LEIDSE GEOLOGISCHE MEDEDELINGEN, Deel 49, Aflevering 3, pp. 379–465, 26-2-1974

THE GEOLOGY OF LIEBANA, CANTABRIAN MOUNTAINS, SPAIN; DEPOSITION AND DEFORMATION IN A FLYSCH AREA

BY

K. MAAS

ABSTRACT

The Nansa-Deva map sheet of the Geological Map of the Southern Cantabrian Mountains is published. The accompanying thesis deals with the stratigraphy and structures of the Devonian, Carboniferous, Permian and Mesozoic rocks which constitute the mapped area. A condensed sequence of nodular limestones and shales was deposited in the larger part of the area during Middle-Upper Devonian and Lower Carboniferous. The rocks of this time interval are preserved badly, occurring for a large part as exotics in Upper Carboniferous 'Wildflysch' deposits. The establishment of a stratigraphic record for these largely allochthonous occurrences was enabled by examination of their conodont content and by comparing them lithostratigraphically and biostratigraphically with the known sequence of the Cardaño area south of the mapped area. The condensed sequence, the thickness of which normally does not exceed 500 m, is followed in the central part of the area by an Upper Carboniferous flysch sequence which may amount to a thickness of 10 km. This sequence becomes intercalated with, and gradually replaced by, a more shallow facies of siliciclastics and biogenetic limestone lenses to the south. Deposition of limestone in a stable shelf environment occurred in the northern part of the area, the Picos de Europa, during the same time interval. The stratigraphy of these Upper Carboniferous rocks was solved mainly by examination of the fusulinid content of the limestone bodies. Biostratigraphical dating of limestone pebbles occurring in conglomerate lenses in the flysch sequence and of exotics from 'Wildflysch' units resulted in a number of maximum ages providing the base for the stratigraphic record of this sequence. A pronounced unconformity separates the Upper Carboniferous and older rocks from the clearly postorogenic Permian and Mesozoic rocks. Upper Carboniferous sedimentary facies boundaries coincide with the boundaries between areas with a different structural development. The structural development in the Picos de Europa, characterized by the formation of large nappes, was independent of, and started much later than, the deformation in the remaining area. There, the Upper Carboniferous was not a time of undisturbed subsidence and sedimentation. Local uplift and erosion, faulting, bending and collapse folding are contemporaneous with the flysch sedimentation in the central part of the area. This contemporaneity is less obvious in the southern area. Several deformation types can be distinguished in the flysch rocks: sedimentary deformation (slumping), early tectonical synsedimentary deformation (bending and collapse) and eutectonical deformation (buckling, kinking and cleavage). All kinds of transitions between these main types were also observed. Unravelling the history of sedimentation and deformation, the spatial limitedness of individual deformation processes, the contemporaneity of deformation and sedimentation and the noncontemporaneity of deformation processes in different parts of the area have to be inferred. Most unconformities that originated during the structural development of the subject area have only a local extent. No unconformity could be traced over the entire area. The establishment of a number of folding phases does not seem an appropriate generalization for the structural history of the subject area.

CONTENTS

CHAPTER I

INTRODUCTION

General statement

This thesis deals with sedimentation and deformation of Palaeozoic and Mesozoic rocks before, during and after the Hercynian Orogeny in an area which nowadays constitutes a part of the Cantabrian Mountains in Spain. The area, represented on the enclosed geological map, lies mainly in the province of Santander and comprises the Liébana valley, the adjacent part of the Picos de Europa, the Polaciones valley and a small part of the Cabuernica area. The map constitutes the Nansa-Deva map sheet of the Geological Map of the Southern Cantabrian Mountains, the result of a geological mapping project, carried out during the last twenty years by the Department of Structural Geology of the Leiden University in Holland.

History of previous work

Geological mapping was carried out south of Liébana, along the main watershed of the Cantabrian Mountains, by students of the above-mentioned department, since the late fifties. The individuals and the areas they worked in are shown in the reliability diagram on the map sheet (Encl. I). Geological mapping in Liébana was started in 1963 under the supervision of Dr. D. Boschma. Four students dealt with the general geology of Liébana and Polaciones in internal reports (Boehmer, 1965; Lanting, 1966; Miedema, 1966; Maas, 1968). These results were summarized and published together with a provisional geological map, scale 1:100 000, by Boschma (1968).

Objective and methods

The field work which resulted in the presented map and thesis, had as primary objective the publication of a 1:50 000 geological map providing the synthesis of the above-mentioned results and also the connection with the already published 1:50 000 geological maps of the southern adjacent areas: the Rio Yuso area (Savage, 1967), the Cardaño area (van Veen, 1965) and the Pisuerga area (de Sitter & Boschma, 1966). To obtain this goal, large areas had to be remapped (see reliability diagram). A considerable amount of fossil sampling had to be done to provide the necessary base for stratigraphical correlation. The largely unknown Picos de Europa were too much of a temptation to be left outside, and anyhow, understanding of the geology of Liébana would be incomplete without some knowledge of this adjacent area.

In the field, use was made of the 1:50 000 topographical maps of the Instituto Geografico y Catastral in Madrid; corresponding aerial photographs were used mainly for the Picos de Europa area, the Mesozoic cover area in the east, and the better exposed areas near the main watershed in the south. The poorly exposed, central part of the area could only be studied by direct observation on small outcrops along road talus, in rivers and brooks. Excepting the Picos de Europa and the Cabuernica part, the area is fairly well accessible. Many of the tracks connecting villages with the few paved main roads have been made trafficable in the last ten years, so that nowadays most locations can be reached within two or three hours of combined driving and walking, when starting from Potes, the chief town in Liébana.

Establishment of a coherent map picture proved to be impossible without a detailed study of the predominant types of sedimentary rocks and their facies relations, as well as of the deformation processes that affected them. So, attention was paid to the flysch associations, in particular to the make-up and spatial occurrences of olistostrome deposits and their relation to the structural development. The structures that dominate are of the gravitational slide and collapse types. The relations of sedimentation and deformation are complex; at least partly, the processes have been contemporaneous.

Since the map constitutes a sheet of an existing series, map units, symbols and colours were, as far as possible, chosen to conform to existing use. The disadvantage of a not always selfevident subdivision in map units is considered to be more than compensated by the obtained correspondence with existing map sheets.

Acknowledgements
This thesis has benefitted greatly from the discussions I had, both in the field and at home, with many colleagues and senior geologists. Their contributions, geological information as well as technical assistance, are gratefully acknowledged. I wish to thank Dr. J. F. Savage for his assistance in the field, for the many discussions we had, for his constant interest and criticism during the progress of the field work and the writing of the thesis, for teaching me to write acceptable English and for his critical reading of the manuscript. I wish to thank Dr. A. C. van Ginkel for all the work he did identifying fusulinids from Liébana, but for his efforts most of the stratigraphy of Liébana would have remained an unsolved problem. I also want to thank him for the long discussions we had and for his critical reading of the stratigraphical part of the manuscript. I wish to thank Dr. R. H. Wagner for confronting me with 'heretical' ideas concerning the geology of the Cantabrian Mountains, for determination of flora samples from Liébana and for his critical reading of the stratigraphical part of the manuscript. I wish to thank the many others not mentioned by name. Last but not least I wish to thank the inhabitants of Pejanda (Polaciones) and Cosgaya (Liébana) for their hospitality and friendship during my stay in those places.

CHAPTER II

STRATIGRAPHY

INTRODUCTION

Before describing the lithostratigraphic units of the Liébana area, some features of the general geologic setting and work of previous authors from surrounding areas will be summarized.

The geologic history of the Cantabrian Mountains can be outlined as follows. During Lower Palaeozoic times continuous and uniform sedimentation took place over a large area, approximately from the Rio Luna in the west to the Rio Pisuerga in the east. The environment was that of a slowly subsiding stable shelf area. Up to the Lower Devonian, rock-stratigraphic units are essentially the same over this whole area. However, from the Middle Devonian onward, two areas, each with a different facies, are distinguished, the Asturo-Leonese facies area and the Palentine facies area (Koopmans, 1962; Brouwer, 1964). The Asturo-Leonese facies, essentially a shallow-marine open shelf, is found in the larger part of the S. Cantabrian Mountains. The Palentine facies, typified by a more pelagic fauna, occurs N of the Cardaño line (van Veen, 1965) in the provinces Palencia, León (Montó) and Santander (Liébana). The deposits of this facies consist mainly of shales and shaly nodular limestones.

Most authors suggest that this sedimentation area was bounded in the north by an area of little or no deposition, the 'Cantabrian Block' or 'Cantabrian geanticline' (Radig, 1962). This block should comprise the Picos de Europa area and the Ponga-nappe Province (Julivert, 1971). In this latter area there exists indeed a lacuna in the stratigraphy, involving Silurian and most of Devonian. In the Picos de Europa, the stratigraphic record is known only from the uppermost Devonian onward; since older rocks are not exposed, evidence of earlier conditions is absent. One might however suppose that the Carboniferous limestone succession of the Picos de Europa is underlain by the Ordovician quartzites, as known in the area to the north and west.

Locally and to a lesser extent, the mentioned lacuna is also present in the Asturo-Leonese area, south of the León line (Rupke, 1965; Evers, 1967). It is not clear whether erosion or nondeposition was the important agent in causing the lacuna. Erosion has been active, as demonstrated by the way the Famennian Ermita Formation rests unconformably upon older strata in the Cantabrian block area (Sjerp, 1967) and in the Asturo-Leonese area (Rupke, 1965; Evers, 1967). Nondeposition also had an influence; various reports of reduced sections of Devonian and older rocks have been made from localities along the León line (Rupke, 1965; Smits, 1965; de Sitter & van den Bosch, 1968).

In the area of the Palentine facies, there is no single marked stratigraphic lacuna. The Upper Devonian nodular limestones of the Vidrieros Formation are overlain by the Lower Carboniferous black shales and nodular limestones of the Vegamián, and Villabellaco/Alba Formations. From scattered evidence two parallel unconformities can be inferred, one at the base of the Vegamián and one at the base of the Villabellaco (cf. Wagner, 1972).

The geologic events that are reflected in the Upper

Devonian and Lower Carboniferous unconformity-levels are generally referred to as the Bretonnic phase. These events caused uplift and nondeposition over a large area, but without any folding and so are strictly not a phase in Stille's original sense (p. 427).

The successions of the Vegamián and Alba Formations are found in almost all the areas mentioned. They mostly occur in patches, in some places after a small amount of erosion. Considering their extent in time they are extremely thin. Usually, they are regarded as a condensed facies.

Towards the end of the Viséan or during the Lower Namurian these conditions came to an end, the sedimentation rate increased, a finely laminated limestone deposition set in over most of the old Asturo-Leonese and Cantabrian block areas. The Caliza de Montaña Formation was formed, the last formation that can be traced over the entire Cantabrian Mountains; this formation reaches into the Upper Namurian. From there on, the lithostratigraphy becomes complex and varied. Rapid sedimentation, repeated uplift, intense tectonic deformation and erosion make up the geologic history during the Upper Carboniferous. Post-Namurian rock-stratigraphic units have only a regional extent and are geographically separated in many cases, due to later uplift and erosion. Correlation is only possible with the extensive use of fossil evidence.

STRATIGRAPHY OF DEVONIAN AND LOWER CARBONIFEROUS IN LIEBANA

Geographic occurrence in Liébana

In contrast to the Picos de Europa area, in Liébana the Lower Carboniferous rocks occur together with Devonian rocks. They are mainly exposed within and around a narrow anticlinal structure that runs from Dobarganes over Mogrovejo to NW until it ends against the thrusted limestone complex of the Picos de Europa. Isolated patches of Devonian and Lower Carboniferous rocks are found in Upper Carboniferous olistostrome deposits, bordering this structure at both sides, but mainly at the northeast flank. In the south, the structure is cut off by the western extension of the Redondo fault (de Sitter & Boschma, 1966). To the east, halfway between Dobarganes and Enterrias, the autochthonous Devonian disappears under the olistostrome deposits which can be traced to the ESE as far as Pesaguero. East of Enterrias between Bores and the Rio Frio we see (Enclosure IV) a lot of small outcrops of Devonian rocks, surrounded by Carboniferous rocks. For most of them an allochthonous position as olistolith in an olistostrome could be ascertained in the field. The two larger exposures, however, may be in autochthonous position. Both dip rather steeply to the south, younging also in this direction. At their north side they certainly have abnormal contacts with patches of Devonian or Carboniferous rocks, although some thrusting seems to have occurred.

Representation of the data

As already stated, the Devonian rocks in Liébana belong to the Palentine facies. The oldest observed rocks are of an Eifelian age (conodont sample X 85-*corniger* Zone). A more complete record of these Palentine deposits is preserved in the Cardaño area and described by van Veen (1965), so we will follow his lithostratigraphic subdivision. Since outcrops of autochthonous Devonian are few and the accessibility is bad, no attempt was made to measure a lithologic section. Formations could not be identified on their lithostratigraphic characteristics only, so that extensive use has been made of determination results of 35 conodont samples. The allochthonous and autochthonous occurrences were grouped in conodont biozones (Enclosure V). The result was compared with the best known autochthonous succession in Liébana (NW of Sebrango at a height of 940 m) and with the known record from the Cardaño area. The resulting chronostratigraphic reconstruction of the Middle-Upper Devonian of Liébana is rendered in the left column of Enclosure V. The conodont samples with their determination results are listed in the Appendix. The corresponding localities are rendered on the 1:25 000 map of Enclosure IV or on the 1:100 000 map of Enclosure VII. Some localities were already known from the work of van Adrichem Boogaert (1967). In these cases his coding was maintained. The new samples are coded '(19)71 C(onodont) 200 (field station)', they were determined by Mr. K. Boersma of the Geological Institute Leiden, and are kept in his collection.

Description of the Devonian formations

Abadia Formation. – The Abadia Formation was first described by Binnenkamp (1963) as a formation consisting of a thick shale sequence in which two limestone members are distinguished. Van Veen (1965, pp. 53 and 54) describes a sequence exposed in the Rio Arauz, which can be considered as a type section. He introduces two formal limestone members of the Abadia Formation, the lower Requejada Limestone Member and the upper Polentinos Limestone Member. "The Polentinos Member is 10–30 m thick and consists of dark grey to black limestones which weather whitish grey and are very well bedded, platy and argillaceous with thin shale laminations and intercalations" (van Veen). According to van Adrichem Boogaert (1967, p. 161), who studied the Abadia Formation in the Arauz area, the Polentinos Member is of Eifelian age. The occurrence of this member is also known from the Montó area, west of Liébana in Valdeon (Kutterink, 1966).

Southwest of La Vega de Liébana, some isolated patches of dark well-bedded, grey-weathering limestone occur. Sample X 85 from this limestone yielded conodonts which belong to the *corniger* Zone (Appendix). Van Adrichem Boogaert (1967) correlated this limestone with the Gustalapiedra Formation despite the difference in age compared to the type section. My investigations have shown the existence of limestones of a Givetian age in the Liébana area, so that it seems clear that a distinction between Gustalapiedra and Abadia can be made. Also the lithologic aspects: "a dark shaly limestone, individual layers varying from 20 cm to 1 m, alternating with thin layers of black shale" (Lanting, 1966), make a correlation with the Polentinos Member of the Abadia Formation more probable. There is an other occurrence of a similar limestone north of Espinama along the old path from Pembes to Aliva, 3 m north of locality 70C172, but this outcrop failed to yield conodonts. Both occurrences are probably allochthonous.

Gustalapiedra Formation. – The Gustalapiedra Formation is described by van Veen (1965, pp. 65 and 66) as a 50–70 m thick sequence of alternating dark-grey and black slates and argillaceous black limestones, conformably overlying the Abadia Formation and passing gradually into a nodular limestone sequence, the Cardaño Formation. As type section the section in the Rio Gustalapiedra east of Barniedo de la Reina was appointed. Cephalopods collected by van Veen and identified by Kullmann (1963) point to an Upper Givetian age, which agrees with conodont faunas collected and identified by van Adrichem Boogaert (1967).

In the Liébana area a wavy-bedded limestone occurrence, west of Mogrovejo, yielded conodonts which point to an Upper Givetian age (sample 70C221, Appendix). North of Bores, some limestone exposures occur in a rather chaotic setting together with quartzitic sandstone patches, in a highly deformed, graywacke-shale matrix. Conodonts (sample 71C206, Appendix) point to a Givetian age. Both these allochthonous occurrences may represent the Gustalapiedra Formation. Autochthonous occurrence of the Gustalapiedra Formation could not be ascertained.

Cardaño Formation. – In its type section (van Veen, 1965, pp. 57 and 58) the Cardaño Formation is described as a nodular limestone development with shale intercalations, overlain conformably by the Murcia Formation, yielding conodonts which point to a Middle to Upper Frasnian age. Such nodular limestones have been encountered in the Liébana area in various places, but in allochthonous settings. South of La Vega de Liébana samples 71C315 and LV 26, north of Enterrias sample X 117, west of Campollo samples X 71 and 71C142, all belong to the olistostrome deposits north and east of the Mid-Liébana ridge. Sample 70C172 north of Espinama in the Remoña Olistostrome may have descended from the autochthonous Devonian of the Montó area west of Liébana. All those samples have the characteristic nodular appearance and a yellow-brown weathering colour. The limestone/shale ratio may vary considerably. Ages of the different samples range from the Frasnian-Givetian boundary (conodonts belonging to the *hermanni cristata* Zone) to the Frasnian-Famennian boundary and possibly to Lower Famennian.*

Autochthonous outcrops of the Cardaño Formation are also known from the Liébana area. In the core of the anticlinal structure west of Mogrovejo between two quartzitic sandstone ridges of the Murcia and stratigraphically below them, a black to yellow-brown shale is exposed. The shale contains nodular iron-rich concretions, in which fossil fragments of trilobites may occur. The base of the shale is not exposed, towards the top it contains thin beds of sandstone and passes gradually into a quartzitic sandstone sequence. In this autochthonous occurrence the Cardaño Formation seems to consist entirely of shales. Van Veen, who first described the Cardaño Formation, stated that its shale ratio increases to the east. According to Ambrose, an entirely shaly development of this formation may happen in some places in the Polentinos area (Mr. T. Ambrose, University of Sheffield, pers. comm.). In its autochthonous occurrence in Liébana, the base of the Cardaño is not exposed. Van Veen states a conformable contact with the Gustalapiedra Formation. The fossil evidence from Liébana, however poor, does not support interruption of sedimentation.

Murcia Formation. – The Murcia Formation was formally proposed by van Veen (1965, pp. 58–61) for the mainly quartzitic sandstone sequence resting conformably on the Cardaño Formation. In the type section (van Veen, 1965, Encl. 2, sect. Id) thickness is about 125 m, in the Cardaño area thickness may range from 60–200 m. According to van Veen, the Murcia is bounded conformably at its top by the nodular limestones and shales of the Vidrieros Formation.

In Liébana, although the Murcia is present in a fair number of exposures, there are but few places where this formation can be studied in some detail. In the exposures S of Pico Jano, E of Dobarganes and NW of Llaves, a combination of weathering, jointing and diagenesis have made it all but impossible to distinguish even bedding planes. The Murcia is there developed as a number of resistant quartzite ridges standing out in topography. Along the old paths from Besoy to Los Llanos and from Vallejo to Sebrango, the Murcia is exposed as a sequence of light quartzitic, sometimes decalcified, sandstone beds of varying thickness (20 cm–1 m), alternating with thin beds of fine black shales. The sequence youngs from Besoy to the north. Alternating with this lithology, three or more prominent banks of quartzite reaching thicknesses up to some 10 m, stand out in topography. Between the quartzites patches of brown to yellow weathering nodular limestone may occur. The quartzitic sandstone layers may be calcareous, reacting with a 0.1N HCl solution, and containing unidentifiable casts of fossils dissolved by

* A further probably allochthonous outcrop of Cardaño Fm. is erroneously shown as Villabellaco Fm. in the detail map, Enclosure IV. This straddles the road from Bores to La Vega about 150 m south of locality X 85. The nodular limestone is in stratigraphic continuity with the quartzite and answers well to the lithological description of the formation. Samples collected by Dr. R.H. Wagner and Dr. J.F. Savage have yielded ostracods, determined by Dr. M.J.M. Bless to be of Middle Devonian age. This dating would appear to conform the correlation suggested by the lithology and the field mapping of Lanting (1966). Further details will be published in due course.

weathering. The bottoms of the sandstone beds are frequently loaded, flute casts and groove casts could be distinguished. Sometimes the beds have a fining-up gradation. The degree of exposure is such that nothing can be said about lateral continuity of individual beds. At the top of the formation some few meters of very black shale may occur, which pass in a nodular limestone-shale sequence. In other places the contact with the overlying Vidrieros Formation is quite sharp. In different places in Liébana, the Murcia seems to be tripartite, consisting of a basal quartzitic division, a middle division containing more shale in which nodular limestone may occur, and a top division similar to the basal division. In the detailed description of the Murcia Quartzite Formation near Mt. Murcia (van Veen, 1965, pp. 58–60), such subdivisions are not mentioned. We will return to this subject when discussing the Vidrieros Formation.

Vidrieros Formation. – The Vidrieros Formation (van Veen, 1965, pp. 60–61) consists of grey to yellow brown, thin to medium bedded nodular limestones, interbedded with shales. The contact with the top of the Murcia Quartzite may be marked by some meters of black shales. The nodular limestones cannot be distinguished from those of the Cardaño Formation in isolated occurrences. Van Veen (op. cit.) reports an average thickness of about twenty meters nodular limestone which seem to pass gradually into the black shales of the Vegamián Formation. In Liébana, in the exposures where an autochthonous position of the Vidrieros could be affirmed, no contact with the Vegamián was preserved or exposed. In one allochthonous exposure, however, about one kilometer N of Enterrias in a fairly good exposed patch of Vegamián, such a contact, consisting of a 20 cm thick breccia layer of nodular limestone material in black silt, proves that the nature of the contact is unconformable. The fact that the conodont samples from the Vidrieros nodular limestones never yielded age determinations younger than Upper Famennian or lowermost Tournaisian (van Adrichem Boogaert, 1967, pp. 157–158), whilst from the overlying Vegamián Formation in the Palentine area only Upper Tournaisian is known (van Adrichem Boogaert, op. cit.), also indicates the presence of an unconformity between these formations.

Now as for the occurrence of nodular limestone between the Murcia Quartzites, we might venture a tectonic explanation, assuming faults running nearly parallel to the bedding plane. In the exposure SE of La Vega along the road to Valcayo, some faulting of that kind certainly is present. For other occurrences, however, notably those S and N of Sebrango, the exposure does not warrant such faulting. Furthermore, if we look at the results of conodont determinations, the grouping of the samples in one Upper Famennian group and a Lower to Middle Famennian group seems possible (cf. Frets, 1965, p. 121, description of an exposure near Santibañez de Resoba). This concept has been expressed in the diagrammatic chronostratigraphic column of Enclosure V. If this concept is right, it is not possible to define the lower boundary of the Vidrieros here, since then the lower part of this formation interfingers with the Murcia Formation.

Description of the Lower Carboniferous formations

Vegamián Formation. – This formation was first referred to as 'couches de Vegamien' by Comte (1959, p. 330), who mentioned its occurrence near Vegamián in the Porma valley. Brouwer and van Ginkel (1964) proposed the Sella Formation, taking the black shales and overlying red 'griotte' limestone together as one formation. Wagner (1963), van Ginkel (1965), van Veen (1965), de Sitter & Boschma (1966), Savage (1967), van Adrichem Boogaert (1967), and other subsequent authors continued the use of the older name. Van Veen (1965, pp. 62–63) gave a detailed description together with a measured section in the Los Lomas river north of Cardaño de Arriba. This section can be considered as a reference section for the Palentine facies area. Such a reference section is necessary, since apart from lithological differences, a different thickness, etc., the basal contact of the Vegamián with the Vidrieros in the Palentine area is not proved equivalent with the Ermita-Vegamián boundary in the Asturo-Leonese area. The lithologic characteristics of the Vegamián Formation are, according to van Veen (op. cit.): a development of some 30 m dense hard black slates, containing phosphate-bearing nodules and lenses; the formation has a characteristic white weathering colour; black chert is found in the slates, in small lenticular occurrences, or finely laminated fine-bedded with a greenish colour in thicknesses of meters; in the lowest and uppermost parts, thin lenses of black argillaceous dense limestones may occur.

In Liébana the Vegamián Formation is known from only three exposures. The most complete exposure is the one mentioned above. Another exposure is found W of Mogrovejo, there the Vegamián may be in autochthonous position. The third exposure is found E of Vallejo, there the allochthonous position of the Vegamián seems obvious since the nearby Murcia is younging away from this exposure. The occurrence of limestone in the Vegamián is known from the exposure north of Enterrias: the basal part of the Vegamián. The other exposures contain only black slates with phosphatic nodules and a whitish weathering colour. No fossil material was encountered. The age of the Vegamián is Upper Tournaisian to Lower Viséan (Higgins et al., 1964; van Adrichem Boogaert, 1967). On top of this formation one usually encounters a development of nodular limestones and shales, sometimes with cherty radiolarite beds. The transition seems to be conformable; in good exposures a gradual passage is observed and no biostratigraphic hiatus exists (Higgins; van Adrichem Boogaert, op. cit.). There is reason, however, to suppose for the Liébana area an unconformity between the Vegamián Formation and the overlying Villabellaco Formation, as will be explained below.

Villabellaco Formation. – The sequence of nodular limestones and shales overlying the Vegamián Formation in the Cantabrian area can be divided in two regionally different developments. In the Asturo-Leonese and Cantabrian block regions, a pink to dark-red colouring is the most conspicuous feature of this sequence. We will describe this development, generally known as the Alba Formation, when describing the Upper Devonian and Lower Carboniferous of the Picos de Europa area.

In the Palentine region, the Viséan nodular limestones have no typical development and are hard to distinguish from the Upper and Middle Devonian nodular limestone formations. Wagner (1955) was the first who described this sequence of grey nodular to wavy-bedded limestones in the Barruelo area (Palencia), for which he proposed the name Villabellaco Formation. This formation, containing goniatite faunas ranging in age from Lower Viséan to Middle Namurian (Wagner-Gentis, 1955, cited in Wagner & Wagner-Gentis, 1963) "represents an extremely condensed succession of strata which are only about 20 m thick. It rests with an angular unconformity on Devonian rocks" (Wagner & Wagner-Gentis, op. cit.).

Van Veen (1965) describes the Viséan limestones in the Cardaño area as "a light gray rock being found as large lenses built up of poorly bedded and massive recrystallized calcarenite and argillaceous limestones" overlying the Vegamián apparently conformably. Van Veen (op. cit.) refers to this sequence as Alba Formation, as does van Adrichem Boogaert (1967). Since there is a clear distinction between the Viséan Alba Formation as described by Comte (1959), and the sequence described above, I prefer the Villabellaco Formation as a distinct unit for the Viséan-Lower Namurian grey-coloured nodular to wavy-bedded limestones in the Palentine facies area.

The Villabellaco in Liébana generally has a more massive character than the Vidrieros. Locally, the nodular character is almost lost, for instance the well-bedded greyish white weathering, dark-grey calcarenitic limestone of loc. 71C304 (an allochthonous occurrence SW of La Vega). A massive development passing vertically into a more shaly development can be seen near Toranzo (Photograph 1). The nodular limestone is exposed in autochthonous position S of Barcena, being almost in contact with the Murcia Quartzites; Vidrieros and Vegamián rocks were not observed here and this occurrence has led me to assume a locally unconformable contact between Villabellaco and older formations. Other but allochthonous occurrences are found W of Campollo, SW of La Vega de Liébana and W of Mogrovejo. Conodont determinations of the samples 71C185, 71C270 and 71C304, all yielded a Viséan age. Sample 71C84 resulted in a Lower Namurian age.

UPPER DEVONIAN AND LOWER CARBONIFEROUS IN THE PICOS DE EUROPA AREA

Ermita Formation. – The Ermita Formation was formally introduced by Evers (1967, p. 97) for a Famennian to Lower Tournaisian sequence of 140 m calcareous sandstones, known to Comte (1959, p. 193) as the 'Grès de l'Ermitage', and mentioned by Wagner (1963, p. 35) as La Ermita formation. The formation "is usually less than 10 m thick and unconformably overlies progressively older strata" (Evers, op. cit.).

From the Picos de Europa area I know only one exposure that might be Ermita: a coarse, cross-bedded sandstone of some 5 m thickness, containing a fair amount of limestone grains, at the top in sharp parallel contact with the Alba Formation, at the base cut off by a thrust fault, and exposed near the entrance of the Espinama–Aliva road into the Picos (las Portillas), E of the road. No fossils were obtained here*. Some 50 m above and E of this exposure, where the old path from Pembes to Aliva enters the Picos, muddy sandstone is exposed below the Alba, containing a restricted fauna of brachiopods and corals. Determinations of the corals (*Cyathaxonia* sp., *Lophophelidium* sp., Dr. G. E. de Groot, Rijksmuseum van Geologie en Mineralogie te Leiden) and brachiopods (*Brachetyra* sp., Dr. C.F. Winkler Prins, Rijksmuseum van Geologie en Mineralogie te Leiden) point to an Upper Devonian to Lower Carboniferous age. With respect to lithologic characteristics, this rock occurrence might also belong to the Vegamián Formation (cf. Higgins et al., 1964, p. 214).

Alba Formation. – The Viséan nodular limestone sequence with its characteristic red colour, is developed in the Asturo-Leonese area and in the Cantabrian block area. This sequence was known to Comte (1959, pp. 40 and 330) as the 'Griotte de Puente de Alba', a red nodular limestone and shale sequence resting paraconformably on the 'couches de Vegamián'. Most subsequent authors used the name Alba. Van Ginkel (1965, p. 184) was the first to mention the Alba Formation as a formal rockstratigraphic unit. Evers (1967, p. 105) offered a subdivision in three members, based on the investigations of Winkler Prins (published in 1968). Wagner et al. (1972) described this formation from a reference locality near Genicera, and proposed that locality as the type locality for the Genicera Formation which they introduced to replace the Alba Formation. To me it seems in accord with the priority rule to maintain the Alba Formation as the formal name.

Due to the development of nappe-like structures in the Picos de Europa area in which the Alba seems to have acted as the 'slide-horizon', exposures of this formation show rocks with strong internal deformation and incomplete stratigraphy. Only in the exposure mentioned above (Espinama–Aliva road), the base of the Alba can be observed. The Alba is mostly developed as a yellow-weathering, greyish-pink to dark-red, nodular limestone,

* A similar calcareous sandstone occurring at the base of the first mappe in the Rio Cares some 10 km to the west has been sampled by Mr. H. Gomez and has yielded a definite Tournaisian conodont fauna according to Drs. M. van den Boogaard, Rijksmuseum van Geologie en Mineralogie, Leiden. This suggests a correction with the Vegamián Formation in which sandstone layers are known (Higgins et al. (1964)).

sometimes interbedded with cherty shales of the same colour; goniatites are common in these rocks, but corals and crinoids also may occur. In places, the nodular character is almost lost. Conodonts obtained from some localities (70A2, 70C257), indicate a Viséan age. The upper boundary of the 'griotte' is gradual, the overlying Caliza de Montaña is at the base very finely laminated, the laminae alternate in colour from grey to almost black and sometimes pink. In some places, the red nodular limestone development is repeated after some twenty to fifty meters of grey laminated limestone, notably at the base of the third nappe (counting them from south to north), as exposed on the top of the Peña Vieja, and at the base of the fourth nappe as exposed between the Tabla de Lechugales and the Peña Santa. In other places the occurrence of Caliza de Montaña under the Alba Formation seems to have a tectonic cause, so for instance at the base of the third nappe in the valley of the Rio Duje near Las Vegas de Sotres and north of Tama, where the Potes–Unquera road enters the Picos de Europa. At the latter site, going from S to N, the succession starts with 80 m dolomitized limestone, and then, in a recess E of the road, some hundred meters of vertical-standing, well-laminated Caliza de Montaña can be observed, younging N, with at the base a four meters thick sequence of pink cherty shale, a lateral equivalent of the nodular limestone.

DISCUSSION AND INTERPRETATION OF DEVONIAN AND LOWER CARBONIFEROUS

The environmental significance of the nodular limestone-shale association, the most distinguishing feature of the Palentine facies, is still unknown. No generally accepted hypothesis concerning the genesis of nodular limestone exists. We can however draw some conclusions, regarding certain differences between the Palentine facies and the more widely spread Asturo-Leonese facies. The rate of sedimentation has been extremely low in the former environment; the total thickness from the Abadia Formation up to the Vidrieros Formation is about 450 m. Considering that the Murcia Formation comprises a third of this thickness, we have a bare 300 m of rock representing limestone-shale sedimentation. The time interval from the beginning of the Eifelian to the end of the Famennian may be estimated at 40 million years, which gives a sedimentation rate of 8 mm per 1000 years, disregarding the Murcia sedimentation. The Asturo-Leonese equivalent of this sequence, from the upper part of the La Vid Formation up to the Nocedo Formation, has an average thickness of 1400 m in the Luna area to 1250 m in the Esla area; about four times as rapid in an environment that is generally regarded as a relatively quiet, shallow shelf. The estimated rate of sedimentation in this facies area in the same time span is about 35 mm per 1000 years.

The deposition of the Murcia sediments can be explained as an influx of mainly immature turbidites in a basin where nodular limestone-shale sedimentation was autochthonous. Sedimentary features as described for the Murcia, point to turbidity current genesis although the C and D intervals of the Bouma sequence (Bouma, 1962), have not been encountered in the field. This influx of allochthonous material interrupted, but did not replace, the autochthonous sedimentation. We have already mentioned the occurrence of nodular limestone between the Murcia Quartzites. We may assume that basin features such as depth and autochthonous sedimentary environment remained unchanged. This suggests that the nodular limestone-shale deposition took place in an environment into which transport of sediment by means of turbidity currents was also possible.

The Upper Famennian unconformity at the base of the Ermita Formation, an important feature in the Asturo-Leonese and Cantabrian block areas, is not present in the Palentine area. The S boundary of the Palentine facies is the Cardaño line (van Veen, 1965) whereas the N boundary can be drawn along the fault contact at the base of the Picos de Europa. Already we have shown reasons to distinguish a different stratigraphic development in Lower Carboniferous times in this area, contrasting with that of the Asturo-Leonese and Cantabrian block areas. Hence, the Palentine facies is extended into the Lower Carboniferous. In the Palentine area, the Lower Carboniferous is distinguished by a relatively thick Vegamián (30 m average, locally more than 50 m), succeeded by the Villabellaco Formation for which a thickness of 25 m is reported from the type area. The Villabellaco reaches into the Lower Namurian (Wagner-Gentis in Wagner & Wagner-Gentis, 1963; this thesis, sample 71C84). In the Asturo-Leonese and Cantabrian block areas the Vegamián is thin to absent. The total thickness of Vegamián and Alba usually does not amount to more than 20 m. It remains doubtful whether the Alba reaches into the Lower Namurian (Higgins et al., 1964; van Adrichem Boogaert, 1967; Wagner, Winkler Prins & Riding, 1972). The Montó area was a transition area during the Lower Carboniferous, a 50 m thick Vegamián rests paraconformably on the Vidrieros and is overlain by a mainly red nodular limestone sequence, the Alba Formation (Kutterink, 1966). In the Asturo-Leonese and Cantabrian block areas the Alba passes gradually into younger sediments. In the Palentine area the Villabellaco is cut off unconformably by the Sta. Maria Formation (Wagner & Wagner-Gentis, 1963); the Alba in Montó may even be absent, due to the unconformable base of the succeeding limestone sequence (Kutterink, 1966, bijlage III, sections IIIB, IIIC). The sedimentation rate of the Vegamián and Villabellaco in the Palentine area is estimated at 4 mm per 1000 years, about the same order of magnitude as the Palentine Devonian. The sedimentation rate of Vegamián and Alba outside the Palentine area was even smaller. In regard to such facts as the unconformable contacts at the bases of the Alba/-Villabellaco and Vegamián Formations, the local absence of Vidrieros and/or Vegamián and the generally patchy appearance of the Lower Carboniferous deposits, we must bear in mind that the sedimentation has been extremely slow, with the possibility of lacunas. The chances that a few meters of sediment, deposited in a large interval of time, remain unaffected and intact, while not being buried by subsequent deposition, are not so great. In other words, it is not necessary to assume even epeirogenetic movements during the Lower Carboniferous to explain the existing unconformities between, and the patchy appearance of, the Lower Carboniferous formations.

STRATIGRAPHY OF THE UPPER CARBONIFEROUS

Review of the historical development of different classification systems

As already stated, lithostratigraphic subdivision of the Upper Carboniferous strata is rendered difficult by the very nature and origin of these rocks. During the same time interval and within a restricted area, entirely different rocks developed in different facies areas, that show a changing pattern with the lapse of time. Sedimentation does not seem to have been continuous; several unconformities can be distinguished. It seems that none of them extends over the whole Cantabrian area; most unconformities are only locally important. To complicate things further, the products of entirely different facies areas may appear remarkably similar to the untrained eye. The question whether certain sandstone-shale alternations should be regarded as belonging to the turbidite facies association, or to a paralic association, can only be solved by a careful sedimentologic analysis; similar problems exist concerning conglomerate deposits. Such problems become important where several authors have, to some extent, used sedimentologic criteria, in distinguishing rock units in those areas.

The exposed Upper Carboniferous in the Cantabrian Mountains is not continuous but divided, by faults and exposures of older palaeozoic rocks, in areally restricted and separated occurrences. Dividing the Upper Carboniferous rocks in siliciclastics and calciclastics, we can make the general statement that correlation of limestone bodies is only possible with the help of palaeontologic evidence, and that the correlation of the siliciclastic rocks is almost entirely based on this former correlation, with only the exception of Upper Westfalian and Stephanian terrestrial strata where palaeobotanic evidence is very helpful. The latter circumstances necessarily induced authors to allow their lithostratigraphic system to depend to some extent on biostratigraphically established relations.

Now, in the subject area, various mutually dependent lithostratigraphic systems from different authors that have influenced each other during the progress of the investigations, are important. The first subdivision was given by Wagner (1955), later revised and completed (Wagner and Wagner-Gentis, 1963; Wagner and Winkler Prins, 1970). This lithostratigraphic subdivision is compiled in the area of Barruelo de Santullan (Palencia), and was primarily not intended for the overall Carboniferous development in the Cantabrian Mountains. Some of its formations, however, were used by other authors in different areas (Frets, 1965; de Sitter and Boschma, 1966; Boschma and van Staalduinen, 1968), sometimes boundaries were changed, and formations redefined, a practice that has created considerable confusion.

The lithostratigraphic units from various authors (Wagner and Wagner-Gentis, 1955, 1963; Kanis, 1956; Nederlof, 1959; Koopmans, 1962; Frets, 1965) and from numerous internal reports by students, were compiled by de Sitter and Boschma (1966), who published a formal lithostratigraphic subdivision for the Carboniferous rocks of the Pisuerga area, together with the first map sheet of the Geological Map of the Southern Cantabrian Mountains. Unfortunately, most of the units thus compiled were informal, lacking in sufficient description, correct definition of boundaries, type section and locality and so on. In spite of such shortcomings, this subdivision, enlarged and completed with units from other areas by Boschma and van Staalduinen (1968), has been used on all the subsequently published map sheets of the above-mentioned series.

Brouwer and van Ginkel (1964) published a lithostratigraphy for a large part of the Carboniferous of the S Cantabrian Mountains. Some of the formation names were taken over by de Sitter and Boschma (op. cit.). Correlation of the geographically separated occurrences, was based mainly on fusulinid biostratigraphy. This system has, to a lesser extent, similar deficiencies as the one mentioned above. Moreover, the proposed subdivisions are too strongly biased by the then incomplete knowledge of certain parts of the Cantabrian Mountains. Van Ginkel (1965) revised and completed this system in a second edition, in which a lot of the formation names were replaced by older names, on account of the priority rule.

Here I shall not enter in detail upon the respective merits, differences and incongruities of the above-mentioned systems. That will be treated, whenever it is necessary, along with the description of the formations I distinguished in my area. One aspect must however be mentioned. In the system of de Sitter and Boschma, the entire Carboniferous is divided into three groups: the Ruesga Group, the Yuso Group and the Cea Group. This division is based on inferred geological history. The Ruesga Group comprises all the formations that belong to the Carboniferous and antedate the beginning of the Sudetic folding phase; the top boundary is drawn at the base of the first unconformable conglomerate that is correlated with this phase. The Yuso Group comprises the formations that postdate the Sudetic phase and antedate the Asturian phase, the Cea Group comprises the remaining younger formations of the Carboniferous. Most authors disagree about the chronostratigraphic level, number, intensity, extent and nature of the different folding phases, a situation which is merely a reflection of the complicated tectonic history. In places where unconformable contacts are lacking, the boundaries of the proposed groups are not mappable. On all the maps of the above-mentioned series, this tripartition has been used. Wherever the group boundary could not be based on an unconformable contact, arbitrary boundaries based on biostratigraphic evidence were drawn. This also is rather unfortunate since biostratigraphic evidence should not form a major criterion in the definition of rock units. The Ruesga, Yuso and Cea Formations and Members are represented in different hues of blue, brown and yellow in this order. Since the herewith published map is part of the series, the same colours and the same group divisions are used. Formations have in consequence been fitted into the group system, which in spite of its incorrect nature and other disadvantages, gives at least an idea of chronostratigraphic equivalence of, lithostratigraphically, entirely different formations.

Presentation of data

The data concerning the stratigraphy of the Upper Carboniferous are represented in a number of composite columns and a schematic block diagram in which the occurrences of formations and members in a certain area are put together in order to give an idea of the lateral and vertical relations (Enclosure III). The subareas which correspond with the different composite columns and the block diagram are given on the index map. In Enclosure VI, the same columns have been set out against the West-European and Russian chronostratigraphic subdivisions of the Carboniferous. The given correlation is largely based on fusulinid biostratigraphy, the correlation of the West-European and Russian standard is based on the conclusions of van Ginkel (1965, and personal communications). The use of the Cantabrian stage in the subject area is based on the decision of the I.U.G.S. Subcommission on Carboniferous Stratigraphy at its meeting in Krefeld 1971.

All localities that yielded determinable fossils are represented on a 1:100 000 index map (Enclosure VII). All fossil samples which have given results with respect to a relative age are listed in the Appendix. Of 67 limestone samples collected during the fieldwork, the fusulinid content was identified by Dr. A. C. van Ginkel of the Geologisch Instituut der Rijksuniversiteit te Leiden. The slides from those samples, identically coded as the localities, are kept in his collection. In some samples, algae have been identified by Dr. J. J. de Meyer, Shell-Rijswijk. From 32 additional localities, fauna determinations were known from existing publications and internal reports. From 9 localities floras were known, or obtained during the fieldwork. Two floras were identified by Dr. F. Stockmans (cited in van Ginkel, 1965, p. 211), the other floras were identified by Dr. R. H. Wagner, Geologic Department of the University of Sheffield.

We will now proceed with the description of rock units. First, the lithostratigraphic subdivision for the Picos de Europa area will be treated. The rock units from that area are entirely different from the units of Liébana, Northern Pisuerga and the N Rio Yuso area, where in spite of different facies regimes, correlation is possible.

Rock units of the Picos de Europa area

Caliza de Montaña Formation. – In the larger part of the Cantabrian Mountains, the Viséan 'griotte' limestone passes gradually into a sequence of dark-grey, finely laminated unfossiliferous limestone which is generally referred to as 'Caliza de Montaña', an old name already used as an informal rock unit by some Spanish authors in the last century. The basal boundary of this unit with the Alba Formation has never been doubtful. Barrois (1882) introduced the name 'Calcaire des Cañons', which was later adopted by Delépine (1943) and Comte (1959). Brouwer and van Ginkel (1964) introduced a formal rock unit, the Escapa Formation, to replace the informal Caliza de Montaña. This formation was based on a general description of the occurrence of the 'Calcaire des Cañons' in the Sierra de Escapa by Barrois (op. cit.). In this description, however, no upper boundary is defined and the chronostratigraphic extent is doubtful. Most subsequent authors retained the old name 'Caliza de Montaña', either as a formal unit (de Sitter and Boschma, 1966; Boschma and van Staalduinen, 1968; and other authors) or as an informal unit (Julivert, 1961; Martinez Alvarez, 1965).

In view of this state of affairs, in this description the name Caliza de Montaña Formation will be used in spite of formal objections. There are several descriptions of this formation by various authors (amongst others Kanis, 1956; Koopmans, 1962; van Veen, 1965; Sjerp, 1966; Evers, 1967; Winkler Prins, 1968; van den Bosch, 1969). Considerable differences in thickness seem to be possible. Fossil content is poor to absent. The top of the formation is bounded by different rock units in different areas, so that it is not clear whether, when this name is used, it always refers to stratigraphically equivalent rock bodies. In view of such facts, the choice of a type section and locality may better wait until a regional study has been made and a representative development can be appointed.

In the present paper, the occurrence of the Caliza de Montaña in the south-eastern part of the Picos de Europa will be described. I cannot supply a type locality and section either; I do not know of any place in the studied area where the formation can be studied in its original lithology from bottom to top. Secondary dolomitization has affected the limestone to such an extent that in most exposures the old lithology is, in a very irregular pattern, for more than 50 % converted into a coarse crystalline dolomite, with yellow to brown, sometimes pink, weathering colours. Moreover, accessibility is generally bad.

Although the field appearance of the Caliza de Montaña is that of massive, largely dolomitized strata, the formation originally consisted of a 250 m to 400 m thick sequence of bedded limestone. The lower part (at least 150 m) of this sequence is thinly laminated. The laminae are about 1 mm thick and vary in colour from dark to light grey, with locally pink laminae in the lowermost part, indicating the gradual passage into the Alba Formation. Beds of graded angular limestone grit (fining upward) and intraformational breccias (Photograph 2) are also typical for this sequence. The gritbeds are only 3 cm thick, the thickness of the breccias amounts to some meters; they consist of angular, flat fragments of the laminated limestone, in a sparry calcite or dolomite matrix. The occurrence of breccias is not restricted to one definite horizon in the sequence. Lens-shaped chert concretions, parallel to the bedding plane of the laminated limestones, occur in some places. Although these limestones have a fetid odour when broken, they did not yield determinable fossils. The upper part of the Caliza de Montaña consists of a more or less bedded bioclastic limestone sequence of variable thickness (100 to 250 m). The limestone is a wackestone or grainstone; boundstones were not observed, although they may be expected, since small fossils are abundant. In slides, biopelmicrites and biopelmicrudites and also the sparry equivalents have been observed. Besides the presence of identifiable foraminifers, there is an abundance of bryozoans, algae, fragments of corals and crinoids, and shell fragments. A mappable boundary between those two parts of the Caliza de Montaña Formation could not be established. This division agrees with the subdivision of this formation in the Curueño area into a lower, Vegacervera Member, and an upper, Valdeteja Member (Winkler Prins, 1968), but a division into two formations as proposed by Wagner, Winkler Prins & Riding (1972) would not result in mappable rock units in the Picos de Europa area. The lower boundary of this formation has already been described. The upper boundary is the sharp contact between the massive appearance of the Caliza de Montaña and the coarse-bedded appearance of the lower member of the Picos de Europa Formation (Photograph 3).

The upper part of the Caliza de Montaña Formation has a Bashkirian age, as indicated by foraminifera (samples and localities: 70215, 70277, 7177, 71146 and 71149) which belong to the *Millerella* Zone and the lower part of subzone A of the *Profusulinella* Zone (van Ginkel, pers. comm.). The lower part did not yield fossil evidence, but considering its stratigraphic position, a Lower Namurian age is assumed.

For the upper part of the formation, the lithology points to the infratidal to intertidal depositional environment of a biogenetic bank (Roehl, 1967; van de Graaff, 1972).

In the lower part of the formation, the synsedimentary 'flat pebble breccias' (Roehl, 1967), the gritbeds and the fetid odour point to a clastic sedimentation in a very shallow environment under anoxic conditions. Autobrecciation is assumed to have happened in the intertidal zone (Winkler Prins, 1968, p. 59, citing Teichert, 1965). There seems no reason to assume chemical precipitation of the limestone as does van den Bosch (1969, p. 176, citing Teichert, 1965). As for the lower part of the formation, the above-described lithologic features can be studied in the Potès–Unquera road section near locality 7177. The upper part is best studied at the southern boundary of the Picos de Europa, north of the village Lon (localities 70259, 70277).

The Caliza de Montaña Formation as observed in the Picos de Europa area, is similar to its occurrence in the Leonides (description by Winkler Prins, 1968; van den Bosch, 1969; and others), and its occurrence in the areas west of the Picos de Europa (description bij Julivert, 1961; and Sjerp, 1966). The lower part seems omnipresent in more or less the same thickness, the upper part may be very thick (Curueño area) or almost absent (some places in the Tarna-San Isidro area). In the Palentine area, the Caliza de Montaña is only known from the southern border, along the Cardaño line (van Veen, 1965) and from a very restricted development in the Montó area (Kutterink, 1966). Occurrences in Liébana seem to be allochthonous, with exception of the Cosgaya limestone unit which may be equivalent to the upper part of the Caliza de Montaña. The idea that the entire limestone development of the Picos de Europa is of a Namurian age and belongs to the Caliza de Montaña (Martinez Alvarez, 1965) cannot be accepted. In the studied area, a limestone sequence with a thickness of about 600 m, varying in age from Lower to Upper Moscovian rests on the Caliza de Montaña with a parallel, but unconformable contact.*

Picos de Europa Formation. – This unit is proposed for the bedded to massive, largely bioclastic limestone sequence overlying the Caliza de Montaña in the studied area.

In the formation two informal members are distinguished, a bedded member and a massive member. Distinction is clear in the S part of the area, but to the north the difference is only locally discernible.

The bedded member in the southern area typically consists of bioclastic limestone layers alternating with thin layers of shale and shaly limestone; to the north, shale is absent. The bioclastic limestone is a grainstone to packstone, biomicrudites and biosparrudites are most frequent, though allochems, pellets and oöliths may occur. The presence of siliceous concretions, lens-shaped and parallel to the bedding plane, and in irregular patches through the entire rock (Photographs 4 and 5) is common in the southern nappe structure. Here the basal part of the bedded member is a typical 'Kieselkalk' (flinty limestone). In the second nappe structure (from south to north) the occurrence of coarse fossil-detritus beds, sometimes with a red matrix, at more or less the same stratigraphic level as the 'Kieselkalk', constitutes the most striking feature of the bedded member. The thickness of the bedded member is about 100 m.

The massive member of the Picos de Europa Formation consists of bioclastic limestones, sometimes bedded, in which packstones, grainstones and, notably in the upper part, boundstones are dominant. There is no

* The presence of the genus *Fusulinella* at various localities in the Picos de Europa was mentioned by Deléphine as early as 1943, after examination of rock samples, collected from that area by Gübler. They concluded that at least the upper 100 m of the 'Calcaire des Cañons' should be of Moscovian age. This result has been confirmed by van Ginkel in 1965.

marked lithologic difference with the bedded member except for the occurrence of shale in the latter. Its thickness is about 400 to 500 m.

The field occurrence of the Picos de Europa Formation is typified by strong dolomitization, recrystallization, and intense jointing, which renders 80 to 90 % of the outcrops worthless for studying the original lithology. The original thickness of the formation is locally strongly reduced by stylolitization (Photograph 6). The preservation of the bedded member as a mappable feature in part of the area, is due to the occurrence of shales and of siliceous concretions parallel to the bedding plane. An accessible type section where the entire formation can be studied in its more original lithology cannot be designated in the subject area. The bedded member can be studied northwest of Tanarrio (Photograph 3) and along the road from Espinama to Aliva, near locality 70183.

The Picos de Europa Formation in its southern development is either bounded at the top by the Aliva Shale Formation or cut off unconformably by the Lebeña Formation. The Aliva Shale Formation is chronostratigraphically equivalent to the upper part of the Picos de Europa Formation in its northern development, but, due to tectonic displacement, no lateral passage is exposed. In the northern nappe structures, the Alba Formation rests with a tectonic contact on top of the Picos de Europa Formation.

Wherever dolomitization is absent, fossils are abundant both in the bedded and the massive member. Algae, brachiopods, bryozoans, crinoids, corals and foraminifera are most common. In some places biostrome-like masses of algae in algal boundstones could be observed in the upper part of the massive member. Sometimes sponge spiculae are an important rock constituent.

Fusulinids have been found throughout this formation, in the bedded member (samples 70183, 70259, 70273, 70775, 70278 and 70295) and in the massive member (samples 70202, 70202A, 70202B, 24-6-70 I, 70205, 70208, 70271, 70322, 71148 and 71144). Some samples from this formation (70206, 70207) could not be assigned to a member on field evidence. Examination of fusulinid content of these samples clearly shows that the Picos de Europa Formation is diachronous. Its base varies rather gradually from *Profusulinella* Zone subzone B to *Fusulinella* Zone subdivision B 1; its top may reach the top of the subdivision B 3 of the *Fusulinella* Zone (sample 70271, a specimen of *Fusulina* cf. *bella* Semikhetova et Melnikova). This indicates a continuation of the limestone sedimentation up to the Moscovian-Kasimovian boundary.

For the massive development of bioclastic limestones and boundstones, intertidal to infratidal conditions are most likely (Roehl, 1967; van de Graaff, 1972). For a 'Kieselkalk' facies, a deeper shelf environment is supposed by Funk (1971), who investigated the very similar 'Helvetische Kieselkalken'. He assumes a quiet shelf-environment at a depth of about 150 m. The red detrital limestone horizons could well figure as a transition facies between the infra- to intertidal sedimentation and the deeper shelf sedimentation of the 'Kieselkalk'. So, the facies concept is that of a shallow shelf area, the Picos de Europa area, with continuous, mainly bioclastic limestone sedimentation (a Bahama-like environment) throughout Lower and Upper Moscovian times. In this area, depth increased to the southern borders (towards the Liébana area). The occurrence in the southernmost nappe of the Aliva Shale Formation, which probably represents a deeper environment, agrees with this concept.

No trace of erosion has been found between the Caliza de Montaña and the Picos de Europa Formation, so that this biostratigraphic gap must be a true hiatus.

The Picos de Europa Formation together with the overlying Aliva Shale Formation, are roughly the chronostratigraphic equivalent of the Escalada Formation together with the overlying Fito Formation in the Beleño basin W of the Picos de Europa (van Ginkel, 1965). Other correlations include the Lois-Ciguera Formation near Riaño (Brouwer and van Ginkel, 1964) and the Corisa Formation in the Pisuerga area (van Ginkel, 1965).

Aliva Formation. – In the first nappe, a sequence of silty and muddy shales with yellow to dark-brown weathering colours, alternating occasionally with beds of detrital limestone, rests on top of the bedded member or the massive member of the Picos de Europa Formation. The basal contact with the Picos de Europa Limestone may seem unconformable (for instance near locality 70192) but this is probably due to the sedimentary fill of irregularities on the limestone surface, caused by partial solution after the limestone arrived in an environment deeper than its depositional environment. The spatial relations between this unit and the Picos de Europa Formation are schematically represented in the stratigraphic diagram of the Picos de Europa area (Enclosure III).

I propose the Aliva Shale Formation as a formal name for this sequence, which is well developed in the Aliva valley, SE Picos de Europa. The best exposed and fairly complete section is to be found on the western slope of the upper course of the Rio Seco. The top of the formation is never seen, since everywhere the second nappe rests with a tectonic contact upon it.

The formation consists of silty shales for more than 50 %. Occasionally, at its base, limonite concretions of an irregular and subrounded shape with violet-brown to yellow weathering colours occur. Two kinds of detrital limestone intercalations in the shale can be distinguished. In the first place, thin continuous beds of about 5–30 cm thickness, which show all the sedimentary characteristics of turbidites: an erosive base with groove casts, lateral continuity (as far as exposed), constant thickness and a sequence of sedimentary structures as described by Bouma (1962). Secondly, thick, lensing beds, consisting of graded breccias of angular limestone fragments (boulder to pebble sizes grading to grit) in a mud matrix. The breccias fine upwards and pass vertically into a graded coarse detrital limestone which may show a partial or complete sequence of 'Bouma' turbidite intervals. These beds vary in thickness between 0.5 and 5 m and they increase in number from W to E. Still further E, across the Rio Seco, even thicker limestone breccia lenses are found, alternating with silty shales and an occasional cross-bedded sandstone lens that may contain fossil wood fragments. The muddy shale north of the Refugio de Alive contains large irregular masses of limestone, seemingly floating, so that they may be considered as olistoliths, although the field evidence is too scanty to prove a mass-slide transport.

All these limestones contain fossils, mostly fragmented and notably abundant in the coarse detrital intervals. The fauna assemblage resembles the fauna of the Picos de Europa Formation.

The age of the Aliva Formation is inferred from fusulinid determinations of the samples 70A1 1, 70A1 5, 70186, 70191, 70192, and sample 70263, the only occurrence of the Aliva Formation outside the first nappe. Since all the samples are from detrital or probably slumped limestone bodies, as shown in the diagram (Enclosure III), only maximum ages can be inferred. However, the determinations show a normal succession from the *Fusulinella* Zone subzone A to the *Fusulinella* Zone subdivision B 2, from the base to the top of the formation. There is one exception, sample 70A1 1 is slightly older than might be expected from its stratigraphic position; this sample is from one of the probable slump bodies, and in a way this deviation can be regarded as additional proof for emplacement by slumping.

The relatively thin detrital beds are best regarded as turbidites. The limestone breccias are almost identical to breccias described from the slope facies of the Dimple Limestone (Marathon Region, Texas) by Thomson and Thomasson (1969, pp. 75 and 76), which were interpreted as proximal turbidites or fluxo-turbidites. In fact, the relations between the different facies types of the Dimple Limestone bear a strong similarity to the relations between the Picos de Europa and Aliva Formations. Similar breccia beds were described by van de Graaff (1971, pp. 167 and 177, facies type Ih). He assumed deposition by grain flow layers for the breccious part and turbidity current deposition for the detrital upper part. As already stated, the big limestone bodies north of the Refugio are regarded as slumped olistoliths. The appearance of very thick breccia lenses and cross-bedded sandstone lenses to the east, points to sedimentation in a more shallow-marine environment. So, the overall picture is that of a marine sedimentation area, shallow in the east, deeper to the west, where mass transport by means of turbidity currents, high-density currents and slumping was frequent. The area was situated at the southern margin of the Picos de Europa shelf area during Upper Moscovian times. Depth can only be stated in relative terms: deeper than the shelf area.

The Aliva Shale Formation is probably equivalent to the Fito Formation in the Beleño basin (van Ginkel, 1965).

Lebeña Formation. – East of the Canal de San Carlos and S of the villages Allende and Lebeña, a sequence consisting mainly of limestone conglomerates, limestone breccias, sandstone beds and shales, rests with a sharp angular unconformity on the already described older formations of the Picos de Europa area. For this sequence the name Lebeña Formation is proposed; this is derived from the Lebeña valley (and village), where this sequence is fairly well exposed along the west side of the road going from the bridge over the Deva northwards up to the next thrusted mass of the limestone complex. Permian and Triassic rocks rest with an angular unconformity on top of this formation, E of the Rio Deva; in other places the formation is cut off tectonically by thrust planes. No conformable contact with younger strata is known from this area.

One formal member is distinguished, the San Carlos Conglomerate Member. This basal member is a very thick-bedded sequence of conglomerate and breccia layers, consisting for 99 % of limestone. The roundness of the pebbles in the conglomerates is astonishing, though partly destroyed by pressure solution. The pebbles are mostly small in size (0.5–2 cm diameter longest axis). Due to its lithologic homogeneity, the San Carlos Member weathers just like a massive limestone, presenting difficulties in mapping. West of the San Carlos fault (see Fig. 2), the conglomerate is present in patches and lenses in shales, not in direct contact with the limestones. Where extensively developed (directly E of the Canal de San Carlos), the San Carlos Member consists of 300–400 m conglomerates and breccias, followed by a 200 m thick sequence of thin- to medium-bedded sandstone with very thin shale intercalations, interbedded with thick- to very thick-bedded limestone breccias and conglomerates (Photograph 7). This sequence passes gradually into the shale. The sandstone beds are almost white, fine-grained, quartz arenites, in which fossil woodfragments may occur. Lithic wacke sandstone beds occur in the road section of Lebeña. The shale member shows almost the same features as the Aliva Shale; even olistolithlike limestone bodies together with patches of limestone breccias are present in the shales, notably north of the village Lebeña.

In the alternation of pelitic, arenitic and rudaceous beds, the breccias occur as lenses, disappearing laterally (Photograph 8). Normally, these breccias fine upwards, a gradation, evident in pebbles and matrix. The ratio of mud in the matrix increases upwards, and in the upper parts of a unit, limestone pebbles float in the mud (Photograph 8). This may also happen in isolated strings without observable relation to a breccia lens (cf. van de Graaff, 1971, facies type IId). Another, less common, type of breccia is typified by coarse calcarenite to grit, in which strings of angular pebbles float; parallel lamination, and traces of cross-bedding are the normal sedimentary features. In most cases a fining-up gradation is present, but the reverse may also be found (Photograph 9). A channelling base does occur. This sediment seems a coarse-grained variant of the type described by van de Graaff (1971) as facies type Ic.

In the real conglomerates, the pebbles are well rounded, with a two-dimensional sphericity of about 0.65 (Rittenhouse, 1943), and are very tightly packed, with almost no matrix. These features would fit well in a conglomerate that originated in a zone of surging. The isolated well-rounded conglomerate occurrences W of the San Carlos fault, surrounded by shales, may be the result of subsequent slumping.

In my opinion the most acceptable hypothesis for a depositional environment in which the above-mentioned facts fit, is as follows. In an initial stage of orogenesis, the limestone shelf area was uplifted and slightly folded, emerging as island ridges open to erosion and abrasion. The breccias may have accumulated as torrential fans which in the instable coastal environment were transported downwards into the sea as mudflows and grainflows. In stable times, abrasion conglomerates originated at the coast, parts of which also would slide down. With the waning of erosion, the deposition of shales became relatively more important.

I cannot give an adequate explanation for the mentioned occurrences of quartz sandstone beds.

All fossils sampled from the conglomerates are allochthonous and essentially the same as described for the older formations. Fusulinids from different pebbles range from the *Millerella* Zone up to the *Fusulinella* Zone subdivision B 2 or B 3 (Samples 70SC and 71SC 212). It is a remarkable fact that fusulinids of such different ages, which must have been distributed over a succession of about 1000 m of limestone, occur together in one spotsample (70SC) of about 3 kg. It is difficult to imagine any other environment but a surge zone to cause this mixing. No determinable fossils have been found in the shales and sandstones. The maximum age inferred from allochthonous fossil content is Upper Moscovian. The age inferred from circumstantial evidence is at least Kasimovian, since the youngest determination from the Picos de Europa Formation was on the Moscovian-Kasimovian boundary.

The Lebeña Formation may be correlated with the Kasimovian strata of the Cabrales area north of the Picos de Europa (van Ginkel, 1971).

Stratigraphy of the Upper Carboniferous in Liébana and southern adjacent areas

Cosgaya limestone unit (Encl. III, columns VIII and IX). – In the steep valleys of the Rio Lera and Rio Cuvo, as well as along the Cosgaya–Espinama road, a limestone-shale sequence is exposed in the core of an anticlinal structure: the Cosgaya limestone unit. Thickness is at least 250 m; the base of the unit is not exposed.

The sequence consists of a dark-grey, often yellowish weathering, muddy limestone, alternating in an irregular fashion with black shales. The limestone has a fetid odour on a fresh-cut surface, is medium to thick bedded, containing thin intercalations of shale. Towards the top, shale dominates, the limestone beds become thicker, up to 10 m, and more lenslike, having a rudaceous character. The shales in this upper part are not pure, but contain thin (2 mm) fairly continuous sandstone laminae, which normally show fining-up gradation and a sharp bottom. Beds and lenses of concretionary limonite are also found in the shales. The top limestone bed contains isolated, well-rounded, quartzite pebbles. The sequence then passes rather abruptly into a sequence of conglomerates (entirely quartzitic in the southern anticlinal flank, containing calcirudite lenses in the northern anticlinal flank) and coarse thick-bedded sandstones, the Cuvo Member of the Cervera Formation. The contact is probably unconformable although parallel; however, no conclusive evidence from exposures has been found.

Fossils are scarce, but in the Cuvo valley, at a topographic height of approximately 800 m, one limestone bed was found, containing many solitary corals in autochthonous growth position. Germs (1966) collected algae from this limestone unit.

According to Germs (1966), the Cosgaya limestone must be of a Bashkirian age; he correlates the algal flora with the base of the *Profusulinella* Zone (Sample Ge 167).

Quiet sedimentation under at times inoxic circumstances in a stable shallow-marine environment within the photic zone, seems most appropriate to explain the above-mentioned features. Towards the top, sedimentation conditions became relatively unstable.

According to its age and stratigraphic position, the Cosgaya limestone unit may be correlated with the older part of the Muda Formation (Wagner & Wagner-Gentis, 1963; Brouwer and van Ginkel, 1964; van de Reijd, 1969), with the Santa Maria Formation (Wagner, op. cit.), and with the upper part of the Caliza de Montaña Formation, notably its development in the Montó area (Kutterink, 1966). A detailed stratigraphic correlation of the above-mentioned units would provide a base for formal denomination. In this thesis an informal name is used to prevent introduction of a possible synonym.

Cervera Formation. – Brouwer and van Ginkel (1964) introduced the Cervera Formation for the rhythmically alternating graywacke-shale sequence, resting either on top of, or, according to the interpretation of Kanis (1956), laterally interfingering with the Caliza de Montaña of the Sierra del Brezo. The formation, in its original definition, is mainly exposed between Cervera de Pisuerga and Mudá, where it rests with a conglomeratic basis unconformably on the Mudá Formation (de Sitter & Boschma, 1966; and others). For the lithologically and stratigraphically similar deposits north of Piedrasluengas and around Potes, the Piedrasluengas Formation was proposed (van Ginkel, 1965). This name caused confusion since a Piedrasluengas Limestone Member had already been introduced before (van Ginkel, 1959).

De Sitter & Boschma (1966) and Boschma & van Staalduinen (1968) ignored the Piedrasluengas Formation. They extended the Cervera Formation to the north and east, including in it the old Piedrasluengas Formation as well as the lithologically similar deposits in the highly complicated Barruelo area, i.e. a part of the Carmen Formation (Wagner & Wagner-Gentis, 1963), the Perapertú Formation (Wagner & Wagner-Gentis, op. cit.), even the Mudá Formation (Wagner & Wagner-Gentis, op. cit.) and the Santa Maria Formation (Wagner & Wagner-Gentis, op. cit.). They accepted a Piedrasluengas Limestone Member and lowered the above-mentioned formations to member rank in the Barruelo area.

In my opinion, the extension of the Cervera into Liébana (i.e. the Piedrasluengas Formation of van Ginkel, 1965) is justified, but the extension to the Barruelo area encounters objection. The Mudá and Santa Maria Formation should not be included, since these limestones are lateral equivalents of the Caliza de Montaña and in unconformable contact with the younger siliciclastic sediments; cf. the Cosgaya limestone unit (Wagner & Wagner-Gentis, 1963; Wagner, 1972).

The occurrence of the Cervera Formation in Liébana in its present configuration is divided into three parts by two fault zones, running approximately WNW–ESE. The occurrence of the Cervera Formation in the Ledantes anticline is separated by the Orcadas fault zone from the occurrence in southern Liébana. The latter deposits are informally named the Upper Deva turbidites. The northwestern extension of the Redondo fault (de Sitter and Boschma, 1966) separates the Upper Deva turbidites from the Potes turbidites (informal name for the Cervera Formation in central Liébana and Polaciones).

The Cervera Formation in the Ledantes area (Encl. III, column VII): The Cervera in the Ledantes anticline has been described by Germs (1966a) as a sequence consisting mainly of rhythmically alternating sandstones and shales, with a subordinate content of breccious conglomerates, gritbeds and scattered limestone lithosomes. The base of the formation is nowhere exposed. At the top the formation is bounded, in this area distinctly unconformable, by the Curavacas Conglomerate. The unconformity cuts down into hundreds of meters of stratigraphic thickness. The formation is intensely folded; its exact thickness is unknown. Savage (1961) estimated that for the upper part at least 1000 m must be involved. Germs, in his attempt to measure the lower part in the Castrejon river, concluded to a thickness of 1200 m for the succession below the first limestone occurrences. This amounts to an approximate minimum thickness of 2200 m. The limestone bodies occur more or less on the same stratigraphic level and are referred to as Ledantes Limestone Member.

The sandstones in the basal 1000 m mostly have a quartzwacke composition (Germs, 1966a), they may be thin- to very thick-bedded, thick-bedding dominates. Most of the individual beds have a constant thickness, as far as the outcrop allows them to be followed. Higher up in the sequence, thin bedding dominates and a lithicwacke composition is most frequent. A strict boundary cannot be drawn, the occurrence of lithicwacke and quartzwacke beds is intermixed. Most of the sandstone beds have a fining-upwards gradation. Elements of the Bouma's turbidite succession of sedimentary structures can be recognized but the sequences are normally incomplete. The C and D intervals are missing in the thick coarse beds, and the A and sometimes B intervals are missing in the finer beds. Bottom structures like flute casts and groove casts are common. Some coarse thick units are structureless and without grading. They may contain contorted shale slabs at the base.

The conglomerates are of different types. On the one hand we see coarse sandstone with floating angular pebbles of the grit fraction, consisting of black chert, quartzite and limestone fragments. In these grits coarse parallel lamination is normal, cross-bedding also occurs. The grits are associated with the turbidites, grading is not always distinct, coarsening-upwards is possible. On the other hand the real conglomerates contain angular limestone and chert fragments but also well-rounded quartzite pebbles, and occasionally contorted slabs of shale, the pebble/matrix ratio is variable, but normally the pebbles do not touch each other; the matrix consists mostly of sandstone, but in some cases of shale. Some of the limestone pebbles could be identified as fragments of the Caliza de Montaña and of the Lower Devonian Lebanza Formation (Germs, 1966a).

The bulk of the Ledantes Limestone bodies are classified (Germs, 1966) as algal biostromal and algal biohermal reefs (classification system of Nelson, 1962). In thin section they may be classified as pellet-containing intramicrites, micrites and algal biolithites (Germs, op. cit.). The limestone lithosomes contain however calciruditic parts too, in which quartzite boulders and lydite fragments occur; even sandstone beds occur in the limestones. The final sedimentary position of the limestone bodies is allochthonous, slumped lenses and knolls, surrounded by a 'matrix' of mud with contorted sandstone beds.

The sedimentation of the rhythmic sandstone-shale alternations is best explained by the turbidity current hypothesis. Though some features (cross-bedding and coarsening-up gradation in grits) seem contradictory, the bulk of the sediment displays all the normal characteristics of a turbidite sequence (Kuenen & Migliorini, 1950; Bouma, 1962; Kuenen, 1967). The grits are very similar to the petromict conglomerates described by Hubert, Lajoie and Leonard (1970) from a flysch sequence in Quebec. The conglomerates with both well-rounded and angular pebbles, floating in a sandstone or shale matrix, can be explained by high-density current resedimentation; they are considered as pebbly mudstones and pebbly sandstones. The occurrence of limestone knolls and lenses in a 'matrix' of contorted siliciclastics is best explained as the result of a chaotic slide and slump resedimentation; an olistostrome deposit (Beneo, 1956, as cited in Marschalko, 1968). Wildflysch is another currently used name for such deposits (Marschalko, op. cit.).

The intermixed occurrence of Caliza de Montaña and Lebanza fragments in conglomerates, probably points to a source area in the south, the Cardaño-Polentinos area, which must have been affected by erosion during the deposition of the Cervera Formation. From this occurrence we may also assume that the Cervera Formation in this area had an unconformable relation to the Caliza de Montaña, and that in order to make erosion of the Lebanza possible, a pre-Cervera uplift, possibly accompanied ty tectonic deformation, must have occurred.

Only the Ledantes Member (sampled by Germs, 1966) has yielded identifiable algae and fusulinids. The fauna of these limestones is relatively poor, but includes crinoids, bryozoans, solitary corals, ostracods, brachiopods, algae and foraminifers. The samples of the different limestone bodies (Ge 1, 2, 3, 4, 5, 6, 7 and Ge 106) all belong to the same biostratigraphic level, *Profusulinella* Zone, top of subzone A, pointing to an uppermost Bashkirian age. This is a maximum age but the identical ages of different olistoliths point to near synchronism of sedimentation and resedimentation.

The correlation of the entire Cervera Formation will be discussed later. The Ledantes Member compares closely with the Perapertú limestones, though the former is slightly older in age.

The Cervera Formation in Southern Liébana, the Upper Deva turbidites (Encl. III, columns V, VI, VIII and IX): The Upper Deva turbidites are distinguished by two features, the occurrence of the Piedrasluengas Limestone Member at the top and the Cuvo Conglomerate Member at the base. The Piedrasluengas Member runs as a topographically pronounced, mappable feature from Los Llazos (Pisuerga area) to the Peña Sopeica (north of Dobres) and is unconformably bounded by the basal horizon of the Curavacas Conglomerate. This is in most cases a parallel unconformity, as is shown in many exposures (columns V and VI). West of Peña Sopeica, we find huge slabs of limestone in the basal part of the Curavacas Conglomerate (column VIII), which are probably eroded and somewhat transported remnants of the Piedrasluengas Limestone. Further west (columns VIII, IX), we lose the Piedrasluengas, the conglomerates form a steep isoclinal syncline in which the northern flank is cut out by the Orcadas fault (Fig. 2) to reappear again W of Peña Orcadas. Still further to the west, the Curavacas Conglomerate thins out and seems to pass laterally into a conformable succession of turbidites, in which no mappable top boundary of the Cervera Formation can be established (column IX). The lower boundary of the Cervera Formation in this area is formed by its, assumed unconformable, contact with the Cosgaya unit (column VIII). The basal part of the Upper Deva turbidites is marked by a conglomerate-coarse sandstone association, the Cuvo Conglomerate Member, which amounts to almost half of the total thickness of the Cervera in the Rio Cuvo section. The conglomerate association interfingers to the east and west with the normal graywacke-shale association, which also in this area constitutes the bulk of the sediment. At about two thirds of the thickness from the base, a string of isolated limestone bodies, conglomerate lenses and even a patch of Murcia Quartzite (south of Valcayo), occur more or less in one level. The thickness of the Upper Deva turbidites could be estimated from a detailed survey along the Rio Cuvo to amount to some 1200 m. However, in that section a part of the sequence may have been cut out by faulting (Encl. II, section $B-B^I$).

The Cuvo Member consists of an irregular succession of mainly quartzitic conglomerates, grits and coarse graywacke beds. Breccious limestone conglomerate lenses and isolated limestone bodies in a pebble-containing mud matrix occur locally. In the conglomerates the quartzite pebbles are well rounded, the limestone pebbles angular to subangular and the matrix consists of sandstone or mud. The pebble/matrix ratio is very variable, pebbles may touch or float. The conglomerates are structureless with exception of a pebble orientation with the longest axes parallel to the bedding plane, occasional gradation of the pebbles (normally fining upwards, but coarsening upwards may also occur), and alternations of the pebble/ matrix ratio, perpendicular to the bedding. The conglomerate bodies are lenslike, interfingering and alternating with breccious grits (similar to those in the Ledantes area) and thick, coarse, structureless sandstone beds. In the graywacke-shale association not only thickness and coarseness of the sandstone beds decrease towards the top of the formation, but also the sandstone/shale ratio changes in favour of the intercalated shale beds. In the basal part of the formation, quartzwackes and quartzarenites dominate, towards the top lithic wackes are more common. The isolated limestone occurrences at two thirds of the formation (see above) are normally breccias, with a large amount of fossil detritus in the grit fraction. The conglomerates at this level are very similar to the Cuvo Conglomerates.

We encounter again conglomerates very similar to the Cuvo Conglomerates under the Piedrasluengas Limestone Member near Dobres.

The Piedrasluengas Limestone has been described in detail by van de Graaff (1972) and is not unlike the bioclastic limestone of the Picos de Europa Formation. The occurrence of the Piedrasluengas is laterally noncontinuous. Individual lenses range from about fifty to three thousand meters. The limestone lithosomes in the east show at the base a gradual vertical passage from the sandstone-shale sequence through a calcareous shale with lenses of fossil debris, to the massive limestone (van de Graaff, op. cit.). The lithosomes E of Sopeica are smaller and more numerous, lying above each other, embedded in a shale with quartzite pebbles, on top of the conglomerates. Synsedimentary distorted strings of sandstone occur in the shales.

As already mentioned, the rhythmic sandstone-shale alternations are explained as turbidites. Considering the lack of internal structures, the coexistence of well-rounded and angular pebbles and the abundance of a structureless shale or sandstone matrix in which the pebbles may float, the conglomerates are explained as a result of resedimentation by high-density currents (fluxoturbidites). In the course of time, this high-density current resedimentation was for the most part replaced by turbidity current transport, possibly an indication that the erosion in the hinterland was waning. The unnamed level at two thirds from the base, with conglomerates and limestone olistoliths, testifies to the recurrence of high-density current resedimentation, accompanied by down-sliding of olistoliths. The mentioned slab of Murcia Quartzite ascertains that in places, probably on the Mid-Liébana ridge, the Murcia was exposed to erosion. The Piedrasluengas Limestone represents a change in sedimentary environment. The depth of the turbidite sedimentation is supposed to exceed 125 m (van de Graaff, 1971). The Piedrasluengas is a shallow-marine autochthonous biogenetic bank deposit, with a subordinate amount of 'lagoonal' deposits and bioclastic grainstones and packstone, formed in agitated water in littoral to sublittoral conditions (van de Graaff, 1972). This shallow environment is supposed to have had only a regional extension. West of the Peña Sopeica it is replaced by a turbidite facies. The occurrence of isolated slabs of limestone, seemingly pushed on top of each other, in a pebbly mudstone-shale environment in the Rio Frio valley east of the Peña Sopeica, is best explained as a result of downsliding of probably already lithified parts of the biogenetic bank in a submarine canyon or in the proximal part of a fan valley (cf. van de Graaff, 1971, p. 196, the resedimentation of the Agujas Limestone). The occurrence of conglomerates near Peña Sopeica, under the Piedrasluengas, makes the existence of a submarine canyon acceptable. The downsliding may have been triggered off by the beginning of tectonic activity.

Limestone pebbles from the Cuvo Conglomerates and the limestone olistoliths associated with the Cuvo Conglomerates, yielded fusulinids belonging to the *Profusulinella* Zone subzone A (samples 70139, 7080 and 70128). Samples Ge 122 (Germs, 1966) and 7136 are either top subzone A or basis subzone B of the *Profusulinella* Zone. Sample 7144 belongs to the *Profusulinella* Zone; further specification is impossible. This renders a maximum age interval of Upper Bashkirian to lowermost Moscovian for the Cuvo Member. From the Piedrasluengas samples P 1 (van Ginkel, 1965) and Ge 118 (Germs, 1966), both belong to the upper part of the subzone B of the *Profusulinella* Zone; age Lower Moscovian, upper Vereyan to lower Kashirian (van Ginkel, op. cit.). Autochthonous fossils are unknown from the conglomerate-turbidite sequences.

The basal part of the Cuvo Member may be correlated with the basal conglomerate of the Cervera Formation that rests unconformably on the Mudá Limestone in the Pisuerga area. The unnamed conglomerate-olistolith level at two thirds from the base may possibly be correlated with the Ledantes Member, but there is no fossil evidence to warrant such a correlation. Comparing the Cervera development in the Upper Deva area with the development in the Ledantes area, the smaller thickness and the occurrence of a larger amount of coarse clastics in the Upper Deva turbidites are significant.

The Cervera Formation in central Liébana and Polaciones, the Potes turbidites (Encl. III, columns II, III, IV, X and XI): The basal boundary of the Potes turbidites is only exposed along the Mid-Liébana ridge. South of the ridge (column XI), the Cervera is developed in a synclinal structure, bounded unconformably by the Valdeon Formation in the NW (column XI), and cut off by the Redondo fault in the S. In N Liébana and Polaciones, the base of the formation is cut out by the San Carlos fault which brings the Upper Moscovian Bedoya olistostrome complex in contact with the basal Potes turbidites. In this part of the area, the upper boundary is formed by the base of the Cabezon Conglomerate Member of the Curavacas Formation. From E to W (column III to II) this contact changes from unconformable to conformable. In the E, a large olistostrome deposit, the Salceda Olistostrome Member (column III) forms the uppermost development of the Cervera. To the W this olistostrome passes into undisturbed turbidite deposits. In the same direction but more to the W, the Cabezon Conglomerate passes into a conglomerate-, grit- and coarse sandstone association which interfingers with the turbidites somewhere N of the Pico Viorna, where the mappable top boundary of the Cervera is consequently lost. At the NE side of the Mid-Liébana ridge between Mogrovejo and Pesaguero, no equivalent unconformity level marks the top of the Cervera. There seems to be a gradual passage into a succession of olistostromes which is only locally interrupted by the much younger Viorna Conglomerate (column X).

In a restricted area south of Barcena, the Cervera Formation rests with a basal conglomerate unconformably on the older Palaeozoic of the Mid-Liébana ridge (column X), the Barcena Conglomerate Member. East and west of Barcena and along the southern border of the ridge, the Enterrias Olistostrome Member rests unconformably on the eroded Devonian. At the south flank of the ridge near Llaves, this unconformity is a bottom-to-bottom contact and probably obscures a fault. In N Liébana the Lon Olistostrome Member, though not developed at the base of the Potes turbidites, possibly constitutes a lateral continuation of the Enterrias Olistostrome. The Tornes Conglomerate Member is developed at more or less the same level in E Liébana.

The bulk of the sediment consists of turbidite sequences, in which deformation by synsedimentary slumping is more frequent than in the Ledantes area and in the Upper Deva turbidites, though not dominant. Thicknesses can only be estimated; estimations from Polaciones and Liébana based on surveys for tectonic sections resulted in minimum thicknesses of 1500 m (Miedema, 1966, Rio Tornes section), 2000 m (Maas, 1968, Collario-Espinal survey) and 2600 m (Lanting, 1966, survey over Turieno and Arguebañes). I suppose, that a thickness of at least 2000 m is a fair approximation for the Potes turbidites in N and E Liébana, but the sequence probably thins towards the Mid-Liébana ridge.

The lithology, depositional environment, fossils, age and correlations of the individual members are detailed below.

1) The Barcena Conglomerate Member. This member is characterized by the simultaneous occurrences of angular limestone pebbles and well-rounded quartzite pebbles; the limestone pebbles dominate, the matrix is usually a coarse sand to grit, in which black chert fragments are frequent; a coarse parallel lamination is frequently visible and together with variations in the matrix/pebble ratio, it gives the conglomerate a bedded appearance (Photographs 10, 11). The conglomerate is very similar to the Cuvo Conglomerates, and the interpretation of the depositional environment is the same. Almost at the top of the member we find a transition sediment of conglomerate to olistostrome. This high-density current-slide deposit is bounded at the top by an intraformational unconformity (Photograph 12). To the E and W the conglomerate passes into the Enterrias Olistostrome in which Barcena Conglomerate bodies occur as olistoliths.

Fossils from the limestone pebbles, samples 71Bar, 7198 and 71101 include foraminifers, determined as belonging to the basal part of subzone A of the *Profusulinella* Zone. This means a maximum age of Upper Bashkirian. Foraminifers of sample 71101 are almost identical to those of sample P 54 (van Ginkel, 1965, appendix 3, map 3) from the Santa Maria Formation (van Ginkel, pers. comm.). The Barcena Conglomerate together with the Enterrias olistostrome complex is correlated with the Cuvo Conglomerate and its supposed stratigraphic equivalents.

2) The Enterrias Olistostrome Member; the Llaves Limestone Bed (Encl. III, columns X and XI). The Enterrias Olistostrome Member consists of a slump-distorted shale-graywacke matrix in which numerous olistoliths occur. Olistoliths vary in size from a big boulder to a block of mappable proportions. The olistostrome extends along both flanks of the Mid-Liébana ridge to Pesaguero in the east. Most of the olistoliths are remnants of Murcia Quartzite, but various Devonian and Carboniferous limestone olistoliths are also present. The olistoliths are as far as possible represented on a 1:25 000 map (Enclosure IV). Most of the Devonian limestone olistoliths were mentioned in the description of the Devonian formations. Large olistoliths of Carboniferous limestone were only encountered along the southern slope of the ridge; these occurrences have been named the Llaves Limestone Bed. They are probably the remnants of a once continuous but restricted biogenetic bank deposit bordering the ridge. They consist of algal boundstones and bioclastic wackestones with a subordinate amount of calciruditic packstones. In thin section, biomicrites, biolithites and intra-micrites were recognized, although the limestone is partly recrystallized. In their resedimented position the limestone bodies may be accompanied and partly enveloped by limestone conglomerate lenses (Photograph 13). Probably, slabs of the outer flank of the biogenetic bank slid down together with parts of the coarse detrital talus.

Fossils are usually fragmented; algae, brachiopods, bryozoans, corals, crinoids, and foraminifers could be recognized. The samples 7060 and 70163 rendered determinable foraminifers, belonging to the lower part of subzone A of the *Profusulinella* Zone. This indicates an Upper Bashkirian age for the Llaves Bed, before the resedimentation. Since fossil evidence from different localities of the Barcena Conglomerate points to the same age, and no younger determinations are known from the entire olistostrome complex, I assume that this maximum age of Upper Bashkirian is also approximately the age of deposition of the Enterrias olistostrome complex. The Llaves Bed compares closely with the Sta. Maria Limestone Formation in the Barruelo area (Wagner & Wagner-Gentis, 1963; Wagner, 1972). However, the latter is not resedimented, and has an unconformable contact with the Villabellaco Formation. The Sta. Maria Limestone Formation should be correlated with the upper part of the Caliza de Montaña Formation (Wagner, pers. comm.).

3) The Lon-Olistostrome Member (Encl. III, column II). North and south of the village Lon, slabs of Caliza de Montaña Limestone occur together with conglomerate lenses, in isolated positions in the basal part of the Cervera Formation. The conglomerate and limestone lithosomes probably constitute one level, owing their present complex exposure pattern to intense tectonic deformation. Due to a lack of sufficient exposures, the olistostrome relation between the lithosomes and the sandstone-shale sediments could not be established beyond doubt. However, the irregular occurrence of the lithosomes, and the fact that they are surrounded by more or less distorted sandstone-shale sediments make an olistostrome genesis acceptable. The conglomerate lenses are lithologically and sedimentologically very similar to the Cuvo Conglomerates. The limestone lithosomes are described by Lanting (1966) as finely laminated, quartz-bearing calcilutites; in the laminae the quartz grains have a fining-upward gradation. Freshly cut, the limestone has a fetid odour. At the base of a limestone olistolith south of Lon, a 1 m thick level of a red limestone breccia, resting on a black cherty shale occurs. This is probably a remnant of the Alba and Vegamián Formations. Together with the Carboniferous limestones, Devonian limestones occur in the Lon Olistostrome. Samples 70C290 and 70C290 II yielded conodonts indicating respectively a Viséan and a Famennian age.

The simultaneous occurrence of remnants of the Vidrieros, Vegamián, Alba and Caliza de Montaña is hard to explain. The most probable source area for the Vegamián-Alba-Caliza de Montaña olistolith, would be a transition zone between the Picos de Europa area and the Liébana basin, which would now be covered, due to southward tectonic transport of the Picos de Europa limestone complex. The Vidrieros remnants could also be derived from this hypothetic transition area, but might as well originate from the Mid-Liébana ridge. Anyhow, these occurrences stress the fact that an unconformable relation of the Cervera with the Caliza de Montaña must also have existed in this area (cf. the Cervera Formation in the Ledantes area).

There is no field evidence for a direct connection of the Lon Olistostrome with the Enterrias Olistostrome. In view of its stratigraphic position a slightly younger age may be assumed for the former.

4) The Tornes Conglomerate Member (Encl. III, column III). The Tornes Conglomerate is best exposed in the NW slope of the Tornes valley NE of Buyezo (E Liébana). This conglomerate consists of rounded quartzitic and limestone pebbles in a matrix of coarse graywacke. Laterally, the conglomerate passes into a graywacke-grit association. The grits are breccious, consisting of angular clasts of limestone, black chert, and quartzite in a graywacke matrix which has also a high content of limestone grains (Miedema, 1966a). The clasts are notably uniform in size, the average diameter is about 3 mm. The grits are bedded, which bedding is marked by a vertical variation in clast/matrix ratio. They normally show a fining-upwards gradation, although coarsening-upwards gradations are also present. The bottom of such a grit unit is sometimes marked by erosive, cut-and-fill structures. The grits pass laterally into a coarse graywacke-shale turbidite association. Consequently, grits and conglomerates are supposed to have been transported by agents such as turbidity currents, high-density currents and grainflow layers; they may be regarded as fluxo-turbidites (cf. the petromict conglomerates described by Hubert, Lajoie & Léonard, 1970). This association of conglomerate lenses, gritbeds and immature turbidites forms an almost continuous horizon throughout the basal part of the Potes turbidites in N Liébana and Polaciones (Encl. III, columns II, III and IV). Limestone pebbles from a conglomerate north of Belmonte (Polaciones) rendered fusulinids and algae (sample 66G6), probably belonging to the upper part of the subzone A of the *Profusulinella* Zone, indicating a maximum age of Upper Bashkirian.

Grits as described above may occur throughout the Potes turbidite sequence, notably in Polaciones and E of Polaciones. Wherever possible, grits and conglomerates have been mapped. They are shown with the same signature although the grits differ strikingly from the well-rounded, quartzite pebble bearing conglomerates (pebbly sandstones and pebbly mudstones).

5) The Salceda Olistostrome Member (Encl. III, columns III and IV). Lithologically this member is characterized by an enormous mass of slump-distorted sediment which originally consisted of thin graywacke beds and shale. The shale content is markedly higher than in the basal part of the Potes turbidites, ranging from 30 to 60 %. The graywackes are usually thinly laminated to very thin bedded, with current ripple lamination in the beds, whereas the laminae have occasionally a fine parallel lamination and fine upwards, sometimes very gradually, into the shales. These 'mature turbidites' are found at the top of the Cervera Formation throughout Polaciones and N Liébana. Near Camaleño and Tanarrio, E of Cabezon de Liébana, and in many other places, the basal part of the overlying Buyon turbidites consists of similar sediments, so that whenever the Cabezon Conglomerate is lacking, and the upper part of the Cervera is not deformed by large-scale slumping, it is impossible to establish an upper boundary to the Cervera Formation. In the slump-deformed turbidites, slabs and boulders of quartzite conglomerates, mixed limestone-quartzite conglomerates, limestone breccias and bioclastic limestones occur. The degree of exposure is generally bad but the characteristics of the lithosomes as resedimented olistoliths are clear in the valley east of Salceda (Polaciones).

The allochthonous limestones are poor in fossil content. Out of many samples only 66H10 (different limestone clasts from the valley east of Salceda) rendered some determinable algae and fusulinids, belonging probably to the *Profusulinella* Zone, subzone B, possibly somewhat younger. At the E border of Polaciones, a limestone breccia with a grit-top, belonging to the upper part of the Cervera and situated below the Salceda Olistostrome level, yielded fusulinids and algae (sample 66H3) belonging to the subzone B of the *Profusulinella* Zone. Since the fusulinids occurred in the matrix, this indicates a *real* age of Lower Moscovian, probably Vereyan (van Ginkel, pers. comm.). Sample 66G16 from the lower part of the Buyon turbidites in S Polaciones, also yielded fusulinids belonging without doubt to the upper part of the subzone B of the *Profusulinella* Zone. In view of these facts we may safely assume a Lower Moscovian age for the Salceda Olistostrome and for the unconformity level at the top of the Potes turbidites in this area. This unconformity is tentatively correlated with the unconformity at the top of the Piedrasluengas Limestone Member.

In the next chapter a regional correlation for the Cervera and other Upper Carboniferous units, distinguished in the subject area, with stratigraphically equivalent units of other parts of the S Cantabrian Mountains will be attempted.

Curavacas, Lechada and Vañes Formations. – Kanis (1956) introduced the 'Curavacas formation', an informal rock unit (Code of Stratigraphic Nomenclature, art. 3 and 13) comprising the mainly terrestrial conglomerate-sandstone sequence of Mt. Los Cintos (N Palencia), which rests unconformably on folded older rocks. The name is derived from the Mt. Curavacas (W of Mt. Los Cintos) where a supposed identical conglomerate sequence was observed. Koopmans (1962) refers to the unconformity at the base of the Curavacas Conglomerate as the lower boundary of the 'post-Sudetic Yuso group', stating moreover that this unconformity is not always marked by a conglomerate development. Brouwer and van Ginkel (1964) formally introduced the Curavacas Conglomerate Formation, comprising the occurrences mentioned by Kanis and Koopmans but also the alternating conglomerate and sandstone-shale sequence that rests unconformably on the Piedrasluengas Member. The Curavacas Formation as defined by Brouwer & van Ginkel (1964) and notably van Ginkel (1965) has no clearly defined upper boundary; in addition, by ignoring the most striking lithologic contrast, it loses the best mappable feature in the terrain. This feature was employed by van Veen (1965), who mapped the conglomerate lithotope as the Curavacas Conglomerate Formation, with a Lechada Formation on top of, and laterally interfingering with, the former. Formally, in such cases the formations should be laterally bounded by an arbitrary cut-off, and distinction should be made between a Curavacas Formation and a Lechada Formation, Curavacas beds in the Lechada Formation, and Lechada beds in the Curavacas Formation. In his description of a section through the Curavacas north of Cardaño de Arriba, van Veen actually mentions the occurrence of Lechada beds between two conglomerate horizons (1965, pp. 168, 169, encl. 2), but he does not make a consequent use of such subdivisions. De Sitter & Boschma (1966), Savage (1967) and Boschma & van Staalduinen (1968) all use the Curavacas Formation sensu van Veen. Although Boschma & van Staalduinen proposed the above-mentioned section (van Veen, 1965, encl. 2, column II) as the type section of the Curavacas Formation, the problem of the interfingering Lechada was not entered into. It is of interest that they indicate possibly equivalent conglomerate occurrences, not only in the Pisuerga area but also in Liébana, as Curavacas Conglomerate Beds. Definitive solution of these classification problems must await the publication of the results of the investigations on the Curavacas Formation carried out by Mr. van Hoeflaken (in prep.). In this paper the Curavacas Formation will be used as proposed by van Veen.

The already mentioned Lechada Formation, named after its occurrence in the Rio Lechada (N León), consists of a turbidite sequence which interfingers laterally with the conglomerates and overlies them conformably. This formation was proposed by van Veen (1965), followed by Savage (1967) and Boschma & van Staalduinen (1968); the latter authors proposed the section described by van Veen as type section (van Veen, 1965, encl. 2, column II). In this section, however, no upper boundary of the Lechada is present.

In the northern Pisuerga area, the Curavacas Conglomerates alternate with, and are apparently conformably overlain by, a sandstone-shale sequence with a subordinate, but important, proportion of limestones; this sequence is generally described as, at least partly, paralic and autochthonous. Boschma & van Staalduinen refer to this occurrence as a part of the Vañes Formation. This formation was formally introduced for a stratigraphically equivalent sequence near the village Vañes by Brouwer and van Ginkel (1964), and originally proposed by Nederlof and de Sitter (1957) as 'série de Vañes'. The formation has been redefined in its type area by van de Graaff (1971), who also studied its occurrence in the Casavegas syncline. According to that author, the lower member of the Vañes Formation in its type area is a lateral equivalent of the Curavacas beds in the Casavegas syncline (van de Graaff, 1971, pp. 194, 210). For reasons that will be explained below, we prefer to include in the Vañes Formation the non-conglomerate rocks of the Palanca syncline and of the syncline between Barrio and Dobres (Encl. I and Fig. 2). These rocks were considered to belong to the Lechada Formation by Boschma & van Staalduinen (1968).

The complex relations of the Curavacas, Lechada and Vañes Formations, illustrated in the lithostratigraphic columns, force their description to be given with reference to a number of subareas, where the formations are present in a number of structures as shown on the Tectonic index map (Fig. 2). The following subareas are distinguished:

1. Southwest Liébana and the northern Rio Yuso area. Here, the Curavacas and Lechada occur in a faulted syncline-anticline-syncline-anticline-syncline structural succession.
2. Northern Liébana. Between the S border of the Picos de Europa and the San Carlos fault, a seemingly continuous succession of slump and olistostrome deposits are considered partly equivalent to the Curavacas and Lechada Formations.
3. Central Liébana and Polaciones. The Curavacas and Lechada are present in a complex synclinal structure, of which the S flank is partly cut out by the Redondo fault.
4. Northern Pisuerga and the area between Caloca and Barrio. This, as already mentioned, is the area where the Curavacas and Vañes Formations are distinguished.

The Curavacas and Lechada Formations in SW Liébana and the N Rio Yuso area (Encl. III, columns VII, VIII and IX, Encl. VI, column VI): The Curavacas in this area consists of two conglomerate horizons. The basal and thickest conglomerate horizon rests with an angular unconformity on the Cervera of the Ledantes anticline (column VII), and N of the Orcadas fault (Fig. 2) with a parallel unconformity on the Upper Deva turbidites, until it wedges out to the west (column V, VI, VII and IX). South of the Ledantes anticline, another important fault, the Naranco fault (Savage, 1967, p. 203), separates the south flank of the Ledantes anticline from the Lechada and Curavacas synclines. In the Lechada syncline the conglomerate rests with a locally well-exposed angular unconformity on the Cervera. The conglomerates of the Curavacas syncline rest unconformably on

Devonian and older Palaeozoic strata, which are exposed as a narrow ridge-like feature, cut off by the E–W Rio Frio fault, north of which the Cervera is exposed again. North of these exposures, tilted almost vertical and cut off by the Naranco fault, younging to the north, the basal Curavacas Conglomerate reappears. On the Pico Cohorra, the conglomerate is partly replaced by the Cohorra Limestone Member (column VII); more to the W several limestone bodies are present in the subsequent Lechada turbidites. Higher in the sequence, on the southeastern spur of Pico Cohorra and on Pico Zamburria, a second conglomerate horizon is present, with isolated limestone bodies on top and at its base (column VII). This conglomerate forms an open syncline on Pico Cohorra and a gently south-dipping monocline on Pico Zamburria. The limestones are named the Frio Limestone Member. In a restricted area south of Cucayo, an equivalent conglomerate-limestone occurrence seems almost entirely bounded by faults; only in one place the conglomerate rests unconformably upon the lower Palaeozoic rocks of the Polentinos area. The limestones of this occurrence are named the Ranas Limestone Member.

The Lechada Formation is present as a turbidite sequence between, and on top of, the conglomerates. In the Coriscao syncline the turbidites are laterally replaced by a huge slid mass, the Coriscao Olistostrome Member. This olistostrome wedges out in the N flank of the Coriscao syncline (columns VIII and XI). In this syncline the Lechada is conformably overlain by the Panda Limestone Member, the basal member of the Pandatrave Formation.

The conglomerates are ill-sorted, polymict and laterally very variable in thickness. Pebble content consists notably of well-rounded quartzitic pebbles, but also angular limestone clasts and even slabs of graywacke and shale occur. In the conglomerates, and laterally between them, we find lenses of coarse cross-bedded quartzitic sandstone (Peña Orcadas), coarse cross-bedded quartzitic grits (SW of the Ranas Limestones), a probably autochthonous biogenetic bank deposit (the Cohorra Limestone), a mud with scattered well-rounded quartzitic pebbles (E of the Cohorra Limestone), and also big olistoliths of Devonian limestone (N of the Rio Frio fault). The whole conglomerate is vertically and in places laterally in contact with the Lechada turbidites. In short, two depositional environments are represented. Part of the conglomerates was obviously resedimented in a marine turbidite environment. Mass transport by high-density currents, slumping and sliding seem the most appropriate mechanisms. On the other hand shallow marine to terrestrial autochthonous sedimentation is represented. In most cases the two environments are difficult to distinguish. There is not necessarily much textural difference between an autochthonous torrential conglomerate deposit and a conglomerate that is a product of resedimentation by mass transport. In view of the proximity of the two kinds of sediment, we assume that resedimentation occurred in a relatively shallow environment.

The Lechada Formation in this area consists mainly of turbidites which have been frequently affected by slumping. The isolated limestone bodies found in this environment, are commonly associated with slump distortion in the surrounding siliciclastics, but resedimentation could not be ascertained for each separate occurrence. The Coriscao Olistostrome Member is the best exposed example in the subject area of large-scale resedimentation by processes of slumping and sliding. Thickness varies from 0–600 m. The size of the – predominantly limestone – clasts varies from boulders to mappable lithosomes. Slump distortion of the siliciclastic 'matrix' is conspicuous. The vertical passage from the olistostrome into the Panda Limestone, an autochthonous bioclastic deposit, is quite rapid but some meters of alternating limestone and shale always precede the massive limestone (Photograph 14).

In the E flank of the Remoña syncline (see Fig. 2), both the Curavacas Conglomerate and the Coriscao have disappeared (column IX). A turbidite sequence reaches from the top of the Cuvo Member (Upper Deva turbidites) to the base of the Panda Limestone. The vertical passage from the turbidites to the massive limestone consists of some 10 m of thick- to very thick-bedded limestone grits in shales. The gritbeds (Photographs 15, 16 and 17) show more or less the Bouma sequence of sedimentary structures as corrected by Allen (1970). In the basal part of a bed, angular limestone fragments of pebble size may be present (Photograph 17). In the finely laminated silt at the top of a bed, *Zoöphycos* 'Spreitenbauten' (Photograph 18) were frequently observed. The presence of coarse clastic grit seems to be incongruous with the *Zoöphycos* burrows, that indicate relatively quiet conditions at shallow to intermediate depths (Seilacher, 1967, as cited by van de Graaff, 1971, p. 179). The grits are considered to be resedimented in a relatively shallow environment; except for the burrows at the top and their place in the stratigraphic succession, they compare well with the grits described from the Potes turbidites. The source area of the grits may have been the talus of a biogenetic bank, possibly the Panda Limestone. Such a configuration would result in a diachronic regressive sequence from turbidites into massive limestone. The sequence is best studied WSW of Pido, at the base of the limestone.

Considering the regional development of the Lechada Formation, there is a striking decrease in thickness from S (Lechada syncline, thickness more than 2000 m; Savage, 1967) to N (northflank of the Coriscao syncline, thickness less than 1000 m).

Most of the limestone lithosomes are bioclastic limestones, in which biogenetic bank deposits predominate (van de Graaff, 1972); such limestones normally abound in fossils. Together with the normal fauna (van de Graaff, 1972), determinable fusulinids and algae have been obtained from various localities. The Cohorra Limestone and two limestone lithosomes just above the basal horizon of the Curavacas Conglomerate more to the west, yielded fusulinids and algae belonging to the upperpart of subzone B of the *Profusulinella* Zone (samples Ge 124, 70455, S 64496, S 131), indicating a Lower Moscovian, upper Vereyan to lower Kashirian, age. The Coriscao Olistostrome yielded fusulinids and algae, some of which belong to subzone A and others to subdivision B 1 of the *Fusulinella* Zone (samples Ge 100, 70515 and 7156). Probably allochthonous fossil plants (sample F 6) point to Upper Westfalian or Lower Stephanian (Stockmans cited in van Ginkel, 1965, see Appendix). Since the overlying Panda Limestone is known to belong also to the *Fusulinella* Zone, subzone B subdivision B 1 (van Ginkel, 1965), the relative age of the Coriscao Olistostrome is near the base of the Upper Moscovian. The Ranas limestone lithosomes (samples 70290, 70481 and 70483) yielded the same mixed occurrence of fusulinids as was encountered in the Coriscao Olistostrome. Of the Frio Limestone, two samples yielded fusulinids belonging to the *Fusulinella* Zone subdivision B 1. Sample 70471 is situated stratigraphically above the second conglomerate horizon, sample X 3 is situated below this horizon. Another sample (S 64493, W of Pico Zamburria) yielded algae pointing also to an age near the base of the Upper Moscovian.

In short we can distinguish a lower Curavacas conglomerate horizon, with an unconformable base, having a Lower Moscovian, upper Vereyan to lower Kashirian, age, and a second horizon consisting of conglomerate occurrences, slid limestone lithosomes and olistostrome deposits, which is situated at the base of the Upper Moscovian. The unconformable nature of this second horizon is often dubious and can only be ascertained in the area of the Ranas Limestones. A regional correlation for these two levels is attempted below.

The Curavacas and Lechada equivalents in northern Liébana (Encl. VI, column II): In the area between the thrusted Picos de Europa complex in the north and the San Carlos fault in the south, a sequence of mainly slump-distorted thin graywacke beds and shales is present. Neither the top nor the base of this generally south-younging sequence is exposed, due to the fault boundaries and the unconformable contact with the Permian Labra Formation in the east. This sequence was named the Bedoya Olistostrome Member, after the Rio Bedoya. In places where the original sedimentary texture is visible despite slump distortion, the sediment is remarkably similar to the 'mature turbidites' at the top of the Potes turbidites. In the oldest part of the sequence, lenses of polymict conglomerates, and mudstones with scattered well-rounded quartzitic pebbles are frequent (Photograph 19); higher in the succession, limestone olistoliths occur.

South of the Bedoya river the sediment is much less deformed by slumping, the turbidite beds are coarser and thicker; this sequence can be followed laterally along the southern slope of the Sierra Sagra until it ends against the San Carlos fault, probably reappearing in the core of an anticlinal structure in the Triassic, NW of the village San Mames in Polaciones. In this badly exposed occurrence, slump distortion is again present. This occurrence is not considered as a part of the Bedoya Olistostrome. We refer to it as the San Mames unit.

Limestone pebbles from a conglomerate lens in the basal part of the Bedoya Member, yielded fusulinids, belonging to the *Fusulinella* Zone, some to the subzone A and others probably to subdivision B 1 (sample 71247). East of locality 71247 at the margin of the Picos de Europa, some Caliza de Montaña remnants occur as olistoliths in the mud. Limestone olistoliths from a stratigraphically higher level yielded different results. Two samples taken from adjacent limestone clasts yielded fusulinids of the *Fusulinella* Zone subdivision B 1 (sample 7178A) and subdivision B 2 (sample 7178). This indicates a probable maximum age of Upper Moscovian (Upper Podolskian to Lower Miachkovian) for that part of the olistostrome.

These data are not equivocal but some attempt at correlation can be made. The oldest part of the olistostrome is correlated with the second conglomerate-olistostrome horizon from SW Liébana. It is probable that several olistostrome and slump units are lying on top of each other. One of the following olistostromes has a maximum age of Upper Podolskian. The olistostrome deposition is followed by a relatively undisturbed turbidite sedimentation. Exposure is such that nothing can be said about the nature of the basal contacts of the olistostrome units. Unconformities might be present. The occurrences of such olistostromes just south of the Picos de Europa margin may point to unstable conditions in a transition area between the flysch basin of Liébana and the Picos de Europa shelf area.

Occurrence of the Curavacas and Lechada Formations in central Liébana and Polaciones (Encl. III, columns II, III, IV and X): First, distribution, boundaries, lithology and depositional environment of the Curavacas Formation are described.

A. Curavacas Formation: In the Curavacas Formation again two conglomerate horizons can be distinguished, a basal horizon, the Cabezon Conglomerate Member, and a second younger horizon comprising the Viorna and Porrera Conglomerate Members.

The Cabezon Conglomerate Member can be followed on the map along the base of the Yuso Group from Polaciones, SE of Tresabuelas, through Liébana to the N flank of Pico Viorna. The member was named after its occurrence in Cabezon de Liébana at the Potes–Piedrasluengas road. It consists of several laterally separated conglomerate lenses E of Cabezon, and a more continuous conglomerate horizon W of Cabezon: this horizon wedges out, interfingering with coarse proximal turbidites N of Pico Viorna. No corresponding conglomerate or unconformity level could be distinguished in the S flank of the Viorna syncline (Fig. 2). The character of the basal contact of this member with older sediments changes markedly from E to W (columns IV, III and II). An angular unconformity can be inferred at the base of the Cabezon in the Collario river (Polaciones SE of

Tresabuelas). The steeply SW dipping Cabezon rests on overturned and refolded strata of the Potes turbidites which form part of a big recumbent structure. The Cabezon is succeeded conformably by a sequence of shales with thin but continuous graywacke laminae, not unlike the mature turbidites at the top of the Potes turbidites. An equivalent unconformity could be observed along the main road through Polaciones, but here the conglomerate is placed between two olistostrome masses in which intraformational unconformities may be common. The unconformable nature of the contact between the Potes turbidites and the Cabezon Conglomerate becomes questionable to the W. No unconformity was observed in Cabezon de Liébana. North of Valmeo in the Quiviesa valley the conglomerate has a basal development of coarse sandstone and grit which has to all appearance a conformable sedimentary contact with the top of the Potes turbidites.

On the Pico Viorna, a much thicker massive conglomerate, the Viorna Conglomerate Member, rests, dipping gently to S, with a distinct angular unconformity on older Lechada turbidites, also cutting off the steeply overturned Cabezon Conglomerate lenses. To the SE the Viorna wedges out but reappears again in the Quiviesa valley, S of Valmeo, much thinner but still with an angular unconformity at the base. This conglomerate lens also wedges out to the east. Conglomerates reappear, more or less at the same stratigraphic level, in, and E of the Buyon valley. These conglomerates are named the Porrera Conglomerate Member after the occurrence on the Peña Porrera. The member consists of a sequence of vertically and sometimes laterally separated lenses, interfingering with turbidites. No corresponding unconformity could be traced.

The conglomerates in this area are polymict, consisting of rounded quartzitic pebbles, an equal amount of mostly angular to subangular limestone pebbles, frequently occurring angular clasts of black chert (mostly in the grit fraction), and a matrix which normally consists of a microbreccia to microconglomerate of the same constituents, though a mud matrix may also occur. Pebble size ranges from grit to boulders. The conglomerate occurrences are not homogeneous; predominantly limestone conglomerates may occur and almost purely quartzitic conglomerate lenses also occur. Frequently, the conglomerates are interbedded with grit breccias and coarse sandstones (Photograph 20). The pebbles may touch each other but just as often they float separately in the matrix. Slices and fragments of graywacke-shale-turbidite alternations may occur as clasts floating in the matrix (Photographs 21 and 22).

All those features are best explained as produced by high-density currents, mudflows, grainflow layers and similar resedimentation mechanisms. No positive indications exist for a fluviatile or littoral origin of these conglomerates. It is impossible to make a lithologic distinction between the Barcena, Cabezon, Viorna and Porrera Conglomerates.

B. Lechada Formation: The Lechada Formation in this area consists predominantly of turbidites, informally named the Buyon turbidites (the Buyon Beds of Boschma & van Staalduinen, 1968). Olistostrome deposits occur in several places. In the turbidite sequences, lenses of homogeneous cross-bedded quartzitic sandstones may occur. Limestone lithosomes are present in the big olistostrome complexes but also in isolated occurrences though frequently associated with a slump facies.

1) The Cabezuela Olistostrome Member (Encl. III, column IV). An olistostrome deposit, having a considerable thickness, is exposed along the main road in Polaciones extending from the Cabezon Conglomerate exposure, southwards to the Mirador de la Cruz de Cabezuela. Its lateral extent is, due to scarce outcrops in this area, uncertain. The Cabezuela Olistostrome consists mainly of slumped turbidites in which lenses of pebbly mudstone and limestone grits occur. Some slump occurrences SE of Cabezon de Liébana may be correlated with this olistostrome.

2) The Sta. Eulalia Olistostrome Member (Encl. III, column X). This olistostrome consists of slump-distorted graywacke and shale in which huge olistoliths occur: remnants of the Murcia and a subordinate amount of Devonian and Lower Carboniferous limestone olistoliths (see Encl. IV). This olistostrome is notably exposed around Toranzo and in the N slope of the Sta. Eulalia brook and seems to replace the Viorna Conglomerate which wedges out E of Picos Jano. The basal boundary of this olistostrome could not be established; in various places there seems to be a gradual passage from the Enterrias Olistostrome into the Sta. Eulalia Olistostrome.

3) The Campollo Olistostrome Member (Encl. III, column X). West of Campollo in the core of the Viorna syncline, another olistostrome rests with a slump contact on generally undisturbed mature Buyon turbidites. This olistostrome is referred to as the Campollo Olistostrome Member. The olistostrome mass as a body is dipping gently to the N. Though its basal contact with the Buyon turbidites west of Campollo seems parallel, its relation to the overturned recumbent S flank of the Viorna syncline on Pico Jano is such that an angular unconformity at its base has to be inferred. The Campollo Olistostrome is made up of more or less the same constituents as the Sta. Eulalia Olistostrome, but the former contains patches of Carboniferous conglomerate (probably Viorna remnants) which occur as olistoliths in the slump. The degree of exposure is generally bad, but at the northern extremity of the member, west of Campollo, the olistostrome nature can be established without doubt. The Sta. Eulalia and Campollo Olistostromes are supposed to have been generated by uplift and subsequent 'gravity-slide-denudation' of the Mid-Liébana ridge, as was presumably the case with the Enterrias Olistostrome. The Campollo Olistostrome was formerly interpreted as a 'tectonic klippe' (Lanting, 1966; Boschma, 1968).

4) The Barreda Olistostrome Member (Encl. III, column III). A sequence of turbidites, mudflow deposits and slumps is developed in E Liébana on top of the Porrera Conglomerates. This succession, its constituent members with their lateral and vertical relations, was mapped and described in detail by Miedema (1966, 1966a). The most conspicuous and probably youngest unit in this succession is the Barreda Olistostrome Member, exposed around and between the villages Barreda and Pesaguero. This olistostrome consists of the normal slump matrix, with olistoliths predominantly of conglomerate and limestone. Although occurrences of shale clasts yielding conodonts indicating Middle and Upper Famennian ages were reported by Miedema (1966), slumped older Carboniferous rocks are the most important constituents of this olistostrome.

5) The Buyon turbidites in Polaciones. The Buyon turbidites are generally to a lesser extent affected by slumping in Polaciones than in central Liébana. Some quartzitic sandstone lenses occur in this sequence, between Tresabuelas and the Peña Labra. These lenses are marked by cross-bedding on a scale of tens of centimeters, the absence of shale, and the presence of well-preserved plant fossils. They are, as far as could be observed, laterally and vertically in normal sedimentary contact with turbidites. The only reasonable explanation seems to be a proximal turbidite with a dune development (Allen, 1970). The Lechada in this area generally youngs to the S and is probably conformably overlain by the Agujas Limestone, the basal member of the Corisa Formation in the Redondo syncline (Fig. 2).

The only fossiliferous rocks are the allochthonous limestone olistoliths and pebbles. Two samples from the Cabezon Conglomerate (70291, L 1) yielded fusulinids belonging to the upper part of subzone B of the *Profusulinella* Zone. Samples 66H10 and 66G16, under and above the Cabezon Conglomerate in Polaciones, belong also to the *Profusulinella* Zone subzone B. Sample 66H5, from a higher level in the Lechada, rendered fusulinids and algae belonging to *Fusulinella* Zone subzone A. An erosion block from the Viorna Conglomerate (sample 71271) rendered fusulinids belonging either to the top of subzone A or the base of subdivision B 1 of the *Fusulinella* Zone. A limestone breccia lens near Soberado (S Liébana) which rests with an unconformable bottom-to-bottom contact on the Upper Deva turbidites, yielded fusulinids belonging to subdivision B 1 of subzone B of the *Fusulinella* Zone.

In view of such determinations the following ages may be inferred. The Cabezon Conglomerate has a maximum and probably real age of Lower Moscovian, upper Vereyan to lower Kashirian. The Viorna Conglomerate should have as maximum age Upper Moscovian, probably lower Podolskian. The limestone olistoliths from the Sta. Eulalia and Campollo Olistostromes, yielded conodonts of Frasnian, Famennian, Viséan and Lower Namurian ages (listed in the Appendix, localities on Encl. IV). Concerning the age of the olistostromes we can only state that the Sta. Eulalia is of about the same age as the Viorna Conglomerate and that the Campollo must be younger.

The Cabezon Conglomerate is correlated with the basal conglomerate horizon of the Curavacas Formation in SW Liébana. The Viorna Conglomerate is correlated with the younger conglomerate olistostrome horizon in SW Liébana. It is remarkable that in central Liébana the second conglomerate is the better developed and has a conspicuous angular unconformity at its base.

The Curavacas and Vañes Formations in the Casavegas syncline and in the synclinal structures between Caloca and Barrio (Encl. III, columns V and VI): The Curavacas in this area consists of three conglomerate horizons, all having a lensing character. The first horizon has been mentioned as the lowermost tongue of the Curavacas, which rests, generally with a parallel unconformity, on top of the Piedrasluengas Limestone. The conglomerate lenses of this horizon are usually thin, passing laterally into graywackes and shales. The conglomerate level strikes SE to W, parallel to the Piedrasluengas until the latter ends W of the Peña Sopeica. Then it follows the tight synclinal structure E of Barrio to reappear in the Palanca syncline, continuing to the E in the S flank of this structure until it is cut off unconformably by the second horizon (Encl. III, column VI). The second horizon is thicker and more continuous. A basal unconformity for this horizon could only be established in the S flank of the Palanca syncline (Encl. III, column VI). It strikes generally parallel to the first horizon; the horizons are separated by graywacke and shale sediments. The presence of this second horizon in the syncline E of Barrio could not be proved. Southeast of Caloca in the S flank of the Palanca syncline, the second horizon is probably cut off by the E boundary fault of the Polentinos block.

In crossing the Orcadas fault, the unconformity at the base of the Curavacas changes from near-parallel to angular; in crossing the probably faulted N boundary of the Polentinos block (the Rio Frio fault, van Veen, 1965), the stratigraphic lacuna increases considerably. Only in one place, in the S flank of the Palanca syncline, NE of Peña Bistruey, the Rio Frio fault is exposed, separating Piedrasluengas Limestone from Lebanza Limestone. There, the second Curavacas horizon rests with a parallel unconformity on the Piedrasluengas Limestone. To explain this parallel unconformity we assume that the Orcadas fault zone has disappeared or merged into the Rio Frio fault, somewhere below the Palanca syncline.

On top of the second Curavacas horizon, the Palanca and Albas Limestone Members of the Vañes Formation are developed in the Palanca and Casavegas synclines, respectively (Encl. III, columns VI and V). The limestones may rest immediately on the conglomerate, in some places, however, a 100–150 m shale intervenes. In the S flank of the Palanca syncline the limestone is fairly continuous, and to all appearance autochthonous. In the north flank of this syncline and in the Casavegas syncline, the limestone occurs in patches, accompanied by small conglomerate lenses, and sometimes slump distortion is visible in the adjacent graywackes and shales.

The third and last conglomerate horizon is present beyond doubt only in the Casavegas syncline. It may be represented in the Palanca syncline by small conglomerate lenses. This third conglomerate might be unconformable, and in that case it can be correlated with an unconformity assumed by van Ginkel (1959) at more or less the same level. The sequence is continued mainly by shales which represent the shale member of the Vañes Formation according to van de Graaff (1971). These shales are conformably overlain by the Camasobres Limestone Member of the Corisa Formation.

According to Breimer (1956) the conglomerates in this area are ill-sorted, structureless, containing almost exclusively well-rounded quartzitic pebbles in a matrix of quartzitic sandstone. Pebble diameters vary from 1 cm to 1 m, decreasing to the east. Breimer investigated this area from the Peña Brez to the east. From our observations in the area between Barrio and Dobres we can add that in the west there is a considerable increase in the proportion of limestone pebbles. East of Peña Brez, the Piedrasluengas Limestone is overlain, with a low angle unconformity, by the Curavacas Conglomerate (first horizon); part of the limestone must have been removed; however, no limestone fragments are present in the conglomerate. East of Barrio (Encl. III, column VII) the conglomerate contains erosion products of this limestone as pebbles and huge slabs which probably have slid down. The conglomerate is well developed where it is laterally adjacent to the limestones of Peña Sopeica, but thins rapidly above these limestones. It is assumed to be deposited at a greater depth than the conglomerates near Peña Brez.

The second conglomerate horizon contains a coal seam at its base SW of Peña Abismo, while at its top it passes upwards into a sequence of marine shales with brachiopods (Breimer, 1956), and the usual bioclastic limestones that consist for a large part of biogenetic bank deposits. In the N flank of the Palanca and Casavegas synclines, the patchy occurrence of the limestones together with the occurrence of slump structures and pebbly mudstone lenses in the adjacent siliciclastics, indicate the activity of resedimentation processes.

The third conglomerate is lithologically similar to the other horizons. The succeeding shale sequence, the shale member of the Vañes Formation, was interpreted by van de Graaff (1971) as a shelf-slope deposit, resting on top of shallow-water sediments and covered at its top in some places by lithic-arenitic turbidites. In general, the Curavacas and Vañes Formations in this area represent a shallow marine environment, in which depth increased with the lapse of time, and, geographically, from SE to NW.

The bioclastic lenses of the Albas and Palanca Members are usually rich in fossils. From several places fusulinids were obtained or were already known. Samples: P 3 (Albas Member), X 1, X 2, Ge 224, 70369, 70371 and 70389 (Palanca Member). All fusulinids belong to the *Profusulinella* Zone subzone B. The Albas and Palanca Members must be younger than the Piedrasluengas Member. Sample P 3 indicates a Lower Moscovian, lower to middle Kashirian age (van Ginkel, 1965).

For the first and second conglomerate horizons, lying between the Piedrasluengas and Palanca-Albas Members, a Lower Moscovian, lower Kashirian age is inferred. South of Cucayo between the first and second conglomerate horizons, a flora (F 5, Appendix), known as the Dobres flora, indicates an age of Upper Namurian-Westfalian A. The shale that contains this flora probably constitutes a large clast in the conglomerate. The age of the third conglomerate horizon is doubtful, the first succeeding limestone, the Camasobres Limestone Member, belongs to the *Fusulinella* Zone subdivision B 1, indicating a lower Podolskian age (Sample P 4, van Ginkel, 1965).

The first and second conglomerate horizons are correlated with the Cabezon Conglomerate and with the basal conglomerate level of the Curavacas Formation in SW Liébana. Correlation of the third conglomerate horizon is not attempted since the available data are insufficient.

Regional correlations of the Curavacas, Lechada and Vañes Formations: The above-described interval comprises the upper part of subzone B of the *Profusulinella* Zone, subzone A and locally the lower part of subdivision B 1 of the *Fusulinella* Zone, apart from the Campollo Olistostrome Member for which no age could be established, and the upper part of the Bedoya Member which is younger. In this interval, a basal unconformity, the Curavacas unconformity could be distinguished locally in SW Liébana, central Liébana, Polaciones and in the northern Pisuerga area. Existence of a second unconformity, the Viorna unconformity, could only be proved in central and SW Liébana. Both unconformities are only of local importance, changing from angular to parallel, and disappearing laterally.

Comparing this interval and its unconformities with the development in the S part of the old Palentine facies area, the following relations can be established. In the Rio Yuso area, the sequence is similar to SW Liébana, consisting mainly of conglomerates and turbidites. A basal Curavacas unconformity is developed. At the top of the conglomerates in the Curavacas and Lechada synclines, a level of partly slumped limestones and limestone conglomerates (the El Ves Limestone Member, van Veen, 1965) and pebble beds is present in a turbidite facies (Savage, 1967). The El Ves contains fusulinids both from subzone A and subdivision B 1 of the *Fusulinella* Zone (van Ginkel, pers. comm.); this level may be correlated with the Viorna unconformity level in Liébana. The lower Kashirian age established for the basal conglomerate horizons in the Casavegas and Palanca synclines, situated between two autochthonous limestones, the Piedrasluengas and the Palanca Members, agrees well with a Westfalian B age (cf. van Ginkel, 1965, pp. 208 and 209). Such an age was established for the Curavacas Conglomerate of Mt. Los Cintos (N of Cervera de Pisuerga) by Wagner (Wagner & Wagner-Gentis, 1963).

A Westfalian A age was inferred for the Curavacas unconformity in the type area of the Curavacas Conglomerate. This dating is based on determination of a flora (Stockmans, in Stockmans & Willière, 1965), found N of Cardaño de Arriba in a conglomerate which should be, according to van Veen (1965), the basal horizon of the Curavacas Conglomerate. In Stockmans and Willière (1965) this flora is mentioned, probably erroneously, as situated above the conglomerate ('au dessus du conglomérat'). Near this flora locality, the Curavacas Conglomerate overlies with a nearly parallel unconformity the older Triollo Conglomerate Member of the Cervera Formation (van Veen, 1965, encl. I). The distinction between Triollo and Curavacas, which may be lithologically identical, is in many places well-nigh impossible, and the conglomerate layers between which the plant fossils were found, might as well belong to the former as to the latter. The age determination favours a situation of the plants in the Triollo Conglomerate below the Curavacas unconformity. Such a situation would agree with the occurrence of Westfalian A plant fossils in a sandstone-shale sequence below the Curavacas Conglomerate N of Triollo (van Veen, 1965, p. 70).

So, a difference in age between the Curavacas Conglomerate on Mt. Los Cintos and on Mt. Curavacas has not been proved. It seems most likely that the latter is a lateral continuation of the former and has about the same age. A distinction between a Curavacas and a younger Los Cintos Conglomerate (Wagner, 1966) seems unnecessary.

Other Upper Carboniferous Formations, younger than those described above, are only locally present in the subject area. They were not studied in detail and will be treated summarily.

Pandatrave Formation (Encl. III and VI, columns VIII and IX). – This formation was informally introduced by Kamerling (1962), described in detail, with a section by Kutterink (1966), and formally proposed by Boschma & van Staalduinen (1968). The Pandatrave Formation comprises a sequence which rests conformably on the Lechada Formation in SW Liébana, comprising a basal bioclastic limestone member, the Panda Limestone Member, a turbidite member and a member of isolated, partly slumped, limestone lithosomes, the Brañas Limestone Member. The turbidite member consists of alternating siliciclastic and calciclastic complexes. The siliciclastics are the true turbidites, the calciclastic complexes consist for the major part of conglomerate and breccia beds, representing high-density current and grain flow layer transport. The limestone members consist of the known bioclastic sediments which are supposed to have originated as biogenetic bank deposits. The Panda Limestone is autochthonous, the Brañas Member is at least partly resedimented by slump-and-slide processes. The vertical passage from the top of the Panda Limestone into the turbidites is very similar to the passage from the Lechada turbidites into the base of the Panda Limestone (cf. Photograph 23). Only the lower part of the Pandatrave Formation is present in the subject area.

Fusulinids from the Panda Limestone (sample Ge 133, and also samples L 21, L 408, L 426 in van Ginkel, 1965) belong to the *Fusulinella* Zone subzone B subdivision B 1, indicating an Upper Moscovian, Podolskian age, probably middle to upper Podolskian (van Ginkel, 1965). Fusulinids from the Brañas Member indicate a younger Upper Moscovian; they belong to the *Fusulinella* Zone, subzone B, subdivision B 2 (van Ginkel, pers. comm.).

As defined above, the Pandatrave Formation has a redefined lower boundary: the base of the Panda Limestone, instead of the top, as defined by Boschma & van Staalduinen. The redefined lower boundary facilitates a correlation of the Pandatrave Formation with the Corisa Formation in the northern Pisuerga area.

Corisa Formation (Encl. III & VI, columns IV, V). – A lithostratigraphic unit containing the geographic name (Sierra) Corisa was employed by the following authors: Nederlof & de Sitter (1957): 'Série de Corisa', Brouwer & van Ginkel (1964): 'Formation de la Sierra Corisa', van Ginkel (1965): Corisa Formation, Boschma & van Staalduinen (1968): Corisa Formation. The authors disagree about the regional extent and boundaries of the unit. In this thesis the Corisa Formation will be used as defined by Boschma & van Staalduinen (1968) with acceptance of the boundary-redefinition by van de Graaff (1971, pp. 210–211). Thus, the Corisa Formation consists of the mainly shallow-marine sequence of siliciclastics and bioclastic limestones, resting conformably on the Vañes Formation in the Casavegas and Castilleria synclines, and on the Buyon turbidites of the Lechada Formation in the Redondo syncline. This sequence is bounded at the top by the Leonian unconformity (van de Graaff, 1971); whenever this level cannot be recognized in the field, the upper boundary of the youngest limestone member is considered as the formational boundary.

In the subject area we are only concerned with the development in the Casavegas and Redondo synclines. In both structures the Corisa Formation is tripartite, consisting of a basal limestone member, the Camasobres Limestone Member in the Casavegas syncline and the Agujas Limestone Member in the Redondo syncline; a siliciclastic member, generally representing shallow-marine sedimentation in the Casavegas syncline and turbidite sedimentation in the Redondo syncline (van de Graaff, 1971); and a top limestone member, the Maldrigo Limestone Member in the Casavegas syncline, the Abismo Limestone Member in the Redondo syncline. The limestones occur in lenses which may display a rather abrupt lateral discontinuity. They have originated as autochthonous biogenetic bank deposits; a part of the

Agujas Limestone Member is obviously resedimented (van de Graaff, 1971).

Fusulinid samples from the different limestone members (samples P 4, P 7, P 72, P 73, all from van Ginkel, 1965) yielded the following results. The Camasobres Member belongs to the *Fusulinella* Zone subdivision B 1 and has an Upper Moscovian, probably lower Podolskian age. The Agujas Member belongs to the same subdivision but has an Upper Moscovian, probably middle to upper Podolskian age (cf. the Panda Limestone). The Maldrigo Limestone and Abismo Limestone both belong to the *Fusulinella* Zone subzone B subdivision B 2; the age is Upper Moscovian, either top of Podolskian or base of Myachkovian (all determinations after van Ginkel, 1965).

The Corisa and Pandatrave Formations are biostratigraphically equivalent to most of the massive member of the Picos de Europa Formation, and to the upper part of the Aliva Formation. In Liébana, only the middle part of the Bedoya Olistostrome Member may represent a stratigraphic equivalent of the Corisa Formation.

Valdeon Formation (Encl. III, column XI). – The Valdeon Formation was introduced by Kutterink (1966) and formally proposed by Boschma & van Staalduinen (1968) for a sequence consisting mainly of turbidites and conglomerates, which rests with an angular unconformity on the folded and thrusted Devonian and Lower Carboniferous strata of the Montó structure in the Valdeon valley, W of the subject area. The Montó structure is thrusted to SE against the Brañas Member of the Pandatrave Formation. To NE the Lower Carboniferous and Devonian strata are covered by the Valdeon Formation. The base of this formation consists in that area of a slump interval which rests unconformably on the Pandatrave Formation. This slump interval is called the Remoña Olistostrome Member; it extends to ENE and is present in Liébana from the Puerto de Remoña to the Espinama–Aliva road. The olistostrome character is very clear in the exposures W of Fuente Dé. Several olistoliths of Devonian origin were found in the olistostrome; nodular limestones (samples 71C30 and 70C172), remnants of the Vidrieros and Cardaño, respectively and quartzitic blocks, presumably remnants of the Murcia. NE of Espinama, the Valdeon Formation has a breccious conglomerate at the base, consisting mainly of subangular limestone pebbles, the Sarques conglomerate. For the Valdeon Formation no upper boundary can be defined since it is cut off in Liébana and E Valdeon by the southern thrust fault of the Picos de Europa complex. A separate occurrence of a similar lithology in the Cares valley N of Cordiñanes, was interpreted by Kutterink as also belonging to the Valdeon Formation. This occurrence has the same relation to the Picos de Europa structures as the Lebeña Formation, resting unconformably on the first nappe and cut off by the thrust plane of the second nappe.

Only one flora has been found in the Valdeon Formation which indicated a Stephanian age (van Amerom, in Kutterink, 1966). From its relationship to other formations it is clear that the Valdeon Formation is younger than the Pandatrave Formation and must be equivalent with the Lebeña Formation.

San Mames unit (Encl. VI, column II). – The San Mames unit consists of a slumped graywacke-shale sequence, probably resting on top of the Bedoya Member, exposed in the core of the faulted anticline NW of San Mames in Polaciones. From this unit a probably allochthonous flora was obtained (F 9, Appendix) which indicated a probable age of lower Cantabrian (Wagner, written comm.). A limestone olistolith yielded fusulinids (sample 65B9) belonging to the *Fusulinella* Zone subdivision B 2/B 3. So, a maximum age of lower Cantabrian is inferred for the San Mames unit.

A very similar rock type is exposed N of Puente Pumar (Polaciones). A flora (F 4, Appendix), recovered from these rocks by Mr. Budding, indicated a Stephanian age (Wagner, pers. comm.). Since these 'Puente Pumar rocks' seem to be a lateral continuation of the San Mames unit, they are provisionally included in it.

'Barruelo Formation' (Encl. III and VI, columns IV and V). – In the cores of the Casavegas and Redondo synclines, younger sediments rest with a parallel unconformity on top of the Corisa Formation. In the Redondo syncline, a turbidite sequence with occasionally wildflysch deposits at its base, is succeeded conformably by a paralic to terrestrial sequence in which several coalbeds are present (Nederlof, 1959). In the Casavegas syncline only paralic to terrestrial sediments are present.

Van Ginkel (1965) distinguished a Caldero Formation, the turbidite sequence in the Redondo syncline, and a Barruelo Formation, the terrestrial sequences of both the Redondo and Casavegas synclines. De Sitter & Boschma (1966) and Boschma & van Staalduinen (1968) proposed the incorporation of all these post-Leonian deposits in the Barruelo Formation. Wagner (in Wagner & Winkler Prins, 1970), who originally defined the Barruelo Formation in the Barruelo subarea, disagreed with such an extension of the formation to stratigraphically non-equivalent sequences. The inclusion of the discussed strata in the Barruelo Formation is obviously incorrect; since I do not want to add to the existing confusion, I will mention them as 'Barruelo Formation'*.

The basal part of the 'Barruelo Formation' in the Casavegas syncline is marked by the occurrence of the Casavegas coalbeds, from which plant fossils yielded an Upper Westfalian D age (Wagner; see Appendix, sample F 1), and the Lores Limestone Member, bearing fusulinids of an Upper Moscovian, Myachkovian age (van

* Wagner & Varker (1972) proposed the following lithostratigraphic subdivision for the discussed strata: the terrestrial deposits in the Casavegas syncline constitute the Ojosa Formation, the turbiditic and terrestrial sequences in the Redondo syncline are assigned to the Brañosera and Barruelo Formations, respectively.

Ginkel, 1965, sample P 10). In the Redondo syncline, the basal part of the paralic to terrestrial sequence contains the Corros Limestone Member, and the Redondo coalbeds, which have yielded a fauna and flora with respective ages of Kasimovian (van Ginkel, 1965, sample P 52) and Stephanian A (Wagner, see Appendix, sample F 2, Stephanian A as redefined after acceptance of the Cantabrian stage).

The 'Barruelo', Valdeon and Lebaña Formations, together with the San Mames unit and possibly the Campollo Olistostrome Member of the Lechada Formation, all belong to uppermost Westfalian or Lower Stephanian, and have, as far as could be observed, an unconformable basal contact with older strata. An exact correlation of these units is, due to the lack of age determinations, not yet possible. But we know at least that the unconformities may differ in age, as well as the superposed strata. For instance the lacunas in the Picos de Europa area and the Casavegas syncline do not overlap (Encl. VI).

Cordel unit (Encl. VI, column IV). – The Cordel unit consists of a probably fluviatile sandstone-conglomerate association, exposed SE of Pico Cordel, bounded at its top with an angular unconformity by the basal Triassic conglomerate, and laterally in faulted contact with the Cervera Formation. This occurrence was first described by Wagner (1970), who mentioned a Stephanian C flora obtained from these rocks (F 3, Appendix).

The unit may be regarded as equivalent with the terrestrial Upper Stephanian B deposits of the Peña Cilda Formation in the Barruelo area (Wagner & Wagner-Gentis, 1963). The latter formation rests with a basal unconformity on the steeply overturned strata of the Barruelo Formation (Wagner & Winkler Prins, 1970).

STRATIGRAPHY OF PERMIAN AND MESOZOIC STRATA

Labra Formation (Maas, 1968 & 1968a; de Jong, 1971). – A sequence consisting of volcanic conglomerates and breccias, tuffs, reworked tuffs and tuffaceous sandstones and shales, is exposed along the south slope of the Peña Labra, in a steep valley NE of the Cueto Ropero, in the area E of the Puerto Sejos, in the faulted anticlinal structure of Puente Pumar and along the SW slope of the Sierra Sagra. For this sequence the Labra Formation is formally proposed. 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 (Photograph 24). A conformable upper boundary is not known from this area. The thickness of the formation ranges from 600 m (below the Peña Labra) to about 1000 m in the Sierra Sagra and east of Puente Pumar (Maas, 1968, 1968a).

The type section designated here is the best exposed section, on the SW slope of the Peña Labra (Encl. I, line a–a') (Photograph 25, Encl. VI). In this section four informal subdivisions are distinguished: a lahar member, a volcanic agglomerate member, a tuff member and red beds.

The lahar member starts the sequence with 75 m of chaotic sediment, large slabs and boulders of different lithology floating in a fine, black, tuffaceous matrix; synsedimentary contortions are frequent; limestone boulders, graywacke slabs and pyroclastic fragments occur together.

The volcanic agglomerate member unconformably overlies the lahar member in the type section; it consists of angular pebbles of lava, pyroclastics, limestone, and other accessory constituents in a coarse pyroclastic matrix. Locally, where erosive channel fill is important, the limestone pebbles dominate; this is notably the case in the gully exposures E of Pico Guillermo. The agglomerate fans alternate with thin layers of fine gray to black tuff. This member passes gradually into the tuff member.

The tuff member consists of a layered sequence of alternating yellow-brown weathering coarse (sand fraction) pyroclastics and fine, gray to black pyroclastics (silt- and mud fractions), with a subordinate amount of volcanic agglomerate lenses. Some reworking must have taken place and in a few cases sedimentary cross-bedding was observed. The tuff member passes gradually into the red beds.

The red beds consist largely of tuffaceous sandstones, siltstones and shales, with a subordinate amount of volcanic agglomerate in erosive channels. Limestone pebbles in these channels contain fusulinids belonging to the *Fusulinella* Zone subdivision B 1/B 2. They may originate from the Agujas or Abismo Limestone (cf. the Brañosera Limestone, de Sitter & Boschma, 1966). The red to purple colour of the rocks is most characteristic. Sedimentary structures are abundant: different types of cross-bedding in the sandstones, a fine parallel lamination in the silts and shales. Caliche horizons and pipes are present throughout this last subdivision.

The type section was chosen for its good exposure and accessibility, but the Labra Formation varies considerably throughout the known outcrop. The lahar member is only locally developed under the Peña Labra and in the Sierra Sagra NE of Aniezo. The volcanic agglomerate is absent in most of the Labra Formation outside the type section area. In the type section area, the Labra Formation has a, probably faulted, lateral contact with an extrusive lava body, which may explain the abundance of volcanic agglomerate there. Outside the type section area, true tuffs have been observed in the anticline E of Puente Pumar, but E of the Puerto Sejos and along the Sierra Sagra most of the tuff member is reworked. The red beds increase enormously in thickness towards N and may attain 600–700 m in the Sierra Sagra. The colouring in this area is not so intense: a pale red to purple alternating with hues of green and grey.

The Labra Formation covered an increasing area with the lapse of time; this is shown for instance E of Lebeña where the red beds rest unconformably on the Carboniferous rocks, whilst the other members are lacking (Photograph 26).

The Labra Formation must be younger than the youngest Carboniferous known in this area. So, a maximum age of Stephanian C can be inferred. A restricted flora from the upper part of the tuff member NW of San Mames (Appendix, sample F 7) yielded a *Walchia piniformis* specimen, which is best interpreted as indicating a Permian age (Wagner, pers. comm.).

Martinez Alvarez (1965) mentioned the occurrence of pyroclastic material in the lower part of the Permo-Trias sequence in E Asturias. Nagtegaal (1968) points to the similarity of the Labra Formation with the Permian Peranera Formation in the S Pyrenees.

Nansa unit. – This unit consists of the Bunter facies (Papa, 1964; Smit, 1966; Maas, 1968; de Jong, 1971), resting unconformably on the Labra Formation. Thickness of the unit varies from about 400 m E of Lebeña to more than 1000 m in the road section of the Nansa valley. Usually, the sequence starts with a quartzitic conglomerate, which is present in all but a few places. 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. In the subject area, the basal conglomerate has its greatest thickness, about 90 m, on the Peña Labra. This thickness decreases rapidly to E and SE, in the N a thickness of 10 to 20 m is normal. 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. In the top part of the Nansa unit, especially in the SE (near Barruelo), evaporation phenomena are present (Papa, 1964; Smit, 1966). Budding, who is preparing a sedimentological analysis of the Buntersection in the Nansa valley, succeeded in collecting a flora (Sample F 8, Appendix) from the upper part of the Nansa unit, indicating a Triassic age (Wagner, written comm.). The Nansa unit passes vertically through a sequence of clays and marls into the calciclastic Tudanca unit (Budding, pers. comm.).

Tudanca unit. – This unit consists largely of well-bedded, grey-weathering micritic, often recrystallized limestones, occasionally alternating with calcareous shales. The Tudanca unit is considered to be of a Jurassic age. Some mollusc-shell beds were discovered near the base of the Tudanca unit, but no determinations resulted in an age indication (Budding, pers. comm.). The Nansa-Tudanca boundary was mapped from aerial photographs according to the different morphologic characteristics of the two units.

CHAPTER III

FACIES DISTRIBUTION IN SPACE AND TIME AND A PALAEOGEOGRAPHIC SUMMARY

INTRODUCTION

There exist rather complex facies relations between the different lithotypes which constitute the above-described lithostratigraphic units of the Upper Carboniferous. To visualize these relations, several facies types are distinguished and shown in their present-day distribution (Encl. VIII). The autochthonous and transitional facies types will be summarized, whereas a more comprehensive description will be given of the allochthonous flysch facies. The distribution in space and time of these facies types in the subject area will be discussed. Finally, I shall attempt to summarize the palaeogeographic relations between the subject area and other parts of the Cantabrian Mountains.

THE DIFFERENT FACIES TYPES OF THE UPPER CARBONIFEROUS STRATA

Three main facies types are distinguished, an autochthonous facies, a transitional facies and an allochthonous facies. We distinguish:

I. Autochthonous facies.

Ia. Autochthonous calciclastics. This facies includes all the limestone units described in the former chapter, except those for which resedimentation could be inferred. The bioclastic limestones consist for the most part of biogenetic bank deposits (van de Graaff, 1972). The non-fossiliferous Caliza de Montaña is probably a shallow-marine limemud deposit.

Ib. Autochthonous, fine-grained siliclastics. This facies comprises all fine-grained, shallow-marine or terrestrial siliciclastic deposits. They may be accompanied by seat-earth levels and coalbeds or by autochthonous limestone lenses.

Ic. Autochthonous, coarse-grained siliciclastics. The conglomerates for which an autochthonous deposition could be inferred, belong to this facies.

II. Transitional facies.

IIa. Transitional, fine-grained siliciclastics. To this type belong the sandstone-shale associations in which both mass transport and grain-for-grain transport phenomena can be observed, while a shallow-marine environment has to be inferred. A variant of this facies will be treated along with the turbidite facies.

IIb. Transitional, coarse-grained siliciclastics. This facies type comprises those conglomerates which are to a certain extent mass-transported but also show characteristics of shallow-marine autochthonous deposition (cf. Savage, 1967, pp. 197, 198). Most of the Curavacas Conglomerates from the N Rio Yuso area, S Liébana and the N Pisuerga area belong to this facies. The sedimentary features observed in these conglomerates, allowing both autochthonous deposition and resedimentation, were already discussed (the Curavacas Conglomerate in SW Liébana and in N Pisuerga and the area between Caloca and Barrio).

III. Allochthonous facies.

IIIa. Turbidites (Photographs 27, 28, 29, 30, 31 and 32). The bulk of the sediments in the Liébana basin belongs to this facies type. As turbidites are considered all rhythmically alternating sandstone-shale associations in which individual sandstone beds are laterally continuous, have a constant thickness, display a fining-up gradation, and have a sequence of structural intervals as distinguished by Bouma (1962). In the subject area several subtypes of turbidites can be distinguished. The Bouma interval sequence is normally incomplete. Turbidites consisting of A, AB, ABC, BC, CD and D intervals are most common (A = graded interval, B = coarse parallel laminated interval, C = ripple current laminated interval, D = fine parallel laminated interval). There are thick- to very thick-bedded turbidites, which consist of a structureless, coarse sandstone, occasionally containing contorted shale slabs, having no gradation. These turbidites may represent an interval more 'proximal' than the A interval. Synsedimentary deformation (convolute bedding or lamination) was observed in the B, C and D intervals, though mostly in the C interval. The D interval may be graded instead of laminated, having a very gradual passage to the shale interval. Distinction between a hypothetic E interval and the autochthonous shales is practically impossible. In a turbidite sequence one type may dominate, but intermixing of different types in one sequence is a common feature (Photographs 29 and 30). The shale ratio tends to be low in 'top truncated' turbidite sequences, and increases in 'bottom truncated' sequences (cf. Photographs 27 and 28). In some places, turbidites were observed, consisting of thin but continuous sandstone laminae, in a sequence where the shale ratio is higher than the sand ratio. These D turbidites show fine parallel lamination or just a fining-up gradation. In such sequences, thin sandstone beds consisting of almost isolated current ripples were also observed (Photograph 31). They may constitute a vertical passage between a turbidite sequence and an autochthonous bioclastic limestone deposit, and are considered as a variant of facies type IIa. They also occur, locally, at the top of the Potes turbidites and at the base of the Buyon turbidites (see Encl. VIII). The isolated current ripples point to a grain-for-grain transport, which cannot be attributed to a turbidity current. The thin laminae on the contrary are hard to explain without a turbidity current mechanism since such laminae were encountered in most turbidite sequences. They are easily overlooked, but in number they constitute in many cases about 40 % of the turbidites (Maas, 1968a). An unusual feature in a turbidite sequence was observed at the top of a BC turbidite, where two interfering ripple patterns were present (Photograph 32). This feature might be explained by assuming a relatively shallow environment in which also normal water currents were possible.

The already mentioned grit beds, which may occur in association with conglomerates and breccia lenses (i.e. those associated with the proximal turbidites in the Aliva and Lebeña Formations, or those in the Tornes Conglomerate), but which may also occur isolated in a normal turbidite succession (notably in the Potes turbidites), constitute a different type of turbidity-current deposit. These grits do not display the Bouma intervals, but the observed sequence of structural intervals, though never complete, fits well in the sequence of structural intervals as proposed by Allen (1970): A, graded interval; B 1, lower interval of parallel laminations; C 1, cross-bedded interval (dunes); B 2, intermediate interval of parallel laminations; C 2, cross-laminated interval (current ripples); D, upper interval of parallel laminations. Photographs 33, 34, 35, 36, 37, 38 and 39, show several features of such turbidite grits, occurring isolated in normal turbidite sequences, exposed in the headwaters of the Rio Saja N of Pico Cordel. In these grits a coarsening upwards gradation was observed more than once (cf. Kelling & Woollands, 1969, p. 261, the multiple graded units). The turbidite grits frequently have a lateral or vertical passage to the next facies type (Photograph 40).

IIIb. High-density current deposits (Photographs 7, 8, 10, 11, 19, 20, 21 and 22). This subfacies includes all breccias, breccious conglomerates, and conglomerates for which a resedimentation, probably by high-density, inertia flows, was inferred. It has been identified in the Aliva and Lebeña Formations in the Picos de Europa area, and in the Liébana area, amongst others, the Cuvo, Tornes, Cabezon, Viorna, and Porrera Conglomerate Members. In some cases there exists a gradual passage from this facies type to the next (Photograph 12).

IIIc. Slump and olistostrome deposits (Photographs 12, 13, 14, 41, 42, 43, 44, 45 and 46).

In the foregoing chapter, all major slump associations and chaotic slide deposits containing large erratic boulders were recorded as olistostrome members. Slump and slide processes are normal phenomena in sedimentary instable areas. The occurrences of turbidites and high-density, inertia flow deposits testify to instable conditions in their source area. However, the deposition of a turbidite must take place in a sedimentary stable environment. The intense slump distortion of mature turbidites (for instance the Bedoya Olistostrome) cannot have been caused by sedimentary conditions as were present during the deposition of the turbidites. The conditions must have been changed and the slumping was probably caused by tectonic activity. This supposition is supported by the fact that most major slump associations can be related with tectonic events as expressed by unconformities, and/or are situated near tectonically instable features as the Mid-Liébana ridge or the Picos de Europa shelf margin.

In the subject area, all kinds of slump products occur and we can distinguish between two extremes: a) slumps in which no original bedding can be distinguished (Photograph 43), the slump has the character of a mudflow, b) slumps in which the original bedding has remained almost continuous, though deformed by gravitational-slide folding (Photographs 44–46). Most slumps, however, have an intermediate character (Photographs 41 and 42); in these slumps the original bedding can be distinguished, but is normally broken up, slabs are pulled apart or pushed upon each other, isolated fold hinges, slump balls, are common. The impression is obtained from the exposures, that there is a gradual passage from resedimentation processes to gravitational-slide tectonic transport: a passage from mudflows to slumps in which the bedding is lost, to slumps in which the bedding is preserved, and to gravitational-slide tectonic deformation. As long as the gravitational-slide deformation can be fixed to one stratigraphic level, and is vertically bounded by undeformed beds having an erosional cut-off at the top, we speak of slump deformation. However, in many cases it is impossible to distinguish between slumping and early post-sedimentary tectonics.

The occurrence of out-size clasts in slump associations has already been discussed in the foregoing chapter; such occurrences also point to a relation of olistostrome deposition with tectonic activity. The olistoliths must have been transported over a considerably longer distance than the slump-distorted host rock. Most of the described olistostromes are probably the product of a great many individual and in time separated slump and slide processes, or else the lateral passage into undisturbed turbidites and the occurrence of undisturbed levels within olistostrome units would be inexplicable.

DISTRIBUTION OF THE FACIES TYPES AND A SUMMARIZED GEOLOGIC HISTORY (Encl. VI, VIII)

Three subareas can be distinguished for the Upper Carboniferous facies types: a stable slowly subsiding shelf area where an autochthonous calciclastic facies dominates, the Picos de Europa area, a rapidly subsiding flysch basin, the Liébana area, and an area where flysch sediments alternate and interfinger with autochthonous shallow-marine and terrestrial sediments, N Pisuerga and the area between Caloca and Barrio. In the last area, subsidence seems to have decreased towards the Polentinos block. This subdivision is a generalization; all kinds of exceptions exist, moreover it is based on the present-day distribution of facies types; it is hardly possible to estimate the spatial relations that existed during the deposition. Nevertheless, a general trend seems to be clear. Throughout the Upper Carboniferous – the Stephanian excepted – there existed a flysch basin which was bounded in the north by a shallow-marine shelf area, and in the south by a shallow-marine to terrestrial area which, with the passing of time, steadily extended to the north. Though several tectonic events, expressed in unconformities, took place, the facies distribution remained essentialy the same from the Upper Bashkirian when the flysch sedimentation started, to the beginning of the Kasimovian when the calciclastic sedimentation in the Picos de Europa came to an end. During this time, the sedimentation rate varied markedly from one facies area to another. The occurrence of lacunas and lack of thickness measurements make an exact comparison impossible. But the sedimentation rate in the flysch basin must have been roughly double that of the Pisuerga area and more than six times that of the Picos de Europa area, of which the sedimentation rate is more or less comparable with that of the Devonian Asturo-Leonese facies.

The transition from Lower Carboniferous to Upper Carboniferous in the Palentine area is not clear. We know that the nodular limestone-shale facies was prolonged into the Lower Namurian. Only the upper part of the Caliza de Montaña is present as a restricted and locally developed, bioclastic limestone facies which rests in one locality (near Revilla) unconformably upon the Villabellaco nodular limestones (Wagner, 1972). The laterally restricted occurrence may be due to a deposition in some restricted, shallow, relatively elevated areas such as ridges, while a flysch sedimentation had already started in deeper parts of the basin. In Liébana, the Cervera Formation is, wherever the relations are clear, unconformable to the Caliza de Montaña equivalents, but we know such relation only from exposures near ridge-like areas. The Cervera flysch sedimentation had almost the same regional extent as the Palentine facies, with the possible exception of a part of the Polentinos area; there, flysch sedimentation seems to start with the Lechada Formation. Comparing the situation during the deposition of the Cervera with that during deposition of the younger formations, we must conclude that, with the lapse of time, the shallow facies gained on the flysch facies from S and W to N and E, though this general trend was several times reversed.

The connection between the limestone shelf area and the flysch basin is hidden by a tectonic overthrust. The flysch in Liébana is almost entirely siliciclastic; the lower part was sedimented within a time of non-deposition on the Picos de Europa shelf (Encl. VI). The younger Aliva Formation presents a lateral passage from a shelf to a flysch basin and the stratigraphically equivalent Pandatrave Formation also contains calciclastic turbidites. However, in the Liébana area a siliciclastic flysch probably remained dominant as indicated by the Bedoya Olistostrome sediments. So, the Picos de Europa shelf

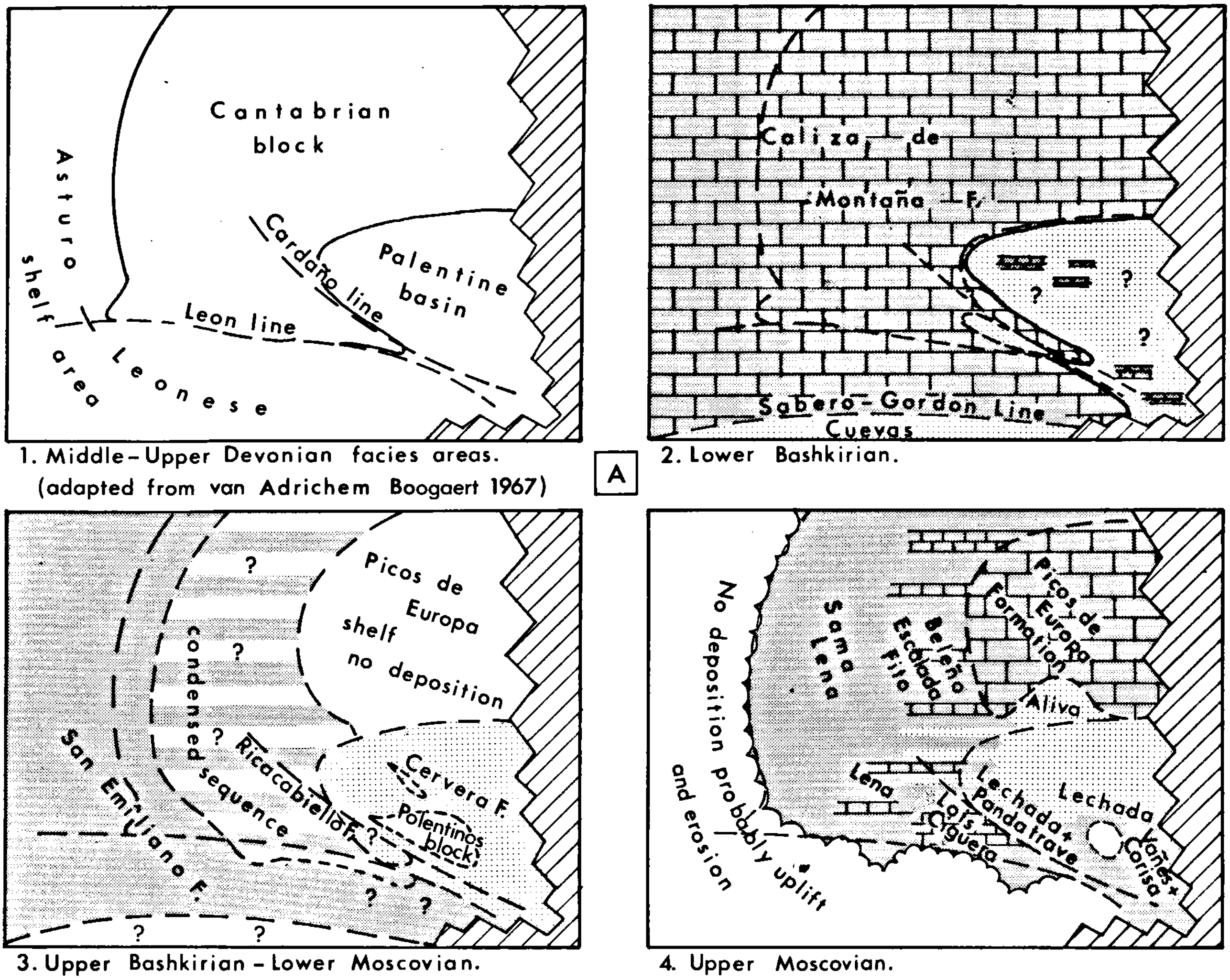
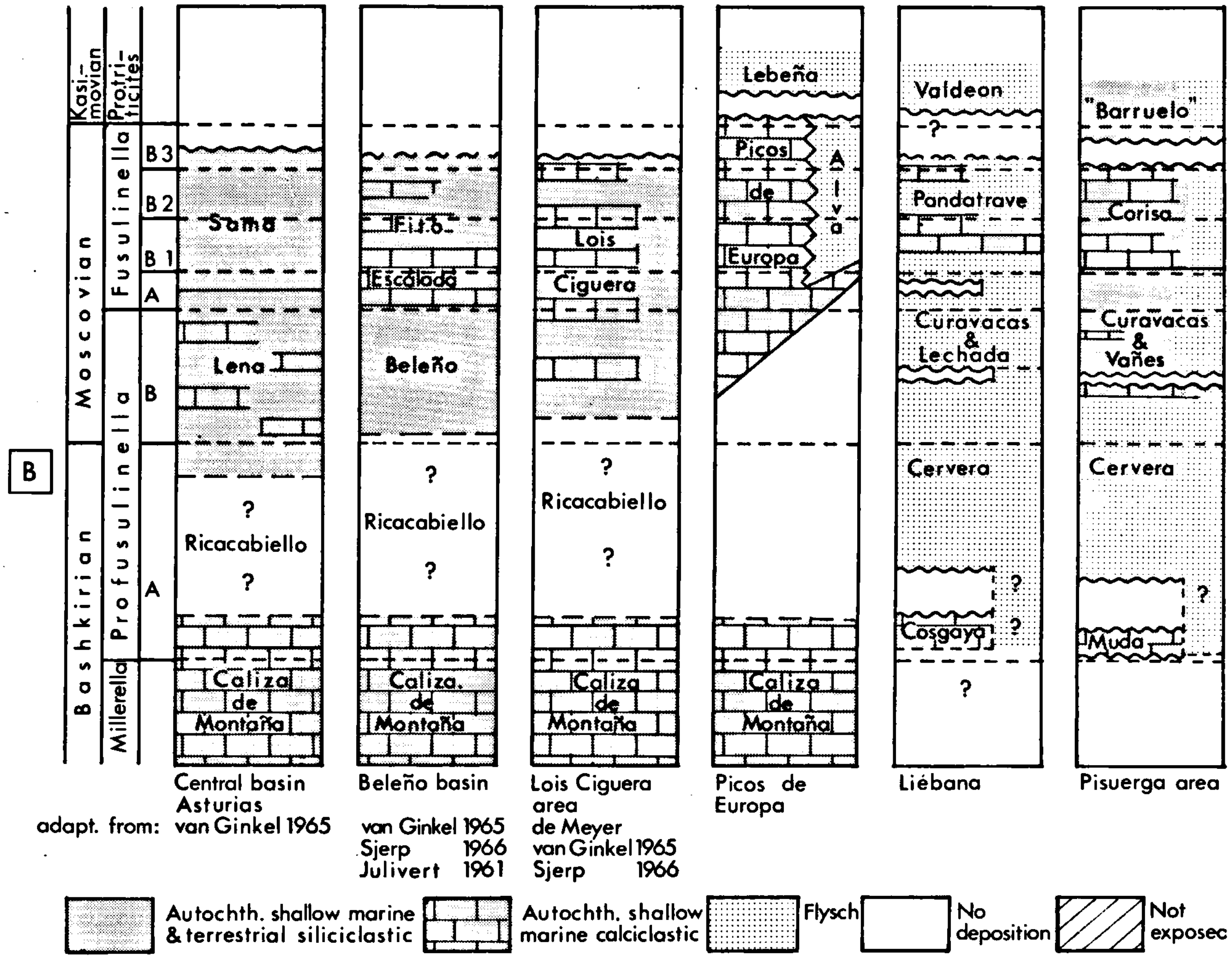


Fig. 1. Generalized representation in maps (A) and columns (B) of the facies distribution in the Cantabrian Mts. during the Upper Carboniferous.

cannot be considered as an important source area and we rather must look to S and W where a shallow-marine to terrestrial facies can be found for the same time span. Such a presentation of palaeogeographic relations agrees well with the facies interpretation of Upper Westfalian strata in the Pisuerga area by van de Graaff (1971).

Within the Liébana basin, flysch sedimentation was neither continuous nor uniform. The thickness of the flysch sediments changes from place to place. The Upper Deva turbidites are markedly thinner than the Potes turbidites; the Potes and probably also the Buyon turbidites thin out towards the Mid-Liébana ridge. The Lechada Formation in S Liébana is much thinner than in the Lechada syncline (Savage, 1967) or in E Liébana and Polaciones. As has been pointed out, most of the olistostrome complexes can be related with unconformity levels, and seem subsequent to tectonic activity. This tectonic activity in the basin and on the basin margins must have been of a local nature; the unconformities may disappear laterally, and correlation of unconformities, at present geographically separated, is hypothetical. There is for instance no proof that the Curavacas unconformity from S Liébana and Pisuerga can be correlated with the Cabezon unconformity in Polaciones. We only know that both unconformities separate strata which belong to the *Profusulinella* Zone subzone B. No equivalent unconformity level was found along the Mid-Liébana ridge. The total picture seems rather a sequence of, in time and space isolated, tectonic events than subsequent tectonic phases which could be expressed in regionally extending unconformity levels. The tectonic events are expressed in the sedimentary record, but in most cases they did not result in major facies changes; the flysch basin remained a flysch basin. The unconformities can be traced to some extent in the Lechada basin (van Veen, 1965; Savage, 1967) and in the Pisuerga area (Wagner & Wagner-Gentis, 1963; de Sitter & Boschma, 1966). But a continuation into the Picos de Europa area or to other parts of the Cantabrian Mountains is in most cases impossible.



REGIONAL CORRELATIONS, THE PALAEOGEOGRAPHIC PICTURE

Regional correlations of the different lithostratigraphic units have been given in the former chapter up to the Cosgaya unit. From then on, correlations have been restricted to the Liébana-Rio Yuso-Pisuerga areas. The reason for this restriction is apparent. The schematic columns of Fig. 1 show clearly that straight correlations for units younger than the Caliza de Montaña and its equivalents, are impossible. Distinct boundaries in one area are non-existent in the other areas. The areas mentioned above, between which correlations are possible to some extent, comprise roughly the former Palentine facies area, which is bounded in the SE approximately by the Cardaño Line and in the N by the Picos de Europa (Cantabrian block) margin. The former Asturo-Leonese shelf area has an altogether different development; the same holds true for the former Cantabrian block area, although the boundaries of these areas are not fixed and fluctuations are common.

The geologic history of these three areas has been represented in four generalized facies distribution maps, and six generalized columns (Fig. 1). The Caliza de Montaña facies covered the entire Cantabrian area with the exception of: a) the area S of the Sabero-Gordon line, b) the area between the Léon line and the Cardaño line E of Riaño, c) possibly, the major part of the Palentine area. In area a) a largely terrigenous sequence, containing turbidites, forms the equivalent of the Caliza de Montaña (Boschma & van Staalduinen, 1968, Cuevas Formation; Wagner, Winkler Prins & Riding, 1972, Olleros Formation). From area b), Koopmans (1962) reported a flysch sequence, resting on top of the Alba Formation, passing laterally into the Caliza de Montaña of the Sierra del Brezo. The possibility of a flysch sedimentation at that time in area c) has been discussed in the former chapter.

The Caliza de Montaña sedimentation comes to an end during the Lower Bashkirian. In the Asturo-Leonese shelf area there is a gradual passage into the San Emiliano Formation (Brouwer & van Ginkel, 1964), a mainly shallow-marine to paralic sequence of limestones, graywackes and shales with occasionally coalseams, which reaches into the uppermost Bashkirian and perhaps into the lowermost Moscovian (van Ginkel, 1965; Rácz, 1965). Younger sediments are not known from the Asturo-Leonese shelf area, except Stephanian strata which rest unconformably on the older rocks which are thrusted and folded. In the Cantabrian block area the Caliza de Montaña is followed by a hiatus (Picos de Europa area) or by an undated thin manganiferous shale sequence which has been interpreted as a condensed sequence (Sjerp, 1966, Ricacabiello Formation). In this area the top of the Caliza de Montaña is also manganiferous in many places. This Ricacabiello facies is also present in several places south of the León line (Evers, 1967; Savage, pers. comm.), where the San Emiliano is lacking. The interpretation as a condensed sequence is doubtful, renewed sedimentation after a time of non-deposition is also possible.

The Ricacabiello is followed by a shallow-marine to terrestrial sedimentation, consisting of shales, graywackes, bioclastic limestones and occasional coalseams, which generally starts in the Lower Moscovian and is continued almost to the upper boundary of the Moscovian (Lena Formation (Barrois, 1882), Sama Formation (Barrois, 1882), Beleño Formation (van Ginkel, 1965), Escalada Formation (van Ginkel, 1965), Fito Formation (van Ginkel, 1965), Lois-Ciguera Formation (Brouwer & van Ginkel, 1964), all formations cited in van Ginkel (1965)). In the Picos de Europa, the hiatus is succeeded by deposition of the bedded member of the Picos de Europa Formation; there the development is entirely marine.

In the Palentine area, as already described, the Cervera flysch sedimentation sets in during the Bashkirian, and has an unconformable relation to the Caliza de Montaña. The Polentinos area and features as the Mid-Liébana ridge seem to have been at least temporarily elevated and open to erosion or submarine denudation; some folding must have occurred there. During the Moscovian, the facies pattern remains more or less the same: no deposition in the former Asturo-Leonese shelf area, shallow-marine to terrestrial deposition in the Cantabrian block area, flysch deposition in the Palentine area. The area of the turbidite facies becomes smaller with the lapse of time. The shallow-marine facies is extended into the Pisuerga area; in the N, however, a part of the Picos de Europa area is covered by flysch sediments (Aliva Formation).

The subsidence of the W part of the Cantabrian block area (Central basin of Asturias), once sedimentation had started, kept up with the subsidence in the flysch basin; more or less the same time interval is represented by some 4000 m of sediment in both areas. The Picos de Europa, however, continued to be an area of slow subsidence with exception of its S margin.

CHAPTER IV

STRUCTURES

INTRODUCTION

General remarks, types of deformation

Within the subject area, several subareas, each with its own distinct structural development, can be distinguished. The distinction is in the style of deformation as well as in the chronologic sequence of tectonic events. Since deformation can be considered as a result of general stress-conditions acting upon the dominant rock type, it is not surprising that the principal boundaries of structural subareas coincide with sedimentary facies boundaries. Moreover, the differences in sedimentary facies, as described in the former chapters, testify to differences in crustal activity in the different subareas (e.g. slow subsidence in a shelf area, rapid subsidence in a flysch basin, intermittent uplift of blocks and ridges). Such different conditions cannot but influence the structural development. Therefore, the final outcome of orogenesis may be considered here as an expression of a complex interaction between processes of deposition and deformation, which both may be regarded as consequences of the crustal activity in that area at that time.

During the study of the different deformation types in the subject area, attention was focussed on folds that were considered to have originated by the force of gravity, in contrast to folds due to lateral stress. Other deformation types as nappe-and-fault structures will not be treated in detail, since they are sufficiently known and described from neighbouring areas. An elaborate analysis of the deformation patterns in the Curavacas and Lechada synclines by Savage (1967), resulted in a deformation model (Savage, 1967, p. 201, fig. 8) that could also be applied to most of the structural development observed in the Liébana area. Savage introduced the concept of collapse folding on the flanks of larger structures which he supposed to have been generated by bending due to vertical forces. This assumption was strengthened by a more recent mathematical fold analysis in the same area, which led to the provisional conclusion that the Lechada syncline should *not* be considered as a fold caused by compressive stress (van de Graaff-Trouwborst, 1971). With the results of these investigations in mind the following types and processes of folding were distinguished in the subject area:

a) folding which is due to vertical movements without appreciable lateral compression.
b) folding which is due to lateral compression.

Folding due to vertical movements
Vertical movements as uplift and subsidence usually result in a number of monoclines, the pattern of which can be described as synclinal and anticlinal structures. The axial orientation of such structures is not dependent on a regional field of lateral stress, but is determined by palaeogeographic features: the configuration of basins and ridges, shelf and shelf slope, deep-seated faults, etc. Such synclines and anticlines must be considered as folds caused by bending instead of buckling (Dennis, 1967, p. 13, 15). Normally, in such cases the anticlines are narrow, having steeply dipping flanks, whereas the synclines are wide with a broad flat bottom, contrasting with the buckle folds in which the anticlines and synclines are mirror images of one another. On the flanks of the bending structures, all kinds of subsidiary collapse structures may develop: cascade folds, flap folds, slip sheets, etc. (cf. Harrison & Falcon, 1934, 1936; de Sitter, 1964; Savage, 1967). Uplift and subsidence movements have also been accompanied by a more superficial sedimentary mass transport by means of slumping and sliding resulting in olistostrome deposits. In these rocks it may be quite impossible to distinguish between sedimentary slump structures and tectonical gravity-slide structures, since there exists a gradual transition between the two. The well-developed collapse folds show several distinguishing characteristics: the axial plane has a horizontal to gently inclined attitude when the fold axis is nearly horizontal; the fold axis may plunge quite steeply in a more or less random direction, but lies in a plane parallel to the flank of the 'host'-structure; the inverted limb of a collapse fold is well preserved, showing no signs of tectonic thinning; the tectonic transport in a collapse fold is downslope in regard to the flank of the 'host'-structure; collapse folds cause a shortening in the vertical, and a dilatation in the horizontal direction. Tectonic transport by collapse on synclinal flanks may generate a stress field in the core of that syncline, causing buckle folds with all the characteristics of folds caused by lateral compression. The axial orientation of such buckle folds is entirely dependent on the axial orientation of the 'host' bending fold.

Folding due to lateral compression
Buckle deformation due to lateral stress also occurs in the larger part of the area. The folds are quite regular, as mentioned above, and are also generally much smaller, having vertical to steeply inclined axial planes and very variably oriented fold axes. Fold limbs are rarely overturned. Generally, these folds are overprinted upon the synsedimentary and other earlier structures. Thus, a typical parasitic relationship, as can be demonstrated from the symmetry and axial plane cleavage relationships, can usually be established in dipping flanks of the earlier folds. The axial plane cleavage of the buckle folds is normally a crenulation cleavage (Rickard, 1961) in the shaly part of the deformed rock. The crenulated older cleavage is a slaty cleavage having an orientation subparallel to the bedding plane; according to Savage (1967), this slaty cleavage is probably a concentric cleavage related to the bending folds.

As stated above, the early processes of deformation, bending with susidiary collapse, are essentially synsedimentary; the subsidence and uplift which caused them, also activated sedimentation (mainly of allochthonous sediments). For younger deformation processes, as buckling due to lateral stress, no such relations could be established.

Alternative formation of the flatlying folds
The assumption that the overturned and recumbent structures have been formed by collapse can only be defended on the grounds of simplicity. On the one hand, essentially similar structures are developed during sedimentation; the style of these synsedimentary structures with pull-apart phenomena and associated olistostromes strongly suggests gravity as the driving force for them. On the other hand, the flatlying structures are subsidiary to, and consequently later than, the development of major folds, but refolded by an even later episode of minor folding. If the now recumbent folds had been first formed with vertical axial planes, we would have to involve a separate folding episode to bring them in a flatlying position before their being refolded by the minor folding. There is no evidence for such a separate episode either from mapping or internal structures.

The collapse hypothesis will be further substantiated by factual evidence in the following sections of this chapter.

Scope of the work, sequence of treatment
To substantiate the above statements, a general description will be given of the structural development in the different subareas. Publication of a 1:50 000 geological map of Liébana was the main object of the field work. The collecting of such an amount of data as would allow a detailed tectonic analysis was beyond the scope of this work. However, sufficient knowledge was obtained to give the general lines of structural history; additional data were available from several unpublished reports (see former chapters and previous work listed in Encl. I). The 1:50 000 map and sections (Encl. I, II) provide a general three-dimensional picture. The following subareas are distinguished and will be discussed separately in this order (see the Tectonic index map, Fig. 2):
a) The Picos de Europa; large nappe structures, collapse structures, local refolding of the nappes.
b) Northern Liébana and Polaciones; large recumbent flap folds and cascade folds refolded by mesoscopic buckling folds.
c) The Mid-Liébana ridge; large-scale sedimentary slumping and sliding, concurring with early tectonic gravitational collapse.

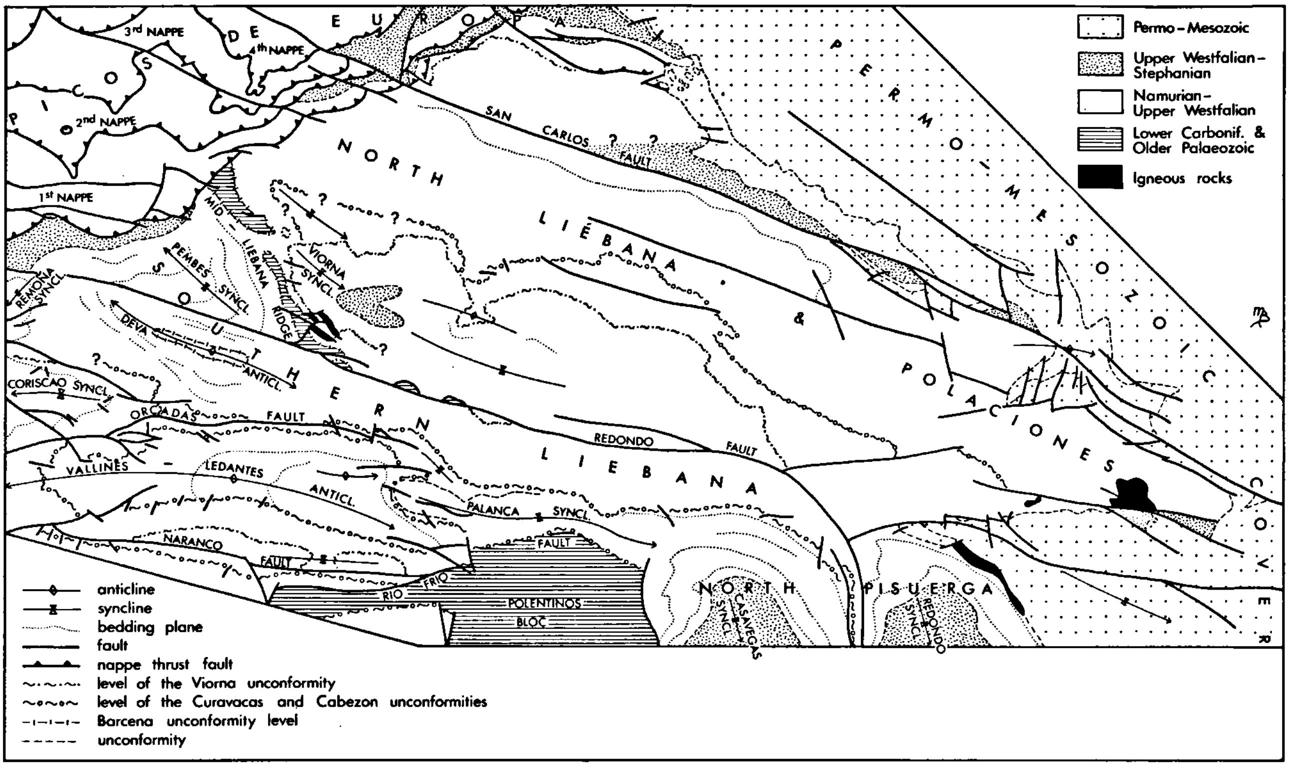


Fig. 2. Tectonic Index Map.

d) Southern Liébana; collapse folds subsidiary to larger structures.
e) Northern Pisuerga; large fold structures considered as a result of both subsidence (bending) and lateral compression (buckling) (de Sitter & Boschma, 1966).
f) The Permo-Mesozoic cover; deformation younger than the Hercynian orogenesis, faulted open synclines and anticlines with gently dipping flanks.

Reconstruction of the structural history is based on two kinds of evidence: 1) changes and breaks in the lithostratigraphic record which can be interpreted as due to tectonic development, 2) overprint relations from which a relative chronology of tectonic events can be deduced. After the treatment of the subareas, the different types of deformation will be summarized once more and the factual evidence for the hypotheses given above will be considered. Special attention will be paid to synsedimentary and early tectonic deformation. The application of the 'tectonic phase' concept will be discussed.

THE STRUCTURAL SUBAREAS

The Picos de Europa (sections A, B, C and D (Encl. II))
Structural history. – Data concerning a pre-Carboniferous structural history are lacking in the Picos de Europa subarea, as in most parts of the Cantabrian Mountains. During the Lower Carboniferous, some uplift and erosion may have occurred; however, the lithostratigraphic record of that timespan is ill preserved due to the later development of nappe structures in which the Alba Nodular Limestones and possibly also the Vegamián Shales acted as a slide horizon. The lacuna between the Caliza de Montaña and the Picos de Europa Formations represents a time of non-deposition without appreciable uplift and erosion in the larger part of the subarea. However, the occurrences of remnants of Caliza de Montaña and even of Alba and Vegamián in the Potès turbidites, and of Caliza de Montaña in the Bedoya Olistostrome point to instable conditions at the S border of the, then, shelf area. The deposition of the Aliva shales and limestone turbidites implies vertical movements (an increasing rate of subsidence) in the S part of the shelf area.

The first tectonic activity that caused considerable horizontal shortening, occurred before and during the deposition of the Lebeña Formation, probably in lower Kasimovian times. This tectonic activity is marked by a drastic change in sedimentary environment. The base of the Lebeña Formation is locally strongly unconformable. A considerable southward transport of the first (southernmost) nappe is inferred, since no trace of a transitional facies can be observed at the thrust contact between the limestone complex and the flysch sediments of N Liébana. This displacement antedates the local tilting of the southernmost nappes to a near-vertical position which caused gravitational collapse both in the limestone complex and in its unconformable cover (section CI–CII, Photograph 47). The gently north-dipping third nappe in section CI–CII is thrusted upon the collapsed Lebeña rocks; consequently its emplacement must be, at least partly, later. So, the genesis of the nappes is a process which preceded, caused and outlasted the deposition of the Lebeña Formation. In the lacuna between Lebeña and Labra, several deformations took place. In this period the nappe transport came to an end. Folding due to lateral compression locally deformed the thrust planes as can be observed along the ridge between the Silla de Caballo and Peña Santa (section B^I-B^{II}) and at the thrust front N of the Puerto de Remoña (Photograph 48). The fold axes have a variable but roughly E–W strike. The exact age of this deformation is unknown; a pre-Labra age is inferred, since no comparable folds were observed in the Labra Formation. Such age relations cannot be assumed for the roughly WNW–ESE striking fault system that probably initiated as a set of tear faults. Vertical displacement also occurred. The fault system can be traced into the Permo-Mesozoic cover. The deposition of the Labra Formation was preceded by uplift and erosion of the entire folded complex of Carboniferous and older rocks; it marks the end of a timespan with strong tectonic activity.

Deformation types and processes. – The nappe structures in the studied area generally consist of the following lithostratigraphic units: the Alba, Caliza de Montaña and Picos de Europa Formations. The Alba Formation acted as a slide horizon at the base of the nappes: it is absent in many places, probably due to tectonic thinning, and, where present, usually shows strong internal deformation. The general thickness of a nappe is about 1000 m. The length (perpendicular to the direction of transport) must be 30–40 km, equalling the length of the Picos de Europa massif. The width (parallel to the direction of transport) cannot be inferred from field evidence but must exceed the relative transport distance. A distance of at least 4 km can be inferred for the transport of the third nappe relative to the second nappe (section A). Similarly, a minimum distance can be inferred for the transport of the second nappe relative to the first nappe. In this latter case the facies relations of the Picos de Europa and Aliva Formations imply a transport distance of about 6 km to discount a missing transitional slope facies (cf. the description of the Aliva Formation in this thesis, and Thomson & Thomasson, 1969). The N–S section through the Picos de Europa massif provided by the Duje river valley is shown to consist of an imbricated structure of at least six nappes (cf. this thesis, Encl. II, and Martinez Alvarez, 1965). The relative transport of each nappe may well exceed 4 km, so that the total NS shortening of the Picos de Europa probably exceeds 20 km, or more than 50 %.

In the area N and W of the Picos de Europa massif, similar nappe structures were mapped (Julivert, 1961; Martinez Alvarez, 1965; Sjerp, 1966). These nappes, however, do not consist of only Carboniferous rocks, but generally include Ordovician and Cambrian formations. This difference can be explained in two ways: stratigraphically by assuming a greater vertical extension for the lacuna of the Cantabrian block in the Picos de Europa area, tectonically by assuming an upcutting of the thrust faults and a translation of the slide horizon to the Alba Formation.

Collapse folds developed in several places of this subarea subsequent to tilting of the nappes into steep attitudes. The tectonic transport caused by this folding is S over N, contrasting to the nappe transport. The collapse folds observed in the limestone complex (section $A-A^I$) are cascade folds as described by Savage (1967); morphologically they can be described as chevron folds (folds on a mesoscopic scale with angular hinges and straight limbs) with horizontal axial planes. Such folding causes vertical compression and horizontal dilatation. The most conspicuous collapse folds are found in the Lebaña Formation, E of the Canal de San Carlos (Photographs 47 and 49, section C^I-C^{II}). These folds also show chevron characteristics. A relation between the thickness of a folded layer and the dimensions of the fold can be observed. The San Carlos Conglomerate (Photograph 47) shows a large synclinal cascade fold; the overturned limb has slid down to N along a subhorizontal N dipping fault plane, cutting through the core of this fold. The complex of thinner conglomerate-breccia layers and turbidites in front of this collapse shows a syncline-anticline-syncline succession. In these folds the inverted limb is much shorter than the subhorizontal normal limb. Only the lowermost syncline looks like a pure cascade fold having a horizontal axial plane and fold limbs dipping in opposite directions but younging in the same direction. The uppermost syncline was probably deformed by the subhorizontal gliding of the inverted limb of the large syncline and this gliding may also have caused the gentle refolding of the normal lying limb. In the turbidites between the conglomerate-breccia layers still smaller collapse folds were observed (Photograph 49); these folds have the same characteristics and also a S over N downslope tectonic transport. The differences in dimensions between the different folds, created fold disharmonies which must have caused intense bedding plane shearing.

Folds due to lateral compression in this subarea originated under a NS stress when the nappe transport was completed. The folds have a variable style, mostly intermediate between chevron and concentric folds (Photograph 48); the axial plane is generally near-vertical. Locally, a parasitic relation of these folds to the tilted nappes could be established.

The faults in this subarea are dealt with now. In the Picos de Europa several joint systems developed in conjugate set patterns as can be observed on aerial photographs (cf. Al Katrib, 1971). The joints are characterized by dolomitization and recrystallization, the occurrence of lead ores seems dependent on the joint systems. One dominant system consists of WNW–ESE and ENE–WSW striking sets. Several large faults strike parallel to the WNW–ESE joint set, along which horizontal displacement has been observed, mostly in a dextral sense which matches the NS compression implied by folds and

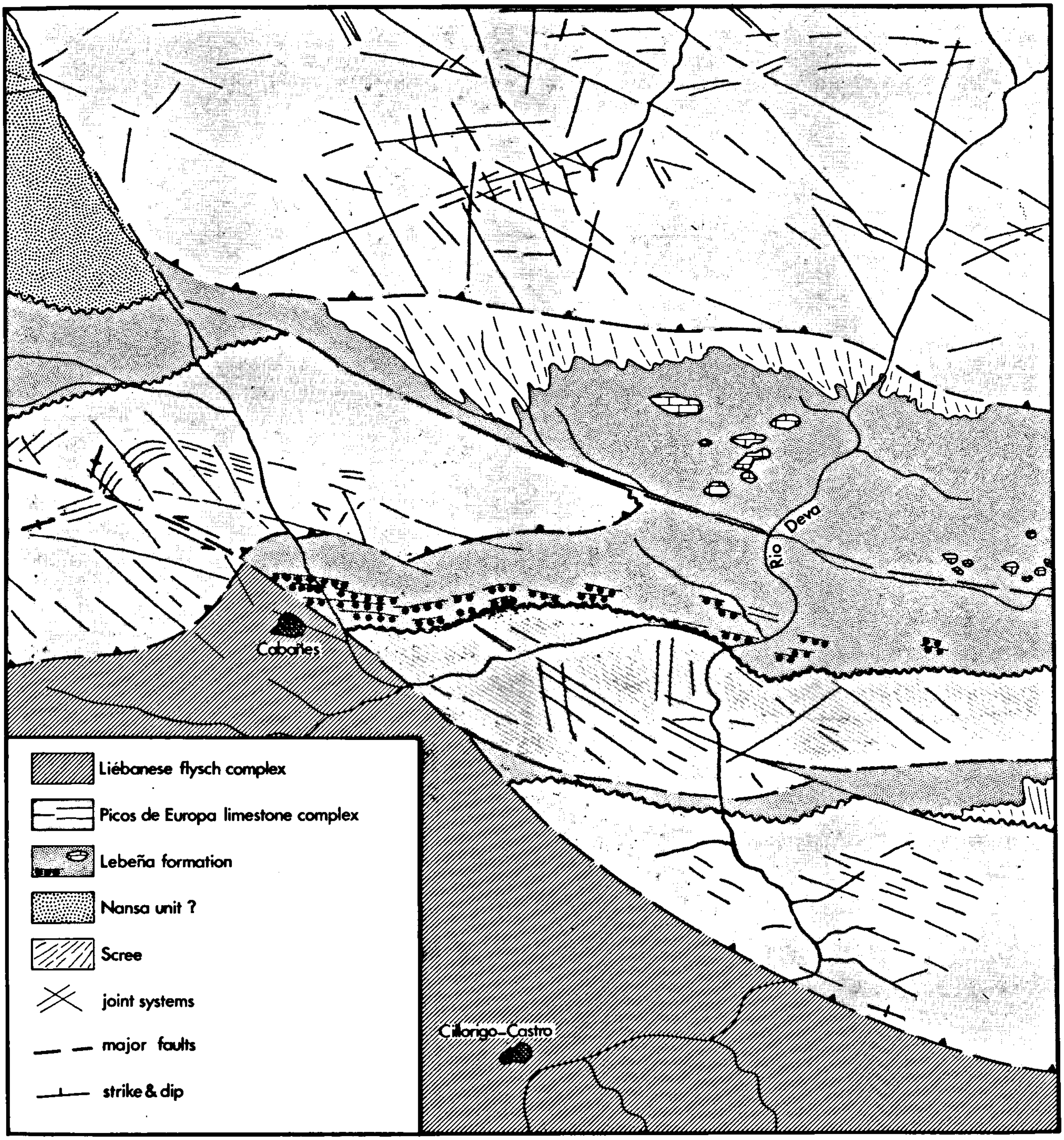


Fig. 3. Aerial photograph interpretation of the Lebeña area.

nappes. However, sinistral displacement also occurred: west of locality 70SC (Encl. VII) the SW part of the faulted second nappe has moved to ESE and is thrust along a steeply W dipping secondary thrust plane against an isolated slab of San Carlos Conglomerate which was tilted and partly overturned to the E. North of Cabañes, partly outside the map, horizontal displacement along a fault caused a folding of a nappe along a vertical axis (Fig. 3). The axial plane is parallel to the fault, and the folding is equivalent to a sinistral displacement. This folding proves the already vertical position of the nappe at the time the faulting started. A similar deformation on a much larger scale affected the nappes WNW of Collio, E of the San Carlos fault (Fig. 2). Apart from horizontal displacement along the faults, vertical displacement took also place. This is notably true for the San Carlos fault which extends into the Permo-Mesozoic cover, continuing to ESE in a fault system that has been active in post-Mesozoic times. Horizontal displacement could not be proved for the faults in the Permo-Mesozoic cover.

Northern Liébana and Polaciones (sections B, C, D, E, F and G (Encl. II))

Evaluation and interpretation of data. – The strongly folded Carboniferous flysch deposits of N Liébana and Polaciones are bounded in the N by the Picos de Europa, in the SW by the Mid-Liébana ridge and the Redondo fault, in the SE, E and NE they are hidden from view by the unconformable Permo-Mesozoic cover. The degree of exposure in this subarea is generally poor, only mesoscopic structures can be observed and the larger structures had to be inferred from scattered observations on isolated outcrops along road talus and steep river valleys. An intensive survey of the available outcrops was made in central Liébana, E Liébana and Polaciones by Lanting (1966), Miedema (1966) and Maas (1968), respectively.

The following critical observations may be noted. The subarea is divided by a continuation of the San Carlos fault; north of this fault, no mappable structures were observed and slump distortion dominates in most exposures. Considering fossil evidence, a southward younging seems likely in this area. South of the San Carlos fault the younging is towards S and SW. The bedding is frequently overturned, especially in the northern half of this part of the subarea. This overturning of the bedding is not due to, but antedates the observable mesoscopic structures. In the lithologically monotonous flysch area, the overturned position of the bedding constitutes one of the few mappable features and was employed as such. It was assumed that the overturning is not restricted to isolated outcrops but covers large parts of the subarea. This assumption could be substantiated by the survey of some road sections with fairly continuous exposure, for instance along the road between Lombraña and Tresabuela (Polaciones). This road section has a length of 2160 m, of which 740 m are not exposed; the bedding is overturned along 1320 m, sometimes twice overturned due to the mesoscopic folding. Along 100 m of the outcrop can there be any doubt that the normal position of the bedding was caused by a second inversion. The entire section is affected by mesoscopic folds, but the enveloping surface of these folds, representing the average dip of the unit, is practically horizontal (section G^{II}–G^{III}), generally dipping gently N. Other less continuously exposed road and river sections yielded similar results, though the lack of continuity made reconstruction of the average dip of the unit in most cases unreliable. Extrapolating from the surveyed sections a map pattern of the occurrence of overturned bedding was drawn and checked against the isolated outcrops occurring between the sections. The resulting map combined with the sections revealed the existence of some major recumbent structures, and the traces of the axial planes of the structures are the map boundaries between overturned and normal bedding. This method, employed to discover and map the large structures, is fairly reliable in central Liébana and Polaciones where roads and rivers run perpendicular to the general strike of the bedding. Such investigation in cross section was impossible in E Liébana between the Quiviesa-Deva and the Tornes valley; consequently, that part of the map pattern is less reliable. The representation of mesoscopic folds in the sections (Encl. II) had to be highly diagrammatical. The large structures observed and mapped, allow different interpretations of their origin and chronologic sequence. Therefore, these structures will be discussed before summarizing the tectonic history of this subarea.

The collapse hypothesis. – The overturned limb of a large recumbent, apparently synclinal structure could be mapped in Polaciones and E Liébana. Chevron folds with subhorizontal axial planes are present in front of (SW), and laterally (WNW) in continuation of, this recumbent structure. All these folds have well-developed unthinned inverted limbs and subhorizontal axial planes, while the younging direction and the direction of tectonic transport (SSE) are identical in the different fold limbs. The tectonic transport can be considered as a transport of material to the axial zone of the sedimentary basin. Contemporaneous unconformities (Cabezon and Viorna unconformities) have a local character, while the sedimentation seems to have been continuous in other parts of the basin. The folding was not followed by a regional break in sedimentation or an important facies change. Such characteristics are best explained when the folding is considered as a large-scale synsedimentary collapse phenomenon. The large recumbent structure is considered a flap fold, the smaller chevron folds with horizontal axial planes are considered cascade folds (Savage, 1967, p. 201, fig. 8). However, accepting this collapse hypothesis, various explanations remain possible for the generation and chronologic sequence of the folding, considering the present configuration of the folds.

Generation of the collapse structures. – Contrary to the collapse structures known from the Curavacas and Lechada synclines (Savage, 1967), no host structure can be distinguished for the collapse structures in N Liébana-Polaciones. Still, the structures must have developed on a slope, and since turbidite deposits are affected by the collapsing, this slope cannot have been a mere palaeogeographic feature. Several models are possible for the generation of such a slope. A tectonic deepening along the axis of the sedimentary basin might have resulted in a steepening of the basin slopes, in which case the entire sedimentary basin must be considered as host structure for subsidiary collapse (cf. Helwig, 1970, p. 182, fig. 7). Such a model would explain slumping, olistostrome deposition and mesoscopic collapse folds, but large recumbent flap folding is not so easily explained. An alternative model is borrowed from Lebedeva (1969). In this model, collapse in a cover rock is subsidiary to differential vertical movements of faulted blocks of a basement. This model, with some

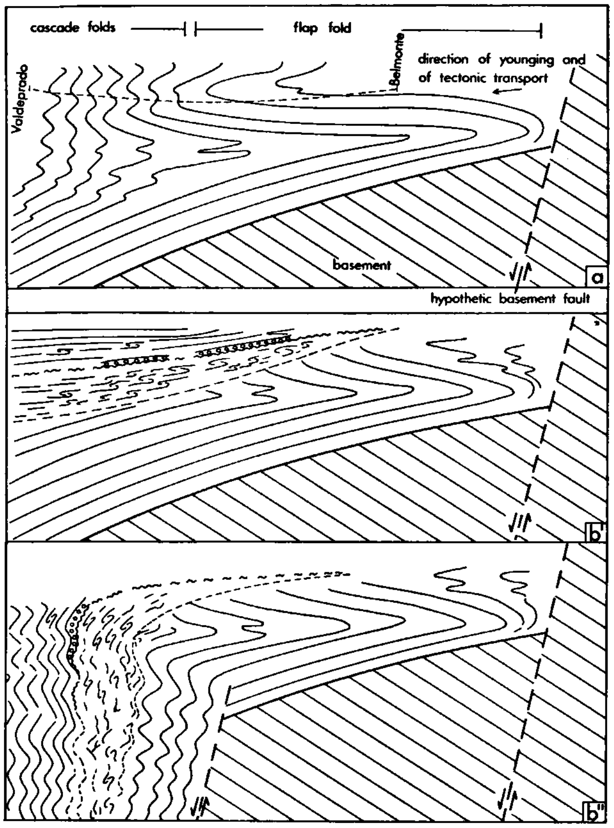


Fig. 4. Alternative hypotheses concerning the gravitational folds in Polaciones.
4a. Contemporaneous development of flap fold and cascade folds.
4b'. Flap folding initiated by basement faulting; contemporaneous olistostrome deposition followed by unconformable deposition of a new flysch sequence.
4b''. Renewed basement faulting initiating cascade folds in a draped structure over the basement fault scarp.

minor changes, is applied to explain the SSW collapse transport in N Liébana and Polaciones. Fig. 4 presents several hypothetical developments for the fold configuration in Polaciones (section F). Flap fold and cascade folds may be considered as contemporaneous and subsidiary to a blockfaulting at the back of the flap fold (Fig. 4a). This hypothesis ignores the presence of the Cabezon unconformity and of the probably syntectonic Salceda Olistostrome. Moreover, a configuration of flap and cascade folds as sketched in Fig. 4a involves a strong internal disharmony which seems incompatible with the known uniform lithology. Therefore, a more complex hypothesis is proposed (Figs. 4b', 4b''): the flap folding was subsidiary to a first block faulting (Fig. 4b') and caused olistostrome resedimentation of superficial sediments; this was followed by a sequence of less chaotic allochthonous sedimentation, starting with the deposition of the unconformable Cabezon Conglomerate; a second block faulting (Fig. 4b'') caused tilting and cascading of this younger sequence, and the deposition of the Barreda Olistostrome (E Liébana, section E) may be connected with this second collapse. The somewhat different fold configuration in section C as compared to section F can be explained by assuming a waning to WNW of the two collapse-inducing fault movements. In central Liébana, collapse to S and SW is marked by cascade folds which are cut off unconformably by the Viorna Conglomerate. This collapse must have occurred in between the collapse events in Polaciones. It was probably likewise fault controlled but seems to have coincided with uplift along the Mid-Liébana ridge.

The hypothesis offered above, explains most structural features that go with it, however, some aspects remain problematic. This applies notably to the dimensions of the overturned limb of the Polaciones flap fold. The minimum width of this limb, perpendicular to tectonic transport, is about 4000–6000 m. The collapse-inducing fault movement should have had a vertical displacement of the same magnitude if the overturning of the limb was effected by rotation from a vertical to a horizontal position; the vertical position being due to a draping of the Potes turbidite cover over the fault escarpment. Such a rate of displacement seems improbable for a fault of only local importance. Besides, one would expect a crumbling-down in cascade folds instead of a rotation to a horizontal position for the vertically tilted Potes turbidites, considering the characteristics of this well-layered rock with alternating competent and incompetent layers allowing to all appearance a fair amount of interlayer gliding. The alternative for the overturning of the middle limb of the Polaciones flap fold, when induced by a much smaller vertical fault movement, offers different but even more serious improbabilities. In such circumstances the flap fold must have developed in a rolling fashion, causing a migration of the fold axis, during the folding, in the direction of the tectonic transport. Such development implies a stretching, subsequent to bending, in the rock material. My own knowledge of rock materials is far too insufficient to estimate the consequences of this implication in so far as the mechanical properties of the rock are concerned and as to whether they would allow such behaviour in a deformation process. However, it seems of importance to know whether or not the process of lithification was completed before the generation of the collapse.

Structural history. – In the N Liébana-Polaciones subarea, nothing is known of the record of structural history for Lower Carboniferous and older times. The frequent occurrence of small-scale slumps, and occasional more extensive olistostrome resedimentation of turbidite deposits, point to crustal activity during the deposition of the Potes turbidites. In Polaciones this activity culminated in the first tectonic collapse folding which seems to be associated with the deposition of the Salceda Olistostrome (see above). This local tectonic event, marked by the Cabezon unconformity, was followed by renewed turbidite sedimentation. In central Liébana, the sedimentation seems to have been continuous. A second, younger collapse, marked by the Viorna unconformity, seems to have been restricted to central Liébana. The collapsing of the Porrera Conglomerate and Buyon turbidites in E Liébana and Polaciones which may be associated with the Barreda Olistostrome was of still later date. In Polaciones, the transport direction of this collapse is locally almost EW. Mesoscopic buckle folding overprinted the collapse structures in the entire subarea. However, a strict chronostratigraphic age relation could not be established. All buckle folds may have the same age, in which case they should have generated during the Stephanian, being younger than the San Mames unit and older than the Cordel unit. However, it is also possible that buckle folds already developed in the flat limbs of the Polaciones flap fold during the initiation of the cascade folds at the SW front of the flap fold. In that case, several generations of buckle folds should be present. This assumption is strengthened by the fact that notably in the Polaciones flap fold, the buckle folds, developed as kink bands, show mutual overprint relations. However, as far as observed the cleavage relations were always the same: an axial plane crenulation cleavage, associated with the buckle folds, deforming a slaty cleavage subparallel to the bedding plane (cf. Savage, 1967). In the flanks of the mesoscopic buckle folds, small-scale subsidiary collapse folds were sometimes observed; they are explained as an effect of a relaxing of the lateral stress which caused the buckle folds. Fault movements as were supposed to have generated the collapse folds, were also active in later times. In some places, post-Mesozoic fault displacement can be observed. The tilting of the Cordel sediments may have occurred prior to the deposition of the Labra Formation, but only its priority to the deposition of the Nansa unit can be proved. Phenomena of contact metamorphism indicate that the granite intrusion of Pico Ijan took place after the deposition of the Cordel unit. The age relation of this intrusion to the deposition of the sediments of the Nansa unit is obscured by faulting.

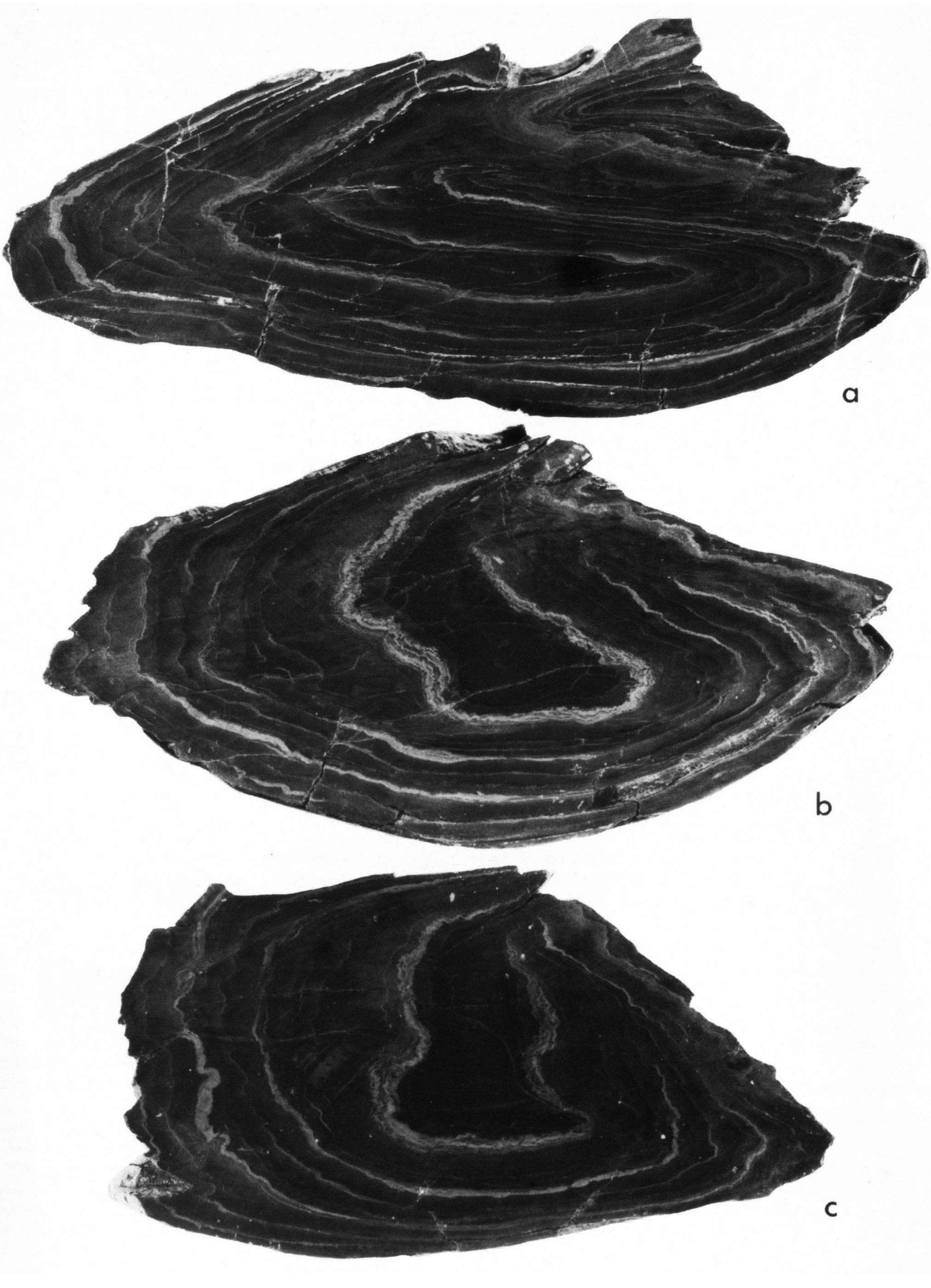
Deformation types observed in slump complexes. — As stated in the foregoing chapter, several types of slump complexes exist: in some, fragmentation and subsequent chaotic transport of rock dominate, in others extensive folding can be observed; furthermore, isolated fold hinges may be observed in a more thoroughly disturbed matrix. Such an isolated slump fold hinge is represented in Fig. 5. The deformed material originally consisted of thin layers of graywacke, grading upwards into shale: turbidites consisting of the D interval only. The reconstruction of the genesis of this slump fold (Fig. 5f) suggests the following sequence of events. At the start of the slumping a wavelike motion developed in the sediment. The 'crest of the wave' toppled over, and by the traction of this motion the underlying sediment was partly elevated into a new wavelike fold. In the further development of the slump process, the rolled slab of sediment was torn loose and entered the slump mass as an elongated body with a more or less circular cross-section and a variable diameter, its axis perpendicular to the general direction of transport in the slump. Eventually, the isolated fold-hinge body was flattened by the load of sediment deposited after the slumping. This flattening caused a buckling of the sandy layers, the buckle folds having a more or less ptygmatic character (de Sitter, 1964, p. 288). Apart from such folds, thin wispy trails of sand can be seen intruding the shale parallel to the plane of flattening. Such characteristics suggest a water-saturated soft condition of the material during deformation.

Another kind of slump folds exists which suggests a more consolidated condition of the slump-affected rock. The so-called rotational slumps, rotated piles of strata bounded by shear planes, belong to this kind. Such rotational slumps are known from various sites in the subject area. The type is thoroughly described by Laird (1968). The two examples represented in Photographs 50 and 51, are both from the talus of the road C627 near road mark 137. A slump genesis is accepted for both structures; however, the numerous small faults, the bounding shear planes and the sometimes kink band-like character of the folds (e.g. Photograph 50; cf. Laird, 1968, fig. 7) suggest a more rigid behaviour of the rock than in the case mentioned above. The transport of material in this type of slump is restricted to transport by folding and shearing. The emplacement of the two examples in the surrounding rock, suggests a genesis within more or less consolidated rock. The direction of transport (to the E) is contrary to the direction of collapse in this area (to the W).

Mesoscopic cascade folds. — Various types of mesoscopic cascade folds developed contemporaneous with the larger collapse structures mentioned above. The symmetry and orientation of such folds is dependent on their place in the larger structure. The folds may be developed as tight chevron folds or kink bands (N.B. no conjugate sets) in the hinge areas of larger collapse folds (e.g. Photograph 52, place: section C–C^{I}, 3 km S of the Picos de Europa boundary). Other more open cascade folds with almost horizontal axial planes developed outside such areas (Photograph 53; place: section C–C^{I}, below Pico Viorna).

One example of a kink band which developed during a collapse deserves special attention (Photograph 54 and its detail: Photograph 55). The kink band has a strong internal disharmony, and minor collapse structures have developed subsidiary to the steep limb of the fold. Photograph 55, the detail, offers a picture of a flap fold resulting from the downsliding of a graywacke layer in the 'depression' caused by the kinking.

Deformation of the collapse structures. — The mesoscopic buckle folds are developed as kink bands and other chevron type folds; a more concentric fold type also occurs but does not dominate. The kink bands are seldom seen as a conjugate set. Interfering kink bands, crossing each other, usually show a distinction in size


a
b
c

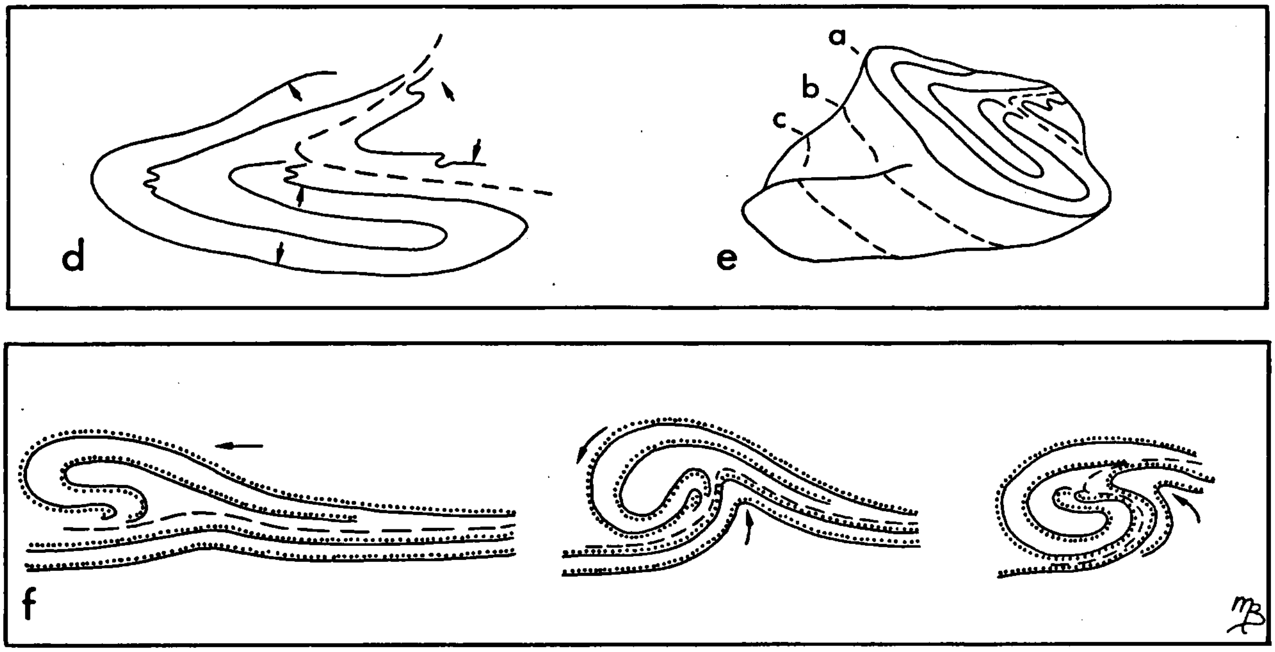


Fig. 5. Characteristics of slump folding; 5a, b and c are saw cuttings perpendicular to the fold axis (photographs 0.8 X); 5d is a sketch of 5a, arrows indicating the direction of younging; 5e is a sketch of the handspecimen that provided the saw cuttings; 5f is a sketch representing the hypothetical development of the slump fold.

and tightness, though it is hard to distinguish different generations. Due to the parasitic character of the folds, all kinds of axial orientation are possible but the E–W orientation clearly dominates. The folds are best developed in the subhorizontal limbs of flap folds. In S Polaciones near Piedrasluengas there exists a distinct cross folding pattern between the roughly NNW–SSE orientated mesoscopic folds generated during the major collapse movements and a roughly E–W orientated system of younger mesoscopic buckle folds with vertical axial planes. In N Polaciones near Belmonte and Pejanda, the refolding of the collapse structures occurred on a more than mesoscopic scale and this deviating size of the folds is probably controlled by the occurrence in that area of coarse proximal turbidites, turbidite grits and conglomerate lenses. In the flanks of these folds, subsidiary mesoscopic cascade folds were frequently observed. These local collapse phenomena were not controlled by a palaeobasin configuration; they are entirely dependent on the orientation of the flanks of relatively small host folds.

The Mid-Liébana ridge (Encl. II, sections C and D; Encl. IV; Fig. 6)

General. – The geology of the Mid-Liébana ridge area has remained problematic in many respects. However, the following generalizations seem valid. In the palaeogeographic configuration, the subarea constitutes a ridgelike feature, separating the turbidite area of N Liébana-Polaciones from the S Liébanese turbidite area. Tectonically, the ridge is a faulted anticlinal structure flanked by and partly submerged in its own 'debris' of syntectonically deposited olistostrome material. Several unconformities testify to repeated uplift and denudation and comparison with the history of other subareas favours a restriction of these processes to the ridge and its immediate surroundings. The subarea offers a fine example of the concurrence of chaotic sedimentary mass transport, large-scale sedimentary slumping and sliding, and early tectonic gravitational collapse.

Structural history. – The Mid-Liébana ridge offers the only outcrops of autochthonous Middle to Upper Devonian and Lower Carboniferous rocks in the entire area of Liébana. Therefore, no conclusions can be drawn concerning the regional occurrence of the unconformities at the base of the Vegamián and Villabellaco Formations respectively, nor concerning the palaeogeographic configuration at that time.

The Upper Carboniferous history of the ridge is marked by at least three unconformities: the observed Barcena and Viorna unconformities and the inferred unconformity at the base of the Campollo Olistostrome. Only the Barcena unconformity is preserved on both flanks of the ridge. The Upper Carboniferous development of the ridge is diagrammatically represented in Fig. 6. The restricted occurrence of the allochthonous Llaves Limestone (Chapter II) suggests that its former autochthonous occurrence was also restricted and so favours the existence of a ridgelike feature preceding the genesis of the Barcena unconformity (Fig. 6a). Structurally, the Mid-Liébana ridge consists of a NW anticlinal and a SE monoclinal part. The gap between these parts, near Sebrango, is filled with material of the Enterrias Olisto-

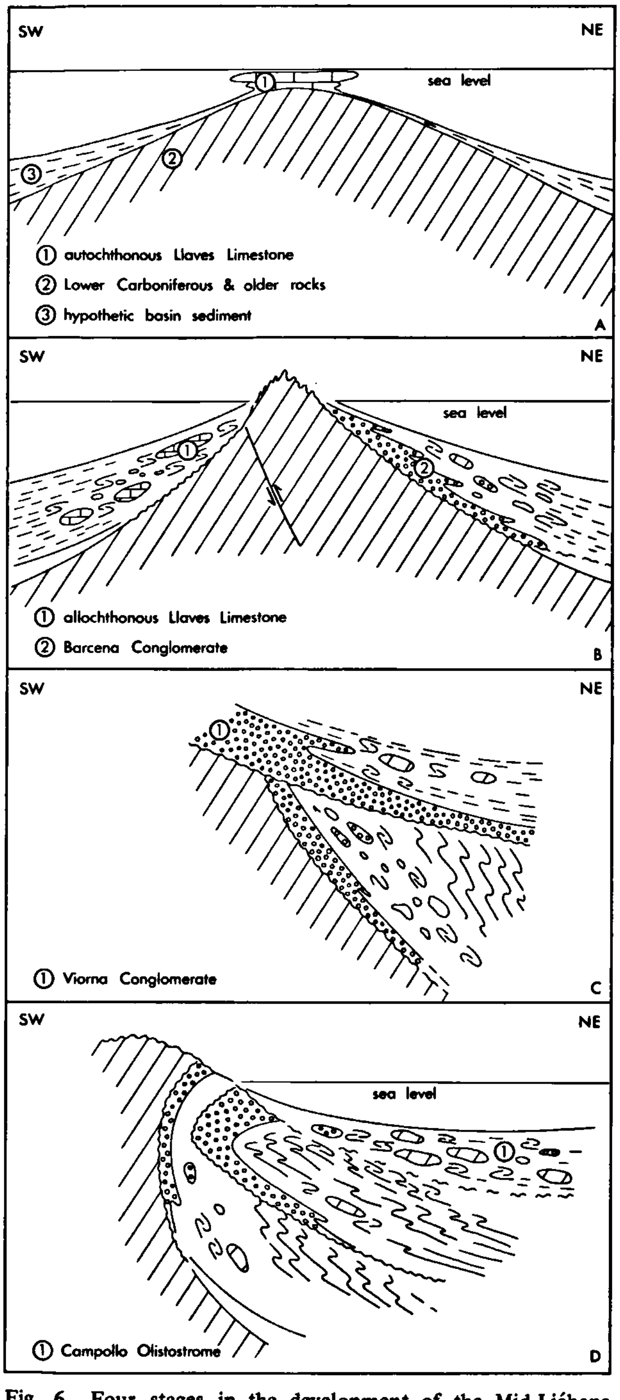


Fig. 6. Four stages in the development of the Mid-Liébana ridge.
A) before the Barcena unconformity.
B) shortly after the Barcena unconformity.
C) shortly after the Viorna unconformity.
D) after the deposition of the Campollo Olistostrome.

strome which has at that place a high-angle-unconformable relation to the extremities of the respective parts of the ridge (Encl. IV). It is clear that the structural event which caused the division of the ridge, must have preceded the Barcena unconformity. The discontinuity of the ridge is most likely the result of a separation of the respective parts by a roughly E–W striking dextral tear fault. The monoclinal SW part of the ridge is explained as a faulted anticline (Encl. II, section B–B^{I}). The bottom-to-bottom contact of Murcia Quartzites and the Enterrias Olistostrome N of Besoy is nowhere visible in outcrop, but must be interpreted as a fault, hidden by an almost contemporaneous unconformity. The mentioned tectonic activity along the ridge caused the unconformable deposition of the Barcena Conglomerate and the Enterrias Olistostrome (Fig. 6b). The two members are contemporaneous in some places and subsequent in other places. In addition to the basal unconformity, other unconformities of minor importance, 'intra-member unconformities', occur (cf. Photograph 12). Remnants of the Barcena Conglomerate constitute olistoliths in the Enterrias Olistostrome (Encl. IV, W of Sebrango and N of Enterrias). The Enterrias Olistostrome is not a homogeneous olistostrome mass but rather an alternation, with several minor unconformities, of pure olistostrome, slump-distorted turbidites and almost undisturbed turbidites. All such facts can be explained by assuming that the processes of relative uplift and deformation of the ridge were intermittently active over a longer period. In the allochthonous material of the unconformable deposits the stratigraphic succession is more or less inverted. The early Upper Carboniferous limestones (olistoliths of the Llaves Limestone and limestone pebbles in the Barcena Conglomerate) are predominant at the base of the unconformable sequence; olistoliths of Middle to Upper Devonian limestones and quartzitic sandstones occur predominantly in the younger part of this sequence. Such a succession would be expected in the case of a slow and rather continuous emergence and denudation of an anticlinal ridge. The occurrence of olistoliths of a Middle Devonian limestone (Detail map, Encl. IV, loc. X 85), which cannot be found outcropping on the present ridge, remains problematic. Later down-throw by faulting, of a hypothetic part of the ridge S of the Redondo fault, is suggested as an explanation. Such movements did occur at any rate a little to the west, as is indicated by the faulted inlier of Viorna equivalent near Soberado (Geol. map, Encl. I).

No equivalent of the Cabezon unconformity (N Liébana and Polaciones) is known from the Mid-Liébana ridge area. The next episode of tectonic activity and unconformable deposition (Fig. 6c) is marked by the observed unconformable Viorna Conglomerate and the assumed unconformable Sta. Eulalia Olistostrome. This episode is more or less a repetition of the former; intermittent uplift of the ridge over a longer period is reflected in the composition of the unconformable deposits. The angular relations between the Viorna Conglomerate and the older Carboniferous, indicate a steepening of the ridge flanks. Cascade folds, observed both S and N of the ridge along the Deva-road, must have been formed at that time. These structures indicate that the steepening of the flanks was caused by vertical bending instead of lateral compression. The origin of the limestone pebbles in the Viorna Conglomerate is still problematical. The map picture favours a source area on or near the ridge. It is possible that at the beginning of its second deformation episode, the ridge was again covered with biogenetic limestone banks which were in due course entirely removed by denudation.

The third definite episode of tectonical activity is marked by the, assumed unconformable, deposition of the Campollo Olistostrome and by the deformation into a cascade fold of the S flank of the Viorna syncline near Pico Jano (Encl. II, section C^{V}–C^{VI}; Encl. IV and Fig. 6d). The collapse and slump structures NW of Toranzo (Fig. 7) must have originated at the same time. A slip sheet of Murcia Sandstone, located at the SW side of the ridge 2 km N of Llaves, (shown as faulted in Encl. I), is also correlated with this last deformation episode. The distinction between this slip sheet mass as an early tectonic phenomenon and certain equally-sized olistoliths of the Campollo Olistostrome as sedimentary phenomena is disputable. In an early orogenetic environment as that of the Mid-Liébana ridge, the distinction between early tectonic transport by gravitational gliding and 'en masse' sedimentary transport in an olistostrome, becomes rather artificial. Similar problems arise when a distinction is made between sedimentary slump folds and tectonic collapse folds in such an environment (cf. Fig. 7).

The Campollo Olistostrome has remained in a subhorizontal position, indicating that the main deformation of the ridge had come to an end. This last deformation episode cannot be dated accurately, but took place sometime between the deposition of the Viorna Conglomerate and the first shaping of nappes in the Picos de Europa area. The tectonic contact between the nappes and the ridge (Geol. map, Encl. I) is such that any later folding seems unlikely to have affected the ridge. The intrusion of the syenitic rocks of Pico Jano (Boehmer, 1965) seems related to a fault system which cannot be dated accurately but which must be younger. Mesoscopic folds with an axial plane, crenulation cleavage and an older slaty cleavage as described from the N Liébana-Polaciones subarea, were also found in the shaly component of well-layered, graywacke-shale associations on the flanks of the ridge. The dating of these folds presents problems similar to those in the former subarea.

Deformation types and processes. – Here, only the concurrence of sedimentary slumps and slides and early tectonic collapse and gliding will be treated. Other deformation types have been discussed sufficiently in the foregoing descriptions of subareas.

One of the most beautiful exposures of the subject area is situated in the steep upper part of an eastfacing hill flank, 1750 m N of the village Enterrias (not to be confused with Enterria) (Fig. 7). In that exposure, all kinds of slump structures were observed: pull-apart structures, imbricated push-structures, slabs of coherent turbidite sediment in a severely contorted matrix of the same origin, isolated slump fold hinges, etc. Typical slump complexes alternate with more coherent turbidite complexes in which well-developed collapse folds were observed (Fig. 7). The contact between such alternatively deformed complexes is generally gradual, but the top boundary of a collapse-folded complex may be quite sharp (Fig. 7). From observations in adjacent exposures, the impression was obtained that the one complex may pass laterally into the other. The definition of a slump structure (Chapter III) included fixation of the deformation to one sedimentary level, bounded vertically by undeformed beds and with an erosional contact at the top. Here, there is not one level of deformed sediment, but in a sequence of more than 200 m thickness, slumped strata and strata folded into fairly large collapse folds, alternate vertically and probably also laterally. The contacts at the top of these folds may be quite sharp, but erosion cannot be proved; offshearing by a downmoving slump mass seems more probable. The exposure is situated somewhat below the E extension of the axial plane of the collapse fold in the Viorna Conglomerate on the E flank of Pico Jano. The mass transport, reflected in the slump and collapse structures of the exposures, seems to be related to the Viorna collapse. Transport directions and the overall younging direction are the same, although the stratigraphic level is slightly

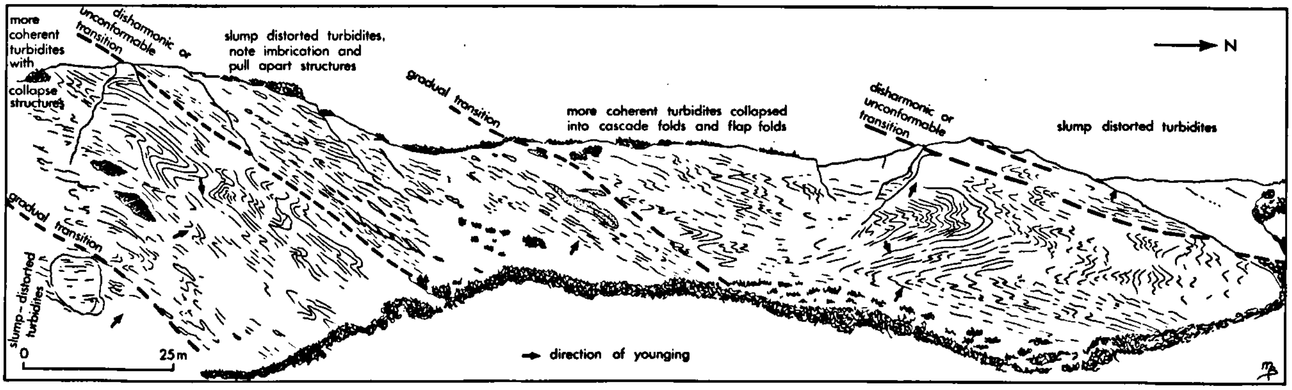


Fig. 7. Exposure of collapse and slump structures, NW of Toranzo, drawn after a panoramic series of overlapping photographs.

higher. It is suggested that one and the same tilting of the NE flank of the Mid-Liébana ridge caused slumping of the then most superficial turbidite sediments, and cascade folding of the lower lying probably already lithified Viorna Conglomerate. Somewhere between the most superficial layers and the Viorna Conglomerate, a transition between sedimentary mass transport and tectonical gravitational collapse must have occurred, just as seems to be shown by the exposure depicted in Fig. 7.

Similar exposures of fairly large collapse folds which seem to be surrounded by a slumped mass are known from the Bedoya Olistostrome Member; however, there slump deformation dominates, and no vertical transition to tectonic gravitational collapse can be observed or inferred. A concurrence of collapse folding and gravitational sedimentary transport of rock material is known from the collapse structures in the Asmari Limestone, SW Iran (Harrison & Falcon, 1936, fig. 4). However, the structures in Liébana are radically different. In SW Iran, collapse folding, degradation of the rock and gravitational sedimentation were subaerial processes; the involved rock complex was entirely lithified long before the collapse folding started. The collapse is not caused by active tilting but is a consequence of loss of support by erosion for layers which were tilted long before. Along the Mid-Liébana ridge, sedimentation, collapse and partial resedimentation constitute a continuous sequence of events. The upper part of the sediment was not yet lithified; the collapse and resedimentation processes were probably entirely subaquatic.

Southern Liébana (sections A, B, C, D, and E (Encl. II))

General. – The Southern Liébana structural subarea is bounded in the N by the Mid-Liébana ridge and the Redondo fault. The S boundary consists of a gradual passage to SE and SW into the structures of N Pisuerga and the N Rio Yuso area, respectively; in the centre there exists a well-defined, partly faulted boundary with the Polentinos block (Encl. I and Fig. 2). The lithological contrast with N Liébana and Polaciones, the occurrence of several competent conglomerates and limestones, has caused a marked structural contrast, too. Southern Liébana is characterized by a number of major bending structures. The occurrence of collapse structures is restricted to the flanks of the bending folds, where subsidiary cascade folds have developed. In these respects the subarea compares closely with the N Rio Yuso area (Savage, 1967). The following major structural units are distinguished (Fig. 2): the Pembes syncline, the Remoña syncline, the Deva anticline, the Coriscao syncline, the syncline between Barrio and Dobres, the Vallines-Ledantes anticline, the half syncline of Naranco river and the Palanca syncline. In some cases, the structures are more adequately described as synclinoria and anticlinoria. Major faults often separate these structural units. The nature of these faults is not clear; they may coincide with sedimentary facies boundaries and the fault movements may have initiated the bending processes.

Structural history. – Several unconformities mark the structural development of Southern Liébana. The sequence of these unconformities, their laterally restricted extent and generally subparallel nature have already been treated in the description of the stratigraphy. A detailed account of the sequence of events will not be repeated here. The structural history, as far as known from the stratigraphic record, can be summarized as follows. Repeated uplift and erosion of the Polentinos block is recorded in the conglomerates of the Cervera Formation (Chapter II: the Cervera Fm. in the Ledantes area), in the older horizons of the Curavacas Conglomerate with the one or two associated unconformities, and in the younger horizons of the Curavacas Conglomerate and the equivalent olistostrome units with the associated observed or inferred unconformities. The Mid-Liébana ridge acted as a comparable tectonically active feature; however, in S Liébana its influence is much less evident in the stratigraphy. The larger bending structures are supposed to have developed gradually in a way analogous to the structural development of the Mid-Liébana ridge. The structures were not studied in such detail as would allow substantiation of the semi-continuous character of their deformation equally well. However, some of the unconformities tend to increase in angularity toward the boundaries of the different structural units, where these units are separated by major faults. Such a tendency would be expected in the case of fault-activated bending. Semi-continuity of deformation seems warranted where the stratigraphic record shows a sequence of subparallel to low-angle unconformities and where olistostrome deposition has occurred repeatedly. In some structures, however, such aspects were not observed. The Remoña syncline, for instance, suggests a more short-lived deformation; the unconformity at the base of the Valdeon Formation must have a strongly angular relation to this structure.

The occurrence of mesoscopic buckle folds with a subvertical axial plane was observed, imposed upon all the larger structural units. The two cleavage systems that go with those folds are the same as in the other subareas: more than one subvertical crenulation cleavage may occur. In some places these buckle folds clearly show an overprint relation to older cascade folds and bending folds: in the overturned N limb of the Palanca syncline, near the source of the Arroyo de Entreovejas, a thin conglomerate lens is refolded into an antiform which has two inverted limbs (Encl. II, section D^{IV}–D^{V} and corresponding place on the Geological Map); in the NNW–SSE striking N flank of the Pembes syncline, several mesoscopic folds with a well-developed axial plane crenulation cleavage show a WSW–ENE orientation (folds observed along a track S of Treviño). In the N Rio Yuso area, the orientation of the subvertical crenulation cleavages is reported to be parallel to the larger faults (Savage, 1967). This conclusion cannot be substantiated for the Liébana area due to lack of data.

Deformation types and processes. – The idea that the

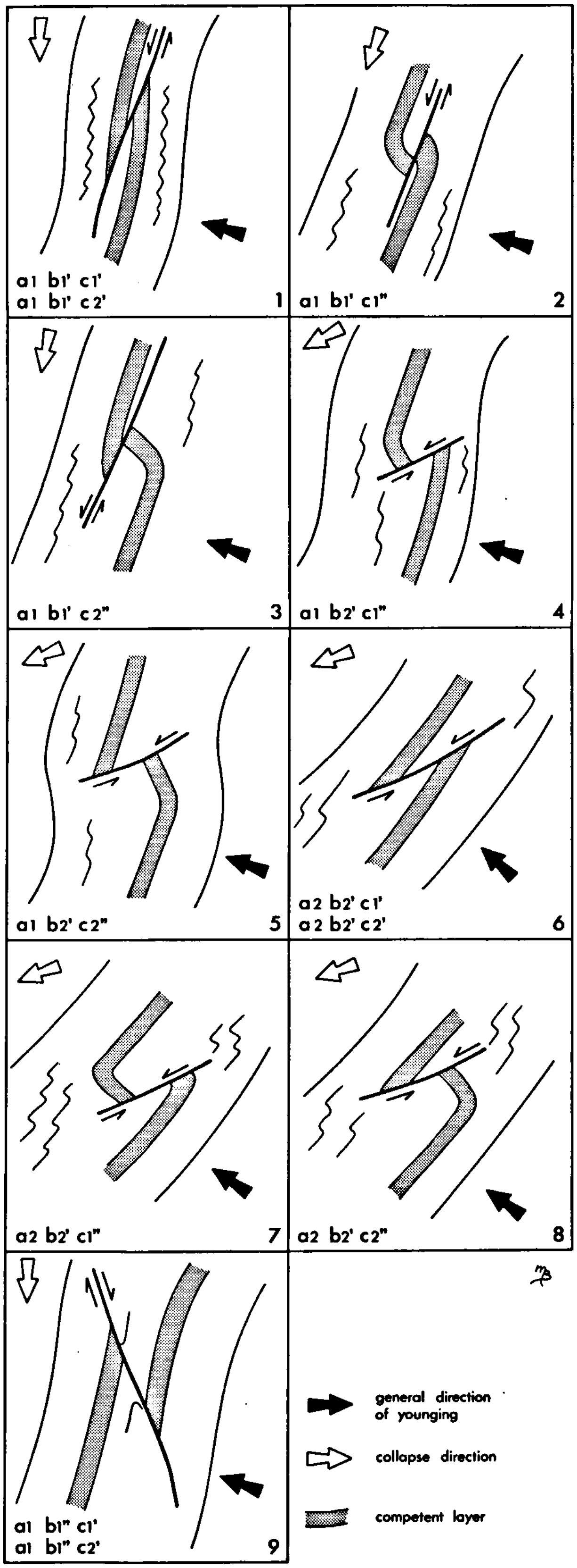


Fig. 8. Different types of collapse fault structures.

larger structures have been generated by bending is hypothetic. Arguments for this idea have been advanced in the section above, as well as in the introduction of this chapter. Summarizing and completing these arguments, the following considerations are mentioned: the occurrence of low angle unconformities and of major faults which seemingly control the structures, suggests vertical tilting instead of lateral compression. The occurrence of subsidiary collapse folds on the flanks of the major structures is incompatible with lateral compression. The almost circular shape in cross-section, of certain synclines, and diapirical shape of certain anticlines, a result of overturning of the upper part of fold limbs, is hard to explain without vertical movements (Encl. II, sections A^{II}–A^{III}, B–B^{I}, D^{IV}–D^{V}, etc.).

Mesoscopic collapse folds, subsidiary to larger host-structures have been treated thoroughly in the tectonical analysis of the N Rio Yuso area (Savage, 1967). The collapse folds observed in the graywacke-shale suites in S Liébana are in no respect different from those of the adjacent area. However, another type of collapse was observed in the more competent rock units. This type of collapse is effected by downthrow due to faulting of part of a bending fold limb after its rotation to an inclined, often nearly vertical, position. Such collapse faulting may result into a local doubling of vertically orientated competent rock units (Encl. I, the Curavacas Conglomerate on Peña Orcadas N of Puerto San Glorio), but more often the fault is developed in the core of an open cascade fold (Encl. II, sections C–C^{I}, D^{IV}–D^{V}, E–E^{I}, etc.). This association of a normal fault with the cascade folding of a competent rock unit may give the fault the appearance of a tilted thrust fault, and sets the fault type apart from the general type of normal faulting. Such collapse by fault movements, from now on to be referred to as collapse faulting, may result into several structural types. The result is controlled by the following factors: a) the original dip of the faulted unit, which may be nearly vertical (a 1), or more gently inclined (a2); b) the orientation of the fault which may be steep (b1), or gentle (b2), dipping in the same direction as the slope of the host fold (b′), or reverse (b″); c) the collapse fold preceding the fault may be synclinal (c1), or anticlinal (c2), tight (c′) or open (c″). Several combinations of these controlling factors are possible. Twelve combinations offer a theoretically possible collapse structure (Fig. 8). Of these twelve combinations, three pairs give identical results; this happens when the fault is equivalent to the middle limb of a tight fold. So, nine different types remain. Of course, all kinds of transitional types are also possible. Most of the collapse fault types represented occur in Liébana, but this kind of deformation is also known from other areas (Savage, 1967, Encl. I, cross-sections; van den Bosch, 1969, p. 195). Collapse fault structures as represented in Figs. 8,6; 8,7 and 8,8, may develop into slip sheets or even into nappes, provided that the host structure is sufficiently large (geanticlinal size), and an adequate detachment horizon is present. The collapse fault type of Fig. 8,9 also deserves attention. This type of fault may induce drag folds which have the misleading appearance of buckle folds caused by lateral compression; the direction of collapse is however vertical and the extension is in the horizontal plane. This kind of collapse faulting may result in an apparent thrust structure with apparently lateral shortening and vertical extension.

Marked disharmonies in the fold structures have been caused by the lithological contrasts existing in this subarea. The disharmonies are generally observed where folds in competent layers as limestones and conglomerates are compared to the folds in the adjacent incompetent graywacke-shale suites. A specific kind of such disharmony may exist between folds in competent layers situated on different levels in a structure where these layers are separated by incompetent layers. Compare for instance the deformation of the Panda Limestone and of the stratigraphically lower Curavacas Conglomerate in the Coriscao syncline. A third kind of fold disharmony may exist where a competent layer has a stratigraphically controlled, lateral discontinuity in extent and/or thickness and/or composition. Such 'lateral fold disharmonies' can be observed in the Curavacas Conglomerate NW of Puerto San Glorio. There, the stratigraphic discontinuity of the conglomerate has caused a rather complicated imbricated structure of three faulted synclines with open collapse folds and well-developed collapse fault structures in the conglomerate. The structural complication is laterally restricted to the stratigraphically discontinuous part of the Curavacas Conglomerate.

North Pisuerga

The larger structures occurring in this subarea, the Casavegas syncline, the faulted Los Llazos anticline and the Redondo syncline were already described by de Sitter and Boschma (1966). They considered the structures as a result of both subsidence (bending) and lateral compression. Since I did not personally investigate this subarea, only a few tentative statements are made here. A beginning of the NW–SE to N–S trend of the structures in N Pisuerga was also observed in S Polaciones and in the Comunidad de Campo y Cabuerniga area. In all those areas this trend is overprinted by roughly E–W orientated buckle folds. This is well demonstrated in the Casavegas syncline by the folds in the Maldrigo Limestone. So, the NW–SE trend is the older trend. This trend was probably caused by the relative position of the Polentinos block. The overturning of the NE flank of the Redondo syncline is considered as a gravitational collapse. The larger structures show more or less the same characteristics as the fault-controlled bending folds of Southern Liébana.

The Permo-Mesozoic cover

This subarea contains the rocks that are younger than the main deformation of the Cantabrian Mountains, the Hercynian Orogeny. Structurally, this subarea is characterized by faulted open synclines and anticlines with generally gently dipping flanks. It is remarkable that 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. The folds seem to be the result of differential tilting of fault-separated blocks. No true fold hinge was observed or mapped; the flexure N of the Pico Tres Mares (E of Piedrasluengas) comes nearest to true fold deformation and this flexure passes laterally into a fault. It is suggested that such 'folds' must have existed in the Liébana area in the initial stages of the bending process and only their position relative to sea level was strikingly different.

The unconformable contact of the basal conglomerate of the Nansa unit with the underlying Labra Formation is such, that besides uplift and erosion, a gentle open folding of the Labra Formation must have taken place before the deposition of the Nansa unit. A continuation of the deformation of the Permo-Mesozoic cover into the underlying older Palaeozoic rocks could nowhere be ascertained. However, such a continuation is likely since the degree of deformation in the Permo-Mesozoic cover diminishes rapidly with increasing distance from the Palaeozoic core of the Cantabrian Mountains.

CHAPTER V

CONCLUSIONS

THE SEQUENCE OF DEFORMATION PROCESSES

The different deformation types, postulated in the introduction of the former chapter, have been treated in the description of the structural subareas. Problems concerning the relations of these types to each other and their place in the general structural development were discussed. Most of the solutions offered had to be hypothetical. Still, the following general ideas emerged.

The deformation types and the sequence of deformation processes are of a local nature, being to a large extent related to the lithology and the depositional history, respectively. Three main structural subareas, each with its own deformational style and history, are distinguished in the subject area: the Picos de Europa (a former shelf area), N Liébana and Polaciones (a former flysch basin), and the transitional area of S Liébana, N Rio Yuso and N Pisuerga (a transitional area both concerning depositional and deformational history). A sharp contrast in structures and structural history exists be-

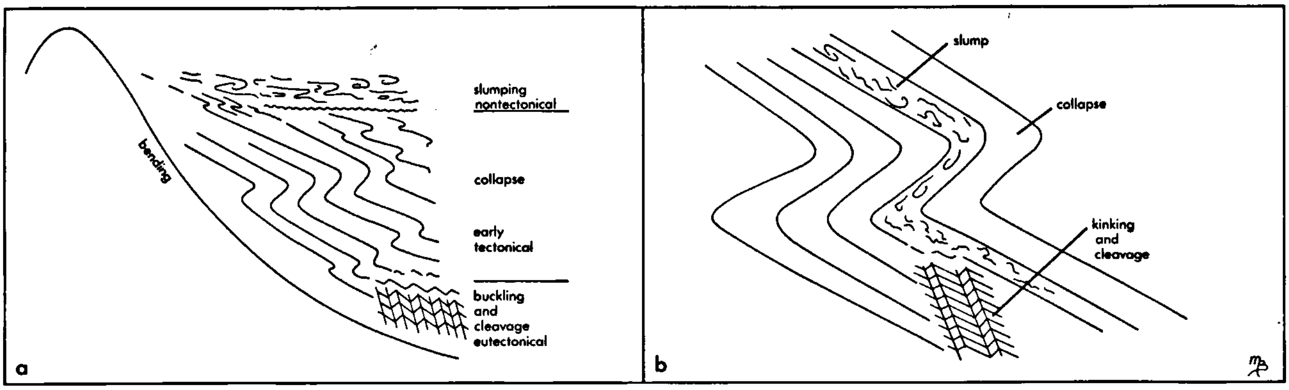


Fig. 9. Concurrence in time of the three main deformation types, while separated in space (a). Concurrence in space of the three main deformation types, while separated in time (b).

tween the Picos de Europa and the other subareas; the differences between the remaining subareas are of a gradual character. For the latter two subareas, a more or less uniform sequence of deformation processes can be established. The deformation processes in those subareas seem to be related to the degree of lithification and the distance to the palaeosurface of the rock that is being deformed. The following generalized sequence of deformation is proposed:

1) Nontectonic sedimentary deformation (slumping). The deformation processes occurred on or very near the sediment surface, affecting rocks not yet or only partly lithified. Folding occurs mainly on a mesoscopic scale. The deformation processes were short-lived. No general stress field existed, since compression and tension structures are concurrent.

2) Early tectonic synsedimentary deformation (bending and collapse). The deformation processes occurred at, or not too far from, the surface in lithified rock. Bending constitutes an exception to this rule, since this process reached further down into the crust. Folding occurs both mesoscopic (collapse) and macroscopic (collapse and bending). The deformation processes, notably the bending, may be traced over a geologically considerable length of time. The dominant forces were due to a vertically orientated gravitational stress field.

3) Eutectonic post- or infrasedimentary deformation (kinking, buckling and cleavage). The deformation processes are considered to have occurred far below the rock surface and far from the sedimentary environment in well-lithified rock. The deformation may be accompanied by recrystallization phenomena. Folding occurs on all scales. The duration of these processes could not be inferred. This type of deformation normally overprints the other types and is due to a laterally compressive stress field.

Nontectonic deformation is generally related to palaeotopographic features; the processes may be triggered by tectonic activity, primarily fault movements. Collapse processes have proved to be conform to palaeoslopes in several cases. Bending is considered to have been generated by such crustal processes as are also responsible for the formation of the larger topographic features. A relation between the eutectonic processes and elements of palaeotopography could not be established. The proposed sequence is mainly based on observation of deformation types in the graywacke-shale suites; it remains a generalization to which of course numerous exceptions exist. Nevertheless, the sequence can be observed or inferred for the larger part of the subject area. The sequence must be considered as existing in space (contemporaneous processes on different levels, Fig. 9a) as well as in time (subsequent processes in the same locality, Fig. 9b). However, the existence in space of the sequence with respect to the contemporaneity of early tectonic and eutectonic processes, is not yet established beyond doubt. The validity of this concept of a deformation sequence is supported by the occurrence of deformation types that seem transitional between nontectonic and early tectonic deformation (e.g. Fig. 7) and of deformation types which combine characteristics of the early tectonic and the eutectonic types (e.g. Photographs 52 and 53). A sequence as proposed should contain a gradual passage from one deformation type to the next, as well as a gradual passage from one extreme to the other within a deformation type (e.g. the different types of slump deformation from the N Liébana-Polaciones subarea).

THE 'OROGENETIC PHASE' CONCEPT

We will now turn our attention to other existing ideas concerning the sequence of sedimentation and deformation during orogenesis. Many investigators confronted with this problem have accepted, with or without modification, the concept of an 'orogenetic phase' as formulated by Stille (1924). So, a short review and discussion of this concept is necessary before continuing our conclusions.

The concept of an 'orogenetic phase' forms a part of the 'orogenetic time-law' ('orogenes Zeitgesetz') enunciated by Stille in his 'Grundfragen der vergleichenden Tektonik' (1924). This 'law' is formulated as follows: "All orogenesis is fixed to relatively few phases which are short-lived and have an approximately worldwide significance". ("Alle Gebirgsbildung ist an verhältnismäszig wenige und zeitlich engbegrenzte Phasen von ± erdweiter Bedeutung gebunden"; Stille, 1924, p. 44). The next statement is added to this 'law' as a second 'law': "It (the orogenesis, K. M.) manifests itself at the same time in the most different regions of the earth" ("Sie tritt gleichzeitig in den verschiedensten Erdgebieten auf"). This statement is presented as the 'orogenetic contemporaneity law' ('orogenes Gleichzeitigkeitsgesetz'). Following these 'laws', Stille distinguishes 'separate phases' ('Einzelphasen') as subdivisions of orogenies. The 'separate phases' are also mentioned as 'folding phases' ('Faltungsphasen'); 'mainphases' ('Hauptphasen') and 'subphases' are distinguished. Several subphases may constitute one mainphase. Stille's phase concept is based on a "comparative method of age determination for orogenetic events" by which is meant a determination of the age of a folding process based on the difference in age of the youngest rocks affected by the folding and the oldest rocks that have an unconformable contact with the folded rocks. Using this method, a record of relative ages of succeeding deformation processes is made for the different orogenies. All tectonic events that appear to have occurred more or less contemporaneous, are considered to belong to one and the same phase, no matter geographic distances, differences in type, style and intensity of the deformation, and differences in place within the different local records of tectonic events. All tectonic events that took place within the time span comprising approximately the Carboniferous and Permian Periods (about 120 million years), are considered to constitute the Hercynian or Variscian Orogenies, following the theoretical frame given above. The Bretonnic, Sudetic, Asturic and Saalic mainphases and the less important Pfälzic phase constitute the subdivision of the Hercynian Orogeny according to Stille (1924, pp. 126–131). The occurrence of the main phases is considered to have been worldwide; for instance, early Permian folding in central and east U.S.A. as well as in E Australia are considered to belong to the Saalic phase. Stille introduced, furthermore, as a consequence of the phase concept, a rigid distinction between epeirogenetic processes (e.g. subsidence) and orogenetic processes (e.g. folding), stating that the one cannot occur simultaneously with the other. The orogenetic processes are considered to be fixed to the short-lived phases in between which only epeirogenetic processes are thought possible.

OBJECTIONS TO THE CONCEPT OF AN OROGENETIC PHASE

It will be clear that many objections can be made to the above reviewed phase concept, on the base of the investigations reported in this thesis. The following objections are made to the following statements of the phase concept.

1) Orogenetic deformation is fixed to short-lived phases.
Objection: the duration of the formation of larger structures has been determined in some cases. It appears that the time for deformation is comparable to the duration of sedimentation processes. The duration of nappe formation in the Picos de Europa included the interval needed for deposition of the Lebeña Formation. The duration of formation of the Mid-Liébana ridge structure included the interval needed for deposition of the sedimentary sequence from the base of the Barcena Conglomerate to the base of the Campollo Olistostrome. A diachronous character was inferred for the formation of large collapse structures in N Liébana and Polaciones, since these structures are truncated from E to W by the in-time subsequent Cabezon and Viorna unconformities. Hence, a deformation process, instead of being fixed to a short-lived phase, may last for a considerable time span.
2) The phase is worldwide in significance.
Objection: The significance of a deformation process can be proved to be essentially local or regional. This will be clear when we compare the sequence in time and the lateral extent of deformation processes from Liébana, with the sequence in time and the lateral extent of deformation processes from the Picos de Europa. The laterally discontinuous character of many unconformities in the subject area (changes from angular to parallel to non-existent) stresses the fact that deformation processes are restricted in space. Contemporaneity alone of deformation processes from geographically separate areas may be a mere coincidence and no adequate reason for the establishment of an orogenetic phase. A tectonic event should be studied in the context of the local (or regional) geological history. The beginning of a sequence of deformation processes somewhere in Europe may be contemporaneous with the final stages of a deformation sequence in some other part of the world. The grouping together of such events in one orogenetic phase makes no sense in terms of understanding orogenic processes.
3) The comparative method of age determination for orogenetic phases.
Objection: Stille's comparative method of age determination is based on the comparison of lacunas in the chronostratigraphical records from different areas; yet such lacunas may show a partial overlap. The dating of concurrent deformation processes as having occurred during the period of the overlap is tempting but not justified. Moreover, the top boundaries of such lacunas, being marked by unconformities, have a diachronous character.
4) The phases of the Hercynian Orogeny.
Objection: Classifying the sequence of tectonic events inferred for the different parts of the subject area into the formal phases of the Hercynian Orogeny would result into a strained, artificial and irrational subdivision which would in no way lead to a better understanding of the geological history of the subject area.
5) The distinction between orogenetic and epeirogenetic processes.

Objection: The rigid distinction between orogenetic and epeirogenetic processes does not agree with the results of the investigations in the Liébana area. Epeirogenetic processes as uplift and subsidence were active throughout the Hercynian Orogeny in this part of the Cantabrian Mountains. Collapse folding subsidiary to bending may be considered as an orogenetic process contemporaneous with, and subsidiary to, an epeirogenetic process.

AN ALTERNATIVE APPROACH TO THE PROBLEM OF OROGENESIS

The offered generalized sequence of deformation processes does not explain the relationship of sedimentation and deformation in terms of cause and effect. However, the results of our investigations provide some reason to assume that processes of deposition and deformation within an orogenesis are linked together as expressions of a fundamental crustal process, controlling both. To elucidate this statement, the following reflections are brought to the attention of the reader.

The geological history of the Cantabrian Mountains, during the older Palaeozoic was marked by the existence of certain fairly large facies regions. Conditions as rate of sedimentation and deformation were as a rule uniform within these regions. The sedimentation rate was moderate to low in the Asturide-Leonide region and low to very low in the Palentine region, whilst sedimentation was largely absent in the Cantabrian block region. The deformation in these regions during that time consisted of slow and continuous processes of uplift and subsidence. The rates of both sedimentation and deformation increased enormously during the Upper Carboniferous, the time of the Hercynian Orogeny. Local sedimentation rates may have increased to ten or even twenty times those in the previous period. On the other hand, strong erosion took place locally. The deformation was no longer confined to epeirogenetic processes; orogenetic processes became of major importance, whilst epeirogenetic processes were drastically intensified. The patterns of sedimentation and deformation were no longer consistent on a regional scale. Facies changes were abrupt and numerous in time as well as in space. The sequence of deformation processes changed from place to place. The rates of sedimentation and deformation seem to have decreased during Permian and Triassic to a level comparable to that before the Hercynian Orogeny. So, in the course of geological history, the rates of sedimentation and deformation increased and slackened more or less simultaneously, reaching a peak during a period of orogenesis. An important change takes place in the palaeogeographic configuration before and after the orogenesis (cf. Rupke, 1965, p. 69, fig. 33). The different facies regions that could be distinguished before the Hercynian Orogeny, remained basic elements of the facies distribution up to Upper Moscovian times (Chapter III of this thesis) when the orogenesis was already well on its way. These basic elements have altogether disappeared when sedimentation was resumed in post-Hercynian times. The orogenesis generated a fundamentally different palaeogeographic configuration.

So, a period of orogenesis is marked by a simultaneous increase and subsequent decrease of sedimentation and deformation as well as by a slow but irreversible fundamental change in palaeogeographic configuration. Such facts can be explained as part of a slow but continuous crustal evolution which is accelerated for a certain period. They fit well into the new concept of crustal plates in slow but continuous movement. The continents constitute only a minor part of these plates so that coincidence of a continent margin and a destructive plate margin is a matter of chance which will be realized only now and then in the course of geological time, manifesting itself as an orogenesis. The fundamental process underlying the orogenesis has a continuous nature so that the orogenetic sequence of deformation processes is likely to have a continuous nature also.

REFERENCES

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

Al Katrib, S. O., 1971. Photogeological study and structural analysis of Picos de Europa – Northern Spain. Unpublished M. Sc. thesis, I.T.C. Enschede.

Allen, J. R. L., 1970. The sequence of sedimentary structures in turbidites with special reference to dunes. Scott. J. of Geol., 6/2, pp. 146–161.

Barrois, Ch., 1882. Recherches sur les terrains anciens des Asturies et de la Galice. Mém. Soc. Géol. Nord, II (1), pp. 1–630.

Binnenkamp, J. G., 1965. Lower Devonian brachiopods and stratigraphy of North Palencia (Cantabrian Mountains, Spain). Leidse Geol. Med., 33, pp. 1–62.

Boehmer, W. K., 1965. De geologie van het Deva gebied. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Bosch, W. J. van den, 1969. Geology of the Luna-Sil region, Cantabrian Mountains (NW Spain). Leidse Geol. Med., 44, pp. 137–225.

–, 1969a. The relationship between orogenesis and sedimentation in the SW part of the Cantabrian Mountains (NW Spain). Leidse Geol. Med., 44, pp. 227–233.

Boschma, D., 1968. Provisional geological map of the Southern Cantabrian Mountains (Spain). Leidse Geol. Med., 43, pp. 217–220.

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

Bouma, A. H., 1962. Sedimentology of some flysch deposits. A graphic approach to facies interpretation. Elsevier, Amsterdam, pp. 1–168.

Breimer, A., 1956. De geologie van het gebied tussen Caloca en Tremaya. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Brouwer, A., 1964. Deux faciès dans le Devonien des Montagnes Cantabriques méridionales. Boletin del Instituto de Estudios Asturianos (Suplemento de Ciencias), 10, pp. 1–10.

Brouwer, A. & Ginkel, A. C. van, 1964. La succession carbonifère dans la partie méridionale des Montagnes Cantabriques (Espagne du Nord-Ouest). Compte rendu Ve Congr. Strat. Géol. Carbonifère, Paris 1963, pp. 307–319.

Comte, P., 1959. Recherches sur les terrains anciens de la Cordillère Cantabrique. Mem. Inst. Geol. y Min. Esp., Vol. 60, pp. 1–440.

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

Dennis, J. G., ed., 1967. International Tectonic Dictionary. A.A.P.G. publication. George Banta Cy. Inc., pp. 1–196.

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

Frets, D. C., 1965. The geology of the southern part of the Pisuerga basin and the adjacent area of Santibañes de Resoba, Palencia, Spain. Leidse Geol. Med.. 31, pp. 113–162.

Funk, H., 1971. Zur Stratigraphie und Lithologie des Helvetischen Kieselkalkes und der Altmannschichten in der Säntis-Churfirsten-Gruppe (Nordostschweiz). Eclogae geol. Helv., 64/2, pp. 345–443.

Germs, G. J. B., 1966. Stratigrafie van het gebied rondom Ledantes en de Kalkalgen van het Liébana gebied. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

–, 1966a. De geologie van het gebied rondom Ledantes. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Ginkel, A. C. van, 1959. The Casavegas section and its fusulinid fauna. Leidse Geol. Med., 24, pp. 705–720.

–, 1965. Spanish Carboniferous fusulinids and their significance for correlation purposes. Leidse Geol. Med., 34, pp. 172–225.

–, 1971. Fusulinids from uppermost Myachkovian and Kasimovian strata of NW Spain. Leidse Geol. Med., 47, pp. 115–161.

Graaff, W. J. E. van de, 1971. Three Upper Carboniferous, limestone-rich, high-destructive, delta systems with submarine fan deposits, Cantabrian Mountains, Spain. Leidse Geol. Med., 46, pp. 157–215.

–, 1972. The Piedrasluengas Limestone, a possible model of limestone facies distribution in the Carboniferous of the Cantabrian Mountains. In: The Carboniferous of Northwest Spain. Trabajos de Geologia, 3, Fac. de Ci. Univ. Oviedo.

Graaff-Trouwborst, T. van de, 1971. Een poging tot mathematische plooi analyse. (An attempt at mathematical fold analysis). Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Harrison, J. V. & Falcon, N. L., 1934. Collapse Structures. Geol. Mag., LXXI, pp. 529–539.

–, 1936. Gravity collapse structures and mountain ranges as exemplified in SW Iran. Quart. J. Geol. Soc., XCII/1, pp. 91–102.

Helwig, J., 1970. Slumpfolds and early structures, NE Newfoundland, Appalachians. Journ. Geol., 78, pp. 172–187.

Higgins, A. C., Wagner-Gentis, C. H. T. & Wagner, R. H., 1964. Basal Carboniferous strata in part of northern León, NW Spain; Stratigraphy, conodont and goniatite faunas. Bull. Soc. Belge de Géol. Pal. Hydrol., 72/2, pp. 205–248.

Hubert, C., Lajoie, J. & Léonard, M. A., 1970. Deepsea sediments in the Lower Palaeozoic Québec Supergroup. The Geol. Ass. of Canada, Special Paper, 7, 1970, pp. 103–125.

Jong, J. D. de, 1971. Molasse and clastic-wedge sediments of the Southern Cantabrian Mountains (NW Spain) as geomorphological and environmental indicators. Geol. en Mijnb., 50/3, pp. 399–416.

Julivert, M., 1961. Estudio geologico de la cuenca de Beleño, Valles altos del Sella, Ponga, Nalon y Esla, de la Cordillera Cantabrica. Bol. Inst. Geol. y Min. de España, Tomo LXXI, pp. 1–346.

–, 1971. Décollement tectonics in the Hercynian Cordillera of Northwest Spain. American J. Sc., 270, pp. 1–29.

Kamerling, P., 1962. De geologie van het Pandatrave gebied. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Kanis, J., 1956. Geology of the eastern zone of the Sierra del Brezo (Palencia, Spain). Leidse Geol. Med., 21, pp. 375–445.

Kelling, G. & Woollands, M. A., 1969. The stratigraphy and sedimentation of the Llandoverian rocks of the Rhayader district. In: Precambrian and Lower Palaeozoic rocks of Wales. Univ. Wales Press, pp. 225–282.

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

Kuenen, Ph. H., 1967. Emplacement of Flysch-type sandbeds. Sedimentology, 9, pp. 203–243.

Kuenen, Ph. H. & Migliorini, C. I., 1950. Turbidity currents as a cause of graded bedding. Journ. Geol., 58, pp. 91–127.

Kullmann, J., 1964. Las series Devónicas y del Carbonífero inferior con ammonoideos de la Cordillera Cantabrica. Est. Geol. de España, Vol. 19, pp. 161–191.

Kutterink, J. A., 1966. Geologie van het Valdeon gebied in Spanje. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Laird, M. G., 1968. Rotational slumps and slumpscars in Silurian rocks, W Ireland. Sedimentology, 10, pp. 111–120.

Lanting, R. W., 1966. De geologie van het Potès gebied. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Lebedeva, N. B., 1969. A model of a folded zone. Tectonophysics, 7/4, pp. 339–351.

Maas, K., 1968. Geologie van het Polaciones-Comunidad gebied, Tektoniek. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

–, 1968a. Geologie van het Polaciones-Comunidad gebied, Sedimentologie. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Marschalko, R., 1968. Term: Wildflysch. Geologické pracé, Spravy, Vol. 48, pp. 175–187.

Martinez Alvarez, J. A., 1965. Rasgos Geologicos de la zona oriental de Asturias. Inst. de Est. Ast., 1965.

Miedema, S. K., 1966. Geologie van het oostelijk Liébana gebied. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

–, 1966a. Stratigrafie en sedimentologie van het oostelijk Liébana gebied. Int. rept., Geol. Inst., Univ. Leiden.

Nagtegaal, P. J. C., 1969. Sedimentology, Palaeoclimatology and Diagenesis of post-Hercynian continental deposits, S. Pyrenees, Spain. Leidse Geol. Med., 42, pp. 143–238.

Nederlof, M. H., 1959. Structure and sedimentology of the Upper Carboniferous of the upper Pisuerga valleys, Cantabrian Mountains, Spain. Leidse Geol. Med., 24, pp. 603–703.

Nederlof, M. H. & Sitter, L. U. de, 1957. La cuenca carbonífera del rio Pisuerga (Palencia). Bol. Inst. Geol. Min. Esp., Vol. 68, pp. 1–44.

Nelson, H. F., Brown, C. W. & Brineman, J. H., 1962. Skeletal limestone classification. In: Classification of carbonate rocks – A Symposium, Ham, W. E. (ed.), Memoir 1 of A.A.P.G., pp. 224–252.

Papa, F., 1964. De Bunter tussen Cervera en Brañosera (N. Palencia). Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Rácz, L., 1965. Carboniferous calcareous algae and their association in the San Emiliano and Lois-Ciguera Formations (Prov. León, NW Spain). Leidse Geol. Med. 31, pp. 1–112.

Radig, F., 1962. Ordovizium/Silurium und die Frage prävariscischer Faltungen in Nordspanien. Geol. Rundschau, 52, pp. 346–357.

Reijd, H. J. van de, 1969. Stratigrafie en petrografie van de Muda Kalk. Int. rept., Geol. Inst., Univ. Leiden, the Netherlands.

Rickard, M. J., 1961. A note on cleavages in crenulated rocks. Geol. Mag., XCVIII/4, pp. 324–332.

Rittenhouse, G., 1943. A visual method of estimating two dimensional sphericity. J. Sediment. Petrol., 13, pp. 80–81.

Roehl, P. O., 1967. Stony Mountain (Ordovician) and Interlake (Silurian) faciès analogs of recent low-energy marine and subaerial carbonates, Bahamas. The A.A.P.G. Bull., 51/10, pp. 1979–2032.

Rupke, J., 1965. The Esla nappe, Cantabrian Mountains (Spain). Leidse Geol. Med., 32, pp. 1–74.

Savage, J. F., 1961. The structural geology of the area round Portilla de la Reina, León, Northwest Spain. Int. rept., Msc. Thesis, Univ. of London.

–, 1967. Tectonic analysis of Lechada and Curavacas synclines, Yuso basin, León, NW Spain. Leidse Geol. Med., 39, pp. 193–247.

Seilacher, A., 1967. Bathymetry of trace fossils. Marine Geology, 5, pp. 413–428.

Sitter, L. U. de, 1964. Structural Geology (2nd ed.). McGraw-Hill, New York, 551 p.

Sitter, L. U. de & Bosch, W. J. van den, 1968. The structures of the SW part of the Cantabrian Mountains. Leidse Geol. Med., 43, pp. 213–216.

Sitter, L. U. de & Boschma, D., 1966. Explanation geological map of the Palaeozoic of the Southern Cantabrian Mountains 1:50 000, sheet 1, Pisuerga. Leidse Geol. Med., 31, pp. 191–238.

Sjerp, N., 1967. The Geology of the San Isidro-Porma area (Cantabrian Mountains, Spain). Leidse Geol. Med., 39, pp. 55–128.

Smit, P. W., 1966. Sedimentpetrografie en milieu analyse van de Permo-Trias afzettingen langs de oostelijke rand van het Cantabrische gebergte (Spanje). Int. rept. Geol. Inst., Univ. Leiden, the Netherlands.

Smits, B. J., 1965. The Caldas Formation, a new Devonian unit in León (Spain). Leidse Geol. Med., 31, pp. 179–187.

Stille, H., 1924. Grundfragen der vergleichenden Tektonik. Borntraeger, Berlin, 443 p.

Stockmans, F. & Willière, Y., 1965. Documents paléobotaniques pour l'étude du houiller dans le Nord-Ouest de l'Espagne. Inst. Royal Sc. Nat. de Belgique, Mém. 2e série, Fasc. 79, pp. 1–106.

Teichert, C., 1965. Devonian rocks and palaeogeography of Central Arizona. U.S. Geol. Survey, Prof. Paper 464, pp. 1–181.

Thomson, A. F. & Thomasson, M. R., 1969. Shallow to deepwater facies development in the Dimple Limestone (Lower Pennsylvanian), Marathon Region, Texas. Soc. of Ec. Paleontologists and Mineralogists, Spec. Publ., No. 14, pp. 57–78.

Veen, J. van, 1965. The tectonic and stratigraphic history of the Cardaño area, Cantabrian Mountains, Northwest Spain. Leidse Geol. Med., 35, pp. 45–104.

Wagner, R. H., 1955. Rasgos estratigrafico-tectonicos del Paleozóico Superior de Barruelo (Palencia). Estudios Geologicos, XI, 26, pp. 145–202.

–, 1963. A general account of the Palaeozoic rocks between the rivers Porma and Bernesga (León, NW Spain). Bol. Inst. Geol. y Min. de Esp., Vol. 74, pp. 1–159.

–, 1966. Palaeobotanical dating of Upper Carboniferous folding phases in NW Spain. Mem. Inst. Geol. y Min. de Esp., Vol. 66, pp. 1–169.

–, 1970. An outline of the Carboniferous stratigraphy of Northwest Spain. In: Colloque sur la stratigraphie du Carbonifère. Congrès et Colloques Univ. Liège, Vol. 55, pp. 429–463.

–, 1972. Carboniferous nappestructures in Northeastern Palencia (Spain). In: The Carboniferous of Northwest Spain. Trabajos de Geologia, 3, Fac. de Ci. Univ. Oviedo.

Wagner, R. H. & Varker, W. J., 1972. The distribution and development of post-Leonian strata (Upper Westphalian D, Cantabrian and Stephanian A) in northern Palencia, Spain. In: The Carboniferous of Northwest Spain. Trabajos de Geologia, 3, Fac. de Ci. Univ. Oviedo.

Wagner, R. H. & Wagner-Gentis, C. H. T., 1963. Summary of the stratigraphy of Upper Palaeozoic rocks in NE Palencia, Spain. Kon. Ned. Akad. Wetenschappen, Proc., Series B, LXVI, pp. 149–163.

Wagner, R. H. & Winkler Prins, C. F., 1970. The stratigraphic succession, flora and fauna of Cantabrian and Stephanian A rocks at Barruelo (Prov. Palencia) NW Spain. In: Colloque sur la stratigraphie du Carbonifère. Congrès et Colloques Univ. Liège, Vol. 55, pp. 487–551.

Wagner, R. H., Winkler Prins, C. F. & Riding, R. E., 1972. Lithostratigraphic units in the lower part of the Carboniferous in Northern León, (Spain), with a 'Note on some goniatite faunas' by C. H. T. Wagner-Gentis. In: The Carboniferous of Northwest Spain. Trabajos de Geologia, 3, Fac. de Ci. Univ. Oviedo.

Winkler Prins, C. F., 1968. Carboniferous Productidina and Chonetidina of the Cantabrian Mountains (NW Spain). Systematics, stratigraphy and palaeogeology. Leidse Geol. Med., 43, pp. 41–126.

APPENDIX

List of fossil samples with the results of biostratigraphic determinations.

All samples are coded according to the localities given on the 1 : 100 000 fossil locality map (Encl. VII), or the localities on the 1 : 25 000 map of Enclosure IV. In the latter case, the map will be mentioned below. Samples and slides are coded identically.

Limestone samples containing fusulinids and incidentally algae. Fusulinids determined by Dr. A. C. van Ginkel (Geological Institute, Rijksuniversiteit Leiden, the Netherlands), algae determined by Dr. J. J. de Meyer (Shell, Rijswijk). Samples collected by the author.

Polaciones Area

66P14	*Fusulinella* Zone subzone B subdivision B 1 or B 2
65B9	*Fusulinella* Zone subzone B subdivision B 2/B 3
66H3	*Profusulinella* Zone subzone B
66H5	*Fusulinella* Zone subzone A
66H10	Probably *Profusulinella* Zone subzone B, possibly younger (fusulinids and algae)
66G6	*Profusulinella* Zone probably subzone A (primitive unknown *Profusulinella* sp. Algae: *Beresella* sp.)
66G16	*Profusulinella* Zone subzone B

Picos de Europa Area

70183	*Profusulinella* Zone subzone B
70186	*Fusulinella* Zone subzone B subdivision B 2
70191	*Fusulinella* Zone subzone B subdivision B 1
70192	*Fusulinella* Zone subzone A
70202	Locality: eastside of the Potes–Unquera road, some 100 m south of the first occurrence of Lebeña shales, going from S to N. *Fusulinella* Zone subzone B subdivision B 2 or B 3
70202A	Locality 70202, *Fusulinella* Zone subzone B subdivision B 2 or B 3
70202B	Locality 70202, *Fusulinella* Zone subzone B subdivision B 2 or B 3
70205	Locality: some 40 m north of 70202 at the same side of the road. *Fusulinella* Zone subzone B subdivision B 2 or B 3
70206	*Fusulinella* Zone probably subzone A
70207	Locality: some 40 m north of 70206. This sample comprises the slides 70207, 70207A and 70207B. *Fusulinella* Zone subzone A
70208	This sample comprises the slides 70208, 70208A and 70208B. *Fusulinella* Zone subzone B subdivision B 1, less probable subzone A
70210	Locality: some 200 m north of 70208. No fusulinids. Algae: *Komia abundans, Macroporella* Ginkeli. Approximation of fusulinid biozone: *Fusulinella* Zone subzone B subdivision B 1 (J. J. de Meijer).
24/6/70I	*Fusulinella* Zone subzone A
70A1,1	This sample comprises the slides A1,1; A1,2; A1,3; A1,4. *Fusulinella* Zone subzone A
70A1,5	*Fusulinella* Zone subzone B subdivision B 1
70215	*Millerella* Zone
70259	*Fusulinella* Zone subzone A or *Profusulinella* Zone subzone B
70263	*Fusulinella* Zone subzone B subdivision B 1
70271	Locality: on the Tabla de Lechugales. *Fusulinella* Zone subzone B subdivision B 3 or *Protriticites* Zone
70273	*Fusulinella* Zone subzone A
70275	*Fusulinella* Zone subzone A
70277	*Millerella* Zone
70278	*Fusulinella* Zone subzone B subdivision B 1
70287	*Fusulinella* Zone subzone B
70295	*Fusulinella* Zone subzone A
70322	*Fusulinella* Zone subzone B subdivision B 2 or B 3
7177	Probably *Millerella* Zone
71144	*Fusulinella* Zone subzone B subdivision B 2
71146	*Profusulinella* Zone subzone A
71148	*Fusulinella* Zone subzone B subdivision B 2
71149	*Profusulinella* Zone subzone A
70SC	Sample of the San Carlos Limestone Conglomerate. Fusulinids from different pebbles belong to different biozones

Millerella Zone:	1 pebble
Profusulinella Zone:	16 pebbles, of
which no further distinction:	6 pebbles
Profusulinella Zone subzone A:	6 pebbles and
Profusulinella Zone subzone B:	4 pebbles
Fusulinella Zone:	6 pebbles, of

	which no further distinction:	3 pebbles
	Fusulinella Zone subzone A:	1 pebble
	Fusulinella Zone subzone B subdivision B 1:	2 pebbles
71SC212	*Fusulinella* Zone subzone B subdivision B 2 or B 3:	1 pebble

Liébana Area

7060	*Profusulinella* Zone subzone A
7080	*Profusulinella* Zone subzone A
70128	*Profusulinella* Zone subzone A
70139	*Profusulinella* Zone subzone A
70163	*Profusulinella* Zone subzone A, possibly *Millerella* Zone
70290	*Fusulinella* Zone subzone A or subzone B subdivision B 1
70291	*Profusulinella* Zone subzone B
70367	*Profusulinella* Zone subzone B
70369	*Profusulinella* Zone subzone B
70371	*Profusulinella* Zone subzone B
70389	*Profusulinella* Zone subzone B
70455	*Profusulinella* Zone subzone B
70471	*Fusulinella* Zone subzone B subdivision B 1
70479	Locality 70290, another limestone olistolith. *Fusulinella* Zone probably subzone A
70481	*Fusulinella* Zone subzone B subdivision B 1
70485	*Fusulinella* Zone
70514	*Profusulinella* Zone subzone B
70515	*Fusulinella* Zone subzone A
7136	*Profusulinella* Zone subzone A or possibly subzone B
7156	Limestone pebbles from a conglomerate lens. Some: *Fusulinella* Zone subzone A. Others: *Fusulinella* Zone subzone B subdivision B 1
7144	*Profusulinella* Zone
7178	*Fusulinella* Zone subzone B subdivision B 2
7178A	Locality 7178, another limestone olistolith 20 m west of 7178. *Fusulinella* Zone subzone B, high in subdivision B 1 or low in subdivision B 2. Comparable with location A 6 in the Cuenca de Beleño (van Ginkel, 1965)
71Bar	*Profusulinella* Zone subzone A, or *Millerella* Zone
7198	*Profusulinella* Zone
71101	*Profusulinella* Zone subzone A
71247	Limestone pebbles from a conglomerate lens. Some: *Fusulinella* Zone subzone A. Others: Possibly younger, *Fusulinella* Zone subzone B subdivision B 1
71271	*Fusulinella* Zone subzone A (the top of subzone A)

Limestone samples with algae and fusulinids, collected by Mr. G. J. B. Germs, determined by Mr. G. J. B. Germs (algae) and Dr. A. C. van Ginkel (fusulinids); data from Germs (1966).

Ge 1	*Profusulinella* Zone subzone A (the top of the subzone)
Ge 2	*Profusulinella* Zone subzone A (the top of the subzone)
Ge 3	*Profusulinella* Zone subzone A (the top of the subzone)
Ge 4	*Profusulinella* Zone subzone A (the top of the subzone)
Ge 5	*Profusulinella* Zone subzone A (the top of the subzone)
Ge 6	*Profusulinella* Zone subzone A (the top of the subzone)
Ge 7	*Profusulinella* Zone subzone A (the top of the subzone)
Ge 100	*Fusulinella* Zone subzone A
Ge 106	*Profusulinella* Zone subzone A (the top of the subzone)
Ge 118	*Profusulinella* Zone subzone B (the top of the subzone)
Ge 122	*Profusulinella* Zone subzone A, or possibly low in the subzone B
Ge 124	*Profusulinella* Zone subzone B (the top of the subzone)
Ge 133	*Fusulinella* Zone subzone B subdivision B 1
Ge 167	(Only algae) *Profusulinella* Zone subzone A (the base of the subzone)

Limestone samples with fusulinids and sometimes algae, collected by Mr. J. F. Savage, determined by Dr. A. C. van Ginkel (fusulinids) and Mr. G. J. B. Germs (algae); data from the private records of Dr. A. C. van Ginkel.

S 63330	*Fusulinella* Zone subzone A
S 131	*Profusulinella* Zone subzone B
S 64495	Only algae, *Fusulinella* Zone subzone B subdivision B 1 or *Fusulinella* Zone subzone A (the top)
S 64496	*Profusulinella* Zone subzone B

Limestone samples with fusulinids, anonymous collectors, determined by Dr. A. C. van Ginkel; data from the private records of Dr. A. C. van Ginkel.

X 1 *Profusulinella* Zone subzone B (the top of the subzone)
X 2 *Profusulinella* Zone subzone B (the top of the subzone)
X 3 *Fusulinella* Zone subzone B subdivision B 1

Limestone samples collected by Mr. R. Lanting (1966), fusulinids determined by Dr. A. C. van Ginkel; data from Lanting (1966).

L 1 Locality not certain, mentioned as "limestone pebbles from the conglomerate near Valmeo".
Profusulinella Zone subzone B
L 2 *Fusulinella* Zone subzone B subdivision B 1

Limestone samples collected and determined by Dr. A. C. van Ginkel; data from van Ginkel (1965).

P 1 *Profusulinella* Zone subzone B
P 3 *Profusulinella* Zone subzone B
P 4 *Fusulinella* Zone subzone B subdivision B 1
P 7 *Fusulinella* Zone subzone B subdivision B 2
P 10 *Fusulinella* Zone subzone B subdivision B 3
P 52 *Protriticites* Zone (Corros Limestone Member)
P 72 *Fusulinella* Zone subzone B subdivision B 1
P 73 *Fusulinella* Zone subzone B subdivision B 2

Samples that yielded conodonts; determinations by Mr. K. Boersma. Samples collected by the author.

70CA1 South of the village Sebrango in autochthonous position
(1 : 25 000 map)
Famennian *styriaca* Zone
70A2 East of Aliva, Picos de Europa Area in autochthonous position
Viséan
70C101 Northwest of Sebrango in autochthonous position
(1 : 25 000 map)
Famennian *costatus* Zone
The sample is contaminated; conodonts that give a Viséan age also occur.
70C141 North of the village Besoy, probably in allochthonous position
(1 : 25 000 map)
Famennian Middle *velifera* to Lower *costatus* Zone
70C146 South of Barcena, probably in autochthonous position
(1 : 25 000 map)
Viséan
70C172 Northwest of Pembes, along the old path to the Aliva valley probably in allochthonous position
Frasnian *asymmetrica* Zone
70C221 West of Mogrovejo, in allochthonous position
(1 : 25 000 map)
Givetian *varca* Zone
70C225 West of Mogrovejo, in allochthonous position
(1 : 25 000 map)
Famennian *costatus* Zone
70C248I North of Tanarrio, probably both in allochthonous position
and 70C248II
Famennian *costatus* Zone
70C257 North of Lon, in autochthonous position
Viséan
70C290 South of Lon, in allochthonous position
Viséan
70C290II South of Lon, in allochthonous position
Famennian *styriaca* Zone to *costatus* Zone
7130 Puerto de Remoña, in allochthonous position, conodont sample lost
Upper Famennian
71C83 West of Campollo, in allochthonous position
(1 : 25 000 map)
Famennian *costatus* Zone
71C84 West of Campollo, in allochthonous position
(1 : 25 000 map)
Lower Namurian (a specimen of *Declinognathus Nodulitrius*)
71C142 West of Campollo, in allochthonous position
(1 : 25 000 map)
Frasnian Upper *asymmetrica* Zone to *gigas* Zone
71C178 Northeast of la Vega de Liébana, in allochthonous position
(1 : 25 000 map)
Famennian *costatus* Zone

71C185 North of Toranzo, in allochthonous position
(1 : 25 000 map)
Viséan

71C206 North of Bores, in allochthonous position
(1 : 25 000 map)
Givetian

71C207 West of Toranzo, in allochthonous position
(1 : 25 000 map)
Famennian Upper *triangularis* Zone to Upper *crepida-crepida* Zone

71C208 Northwest of la Vega de Liébana, in allochthonous position
(1 : 25 000 map)
Givetian-Frasnian boundary *hermanni-cristata* Zone

71C270 West of Mogrovejo, probably in allochthonous position
(1 : 25 000 map)
Viséan

71C304 East of Bores, probably in allochthonous position
(1 : 25 000 map)
Viséan

71C315 South of la Vega de Liébana, probably in allochthonous position
(1 : 25 000 map)
Frasnian Upper *asymmetrica* Zone to Lower *triangularis* Zone

71C321 North of Valcayo, probably in autochthonous position
(1 : 25 000 map)
Famennian

71C323 North of Valcayo, in autochthonous position
(1 : 25 000 map)
Famennian Upper *quadrantinodosa* Zone to Upper *styriaca* Zone

71C327 North of Valcayo, in allochthonous position
(1 : 25 000 map)
Famennian *costatus* Zone

71C333 North of Vendejo, in allochthonous position
Famennian *costatus* Zone

Conodont samples collected by Mr. R. Lanting and determined by Dr. H. A. van Adrichem Boogaert (1967)

X 92 = 71C185
(1 : 25 000 map)
Viséan *bilineatus-commutatus nodosus* Zone

X 116 North of Enterrias, in allochthonous position
(1 : 25 000 map)
Famennian *costatus* Zone

LV 435 North of Valcayo, in allochthonous position
(1 : 25 000 map)
Famennian Lower *velifera* Zone

X 158 North of Valcayo, in autochthonous position
(1 : 25 000 map)
Famennian *quadrantinodosa* Zone

X 117 North of Enterrias, in allochthonous position
(1 : 25 000 map)
Frasnian Upper *gigas* Zone

X 71 West of Campollo, in allochthonous position
(1 : 25 000 map)
Frasnian *gigas* Zone

LV 26 South of la Vega de Liébana, in allochthonous position
(1 : 25 000 map)
Frasnian Lower *asymmetrica* Zone to Middle *triangularis* Zone

X 85 East of Bores, in allochthonous position
(1 : 25 000 map)
Eifelian *corniger* Zone

Flora collections

F 1 The Casavegas coalbeds yielded an associated flora of Upper Westfalian D age (Wagner in Wagner & Varker, 1972).

F 2 The Redondo coalbeds are a part of the Redondo Coal Member of the Barruelo Formation. The floras from the Barruelo Formation in this subbasin are correlated with the Stephanian A (Wagner 1955, p. 164; Wagner in Wagner & Varker, 1972).

F 3 East of the Pico Cordel.
Stephanian C (Wagner, 1970).

F 4 North of Puente Pumar.
Flora remnants collected by Mr. M. Budding, determined by Dr. R. H. Wagner; a poor collection which gives only evidence to Stephanian (Wagner, pers. comm.).

F 5 South of the village Cucayo, east of the Rio Frio, along the track that goes from Cucayo to Lores at an approximate topographic height of 1200 m.

The collection was made by Mr. J. A. van Hoeflaken and is known as the Dobres flora; it derives from a small exposure of muddy rock with rootlets, right under a thick conglomerate bed of the Curavacas Formation and on top of, or part of, some thinner conglomerate bands that seem to be cut off by the former. The position of this flora may be allochthonous.
Age: Namurian C to Westfalian A (Wagner, 1959).

F 6 On the top of the Pico Coriscao, at the base of the Panda Limestone. Collected by Mr. J. F. Savage, determination by Dr. F. Stockmans, "this flora has affinities to higher Westfalian as well as to Stephanian floras" (Stockmans, cited by van Ginkel, 1965).

F 7 North of San Mames, SW of Cueto Cucon.
Some plant fossils were recovered from the Labra Formation one of which could be identified as *'Walchia piniformis'*.
Age: probably Permian (Stockmans, oral comm. by Budding).

F 8 At the eastside of the road along the Rio Nansa, south of La Lastra.
Flora remnants collected by Mr. M. Budding, determined by Dr. R. H. Wagner. The plant fragments belong to the Voltziaceae group and are comparable with specimens from the 'Grès à Voltzia'.
Age: Triassic (Wagner, written comm.).

F 9 Northwest of the village San Mames in the core of an anticlinal structure.
Floral remains collected by Sr. Florencio Fernandez; the position of the flora may be allochthonous.
Identifications by Dr. R. H. Wagner and Dr. J. A. Knight: cf. *Alethopteris ambigua* Lesquereux; *Polymorphopteris* cf. *polymorfa* (Broignart) Wagner; *Lobatopteris alloiopteroides* Wagner; *Pecopteris hemitelioides* Broignart; *Asterotheca* sp.
Probable age: either Upper Westfalian D or Lower Cantabrian, and most likely the latter, according to Wagner and Knight (written comm.).

ERRATA

Geology of Liébana

ENCLOSURE I

1 Loc. 43°12'03"-01°02'17" — Scree signature not indicated.

2 Loc. 43°06'41"-01°05'32" — Cervera limestone lithosome indicated as conglomerate.

3 Loc. 43°04'35"-01°04'59" — Small Yuso limestone lithosomes in the core of the syncline not indicated.

4 Loc. 43°08'07"-01°01'53" — NW strike and vertical dip of the bedding in Barcena conglomerate not indicated.

5 Loc. 43°02'53"-00°58'21" — Cervera limestone lithosome indicated as Villabellaco limestone.

6 Loc. 43°01'06"-01°00'12" — ENE strike and 55 degrees SSE dip should be indicated in the Yuso conglomerate.

7 Loc. 43°01'53"-00°59'10" — The small lithosome in the Yuso conglomerate is probably Lebanza limestone.

8 Peña Brez is indicated on Loc. 43°02'55"-00°48'40" but should be indicated on Loc. 43°03'05"-00°54'15".

9 Loc. 43°03'44"-00°56'05" — The three lenses indicated as Villabellaco limestone are Cervera conglomerates.

10 Loc. 43°03'37"-00°56'01" — The two Cervera limestone lithosomes which are separated by an EW fault are also separated by a lens of shale which is indicated as limestone.

11 Loc. 43°03'39"-00°56'31" — Two very small limestone lithosomes should be replaced by a slump deformation sign.

12 Loc. 43°06'46"-00°57'06" — ESE strike and 30 degrees SSW dip is not indicated in the Viorna conglomerate.

13 Loc. 43°06'40"-01°00'21" — ESE strike and 80 degrees NNW dip is not indicated in the Viorna conglomerate N of Dobarganes.

14 Loc. 43°09'04"-00°46'54" — The unconformity near Peña Sagra should be indicated at the base of the conglomerate.

15 Legenda; the limestone lenses of the Cervera Formation should be indicated with a NS and EW striking blue cross lineation.

ENCLOSURE II

Localities are given in mm distance from the left side of the section.

1 Section CC^{I} on 100 mm — Vertical fault not indicated.

2 Section $C^{V}C^{VI}$ on 184 mm — The red patch of intrusive rock is dislocated to section CC^{I} on 174 mm.

3 Section $D^{II}D^{III}$ on 50 mm — A Yuso limestone lithosome below the conglomerate does not have the right colour.

4 Section DD^{I} on 135 mm — The lens of Yuso conglomerate does not have the right colour.

5 Section EE^{I} on 194 mm — The indicated lens of Villabellaco limestone should be Cervera conglomerate.

6 Legend of the index map: — The word Murcia should be omitted.

ENCLOSURE IV

Localities are given in mm distance from the lower left corner of the map, measured along the left side and along the lower side of the map.

1 Locality 142 - 0 — Vidrieros limestone and shale should be indicated instead of Cervera limestone and shale.

2 Locality 107 - 59,5 — Vidrieros limestone should be indicated instead of Cervera limestone.

3 Locality 92 - 64 — The limestone lithosome NE of Besoy (C141) is Vidrieros limestone and not Cervera limestone.

4 Locality 52 - 259 — The limestone lithosome W of Toranzo (C207) is Vidrieros limestone and not Cervera limestone.

5 Locality 22 - 307 — The limestone lithosome exposed along the road belongs probably to the Gustalapiedra or to the Cardaño Formation (compare note on p. 383).