic represented in its germanic red-bed facies reaches a considerable thickness, and the character does not change. Towards the top the red mudstones may contain cavernous dolomites and gypsiferous marls, which are supposed to belong to the Keuper. Even salt has been encountered (Salinas de Pisuerga).

Maas (1974) included the Triassic deposits in the north-east in the Nansa Unit.

TRIASSIC

Nansa Unit
This unit consists of the Bunter facies, resting unconformably on the Labra Formation. Thickness of the unit varies from about 400 m E of Lebena to more than 1000 m in the road section of the Nansa Valley. Usually, the sequence starts with a quartzitic conglomerate. The conglomerate is lithologically very homogeneous, consisting of well-rounded quartzitic pebbles and boulders in a coarse quartzitic-sandstone matrix. Pebble beds alternate with layers of usually cross-bedded sandstone, and shale. The pebbles frequently show imbrication. At the top, the conglomerate passes into cross-bedded coarse-grained sandstones, alternating with gritbeds and thin shale intercalations. About 100 m above the basal conglomerate, a thinner and laterally less extensive conglomerate is locally developed. The total sequence is fining upwards, the upper part of the Nansa Unit consists predominantly of silts and shales. The colour – pale red or neutral in the coarser parts – is deep purple in the silts and shales, which are very similar to the silts and shales of the red beds of the Labra. The Nansa Unit passes vertically through a sequence of clays and marls into the calciclastic Tudanca Unit.

JURASSIC

Tudanca Unit
This unit consists largely of well-bedded, grey-weathering micritic, often recrystallized limestones, occasionally alternating with calcareous shales. Some mollusc-shell beds were discovered near the base of the Tudanca Unit, but no determinations resulted in an age indication.

Maas (1974) considered his Tudanca Unit to be of Jurassic age.

The Jurassic limestones along the southern border between Cervera de Pisuerga and Cillamayor lie conformably and slightly transgressively over the Triassic. The whole Jurassic is built up of limestones, shales and marls. At the base lies a series of massive, grey, and slightly breccious limestones in which the bedding plane can hardly be observed. Next follows a sequence of well-bedded, grey marls and limestones which contain ammonites and brachiopods.

Wagner (1955) and de Sitter (1955, 1957) mapped the Villanueva Limestone as a Jurassic rock, but determinations of conodonts and goniatites collected by Frets (1965) indicate a Lower Carboniferous age.

CRETACEOUS

The Cretaceous rocks lie conformably upon the Jurassic and unconformably upon the Palaeozoic rocks.

Wealden facies (Helmig, 1965)
A great part of the Palaeozoic core of the Cantabrian Mountains is bordered to the south by continental sediments which are comparable to the Wealden facies of the English Cretaceous. They are red to pink clays, white to pink, kaolinitic sands and gravels with lignitic streaks and lenses.

The age determinations of the Wealden clearly indicate that it is a diachronous deposit. In the east, in the province of Burgos, it is reported to be of Albian age (Ciry, 1939), while in the west, near Aviados in the province of León, pollen from lignitic lenses point to a Turonian age (van Amerom, 1965).

Cretaceous rocks crop out in a continuous band of 1 km width from Cervera de Pisuerga in the east to Guardo. Further to the west this strip becomes discontinuous. The oldest rocks of this zone in our area are developed in a Wealden facies; the limestones and marls of the marine Cretaceous are stratigraphically superposed on the Wealden deposits. The individual limestone or marl beds are separated by intercalations of more clastic sediment.

Evers (1967) distinguished a Voznuevo Formation to be correlated to the Wealden and a Boñar Formation.

Voznuevo Formation
This sandstone formation is named after the village of Voznuevo 2 km E of Boñar. Along the road from Boñar to Collé, many exposures can be found in the clay and gravel pits in these gently south-dipping strata.

The multicoloured sandstones with lignitic clay and gravel lenses resemble the Wealden-facies. The well-rounded quartz pebbles are generally coated purplish-red by a hematite film. Iron concentrations, lignite, large muscovite flakes, kaolinite and the vivid colours, shading from white to purple, are the most characteristic features. The upper part of the Voznuevo Formation is composed of finer grained, micaceous sandstones with carbonaceous clay lenses, but gravel streaks still occur. Pollen assemblages from the carbonaceous clay lenses indicate the lower part of the Upper Cretaceous, probably only the Cenomanian and Turonian (van Amerom, 1965).

Boñar Formation (Evers, 1967)
The Boñar Formation is composed of arenaceous limestones and marls. The thickness, composition and age change gradually from east to west. The type-section was measured along the path from Voznuevo to Las Bodas; the village of Boñar is built on the western extension of these limestones. Overlying the glauconitic transition zone, which according to Ciry (1939) might represent the top of the Coniacian in the Porma area, the following members are found from base to top:
a. Arenaceous marls and micaceous sandstones grading into coarse-grained, thickly bedded and well-cemented calcareous sandstones with yellow to reddish colours.
b. An alternation of light-coloured limestones and arenaceous marls.
c. Argillaceous sandstones grading into calcareous sandstones and arenaceous limestones.
d. Thickly bedded arenaceous and dolomitic limestones and sandy marls.

TERTIARY

Along the southern boundary of the area the Cretaceous is covered by Tertiary sediments of the Duero Basin. They are mostly reddish continental deposits. The subdivision into Lower and Upper Tertiary is based on the existence of an angular unconformity between those two units. Mabesoone (1959) based this conclusion on lithologic correlations with other areas where fossils have been found. Rupke (1965) mapped a lower series consisting of mostly well-lithified conglomerates with sandstone-shale lenses. Pebbles consist of Cretaceous and Palaeozoic material. For a detailed sedimentpetrological description the reader should consult Mabesoone (1959), who describes the two above-mentioned conglomerates as rocks of the Cuevas facies.

Vegaquemada Formation (Evers, 1967; van Staalduinen, 1973)
Evers defined a Vegaquemada Formation, a Candanedo Formation and a Riacos Formation.

The type exposure is situated on the left bank of the Porma River, east of the village of Vegaquemada. Reddish brown argillaceous sandstones, thin conglomerate beds, claybeds and micaflakes all through the sequence are the most characteristic features. The lower part is mainly composed of argillaceous sandstones. Only when unconformable contacts with the underlying Voznuevo Formation or the limestones of the Boñar Formation are present, could the sandstones be distinguished from those of the Voznuevo Formation. The upper part is characterized by increasing grainsize of the sandstones, and channelling of coarse-grained sandstones in finer sandstones.

Candanedo Formation (Evers, 1967; van Staalduinen, 1973)
This formation is named after the village of Candanedo

in the lower Porma Valley, where the limestone conglomerate beds are best exposed.

The transition from the Vegaquemada Formation is very gradual; the first thick limestone conglomerate bank is regarded as the base of the Candanedo Formation. In the lower part of the formation, the poorly sorted limestone boulders may reach nearly 1 m^3 in size, but are smaller and more rounded rowards the top. The matrix consists of coarse sand or gravelly material well-cemented by calcite. Thin red sandstone lenses that intercalate the thick limestone conglomerate banks mark the stratification. Channels cutting through several underlying strata and filled with finer-grained fluviatile deposits are due to the shifting courses of torrential rivers, but the limestone boulder banks are best explained by gravitational sliding down tectonically steepened slopes. The frequency and dimensions of the boulders derived from the Boñar Formation decrease towards the top, while the proportion of Palaeozoic limestone and quartz-sandstone pebbles increases.

Riacos Formation (van Staalduinen, 1973; van den Bosch, 1969)

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.

The post-Hercynian sediments, in the Riello Basin and north of Robles de Laciana, closely resemble the Miocene deposits further to the south and hence have been provisionally included in the Riacos Formation.

Coarse boulders of quartzites alternating with sandstones and sandy clays are the most characteristic rocks of the Riacos Formation. The beds lie in an almost horizontal position resting slightly unconformable upon the underlying rocks. 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; locally clay balls were found. Coarse cross-bedding and channels are common.

CHAPTER 3

STRUCTURES

Real insight into the structures of Southern Cantabria was first possible after Comte's monumental work following on that of early Spanish and French pioneers like Barrois, Casiado de Prado and Délépine. Although summarily expressed, Comte demonstrated the multiplicity of the folded thrust sheets here with the greatest clarity. The Leiden school has elaborated and enlarged the picture with semidetailed mapping especially by extending it into areas covered by Upper Carboniferous rocks.

The regional structure has been discussed in a number of papers, e.g.: Julivert (1971), Wagner & Martinez-García (1974), and Savage (1979) but the discussion will not be entered into here. Rather, it is hoped to present the geometry of the rocks as expressed in the maps and sections. The kinematic analysis will not go further than the interrelationships between the structures and textures to be found in the area covered by the maps.

The major subdivisions of these mountains, revealed in the stratigraphy of the Palaeozoic rocks, are evident in their structure too. The Leonides, south of the León Line have folding and thrusting with northerly vergence as typical structural style, although backfolding (Rückfaltung) and faulting complicates the picture. North of the León Line the general picture is less easy to specify. A large part is covered by Upper Carboniferous sediments, mostly disharmonically folded with respect to the underlying older Palaeozoic. In the north, the Picos de Europa, we see complete detachment of the, largely limestone, Carboniferous sequence which are in stacks of thrust sheets with a southerly vergence. The older Palaeozoic sequences are also present in folded multiple thrust-sheets in the San Isidro-Tarna area (Ponga Nappes) with an easterly vergence and in thrust-folds in the Carrión area again with a southerly vergence.

Together these structures represent all parts of the wellknown 'Knee of Asturias'. The Leonide thrust sheets are the southern segment of the arc, the Ponga Nappes the N–S segment (despite the extensive E–W folding), and the Picos de Europa represents the northern segment of the arc. Hence our map area contains all the elements of this enigmatic structural phenomenon.

STRUCTURES OF THE LEONIDES

The most westerly structures are the simplest, so we shall begin there, in the area of the Luna-Sil Sheet (Encl. I; van den Bosch, 1969).

Narcea Anticlinorium

The Narcea Anticlinorium is a very large structure dividing the Cantabrian Zone from the W Asturian/Leonese Zone of Lotze (1945). Only a small part of the Precambrian core and the northeastern flank are covered by the map, and no real structural mapping has been done in the former. Nevertheless such evidence as there is shows that bedding and mesoscopic structures are cut off by the unconformable base of the Herrería Formation. The multiple cleavage development in the Mora Slates is not to be found in the Palaeozoic sediments so that an essentially Precambrian deformation must have affected this area.

The attitude of the Palaeozoic sequence demonstrates the later development of the major structure while the large angle of discordance between the base of the Stephanian Prado Formation and that of the Cambrian Herrería illustrates the Hercynian age of at least much of the folding (the Palaeozoic sequence is otherwise conformable (p. 78)).

Salce Unit

The Salce Unit is essentially the SE extension of the core of the Narcea Anticlinorium which has been extensively blockfaulted. These faults not only control the present outcrop pattern of the unconformable Prado Formation outliers but also dictated the sedimentary development during the Stephanian. E–W normal faulting dominates, but along the northern boundary bedding plane faults appear, especially where the Lancara Formation is involved.

The Precambrian core of the Narcea Anticlinorium disappears rapidly under the unconformable Herrería and Prado Formations in the Bernesga Sheet. The structure must plunge eastwards although the Cambrian unconformity is often near vertical or even overturned locally (de Sitter, 1961). As in the Babia Alta a set of relatively simple, plunging synclines – the Alba Synclinorium – have been formed in the otherwise little disturbed NE flank of the Narcea Anticlinorium (Encl. IA: Sections 1, 2 and 3).

Babia Alta Unit

The Babia Alta Unit is the Palaeozoic NE flank of the Narcea Anticlinorium, north of Piedrafita de Babia. It is characterized by two folding regimes disharmonically developed but probably synchronous or at least continuous. The upper part of the sequence, younger than the La Vid Formation is folded into broad synclines with narrow, even faulted anticlines between them (Encl. IA: Sections 9, 10, 11 and 12). The individual Devonian limestone formations have developed complex systems of minor folds and faults in places, but in general they are folded harmonically. The strong variation in width of the outcrop of the La Vid Shales demonstrates their role in decoupling these structures from those below.

In the lower Palaeozoic sequence the fold system is dominated by the forms interpreted as an anticline and a syncline with curving axial planes. In fact the thrusts north of Cospedal, telescope another pair of folds into the northern syncline, so the WSW axial trend shown is an artefact of those movements, probably accentuated by later refolding. The true fold axes plunge to the west with the angle of plunge increasing in that direction. The thrusting along the base of the Lancara Formation into the Carboniferous San Emiliano Formation certainly had a strong easterly component of movement but other components may also have been important.

Babia Baja Unit

Beneath the folded Babia Thrust, the Babia Baja Unit, in which the Devonian sequence above the La Vid Formation has thinned to the distinctive Caldas Formation, has folds of relatively long wavelength incorporating harmonically almost all of the Palaeozoic sequence. The structural pattern is dominated by the narrow, E–W Villasecino Anticline which is much faulted. The other folds are not so tight, and a general westerly plunge allows the development of open, divergent folds in the San Emiliano Formation. The León Line probably peters out against the Narcea Anticlinorium here, although an effect can be traced on through the Villablino Basin. In this sense the Babia Alta and Babia Baja Units might be excluded from the Leonides. However, both stratigraphically and structurally the westernmost West Asturian units resemble the Leonides very closely.

To the north of the Villasecino Anticline the broad, triangular-shaped Tesa Syncline illustrates the interference of the N–S trends of W Asturias with the E–W trend of the Leonides, which has caused the double-plunge of the synclinal axis. This structure is also asymmetric probably as a result of the differences in the stratigraphic sequence between the north and south flanks. Parasitic folds are common in the south limb, normal faults around the eastern termination, and indeterminate tectonic thickening of the La Vid Shales in the south again.

The Ubiña Structure is an overthrust anticline identical to those in the Babia Alta Unit. The major anticline plunges quite steeply SSW and would seem to belong to the West Asturian fold system (de Sitter & van den Bosch, 1968). The axial thrust continues into wrench faults in the eastern and western (not on 1 : 50 000 maps) limbs, which supports a kinematic model with movements from west to east here.

Luna Unit

The Luna Unit, wedged between the Narcea Anticlinorium and the Babia Baja Unit, is largely made up by the long complex Abelgas Syncline and the thrust slices of the Aralla Zone. The Rozo Thrust at the base of the Lancara Formation has been thrust over and refolded with the San Emiliano rocks of the Babia Baja Unit. Numerous partial repetitions of the sequence have been caused by small upthrusts, and the whole zone is overridden by the north flank of the Abelgas Syncline along the Abelgas Thrust. Similar thrusts are to be seen in the syncline, back-limb thrusting along the north flank, and relief for the space problem in the core of the structure. The role of the Formigoso, La Vid and Huergas Formations as incompetent, detachment horizons is particularly clear in Sections 1 to 7 (Encl. IA).

The opposition of the thrust relations of the Babia and Rozo Thrusts on either side of the westerly extension of the Babia Baja Unit is the Knee of Asturias in a nutshell. Much of the difficulty in understanding the structures here lies in the refolding of the thrust sheets and the early folds. If all the structural correlations suggested by van den Bosch in the map are accepted, the geometric problems would appear to be enormous. However, the extremely fractured nature of the strata SW of Huergas suggests that the structures may be discontinuous along a line somewhere between the trend of the Villasecino Anticline and the Granjos Fault. This would then be the westerly extension or a splay of, the León Line.

The other feature of the structures here is their strict compliance with the lithologic sequence, so that the tightening of the concentric folds leads to predictable disharmony facilitated by incompetent levels in the Lancara, Formigoso, La Vid and Huergas Formations. This decollement is a feature characteristic of the Leonides and, in fact, of the mountain chain (Julivert, 1971; Savage, 1979). It is also important to realize that despite the apparent concordance of the Palaeozoic sequence directly overlying the Precambrian core of the Narcea Anticlinorium, very long E–W segments are interrupted by strike-faulting. These structures disappear under the Stephanian cover of the Villablino Basin and are not evident in the strata striking NW in the Babia Alta Unit on the other side of the Basin. These faults could imply that the southerly Mora and Herrería Formations have not been subjected to the thrusting movements evident in the younger rocks.

Aralla Zone

The Aralla Zone continues into the area of the Bernesga Sheet with folds and faults repeating the Devonian sequence. The Abelgas Thrust is mapped as a continuation of the Bregon Thrust further east, but the connection is by no means clear, since west of the Bernesga Valley San Pedro is in contact with La Vid so that little throw can be proved. This zone is characterized by very varied folding, usually tight, mostly involving only the La Vid and San Pedro Formations; where the higher Santa Lucía Formation is seen the folds are clearly disharmonic. The extreme development of disharmonic folding occurs in the folds at the boundary between the La Vid Shale and limestone members. These intraformational folds are inclined to the south, even recumbent, mostly tight but occasionally isoclinal. This unit is bounded to the north by the steep Rozo Thrust mostly at the base of the Lancara Formation. The cross-sections do not extend through this front which is in fact much more complex than is possible to present on a 1:50 000 scale (see van Staalduinen, 1973, figs. 20 & 21).

In the eastern part of the Luna Unit the Abelgas Syncline splits into three distinct structures: the Alba Synclinorium, the Mirantes Anticline and the Pedroso Syncline from south to north.

Alba Synclinorium

The Alba Synclinorium is simply and clearly expressed in the uppermost Devonian formations. Subsidiary strike faulting causes some repetition in the SW flank of the syncline, but complex structures are only present in the core of the structure south of Mirantes along the shore of the Embalse de Luna, north of Cuevas, and from the Río Bernesga east to the edge of the sheet. As shown by the sequence of cross-sections (Encl. IIA) the double axes in the west (Section 5) are asymmetrically developed with respect to the major structure, apparently due to variations in the thickness of competent members within the Cuevas Formation. Complexity and asymmetry increase to the east, culminating in a convergent fan of minor folds in the valley of the Río Torío (Encl. IIA: Section 13). The rocks of the Cuevas Formation have also developed a series of intraformational minor folds which are not evident on the map. Van Staalduinen's analysis (1973) shows that the majority may be regarded as parasitic folds with a few cascade folds, probably associated with a late stage of the development of the structure.

Mirantes Anticline

The Mirantes Anticline is indicated as the Sabero-Gordón Fault Zone on the cross-sections 1 to 4 (Encl. IIA) but in fact represents the anticline between the Alba and Pedroso Synclines. The major structure is tight and frequently faulted as well as complexly folded. The broad axial zone is clearest just N of Mirantes as well as between Los Barrios de Gordón and San Mateo, 2 km E of La Pola de Gordón (Encl. IA: Sections 7, 8, 9, 10) (note that San Mateo has been remapped because of appreciable errors in the coloured map sheet; see van Staalduinen, 1973, fig. 18). Of the parasitic and other minor folds only those of Mirantes are fairly complete, the structures to the east being cut by a number of strike faults. The Santa Lucía Limestone at San Mateo has the form of a faulted asymmetric syncline in the axial zone of the anticline so that the latter has the form of a box-fold (Encl. IA: Section 10; van den Bosch, 1969, p. 201; van Staalduinen, 1973, p. 183).

Pedroso Syncline

The Pedroso Syncline to the north of the Mirantes Anticline is a tight fold with a slight easterly plunge and its axial plane inclined to the south. The flanks are marked by two strike faults which have no obvious relation to the fold but rather seem to be synthetic step-faults related to the Southern Boundary Flexure (p. 111), with downthrow to the south. At the western nose, near Pedroso, the axial plunge is steeper and the splay folds along the north flank become practically isoclinal. The Aralla Zone continues further eastwards through the area covered by the Bernesga Sheet in a broad belt within which a number of distinct structural elements can be distinguished – the Rozo, Pozo and Bregón Units.

Rozo Unit

The Rozo Unit is the complex of folded thrusts to the south and west of Villamanín stretching as far as El Rozo Mountain and possibly finding a further extension in the La Plata structures. In the east two distinct thrusts are evident, the southernmost being interpreted as the continuation of the bounding thrust to the west, here cutting upwards in the stratigraphy of another, earlier sheet which only outcrops further to the east. This Correcilla Nappe reaches much fuller development in the area covered by the Torío-Curueño Sheet (see below, p. 103). Kinematically, if the southerly origin of the thrust sheets has been correctly interpreted then this implies the Rozo Thrust cut downwards into the underlying Correcilla Nappe as the former structure moved north.

Where the two thrusts only effect a double outcrop of the Lancara Formation, together with portions of the

Oville south of Rodiezmo, complex folding, refolding and faulting of the slices together with parts of the underlying autochthonous Lena Formation has taken place. Inverted thrust slices have been folded mostly on too small a scale to be shown on 1 : 50 000 maps but this has been adequately illustrated by van Staalduinen (1973, figs. 20 & 21). These folds seem to correspond to the isoclinal folds in the Barrios and Oville Formations of El Pozo to the west. It may well be that only a part of the usually competent Barrios Formation is involved as the wavelength of the folds may be less than 300 m (cf. the absence of the upper part of the Barrios Formation in the Aralla Zone in van den Bosch, 1969, pp. 150–151). The curving fold trends shown on the map seem to be the result of later refolding of WNW–ESE trending folds such as occur near Aralla, possibly about NNE–SSW trending axes so common to the south. Almost all measured axes plunge to the west so that the inverted thrust sheets would be expected to disappear beneath the autochthonous Lena rocks in that direction, as indeed can be seen on the map.

The isolated strip of Lancara and Oville rocks through Peña de la Plaza has been interpreted as a down-faulted slice of the Rozo Nappe but there is no evidence of such a structure in the surrounding Carboniferous rocks. The curious Lower Carboniferous/Upper Devonian inlier of Collada de Alonga must be connected with the Plaza Structure but simple faulting is inadequate to explain the outcrop pattern. It seems more likely that synsedimentary thrusting/gliding movements have been responsible.

Pozo Structure

The Pozo structure at Peña del Pozo is essentially an overthrust periclinal nose which has to be related to the overridden syncline of Villasimpliz to the north because of the way they both die out to the west. (N.B. The Oville Formation in the core of the anticline has not been distinguished from the San Pedro Formation of the northern flank through a colourprinting error. This is clear by comparing the Bernesga (Encl. II) with the Torío-Curueño (Encl. III) Sheet).

The anticline is inclined to the south, and the overthrust inverted by refolding with south vergence but the fold axis still plunges quite steeply NW almost down the dip of the axial plane. The divergent axial traces shown on the map imply variations in axial dip rather than direction. The full development of this structure is to be seen further east in the Torío-Curueño Sheet.

Between the Ríos Bernesga and Torío the fault plane cuts through many folds in the Palaeozoic of the Correcilla Unit as well as some folds in the basal Lancara of the Pozo Nappe itself. In the Torío Valley the Cambrian Lancara Formation has been brought into contact with Upper Carboniferous limestones. Further evidence is obscured by the unconformable Upper Carboniferous Prado rocks, until just west of Correcillas village the Pozo Thrust reappears and continues eastwards through Valdorría cutting off the Correcilla Unit entirely again. This results in a bilateral symmetry for the latter, coincident with the variation in axial plunges found (p. 103).

The Montuerto Syncline at the eastern termination of the Pozo Nappe is tight, with a steeply plunging, curving axis and a number of complications on the south flank. The complex folding and brecciation near Valdepiélago suggest more severe deformation here, and the complex redoubling of the Lancara may be an indication of the proximity of a root zone.

Bregón Unit

The Bregón Unit is the most southerly of the Leonide thrust sheets, although it is very unimposing because the thrust is involved in folded alternations of San Pedro and La Vid rocks between the Ríos Bernesga and Luna. Further east in the Bernesga Valley the structure is closely similar to that of the Pozo Unit with a periclinal nose plunging west above the thrust plane as far as Bregón Mountain. There is no sign of a corresponding syncline below the thrust east of the Río Bernesga, but westwards, between Buiza and Tolledo one is completely developed in the La Vid and San Pedro Formations. Where the thrust cuts through the Barrios Formation the Santa Lucía Limestones have been thrown into tight, steeply plunging folds, others (not all shown on the map), occur between Bregón and La Campa. At the easternmost tip of the Matallana Basin a short outcrop of Lancara Formation is thought to represent the extension of the Bregón Nappe since these rocks rest on the Huergas Formation just as they do where the Bregón Thrust is last seen in the west and the Esla Nappe even further east. There can be little more of the Bregón Unit to the south because the pre-Stephanian rocks which appear there from beneath the Matallana Basin form a continuation of the Pedroso Syncline and the Mirantes Anticline.

It must be assumed that all structures to the south have taken part in the northerly movement of the thrusts but there is no corresponding change in them westwards which could accommodate the absence of the thrust there. We have to conclude that the total deformation varied little if anything, and that we see differences in structural style to be correlated with the lithological sequence involved. This implies that the thrust faults should be regarded as having been generated in the axial planes of the folds such as Pozo and Bregón. The way they die out upwards in the incompetent La Vid Shales makes the form of a classic nappe much more unlikely.

Cueto Negro Unit

North of the Pozo Thrust front the Cueto Negro Unit has the form of a large anticlinorium dominated by the Lower Cambrian Herrería Formation core. The whole of the Palaeozoic sequence, in its north Leonide facies (Middle-Upper Devonian Caldas Formation) is seen on the flanks and the western termination. Faults are almost impossible to map in the Herrería rocks but they must be present in the Sierra de Cueto Negro and in the tight to synformal extension of the Villasecino Anticline, because of the relations with the Palaeozoic cover rocks.

The contact between the Lancara and the Herrería Formations has been mapped as stratigraphic so that the latter should take part in the structures that plunge to the west. However, it is geometrically impossible for the latter to conform entirely in view of the space problem in the core of such tight, concentric folds. Faults postulated in the axial planes of the Villasecino and Robledo Anticlines express recognition of this but the lack of marker beds in the Herrería Formation renders more detailed mapping impossible. The inlier of younger Palaeozoic rocks in the Sierra del Cueto Negro is a confirmation of the presence of the sort of complications to be expected (see also Martínez Alvarez et al., 1968). A cross section (Encl. X: Section 1) constructed from the little data that we do have shows that the anticlinorium can be interpreted as, at least two, thrusted anticlines largely similar to those of the Pozo and Bregón Units. The involvement of the Herrería Formation, makes the suggested structures much more comparable to the Forcada and Bodón Units further east (see p. 104), and since the structural position along the northern boundary of the Leonides is also very similar, a comparable role if not a direct connection may be entertained.

Correcilla Unit

The Correcilla Unit forms a striking massif on the map as well as in the terrain. This is the result of refolding of the thrust sheet into a large synclinorium with a double-plunging axis. Rocks from the Middle Cambrian Lancara Formation up to the Namurian Caliza de Montaña have thus been preserved between the Ríos Bernesga and Curueño. At Villamanín, in the Bernesga Valley, the overthrust, where it appears from beneath the Rozo Thrust, is developed at the base of the Lancara Formation. In the Torío Valley, however, the thrust cuts rapidly through the sequence up to the Santa Lucía Formation which, in the form of a frontal syncline, is developed along the northern edge of the unit in the Fuentes de Sancenas. The thrust swings south from here cutting into the axial plane of the anticline that can be traced from Gete near the Río Torío. This fold must therefore predate the fault although a genetic connexion between the two as Evers (1967, p. 138) would have it, may well have existed. The easternmost extension of the Correcilla Unit in the Curueño Valley becomes more and more complex, largely through intraformational folding and faulting of the Lancara, with occasionally the Oville Formation. East of Montuerto the closure of the synclinorium is very clear in the map: the fold axes having quite a steep west plunge here.

While the overthrust is usually overturned in conformity with the synclinorium, it only reveals minor folding at a limited number of places; in contrast the interior of this unit is strongly folded over most of its outcrop. These folds are typically disharmonic through the varied stratigraphy of the nappe. The largely intraformational nature of the folds makes it theoretically unlikely that the thrust itself has been so affected, which is in line with the mapping. The smaller folds are concentrated in the axial zone of the synclinorium, and the variation in form revealed along the axial traces demonstrates clearly their disharmonic nature. The shales of the Formigoso, upper La Vid and Huergas Formations have played the principal role as incompetent and detachment horizons. In the west, near Horzal, the plunges are gentle both east and west, but beyond Correcilla they are very steep to the west. Here the map picture is an almost exact cross-section, and the variations in form and the southern vergence typical for all are very clear. In general they are asymmetric with longer limbs dipping towards the axis of the synclinorium suggesting a parasitic relation with the latter. The axial directions vary much more than would result from simply refolding a syncline with pre-existing minor folds but no real analysis has yet been made.

The Namurian Caliza de Montaña Formation is folded together with the Lower Carboniferous Vegamián and Alba Formations, as well as the Upper Devonian Ermita Formation although the extent of these types of folds cannot be traced very far because of the lack of markers in the Caliza de Montaña. These rocks are often deformed by mesoscopic folds as well as extensive tectonic dolomitization.

Gayo Unit

The Gayo Unit is only discernable east of the Río Bernesga in Peña Fontón, and terminates against the Porma Fault in the Porma Valley. The sheet is thinner than most, despite the inclusion of Carboniferous rocks because much of the Devonian sequence has been eroded. In the west we see one of the most characteristic plunging faulted pericline terminations. The axis plunges steeply NNW down the dip of the south-inclined axial plane (the symbols on the map only refer to a few small folds in the core of the structure where all the beds are overturned). From the core of this fold the thrust plane follows the base of the Lancara Formation for almost the entire length of outcrop. For the most part the thrust sheet rests against Upper Carboniferous San Emiliano rocks but, near Valdeteja and W of Almuzara, lenses of older formations appear in the southern flank of synclines often to be found near the thrust front. The frontal and subsidiary faults have developed striking mylonitic and pseudotachylitic rocks at a number of places, notably where the Barrios Formation is cut near Peña Fontón and Valdeteja. At the former location the brecciation extends far into the Carboniferous rocks and has been accompanied by considerable lead-zinc mineralization. Mylonitization is also evident near the eastern termination of the nappe at Valdecastillo where the Lancara Formation has been doubled and some minor disturbance affects the Barrios Formation. The higher parts of the unit follow the form of the Montuerto Syncline in the Pozo Unit although the tightness of the post-thrusting fold and the steep NNE to NNW plunge of the axes mean that these folds must have developed disharmonically in relation to those further west. In the last few hundred metres of its easterly extent the fault shows the tendency to cut upwards into the Gayo Unit and downwards into the Caliza de Montaña of the Bodón Unit to the north. It is tempting to imagine that this indicates the earlier existence of E–W symmetry similar to that of the Correcilla Unit (p. 104). The dips are generally steep, overturned to the north although in some places small folds in the San Emiliano Formation result in flatlying, overturned beds. The inverted attitude is primarily attributed to the post-thrusting refolding considering the attitude of the thrust plane; however, the folds cut off by the thrust must have been at least contemporaneous if not older.

Bodón Unit

The Bodón Unit is only clearly definable east of the Río Torío, although as has been suggested it may well be that this unit is the continuation of a structure in the Cueto Negro (Bernesga Sheet). At any rate the formations younger than the Barrios can be mapped through as continuous outcrop. To the east this unit also terminates against the great Porma Fault in the Porma Valley. There are two main unusual features: the thick sequence of the Herrería Formation and the even thicker Upper Carboniferous, San Emiliano and Caliza de Montaña Formations. In this unit therefore the basal glide horizon is no longer that of the Lancara but one of the many shaly intervals within the Herrería Formation.

Near Bodón Mountain a major part of the sequence, from the Oville to the Alba Formation inclusive, is missing, probably due to faulting since the Caliza de Montaña here is strongly dolomitized. The faulted-out section will not have been so great as in the Cueto Negro because Upper Devonian erosion had reached well into the Barrios Formation as can be seen near the eastern of the two Bodóns N of Valdeteja.

The great Z-form exhibited by the Bodón Unit between the Ríos Curueño and Porma is a further witness to the refolding of the Leonide Nappes. Since the bedding almost never dips at less than 60° around the Z-fold axes, it is clear that these plunge steeply, mainly to the west but varying considerably especially where flanks are overturned. The map picture is essentially a cross-section of the folds, and it is therefore likely that the folds seen in the San Emiliano Formation are detached from the former. The middle limb of the Z-fold is cut by several vertical faults with a component of dextral apparent throw which would fit the asymmetry of the fold.

The eastern termination against the Porma Fault near Valdecastillo is strongly deformed by many small folds and faults (not all on the map) as well as extensive dolomitization of the Caliza de Montaña Limestones. The Barrios Formation is also altered from the usual resistant quartzites to a loose sand quarried for glass-making just west of Valdecastillo. The majority of the beds are steeply overturned here so that most fold axes plunge steeply NNE with a few shallower to the east and west associated with beds the right way up. This extra deformation has clearly been associated with the fault and indicates at least a considerable component of strike-slip movement although the folds are not clear in suggesting the sense of movement.

Forcada Unit

The Forcada Unit forms the most northerly of the Leonides between the Ríos Torío and Porma especially if we include the Armada and Pallide Subunits. The lower Palaeozoic section is much thinner due to Upper Devonian erosion which usually reached deep into the Oville Formation here. Occasionally the Lancara Formation has also been eroded but this is not shown on the present maps. In addition the Caliza de Montaña is stratigraphically much thinner than in the Bodón Unit, and there is only a very thin sequence of siliciclastic deposits preserved above it. The outcrop pattern follows that of the Bodón Unit in the Z-fold but the internal structures in the N–S segment are much more complex than to the east or west. Occasional small folds plunging to the north as well as NW and NE have been mapped in the latter. In the east slopes of Peña Forcada, complications abound; the formations are generally doubled although the Oville reappears in all three thrust slices shown. Not only is the unit sliced-up, the slices have been subsequently folded around plunging axes in the frequently overturned beds. The breadth of the synclinorium mapped between Canseco and Redilluera is very much less than the N–S limb of the Z-fold, suggesting that, once again, these large deviations in the fault trend are not simply a question of refolding affording a downplunge view of the unit. The tectonic style is reminiscent of that to the south of Montuerto at the termination of the Pozo Unit. As has been suggested, it may well be that this sort of imbrication indicates the proximity of the root of an overthrust.

Armada and Pallide Subunits

The Armada and Pallide Subunits appear on the Esla Sheet as klippe belonging to the Forcada and Bodón Thrust Sheets, respectively (cf. Rupke, 1965). These correlations are suggested by the levels reached by the Devonian erosion – the Oville in the first of the pairs and the Barrios in the last – together with the involvement of the Herrería Formation exclusively in the last two. The relation to the surrounding Westphalian and Stephanian rocks is equivocal; faulted contacts have been observed whereas the map picture strongly implies unconformities. The correlation with the westerly units is so close that these rocks must have been thrust in the same way although the extensive unconformity would hardly allow of this. Almost certainly post-Westphalian movements have disturbed many of the contacts with the thrust units as suggested by Rupke (1965, p. 59). Incidentally this phenomenon must be expected along the whole of the Leonide Thrust front, rendering general conclusions about the date of movements from relations there quite limited in significance.

The beds in the Armada and Pallide Subunits dip steeply almost everywhere and axes plunging steeply 50 to 60 degrees to between N and W have folded them to yield the existing outcrop pattern. Revision mapping has revealed that the synclinal structures further east near Reyero and Viego are much more extensive than shown on the 1 : 50 000 map although largely concealed by the unconformable Upper Carboniferous cover. All of these structures appear to be faulted against one another and are probably detached at the base of the Lancara Formation; however, the involvement of the Barrios Formation suggests a correlation of the Bodón and Pallide Units. The absence of the Herrería Formation is probably not due to concealment under the unconformable cover but more likely to an upward slicing of the thrust plane here. All these complications suggest an original termination although later movements along the Porma Fault will have disturbed the relations now observable.

Esla Unit
The Esla Unit is indicated on the structural sketch map (Encl. X: 8) as comprising the Pardomino Ridge, the Esla or Valdoré autochthonous, the Las Salas Zone and the Esla Nappe itself in two segments, the Felechas and the Agua Salio Synclines.

Pardomino Ridge
The Pardomino Ridge of the Lower Cambrian Herrería Formation interrupts the E–W trend of the Leonides in a particularly striking way on its northwest flank since the Porma Fault cuts through the five structural units there. Due to the absence of truly correlable markers the structures within the Herrería Formation are not exactly known although it seems likely that they are more complex than indicated on the Torío Sheet.

In contrast to the N and W boundaries, those to the S and E seem to be stratigraphic, if not completely conformable – the shaly transition beds have even been indicated along the E–W stretch of the contact. The SW and NE extremities of the Ridge are structurally complex: although many of the folds are intraformational and only involve the Lancara Formation it is evident that others involve considerable thicknesses of Herrería and Oville Formations.

The Palaeozoic sequence conformably overlying the Pardomino Ridge is the autochthonous of the Esla Nappe par excellence. The contact north of the Felechas Syncline is more complicated than the maps would suggest but the contrast with the structures to the NE is still strong. There the development of decollement structures, doubling and redoubling the uppermost Herrería, Lancara and lower Oville Formations was probably preceded by folding. The faults increase in apparent throw away from the Ridge so that, near Valbuena, the Lancara is in contact with the La Vid Formation. This fault probably has at least some component of strike-slip and it cuts off the Valbuena Syncline which is perforce an earlier structure. Drag folding and mylonite forming is very well developed along this fault; while these may also be observed along other faults near the Pardomino Ridge they are typically absent from the Esla and other larger thrusts.

The rocks below the Esla Thrust and east of the Valbuena Syncline must be considered parautochthonous. They are mostly strongly folded and faulted although faulting is even more severe in the Las Salas Zone to the north. The folds are small and numerous in the upper parts of the sequence because of the marked stratigraphic thinning of many of the units here. To the south the beds, while inverted, are folded together with the Esla Nappe in the Agua Salio Syncline as far as Vellila de Valdoré. Here we enter the window of Valdoré in which the autochthonous Devonian rocks have been disharmonically folded in respect to both the overlying nappe and the underlying older Palaeozoic. The former relation is clear from the outcrop pattern, and the latter from the construction of cross sections (Encl. X: Sections 5, 6). In fact these constructions also imply that the older Palaeozoic in the unnamed anticline between the Agua Salio and Felechas Synclines must be detached at the level of the Lancara Formation, so that we may expect structures similar to those along the Pardomino Ridge at depth here.

Las Salas Zone
The Las Salas Zone is typified by the large number of faults (many more than shown on the map) although there are also many small folds. The latter are small and tight as a result of the stratigraphic thinning of many of the formations here. The axial planes of these folds are subvertical and the axes often steeply plunging while the usual asymmetry suggests a relation to sinistral strike-slip fault movements. To the west the structures disappear beneath the unconformable Stephanian Cea Formation around Salamon, and to the east beneath the Westphalian Lechada Formation whose unconformable relation has recently been established (compare and contrast the mapping on the Esla and Yuso Sheets (Encl. V and VI)). The Barrios and Oville Formation are little folded since the former is still quite thick, whereas it is heavily fractured by small faults that cannot be mapped.

The zone coincides with the regionally important León Line which is usually covered by youngest Carboniferous deposits. Here the evidence of a long history of tectonic activity can be read from the stratigraphic sequence, and the faults and folds recorded may well have a complex history of development, extending far back beyond the Hercynian orogeny (Koopmans, 1962).

Esla Nappe s.s.
The Esla Nappe s.s. is a unit defined as covering an area between the Ríos Porma and Cea with its maximum extent in the Esla Valley. This definition excludes the possible correlation with the Bregón Unit mentioned above (p. 102). The outcrop is divided into three segments by the autochthonous Unit of Valdoré and the unconformable Stephanian molasse of Sabero; these are the Felechas Syncline, the Agua Salio Syncline and the Peña Corada Unit (see Encl. X: Section 6) (Rupke, 1965; de Sitter, 1957).

Felechas Syncline
The Felechas Syncline to the west is cut off in that direction by the unconformable Cretaceous Voznuevo Formation. The map shows the thrust outlining the syncline, following, but not always exactly, the base of the Lancara Formation. The structure is more asymmetric than the map suggests, the north flank dipping as little as 50° N over much of its length (ref. overlap Torío

Sheet). This opposing very gentle dip along the south flank implies an axial plane dipping north, and an axis plunging NNW, that is an attitude typical for the folds of the thrust sheets in the Leonides.

Beneath the thrust, consecutively younger formations gradually appear towards the east and south. Complications are evident around the nose of the structure, but their elucidation is hampered by extensive tectonic dolomitization of the limestones here. The core of the syncline comprises mainly the La Vid Formation, intricately folded and faulted. The upper beds of the San Pedro Formation are also involved in these folds and the map pattern therefore shows something of their nature. They are disharmonic probably even with respect to the basal San Pedro, but certainly with the massive Barrios Formation, which would appear to have dictated the dimensions of the major fold. There is no obvious parasitic relationship between these smaller folds and the major structure.

The eastern nose of the syncline would seem to have been considerably modified by N–S folding comparable, but not contiguous, with those of the Valdoré window.

Agua Salio Syncline

The Agua Salio Syncline is perhaps the most striking example of a folded nappe to be seen in the Leonides (Encl. V: Esla Sheet; Encl. X: Section 6). The thrust plane is often vertical but varies to quite shallow dips opposite the Felechas Syncline. The structure involves the full Lower Palaeozoic sequence as developed in the Leonides: it is asymmetric and inclined to the south. The thrust is indicated as a simple planar fault everywhere except along the northern flank of the syncline, where complications of minor folding and faulting are perhaps more frequent than indicated on the map. East of the Río Esla the main fault slices upwards through the thrust sheet which is here folded into a relatively sharp anticline, the fault remaining essentially planar and vertical. It must be accepted that this fold was an original feature of the structure – a ramp fault – developed during thrusting as proposed by de Sitter (1957) and Rupke (1965). In essence the north flank of the Agua Salio Syncline was initiated during the thrusting, and it is tempting to couple the rest of the structure as a consequent result.

The frontal anticline bears a number of intraformational, parasitic folds in the various formations above the Barrios. These folds plunge at various angles to the east with the result that the whole sequence disappears beneath the unconformable cover of the Cea Formation. The map sheet suggests that the thrust plane curves southward beneath this cover, but the evidence is very tenuous. What is clear is that the eastern part of the Esla Nappe has been folded along near N–S axes, which may be later than the E–W directions typified by the Agua Salio Syncline. The interference of these two sets may be responsible for the vertical axes found south of Remolina. Further south these N–S folds are overturned and inclined to the east near Oceja de la Peña. The exact extent and significance of these folds cannot be elucidated, because so much is covered by the Cea rocks of the Valderrueda Basin.

The suggested connection of the Felechas and Agua Salio Synclines is very logical in terms of the constructed cross-sections and also the isolated patch of exposed Lancara below the Cea unconformity just north of Sabero. The lateral offset of the axial planes and the change of plunge would then have to be accounted for by refolding along N–S axes as has already been suggested.

Valsurvio Dome

The older structures of the Leonides east of the Esla Unit, are concealed for some distance by the unconformable Stephanian Cea Formation of the Valderrueda Basin. Where Devonian rocks reappear in the Valsurvio Dome the sequence is sufficiently different to maintain individual formation names (see p. 85 & Fig. 1), and the structures are not the typical thrust-folds of the Leonides. Nevertheless the lithologic correlation is fairly complete, and the northerly vergence of the structures makes it reasonable to consider this assignment as correct and befitting a unit south of the León Line. Koopmans (1962, p. 220) summarizes the structures as follows:

"Several folding phases have effected different types of deformation in these structural units, according to the lithological characteristics of the outcropping rocks.

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

First generation structures.

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

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

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

Second generation structures.

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

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

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

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

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

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

Sierra del Brezo
Further east the last structural unit in the Leonides (the first mapped incidentally) is found in the Sierra del Brezo. Kanis (1956, p. 423) summarized the structures as:

"Broadly outlined, one has to consider the Devonian between Ventanilla and San Martin [Carrión Sheet] as a great overturned anticlinal fold... This isoclinal fold in the Devonian, overturned towards the south, lies on Lower Carboniferous rocks and is covered by them... The axial plane of the major Devonian fold between Ventanilla and San Martin is folded again,... perpendicularly to the older axis. This Devonian area has, therefore, been influenced at least by two deformations, one having a northeastern direction and a second, later one, causing the folded axial plane... The structure near Ventanilla is also complicated. Here the most important features are folds in the axial plane of the overturned anticline, caused by compression (alpine?) from east–southeast...

Structure of the Caliza de Montaña Series.
This structural unit between the great thrust fault [Southern Boundary Fault] and the fault zone of Ruesga contains, besides limestones and shales with subgraywacke beds, griotte and radiolarian rock... one can detect the following:
1. great, almost isoclinal folds form the principal structure,
2. these folds are overturned towards the SW in Peña Redonda and Orocada and towards the SE in Picos de Las Cruces and Almonga,
3. this change in direction of the axial plane gives rise to structural complications round Peña Celada [4 km east of Peña Redonda],
4. it is probable that this change of direction is connected with the compressed structure in the Devonian between Ventanilla and San Martin,
5. the thrust fault bordering the Carrión basin bends to the west, southeast of Peña Redonda but another fault continues in a northwesterly direction along the projected extension of the thrust."

Post-Stephanian Hercynian structures in the Leonides
The unconformable cover of the thrust-fold complex is also considerably deformed. There are many examples of deposition upon accidented terrain near the unconformity, and the facies seen also make it likely that many strata were laid down with an original dip. Structures such as those of Peña Rionda (4 km E of Valdoré, Encl. V: Esla Sheet) are clearly disharmonic with the underlying older Palaeozoic in a style and to a greater extent than can be explained by original dip. In his discussion and interpretation of these structures Helmig (1965) emphasizes a direct control through 'basement' faulting, and he cites examples such as near Morgovejo (Valderrueda Basin, Esla Sheet). That such structures have developed is almost certainly true, but it is probably going too far to look for such an origin in every case.

In general the structural control of the deposition of the unconformable cover is indisputable having regard to the correlation between them and the major structural features such as the León Line, the Sabero-Gordón Line and the Southern Boundary Flexure (Evers, 1967). The fault control of smaller basins south of Villablino has also been emphasized by van den Bosch (1969). The subsequently developing structures will have been influenced not only by basement structures and topography but also by the variations in lithology as well as the confining boundaries of the basins.

STRUCTURES OF THE ASTURIDES

North of the León Line structural development has been much more heterogeneous than in the Leonides. Not only is the style of deformation extremely variable, as might be expected in view of the lateral changes in the stratigraphic sequence, already noted, but the orientations are also much more variable. A very large part of the area is underlain by rocks of Upper Carboniferous, in particular Westphalian (or Moscovian) age, which are often paraconformable with older rocks. Younger Stephanian deposits are relatively rare and often occur in structurally complex situations here. This means that distinction between various folding phases as conceived by Kanis (1956) and Koopmans (1962) is practically impossible.

The style of the major structural developments may be roughly described under five main types – once again related to the stratigraphic sequence involved:
– Upper Carboniferous flysch and lowermost Palaeozoic strata thrust and folded together – Ponga Nappes (San Isidro-Porma Sheet),
– Thick, Upper Carboniferous Carbonates detached and mutually thrust – Picos de Europa Nappes,
– Thin, Lower Palaeozoic sequence (Palentian facies) folded and thrust independently – Monto, Carrión, San Julian,
– Thin, Lower Palaeozoic (Palentian facies) folded and thrust together with Carboniferous flysch – Cardaño, Liébana,
– Thick, Upper Carboniferous flysch folded with little evidence of involvement of older rocks – Valdeón, Lechada, Curavacas, Pisuerga, Liébana, Lois-Ciguera, Maraña (see Structural Sketchmap, Encl. X).

Ponga Nappes

The Ponga Nappes are only partly covered by our maps, notably the San Isidro-Porma Sheet with a small extension into the Yuso Sheet (Encl. X: 8). The major part of this structural unit has been investigated by the University of Oviedo (Martínez Alvarez, 1962; Julivert, 1965). These nappes are typified by a truncated older Palaeozoic sequence – the uppermost Devonian and lowermost Carboniferous deposits rest on the eroded Ordovician Barrios Formation for the most part. The sequence down to the base of the Lancara Formation which again acts as the detachment horizon, is usually normal, and higher in the sequence the Carboniferous flysch only differs in the probable existence of an hiatus during much of the Upper Namurian (Sjerp, 1967).

The maps show clearly the repetition of the succession from east to west, which would seem to have been the direction of the largest component of movement. The thrust sheets have been refolded about E–W axes of very variable plunges, usually steep. The thrust faults frequently cut upwards through the bedding but the lack of real variation in much of the sequence makes them difficult to follow so that the traces on the maps are often uncertain. Multiple repetitions of the Lancara Formation near Puebla de Lillo resemble closely those of the Pardomino Ridge, and it is considered probable that they too are associated with a basement high. The older Palaeozoic sequence of the Tarna and Peña de la Cruz, further east is very much thinner than anywhere else and, except near Valdehuesca (Torío-Curueño Sheet), is only found in the folded klippen and olistoliths of the Maraña area. The latter demonstrate the emplacement of the nappes in the Lower Westphalian (Sjerp, 1967).

The folds of the thrust sheets are usually tight, sometimes inclined to the south as near Peña del Aguila. The outcrop of the axial planes of those folds in the map does not indicate the true trend of the axes, because of the prevalence of steep plunges – up to 70° – there is probably less than 20° variation from an E–W trend.

Several large wrench faults are evident in the San Isidro Sheet, clearly visible due to the offsets of the Barrios Formation. Probably the most important is the Ventanilla Fault coinciding with the NW–SE Cardaño Line here. A dextral offset of about 5 km can be read from the displacement of the axis of the Zalambral syncline from the Valdosin (Encl. X: 8), and Sjerp (1967) deduces a vertical component of throw too. At Cofinal an almost E–W fault with an apparent displacement of some 3 km could not be traced further into the Carboniferous flysch. Since these faults offset the already folded thrust sheets they postdate that part of the deformation.

The limestone-bearing Carboniferous flysch of the thrust sheets has been thrust and folded together with the older Palaeozoic rocks, but there are also many small disharmonic folds developed within the sequence especially in the parautochthonous to the south. Some interference patterns due to the interaction of different axial directions are evident. Near Peña del Alba on the Río de San Isidro, N–S axial traces have been mapped but these axial planes are almost horizontal and the axes not far from E–W. No analysis has been made of the fold patterns mapped although Julivert and Marcos (1973) have shown that it is possible to do so.

The large nappes become less obvious to the northeast (Encl. VI: Yuso Sheet). Many of the faults curve northwards and even round to northwest in a broad, steep, imbricate zone NW of Oceja de Sajambre. Most of them involve only the repetition of Carboniferous rocks but Neoncito has the typical lower Palaeozoic section finally cut off to the NE by faults associated with the Picos de Europa thrusts partially concealed by the Upper Carboniferous unconformity there.

Picos de Europa

The Picos de Europa lie mainly north of our maps but important segments can be seen in the northern parts of the Yuso and Nansa-Deva Sheets. The almost exclusively carbonate sequence of the Upper Carboniferous here has enabled the development of thick thrust sheets, which have been thrust southwards over one another and the flysch basin of Liébana (Encl. VIIIA: Sections A, B, C, D). These sheets are stacked up against one another, usually dipping north, and extensive refolding has often accentuated many dips especially along the southern margin of the Picos. Here the general steepening of the dips has been accompanied by the development of cascade folds with a northerly vergence – the mirror image of the refolding of the Leonide Nappes (Encl. VIIIA; Maas, 1974).

The general trend of the thrust sheets is E–W al-

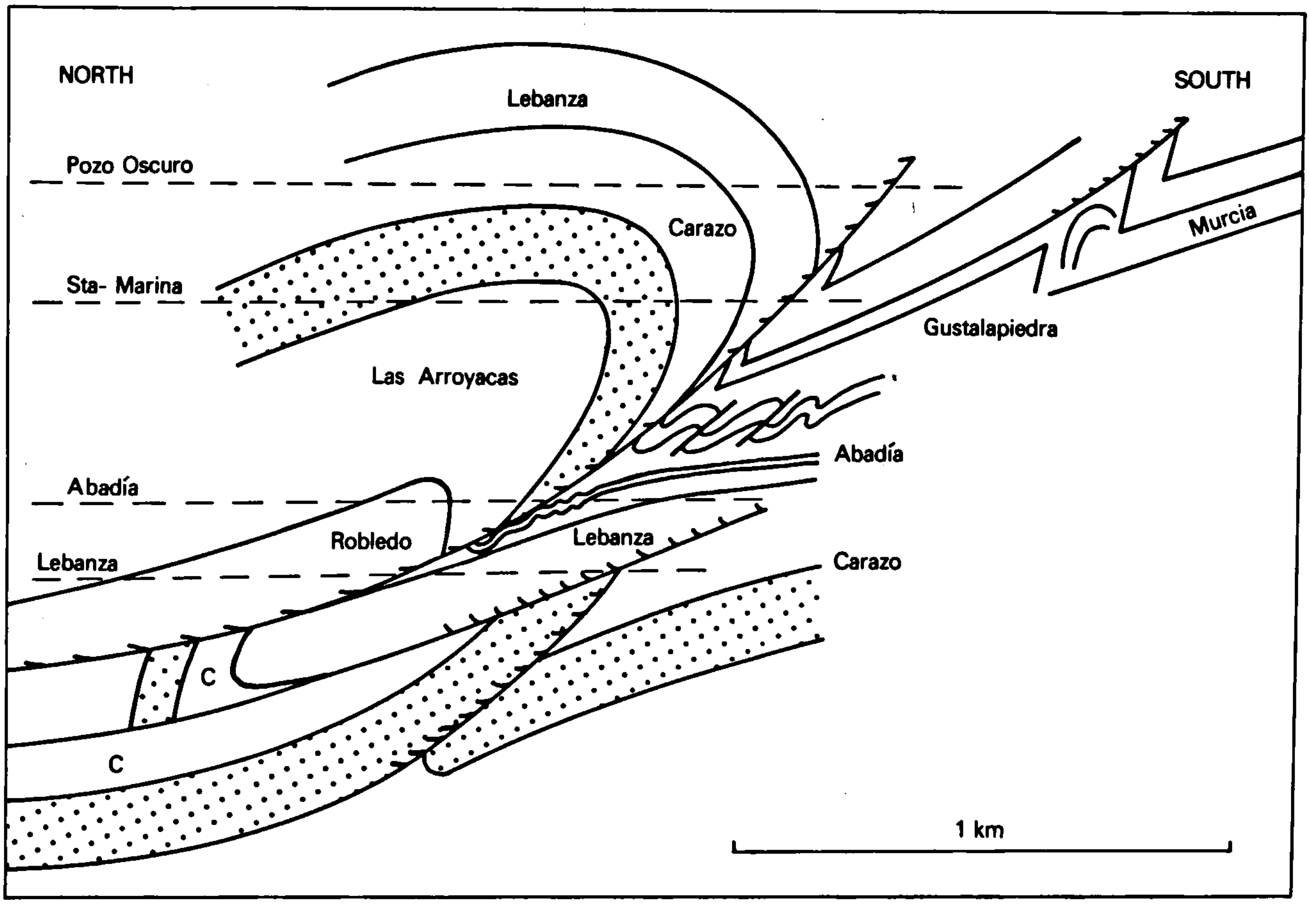

Fig. 6. Composite diagram of the Lebanza Thrust (after Ambrose, 1972).

though the outcrop curves to NW on the Yuso Sheet and to the ENE on the Nansa-Deva Sheet. They are cut by a number of WNW trending sinistral wrench faults which probably also have a component of vertical movement along them too.

Monto, San Julian and Carrión

The inliers of Liébana, Monto, Cardaño and Carrión comprise structures developed in the Palentian facies of the Silurian and Devonian with only the smallest of segments of Lower to Upper Carboniferous rocks.

The Carrión Unit is areally and stratigraphically the most extensive of these structural entities, having been described by van Veen (1965) and Ambrose (1974).

The most striking feature in the field is the extensive faulting, much of very small apparent throw. These are troublesome in mapping and can never be shown fully especially on the scale we present. The most important faults are thrusts which have developed out of faulted folds due to the disharmony of folding (N.B. the largest thrust fault striking E–W just N of Lebanza has, in error, no barbed decoration on the Carrión and Pisuerga Sheets).

Three major anticlines – through Tangua, Peña Carazo and Polentinos are evident from the outcrop of the Silurian/Devonian, Arroyacas Formation. These have typically overthrust flanks all verging south (Fig. 6 and Encl. VIA, VIIA) while local steeper plunges as, for example, near Arras and Polentinos, result in variations of the outcrop pattern.

The Upper Devonian rocks in the core of the Bistruey, Cortes and Curueño Synclines are either tightly compressed or completely cut out by the southward thrusting.

As can be seen at the eastern end of its outcrop, the Lebanza Thrust originates in the shales at the base of the Robledo Formation (Fig. 6). This style of detachment has been used in the construction of the rest of the cross-sections (Encl. VIIIA).

Axial plane cleavage has been formed in shales in the cores of some tightly compressed folds. It is usually quite penetrative although crenulation cleavage can also be identified in some places. The nature of the crenulated foliation has not yet been established.

The Monto inlier, comprising only Upper Devonian rocks together with some Lower Carboniferous, has only been briefly described up to now by Kullmann (1962) and van Adrichem Boogaert (1967). The sequence is quite thin, and the north-dipping pile of thrust sheets have been subsequently folded, resulting in structures too intricate to present on a 1:50 000 scale. The relationship with the surrounding Upper Carboniferous sediments is unclear but suggestive of an unconformity. Olistostromes in the latter sequence bear olistoliths of Palentian Devonian rocks confirming the uplift of the latter late in the Westphalian. There are no obvious roots for these thrust sheets exposed at present. The succession is not found in the Picos Nappes nor in the inliers where older rocks occur further north (Gomez, 1974; Maas, 1974; Martínez Alvarez, 1962). Hence these structures must have been rooted locally and, like the neighbouring Liébana structures, have to be regarded as parautochthonous. Since the frontal thrust of the Picos overrides the olistostrome-bearing sequences of Puerto de Remona and Fuente Dé (mapped as Valdeón Formation), we conclude that the former are younger structures.

The San Julian inlier also forms an apparently isolated inlier of Palentian facies Devonian rocks within Carboniferous flysch. The cross-section E–F (Encl. IXA) shows two interpretations explaining the positive relief by pre-Carboniferous erosion and by faulting. Some contacts certainly suggest the second alternative but the widespread occurrence of isolated blocks of Devonian rocks in the Carboniferous flysch (many not shown on the map) imply that the older series was being eroded at the same time as sedimentation of the latter. The series of tight folds inclined to the south matches the asymmetry of the structures of the Carrión Unit and implies that their genesis was probably comparable.

Liébana and Cardaño

The Liébana and Cardaño structures differ from those above by the involvement of more or less Carboniferous flysch. This is clearest in the Cardaño Unit, which is in fact a continuation to the west of the southernmost Carrión structures. These have also been described by van Veen (1965), and the sequences here provide the majority of the type localities of the Palentian facies for the Upper and Middle Devonian. This follows from the absence of the intense faulting which affects the Carrión area. The sequence is tightly folded and generally detached in the shaly Middle Devonian Gustalapiedra Formation (Encl. VIA: Section C^{III}-C^{IV}). Thrust faults may cut some if not all of the lower limbs of the folds.

The unconformity at the base of the Westphalian Curavacas Formation is very gentle here, in strong contrast with the relation some 5 km N near Peña Prieta, where the erosion is very evident and typical Carrión structures are cut off. The conclusion that these structures had a long history of development seems inescapable.

The Liébana inlier – the mid-Liébana ridge of Maas (1974) – is much more complex than our 1:50 000 map can show. In fact even the 1:25 000 scale map in Maas (1974, Encl. IV) is generalized and due to poor exposure very interpretive. The thin Upper Devonian of the Palentian facies is repeated in structures possibly similar to those of Monto, but it is not clear just how much has been disturbed and resedimented in the surrounding olistostromes.

The inlier terminates against the Picos de Europa Thrust north of Mogrovejo, and it is again concluded that these structures predate the final emplacement of the Picos Nappes.

Upper Carboniferous flysch

The structures in the Upper Carboniferous flysch of the Cantabrian Mountains north of the Leonides are varied in the extreme. To the west, in the area of the Ponga Nappes they form an integral part of those structures, as described above. To the east we have very divergent fold trends, e.g. contrast the Lechada/Curavacas Synclines with the folds of the Pisuerga Basin (Encl. X: 8). It has not proved possible to establish a regional pattern of trends with a fixed sequence as suggested by de Sitter (1957).

The Pisuerga Basin is an almost equidimensional unit, primarily structural as emphasized by van de Graaff (1971a), and owing nothing to simplicity. The cross-sections to the Pisuerga Sheet (Encl. IXA) show how the eastern flank dips steeply or is strongly overturned. This attitude is continued to the SE, through Barruelo which seems to imply some structural control separate from the basin development itself. The west flank is terminated by a N–S-fault, whose sinuous trace may well be the surface expression of a zone of faults. Although the fault we see cuts the structures in the very youngest sediments of the Pisuerga Basin (early Stephanian) it is likely that a fundamental feature exists at depth since there is evidence of synsedimentary movements in the sequence nearby (de Sitter & Boschma, 1966; Wagner & Varker, 1971).

Within the basin there are a number of open to tight folds with steeply dipping axial planes trending NNW. It seems most likely that these are accommodation folds resulting from the confinement of the sequence between the two bounding features.

The Liébana Unit of Maas (1974) and Boschma (1968) is the most extensive Carboniferous flysch area in our maps. It is divided into the North Liébana/Polaciones and Southern Liébana Subunits by the Liébana Ridge (Encl. X: 8). As can be seen from the cross-sections (Encl. VIIIA), the whole unit is interpreted as a great synclinorium overturned towards the south. In general the large scale folding and overturning preceded a later phase of refolding by a system of folds with subvertical axial planes. Cleavage is too sporadic in its development to be of use in a regional analysis but the fold sequence is reasonably clear (Rupke, 1977).

In North Liébana/Polaciones a large proportion of the strata are overturned, only becoming normal in the younger Westphalian rocks in the Viorno Syncline (Encl. VIIIA: Sections B and C; see also Nansa-Deva Sheet). An angular unconformity has been recorded within the Westphalian sequence but it has not proved possible to distinguish folding phases on the basis of it (Maas, 1974). The deformation would seem to decrease southwards which may be interpreted as implying the influence of the advancing nappes of the Picos de Europa.

The Southern Liébana Subunit is characterized by upright folds mostly tight and slightly inclined to the south. The differences between the folding styles of the Westphalian and Namurian sequences are due to the inclusion of the massive conglomerate and limestone layers in the former.

The Lechada and Curavacas Synclines are essentially similar to the Southern Liébana structures, being controlled by the Curavacas Conglomerate, but distinctive in having an extensive cleavage development enabling detailed analysis (Savage, 1967). This has led to the conclusion that these are drape folds, bending to accommodate vertical movements in the underlying beds. This is illustrated by the opposing vergences found on either flank towards the axis of the structure in the higher structural levels of the Lechada Syncline (Encl. VIA: cross-section Z–Z′). The generally flatlying attitude of the slaty cleavage is also witness to a considerable vertical component of the tectonic transport.

The two structures are faulted en echelon, and the eastern termination of the Curavacas Syncline is also faulted against the Carrión Unit. The apparent throw of this N–S fault where it cuts Devonian strata near El Tejo does not match that required to cut out the Curavacas Formation. It seems certain that this fault has also had a long history of movement probably including the times of folding.

The whole set of structures plunges more or less gently westwards, but would seem to die out due to the interfingering and pinching out of the Curavacas Conglomerates in that direction.

The Lois-Ciguera Unit to the west of Riaño, on the Esla and Yuso Sheets, is distinguished by the pattern of tight folds revealed by the layers of limestone many there. The fold trends and plunges are very variable, and no simple analysis seems possible. Stratigraphic variations are probably the most important controlling factor because, as can be seen on the Yuso Sheet, where the limestone bands grade eastwards into massive bodies, the lateral equivalents, in Peñas Pintas and Pico Yordas are little folded. The folds also terminate abruptly to the south and west where they disappear beneath unconformable Stephanian Cea Formation along the León Line. Beyond the Cea outcrop the character of the Westphalian succession is very different and no comparable structures have been found. The change in structural style as well as the sequence within the Cea Formation outcrop make it likely that faulting is present here – probably more complex at depth than we can determine at the surface.

The remaining areas of Upper Carboniferous flysch such as the Valdeón, the upper Esla Valley and the Riaño-Cervera Zone are relatively poorly understood. Exposures are generally inadequate to allow thorough analysis while the sedimentology of the deposits forms another unknown factor. There is evidence of synsedimentary deformation, and inverted limbs to folds are not uncommon. Structures with opposing vergence on either flank are also known here e.g. near Horcadas (Encl. VIA: Section Z), which involves Stephanian rocks. Cleavage, though present, is poorly and sporadically developed.

POST-HERCYNIAN STRUCTURES

Post-Hercynian deformation has probably affected some if not much of Cantabria. It is, however, impossible to distinguish their effects except where younger rocks are present. Since the prime object of the mapping has been the Hercynian core, such evidence of later orogenic movements has only accumulated incidentally, and the subject perhaps deserves much more attention. The opinion of most workers is that the influence of later movements has been restricted to the margins of the core despite the lack of any real proof.

Of the structures in the Permo-Mesozoic cover to be seen in the Nansa-Deva Sheet, Maas (1974, p. 426) writes "...[they are] characterized by faulted open synclines and anticlines with generally gently dipping flanks... the faults are not subsidiary to the folds but follow a general pattern which agrees with the pattern of the larger faults in the Palaeozoic rocks."

Along most of the Southern Boundary Flexure the Mesozoic and Tertiary strata dip steeply or are overturned near the contact with the Palaeozoic core, giving rise to this term. Most of these structures have been interpreted as 'mountain fold-thrust uplift' types of Berg (1962) together with rejuvenation of older structures (Evers, 1967, pp. 143–144; Rupke, 1965, p. 68).

The steeply plunging fold near Villanueva de la Peña (Encl. VII: Carrión Sheet) affects all except higher Miocene strata and has been interpreted as implying Tertiary (Alpine) dextral movement along the Cotolorno Fault (Koopmans, 1962, p. 209). The steep axial plunge implies that the strata dipped steeply before the faulting, suggesting a complex, extended period of Alpine tectonic events.

CONCLUSIONS

The Southern Cantabrian Mountains afford a view of a wide range of structural types, ranging from classic refolded nappes through multiple cleavage developments to drape folding and slump and slide deformation. These features are comparable to many in other mountain chains but the centripetal symmetry illustrated, if only partly, seems symptomatic of a fundamental difference. De Sitter's 1962 comments, that we only see a chance fragment of the Hercynian orogene (see p. 77), are even more significant today in view of the growing evidence of the fragmentary nature of such orogenes elsewhere in the world. It must be concluded that de Sitter's original purpose in issuing these maps to stimulate discussion and further work has not only been realized in his own time but looks like bearing fruit for some time to come.

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