LEIDSE GEOLOGISCHE MEDEDELINGEN, Deel 52, Aflevering 2, pp. 265–339, 6-10-1983

CONODONT BIOSTRATIGRAPHY AND DEPOSITIONAL HISTORY OF THE MIDDLE DEVONIAN TO LOWER CARBONIFEROUS IN THE CANTABRIAN ZONE (CANTABRIAN MOUNTAINS, SPAIN)

BY

J.G.M. RAVEN *)

SUMMARY

Within the Cantabrian zone during the Devonian and Early Carboniferous three large palaeogeographical units were of importance: the Asturian geanticline, the Palencian basin and the Asturo-Leonesian basin. These units have a different history of sedimentation, particularly the Palencian basin. Further, in the Asturo-Leonesian basin there were some important structural elements: the Intra-Asturo-Leonesian facies line, formed by an active, normal fault parallel to the border of the basin, and two structural highs: the Pardomino high and the Somiedo high which divided the basin into three pieces. The Asturo-Leonesian basin was a narrow, shallow continental shelf which in the south and west passed into the deeper part which extended over the West Asturian-Leonese zone. With help of biostratigraphic correlations of the Middle Devonian to Early Carboniferous deposits in the Cantabrian zone, based on conodonts, an overview is given of the depositional history of the entire area.

Several times during the Givetian and Frasnian a biostromal platform developed in the Asturo-Leonesian basin with small bioherms of stromatoporoids and corals at the southern border along the facies line and in the east along the León line, with a lagoon behind. In the Palencian basin the sediment supply was always smaller. There, shales and nodular limestones with pelagic faunas (Gustalapiedra Formation) were deposited. Carbonate sedimentation started simultaneously with the formation of the first carbonate platform of the Portilla Formation in the Asturo-Leonesian basin. Repeatedly the Asturian geanticline was uplifted leading to tilting of the Asturo-Leonesian basin and to erosion of the uplifted parts. The siliciclastic erosion products were deposited in the subsided parts leading to progradation of the coast with the formation of coastal barriers notably along the facies line. As soon as the supply of siliciclastics decreased, a new carbonate platform could form. The last stromatoporoid-coral biostromes formed such a platform at the end of the Frasnian in the Esla area (Crémenes Limestone in the Nocedo Formation). At the same time, in the west of the Asturo-Leonesian basin fan-deltas formed along the facies line, with conglomerates which originated from erosion at the geanticline, which then apparently extended to the facies line.

The Asturian geanticline had extended gradually during the Devonian and the differences between the basins had increased. Uplift of the geanticline resulted in the emergence of the entire Asturo-Leonesian basin during the early Famennian. Then, erosion products from the geanticline were also transported into the Palencian basin by turbidity currents (Murcia Formation) interrupting the sedimentation of nodular limestones which recovered later on (Vidrieros Formation). At the end of the Famennian, due to a transgression the sea spread rapidly over the truncated geanticline. South of the Intra-Asturo-Leonesian facies line initially turbiditic storm deposits (Fueyo Formation) formed while a thin layer of sands and crinoidal grainstones (Ermita Formation) was deposited on the geanticline and in the major part of the Asturo-Leonesian basin. The entire Cantabrian zone may have been emergent during the early Tournaisian and, together with deformations in the West Asturian-Leonese zone movements lead to an inversion of the relief in the Cantabrian zone with the formation of a number of small basins in the platform. When, during the late Tournaisian a transgression lead to the spread of cold, nutrient-rich water over the entire area, some of this water stagnated in the basins so that black shales, radiolarites and phosphatic nodules (Vegamián Formation) could be deposited at the same time as the carbonate sedimentation continued on other parts of the platform. The continuation of the transgression lead to the formation of nodular limestones (Alba Formation), first only on the shallow platforms but later also in the deeper parts because circulation in the basins improved.

The Variscan orogenesis lead to deformations of which some are discussed in this paper. Lithological data indicate that the Esla nappe and the Valsurvio area were displaced towards the north and later the South Cantabrian block was shifted about 15 to 20 km towards the east along the Sabero-Gordón line.

The different formations and smaller lithological units are described and an informal subdivision of the Portilla Formation is proposed. The conodont faunas which were found are described. A number of interesting or problematic species is treated more extensively. The conodont biozonation is discussed. For the early Carboniferous the regional zonation of Higgins (1974) is adapted: the *Polygnathus* fauna and *Gnathodus pseudosemiglaber* Zone are introduced. Rather extensively the palaeoecology of the conodonts will be discussed.

*) Address: Dr. J.G.M. Raven, Binnenweg 46, 2264 MK Leidschendam, The Netherlands

SUMARIO

Dentro de la zona Cantábrica durante el Devónico y Alto Carbonífero tres grandes unidades paleogeográficas fueron de importancia: el geoanticlinal Asturiano, la cuenca Palentina y la cuenca Astur-Leonesa. Dichas unidades tienen una distinta historia de sedimentación en particular la cuenca Palentina. Además, dentro de la cuenca Astur-Leonesa había unas unidades estructurales importantes: le linea facial Intra-Astur-Leonesa, formada por una falla normal activa paralela al borde de la cuenca, y dos elevaciones estructurales: el alto de Pardomino y el alto de Somiedo, que dividieron la cuenca en tres partes. La cuenca Astur-Leonesa era la parte estrecha y poco profunda de una plataforma continental que en los bordes sur y oeste pasaba a una plataforma más profunda (zona Asturoccidental-Leonesa). Con la ayuda de correlaciones bioestratigráficas de los depositos del Devónico Medio hasta el Carbonífero Inferior en la zona Cantábrica, basado en conodontos, se da un resumen de la historia de sedimentación en todo el area.

Unas veces durante el Givetiense y Frasniense se desarrollaba una plataforma biostromal en la cuenca Astur-Leonesa sobra la cual, en el borde meridional a lo largo de la linea facial y en el este sobre la linea de León había biohermas pequeñas de estromatoporoideos y corales con una lagoon detrás. En la cuenca Palentina en todo momento el aporte de sedimentos era más pequeño. Allí se depositaban arcillas y calizas nodulosas con faunas pelagicas (Formación Gustalapiedra). La sedimentación de carbonatos empezó al mismo tiempo que la formación de la primera plataforma calcárea de la Formación Portilla en la cuenca Astur-Leonesa. De vez en cuando el geoanticlinal Asturiano fue levantado causando la volcación de la cuenca Astur-Leonesa y la erosión de las partes elevadas. Los productos de erosión (siliciclasticos) se depositaban en las partes bajas, de modo que la costa se extendió gradualmente con la formación de barras en particular a lo largo de la linea facial. Una vez disminuyó el aporte de siliciclasticos, de nuevo podía formarse una plataforma calcárea. La última biostroma de estromatoporoideos y corales formó una plataforma a finales del Frasniense en el area del Esla (Caliza de Crémenes en la Formación Nocedo) Al mismo tiempo en el oeste de la cuenca Astur-Leonesa a lo largo de la linea facial se formaban conos aluviales con conglomerados provenientos de la erosión en el geoanticlinal que entonces evidentemente se extendió hasta la linea facial.

El geoanticlinal Asturiano se había extendido gradualmente durante el Devónico y también la diferencia entre las cuencas había aumentado. Durante el bajo Fameniense el geoanticlinal aún se extendió tanto que la cuenca Astur-Leonesa se emergió entera. Productos de erosión de la geoanticlinal entonces fueron introducidos también en la cuenca Palentina por corrientes de turbiditas (Formación Murcia) interrumpiendo la sedimentación de calizas nodulosas que después de restableció (Formación Vidrieros). A finales del Fameniense, por una transgresión, el mar se extendió rápidamente sobre el geoanticlinal aplanado. Inicialmente al sur de la linea facial se depositaron turbiditas causadas por tormentas (Formación Fueyo) mientras que un estrato delgado de areniscas y calizas de crinoídeos (Formación Ermita) se depositó sobre el geoanticlinal y en la mayor parte de la cuenca Astur-Leonesa. Probablemente toda la zona Cantábrica se emergió durante poco tiempo en el alto Tournaisiense. A la vez se originaban deformaciones en la zona Asturoccidental-Leonesa y movimientos que llevaban a una inversión del relieve en la zona Cantábrica con la formación de unas cuencas poco profundas en la plataforma. Cuando, en el bajo Tournaisiense una transgresión llevaba a la extensión del agua fria y rica en alimentos sobre todo el area, parte del agua se estancaba en las cuencas de modo que arcillas negras, radiolaritas y nodulas fosfáticas (Formación Vegamián) podían ser depositadas a la vez que la deposición de calizas continuó sobre otras partes de la plataforma. La continuación de la transgresión llevó a que empezaran a formarse calizas nodulosas (Formación Alba), inicialmente solo sobre las plataformas poco profundas pero después también en las partes más profundas por incremento de la circulación en las cuencas.

La orogenesis Varistica llevó a deformaciones de las cuales se discute unas en esta publicación. Datos litologicos indican que el manto del Esla y el area de Valsurvio fueron desplazados al norte y después el bloque Cantábrico Meridional fue desplazado unos 15 o 20 km al este a lo largo de la linea de Sabero-Gordón.

Se describen las distintas formaciones y unidades litológicas más pequeñas y se propone una división informal de la Formación Portilla. Se describen las faunas de conodontos halladas. Más extensamente se trata unas especies interesantes o problematicas. Se discute la biozonación. Para el alto Carbonífero se adapta la zonación regional de Higgins (1974): la fauna de *Polygnathus* y la Zona de *Gnathodus pseudosemiglaber* serán introducidas. Bastante detallada también se trata la paleoecología de los conodontos.

SAMENVATTING

Binnen de Cantabrische zone waren tijdens het Devoon en Vroeg-Karboon drie grote paleogeografische eenheden van belang: de Asturische geanticlinaal, het Palentijnse bekken en het Asturo-Leonese bekken. Deze eenheden hebben een verschillende sedimentatie-geschiedenis, vooral het Palentijnse bekken. Verder waren er binnen het Asturo-Leonese bekken enkele belangrijke strukturele elementen: de Intra-Asturo-Leonese faciëslijn, gevormd door een aktieve normale breuk parallel aan de rand van het bekken, en twee strukturele hogen: het Pardomino hoog en het Somiedo hoog, die het bekken verdeelden in drie delen. Het Asturo-Leonese bekken was het smalle, ondiepe deel van een continentaal plat dat aan de zuid- en westrand overging in de diepere shelf (Westasturisch-Leonese zone). Met behulp van biostratigrafische correlaties van de middendevonische tot onderkarbonische afzettingen in de Cantabrische zone, met behulp van conodonten, wordt een overzicht gegeven van de afzettingsgeschiedenis van het hele gebied.

Enkele keren tijdens het Givetien en Frasnien ontwikkelde zich een biostromaal kalkplatform in het Asturo-Leonese bekken met aan de zuidelijke rand langs de Intra-Asturo-Leonese faciëslijn en in het oosten langs de Leónlijn kleine biohermen van stromatoporen en koralen waarachter zich een lagune bevond. In het Palentijnse bekken was de sedimenttoevoer steeds geringer. Daar werden schalies en knobbelkalken met pelagische fauna's afgezet (Gustalapiedra-Formatie). De kalksedimentatie begon tegelijk met de vorming van het eerste kalkplatform van de Portilla-Formatie in het Asturo-Leonese bekken. Herhaalde malen werd de Asturische geanticline opgeheven waardoor het Asturo-Leonese bekken kantelde en de hooggelegen delen werden geërodeerd. De siliciklastische erosieprodukten werden afgezet in de gedaalde delen zodat de kust zich geleidelijk uitbouwde en strandwallen gevormd werden, vooral langs de faciëslijn. Zodra de aanvoer van siliciklasten afnam kon zich opnieuw een kalkplatform vormen. Het laatste stromatoporen-koralen biostroom vormde een dergelijk platform aan het eind van het Frasnien in het Esla

gebied (Crémeneskalk in de Nocedo-Formatie). Tegelijkertijd werden in het westen van het Asturo-Leonese bekken langs de faciëslijn puinwaaiers gevormd met conglomeraten die afkomstig waren van erosie op de geanticlinaal die toen blijkbaar tot aan de faciëslijn reikte.

De Asturische geanticlinaal had zich tijdens het Devoon geleidelijk uitgebreid en ook het verschil tussen de bekkens was toegenomen. Opheffing van de geanticline tijdens het vroeg-Famennien leidde tot het droogvallen van het gehele Asturo-Leonese bekken. Erosieprodukten van de geanticlinaal werden toen ook het Palentijnse bekken ingevoerd door turbidietstromen (Murcia-Formatie); daardoor werd de knobbelkalksedimentatie onderbroken maar later herstelde deze zich weer (Vidrieros-Formatie). Aan het eind van het Famennien breidde de zee zich door een transgressie snel uit over de afgeplatte geanticlinaal. Ten zuiden van de faciëslijn werden aanvankelijk turbiditische stormafzettingen (Fueyo-Formatie) gevormd terwijl een dunne laag zanden en crinoïdenkalken (Ermita-Formatie) werd afgezet op de geanticlinaal en in het grootste deel van het Asturo-Leonese bekken. De gehele Cantabrische zone viel misschien korte tijd tijdens het vroeg-Tournaisien droog en tegelijk met deformaties in de West Asturisch-Leonese zone leidden bewegingen tot een omkering van het relief in de Cantabrische zone waarbij een aantal ondiepe bekkens in het platform werden gevormd. Toen tijdens het laat-Tournaisien een transgressie leidde tot de uitbreiding van koud, voedselrijk water over het gehele gebied stagneerde het water in die bekkens, en zwarte schalies, radiolarieten en fosfaatknollen (Vegamián-Formatie) werden afgezet terwijl op andere delen van het platform de kalksedimentatie doorging. De voortgang van de transgressie leidde ertoe dat zich knobbelkalken (Alba-Formatie) begonnen te vormen, eerst alleen op het ondiepe platform maar later ook in de diepere delen door de toename van de circulatie in de bekkens.

De Varistische orogenese leidde tot deformaties waarvan er sommige in dit artikel worden besproken. Lithologische gegevens geven aan dat het Esla dekblad en het Valsurviogebied naar het noorden werden verplaatst en dat later het Zuidcantabrische blok langs de Sabero-Gordónlijn ongeveer 15 à 20 km naar het oosten werd geschoven.

De verschillende formaties en kleinere lithologische eenheden worden beschreven en een informele indeling van de Portilla-Formatie wordt voorgesteld. De aangetroffen conodontfauna's worden beschreven. Een aantal interessante of problematische soorten wordt uitgebreider behandeld. De biozonering wordt besproken. Voor het vroeg-Karboon wordt de regionale zonering van Higgins (1974) aangepast: de *Polygnathus*-fauna en de *Gnathodus pseudosemiglaber*-Zone worden ingevoerd. Vrij uitgebreid wordt ook de paleoecologie van de conodonten behandeld.

CONTENTS

1. INTRODUCTION

1.1. INTRODUCTION

This thesis deals with the geological history of the Cantabrian Mountains from Middle Devonian to Early Carboniferous time. Lithostratigraphical and biostratigraphical data (mainly from conodonts) are given and form the fundament on which an interpretation is based.

The present study is part of the geological investigation of the Cantabrian Mountains which has been carried out since 1950 by staff members and students of the University of Leiden, based on the work of Comte (1959). First much emphasis was laid on geological mapping of the area (see Savage & Boschma, 1980), later also detailed studies on individual formations appeared (e.g. Reijers, 1972; de Coo, 1974). Relatively few biostratigraphical data were at hand, the main study being that of conodonts by van Adrichem Boogaert (1967) which concentrated on the Lower Carboniferous rocks and the Palencian basin.

In 1977 I started an investigation on the stratigraphy of the Upper Devonian rocks in the Esla area as well as a study of Middle and Late Devonian conodonts (Raven, 1980 a, b). From 1980 to 1982 this study was extended to nearly all Middle Devonian to Lower Carboniferous rocks in the Cantabrian Mountains. Numerous sections were measured, sampled or at least visited. This thesis is based on that research in the field combined with a study of the conodonts collected by myself, by van Adrichem Boogaert (1967, in part), Maas (1974), van Tongeren (1975) and students of the Department of Stratigraphy and Palaeontology. Ms. Jolanda Huisman and Mr. Peter van der Ark studied conodonts of the Santa Lucía Formation and the base of the Alba Formation respectively. Also all the lithostratigraphical data available in the department were gathered. All this information was used in the present study. Of course, I am responsible for the given interpretations. Only where necessary I will discuss other opinions or errors.

In order to facilitate comparisons the terminology in this publication follows some standard classifications. The limestones were classified according to Dunham (1962) and Embry & Klovan (1972), the sandstones according to Pettijohn (1957), the sedimentary structures according to Reineck & Singh (1973), the grain size according to the Udden-Wentworth scale. The sorting was estimated with the aid of the figures in Beard & Weyl (1973).

The numbers between brackets refer to the paragraphs in this paper. The cross-sections and the sections in the southern part of the studied area are indicated in Encl. 1: Fig. 1.

1.2. ACKNOWLEDGEMENTS

This publication combines results of investigations by many persons. I am grateful to all the students who wrote the reports which I used. I am even more grateful to those students who collected samples in the field and put them at my disposal. Of them I mention Ton ten Have, Franek Mrozek, Gerard van Loevezijn, Gerald van der Sluijs, Anthonie van der Marel, Joop Witte, Lucie Wassink, Jaap-Jan Vermeulen, Frans Kunst and Niek Schelling. Samples were prepared at the laboratory by Mrs. Liesbeth van der Poort-de Jong who indefatigably continued her careful work during five years. Dr. M. van den Boogaard (Leiden) and Dr. P. Bultynck (Brussels) advised me about the identification of the conodonts. Discussions with Dr. J.F. Savage (Utrecht), Dr. M. van den Boogaard (Leiden), Dr. G. Klapper (Iowa), Dr. T.J.A. Reijers (Rijswijk), Mr. P. Seibert (Tübingen) and others were very useful. I thank Prof.Dr. P. Carls (Brunswick) and Dr. G. Becker (Frankfurt am Main) for their written communications.

At the Geological Institute much work was done by the technical staff. I mention Mr. C.J. van Leeuwen who made the thin sections, Mr. J. Bult who drew the figures and Mr. J. Verhoeven who made the photo prints. During the last year of preparation of the manuscript the Rijksmuseum van Geologie en Mineralogie (Leiden) provided technical assistance. It further helped me by putting at my disposal the collections of van Adrichem Boogaert and others. Dr. W. de Priester (Leiden) accompanied me in taking the photographs with the scanning electron microscope at the Zoölogisch Instituut (Leiden). Dr. J.F. Savage (Utrecht) critically read the manuscript which thus improved considerably. Further Dr. M. van den Boogaard (Leiden) critically read Chapter 4 and Ms. Jane Tremaine (Leidschendam) corrected some errors in the English text.

I express my sincere thanks to all of them, also to those not mentioned personally. Further thanks are due to the family Gonzalez living near the Porma press (León, Spain) and to Mrss. Manuela and Pilar Calleja in Pola de Gordón (León) for their great hospitality. I also thank the Ministerie van Onderwijs en Wetenschappen for the grant which permitted the preparation of this thesis.

All the measured sections, a list of the conodont samples and their contents and the samples are stored at the Rijksmuseum van Geologie en Mineralogie (Leiden). A copy of the list, as well as simplified sections, have been sent to Dr. P. Bultynck (Brussels), Dr. G. Klapper (Iowa) and Dr. S. García López (Oviedo). The internal reports of the Department of Stratigraphy and Palaeontology are now stored at the Rijksmuseum van Geologie en Mineralogie.

1.3. INTRODUCTION TO THE STRUCTURES

1.3.1. Structural zones

In analogy to Kossmat's subdivision of the Variscides of central Europe, Lotze (1945) divided the Iberian Variscides into a number of structural zones. Later Julivert et al. (1974) proposed a modified version. In this paper reference will be made to the three northern zones, from southwest to northeast the Central-Iberian, the West Asturian-Leonese and the Cantabrian zone. Many authors interprete the Cantabrian zone as miogeo(syn)cline and the West Asturian-Leonese zone as eugeosyncline, e.g. Lotze (1945), Matte (1968), de Coo (1974) and Wagner & Martínez García (1974). Most data, however, favour the alternative model of Kullman & Schönenberg (1975, 1982), Savage (1981) and Brouwer (1982) who interprete both zones as part of a cratonic basin: there is no evidence of a nearby oceanic realm (Brouwer, 1982).

Already before the description of the zones it was recognized that the Variscides in northwestern Spain form part of a large curvature, the Asturian arc. Palaeomagnetic and structural data are used as evidence for the gradual compression of the arc during the Variscan orogeny but at least a part of the curvature may be a primary structure (Ries et al., 1980) as is further indicated by the importance of décollement tectonics (Savage, 1979, 1981).

1.3.2. Palaeogeographical units

Brouwer (1962) distinguished two facies in the Cantabrian zone, the Asturo-Leonesian facies representing subtidal to tidal environments on a relatively stable shelf, characterized by an abundant benthonic fauna, and the Palencian facies deposited in a slightly deeper basin which during the greater part of its history was devoid of coarse detrital material and was characterized by a fauna in which pelagic elements take a predominant position. Later the areas were

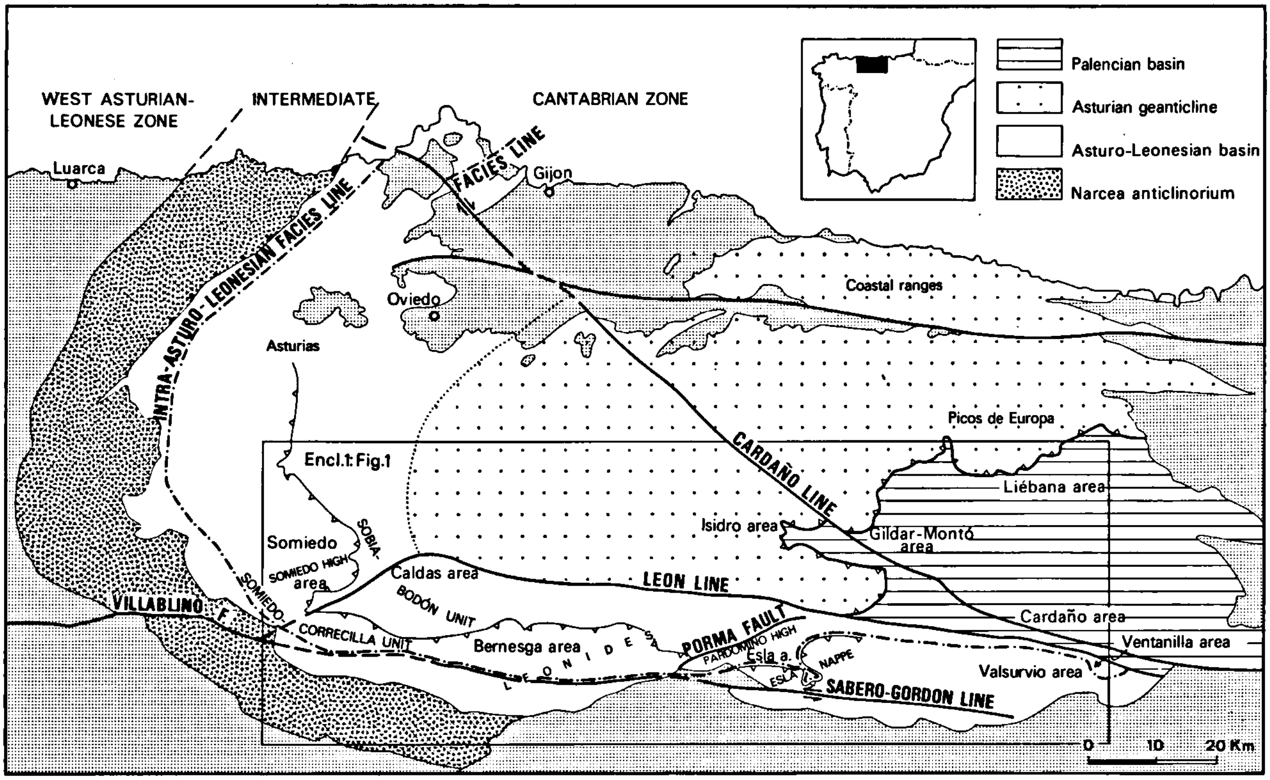

Fig. 1. Map of the major palaeogeographical units which were of importance during the Devonian and Early Carboniferous. Further the areas are indicated which are mentioned in the text.

considered as separate basins: the Asturo-Leonesian basin and the Palencian basin (Fig. 1; van Adrichem Boogaert, 1967; Brouwer, 1968). The large area between these two basins was first mentioned by Radig (1966) and later treated more extensively by van Adrichem Boogaert (1967) who called it the Asturian geanticline (Fig. 1). The sediments which were deposited on the geanticline belong to the Asturo-Leonesian facies. To the west and south the Asturo-Leonesian basin is bordered by the Narcea anticlinorium which separates the Cantabrian zone from the West Asturian-Leonese zone (Fig. 1).

1.3.3. Nappes

Nappe structures are very frequent in Cantabria. They were formed by décollement at different levels within the stratigraphic column but true metamorphic basement is not included in the nappes. The displacement of the nappes was probably caused by gravity sliding and restricted to several or several tens of kilometers (Savage, 1979; Heward & Reading, 1980). The nappes were displaced in a centripetal pattern. Also the distinct structural provinces are bordered by thrust planes and therefore the original distance between them has been much reduced (Julivert, 1971; Heward & Reading, 1980). In this way some of the sudden facies changes along some of the boundaries between these provinces may be explained.

As an example the structure of the Asturo-Leonesian basin (Fig. 1) is discussed in more detail (based on Julivert, 1971). In the west and south the Narcea anticlinorium is thrust over the basin, slightly reducing the width of the basin. The width of the basin is reduced much more by folds, upthrusts and nappes within the basin: this reduction is estimated at about 50 % for the Leonese part of the basin (Savage, 1979). The intensity of deformation increases from the northwest towards the southeast: in the northwest there are large open folds, more southward two large units are recognized: the northern Sobía-Bodón unit and the southern Somiedo-Correcilla unit, each of them further subdivided by internal upthrusts (e.g. Evers, 1967), and a nappe, the Somiedo nappe. East of the Porma fault some upthrusts are present in the northwest but the most important structure is a single nappe, the Esla nappe (Rupke, 1965). This nappe was originally situated south of the Sabero-Gordón line (5.2.). Still farther east the structures become more complicated. In this paper generally no palinspastic reconstructions are made. A reconstruction of the Asturo-Leonesian part of the shelf would show that originally it was about 40 km broad and at least about 220 km long.

1.3.4. Fundamental lines

Several fundamental lines were discovered in the southern part of the Cantabrian zone, the León line (de Sitter, 1962), Sabero-Gordón line (Rupke, 1965), Cardaño line (van Veen, 1965) and Pontón line (Sjerp, 1967) (Fig. 1), which were considered to be fundamental features with a deep-seated origin (e.g. Kullman & Schönenberg, 1978). Due to the fundamental lines, during the Devonian and Carboniferous deep-seated faults separated blocks with a distinct history (Krans, 1982) and with different facies. Since the position of the faults shifted, actually not all of them still form an active fault. Along the León line and Sabero-Gordón line wrench faults developed during the Stephanian. Parallel to the fault along the Sabero-Gordón line another fault was present during Devonian and Carboniferous which divided the Asturo-Leonesian basin into a shallow inner area and a slightly deeper outer area (compare van Loevezijn, 1983). Because this facies line is recognized not only in the Bernesga and Esla areas where the Sabero-Gordón line was defined (Fig. 1; 5.2) I propose a separate name for it: Intra-Asturo-Leonesian facies line. In the west the León line more or less separated the Bernesga and Esla areas in the Asturo-Leonesian basin from the Asturian geanticline; in the east it separated the Asturo-Leonesian and Palencian basins. The transitional area between the basins was during the Devonian a shallow area as may be concluded from the rapid decrease in thickness and the increase of hiatuses within the Devonian deposits in the Asturo-Leonesian basin towards this transitional area (compare Koopmans, 1962). Due to later deformations the width of this zone was greatly reduced and actually the Asturo-Leonesian facies and Palencian facies are partly overlapping.

The Asturo-Leonesian basin is divided into two parts by the southwest-northeast running Porma fault (Fig. 1; Rupke, 1965; Evers, 1967). Due to an early tilting of the eastern block Cambrian deposits are exposed here, actually forming the most important outcrop of the Cambrian in this basin. The area was already a topographical high during the Devonian (Pardomino high, Rupke, 1965). Van Adrichem Boogaert (1967: fig. 58) used the name in another way; I will follow Rupke's original definition. The distribution of facies and pattern of isopachs lead ten Have (1979) to the assumption that another structural high was present in the western part of the basin, the Somiedo high (Fig. 1). This high has the same orientation as the Pardomino high and is bounded by the León line: Savage (1979) assumes that the León line merges into the Villablino fault.

1.4. INTRODUCTION TO THE GEOLOGICAL HISTORY

A number of deep-seated faults originated during the Proterozoic, separating blocks with a distinct history (Krans, 1982). Sedimentation had a maximum in the West Asturian-Leonese zone where many km thick Lower Palaeozoic deposits were formed while the Cantabrian zone had more or less a geanticlinal character resulting in thinner deposits with different facies and important hiatuses. The central part of the Cantabrian zone maintained its geanticlinal character during the Devonian, initially it was bordered by a shallow shelf with reefs but since in the course of the Devonian the central area tended to emerge the importance of subtidal to intertidal siliciclastics increased. However, another part of the Cantabrian zone (the Palencian basin) became a starved basin with condensed sequences. The topography was reversed during the Early Carboniferous and from then on the main sedimentation took place on the former geanticline while the West Asturian-Leonese zone was uplifted. From Namurian to Stephanian the Variscan orogenesis took place, thick sequences were deposited in rapidly changing basins. Several folding phases took place, décollement nappes slid down centripetally and wrench-faults became active. Thick Stephanian deposits were deposited in basins bordered by large faults, in the Cantabrian zone as well as in the West Asturian-Leonese zone.

It is difficult to reconstruct the former extension of the Devonian and Carboniferous seas. In general in the West Asturian-Leonese zone there is a hiatus between the Silurian and Stephanian deposits and only very locally Lower Devonian deposits have been preserved in northwestern Spain. It seems probable, however, that the Devonian and Carboniferous seas had a much greater extension than is suggested by the actual distribution of their deposits. In the Celtiberic Chain, in the easternward extension of the West Asturian-Leonese zone, the Devonian is more complete and may give an impression which facies during the Devonian may have been present southwest of the Cantabrian zone. In the Celtiberic Chain there are at least some hundreds of meters (close to 1,000 m?) of Givetian to lower Frasnian strata. Sandstones and shales predominate but thin limestones occur. During the Frasnian the deposits became more and more distal turbiditic sandstones together with dark shales without shelly benthos. The late Frasnian is represented by dark to black shales with levels full of *Styliolina* and a few entomozoans. Shales were deposited during the Famennian (P. Carls, Braunschweig, pers. comm.). A thick turbidite sequence with dark fissile shales and greywackes was deposited during the early Carboniferous (Villena et al., 1981). These data suggest a stronger subsidence during Devonian and early Carboniferous in the West Asturian-

Leonese zone as compared with the Cantabrian zone, the sediments generally being siliciclastics derived from the Central-Iberian zone while thick reefal limestones are lacking. In the north the West Asturian-Leonese zone was bordered by the Cantabrian-Ebro massive, with reef limestones along the margin of the Cantabrian part, behind which there was the deeper Pyrenean trough (Carls, 1982). Thus, southward the Asturo-Leonesian inner shelf passed into the deeper shelf of the West Asturian-Leonese zone, which received sediment both from northern and southern sources.

2. ASTURO-LEONESIAN BASIN AND ASTURIAN GEANTICLINE

2.1. INTRODUCTION

In the Asturo-Leonesian basin (Fig. 1) the Devonian is rather thick with a predominantly calcareous lower and middle part and a siliciclastic upper part. During the Devonian period several large biostromal carbonate platforms formed (Santa Lucía Formation, Emsian and Eifelian; Portilla Formation, Givetian and Frasnian) separated by the siliciclastic Huergas Formation (Eifelian and Givetian). On the Asturian geanticline (Fig. 1) most of the Devonian is lacking with only a thin Famennian (Ermita Formation). During Tournaisian and Viséan times condensed sequences were formed in both areas (Ermita, Vegamián and Alba Formation). The principal divisions of the Devonian and Lower Carboniferous in the Asturo-Leonesian basin were introduced by Barrois (1882) for Asturias and by Comte (1959) for León. Their classifications were modified by later authors into formal formational names. New names were introduced for the sediments in particular areas, however, now that these deposits have been studied in more detail, more precise correlations have made many names superfluous. The lithostratigraphic nomenclature will be discussed in this chapter (Table 1).

2.2. PORTILLA AND CANDÁS FORMATIONS (GIVETIAN AND FRASNIAN)

2.2.1. Introduction

Barrois (1882, pp. 481-482) gave a description of the section Perán near Candás (Asturias; the areas are indicated in Fig. 1) including the stratotype of his "Calcaire de Candás" (Table 1).

Asturo-Leonesian basin and Asturian geanticline			Palencian basin
Asturias	León		
ALBA FORMATION Genicera Formation	ALBA FORMATION Griotte à *Goniatites crenistriata* Griotte de Puente de Alba Genicera Formation	Sella Formation Getino Formation	ALBA FORMATION Villabellaco Formation
	VEGAMIAN FORMATION Couches de Vegamián		VEGAMIAN FORMATION
ERMITA FORMATION Candamo Formation Caliza Blanca Baleas Formation	ERMITA FORMATION Grès de l'Ermitage Baleas Formation Caliza de Las Portillas	CAMPORREDONDO FORMATION Aguasalio Formation Piedrasecha Formation	VIDRIEROS FORMATION Montó-Schichten Verbios Formation
	FUEYO FORMATION Schistes de Fueyo		
			MURCIA FORMATION Moradillo-Sandstein
NOCEDO FORMATION Grès de Candás Arenisca del Naranco Piñeres-Sandstein	NOCEDO FORMATION Grès de Nocedo Calcaires de Valdoré		
CANDAS FORMATION Calcaire de Candás	PORTILLA FORMATION Calcaire de la Portilla Valcovero Formation		GUSTALAPIEDRA FORMATION

Table 1. The formations in the different areas (in capitals) and names which were used for the whole or part of the formations and are not used in this paper.

Comte (1959) described the "Calcaire de Portilla" with its stratotype on the right bank of the Arroyo de la Portilla, north of Matallana estación (Bernesga area; the sections in the southern part of the studied area are indicated in Encl. 1: Fig. 1). Reijers (1972) described the stratotype and some reference sections. Although the Candás and Portilla are equivalents for which one name would be sufficient, the names will be used for occurrences in the provinces of Asturias and León + Palencia respectively, because both names are widely accepted. In both stratotypes the middle siliciclastic member is lacking. When speaking in general terms I will refer to both formations together as Portilla Formation. Also other names have been introduced for the same deposits (Table 1) but these are not used in this paper.

Facies and correlations. - The limestones of the Portilla and the Candás are described in detail in a number of published and unpublished reports (amongst other Reijers, 1972, 1973, 1974, 1980; Mohanti, 1972; van der Baan, 1970; ten Have, 1979; Sanchez de la Torre, 1977) and therefore the sediments will not be described in detail here. Emphasis will be laid on biostratigraphical and lithological correlations and facies maps because these may give more insight in the history of sedimentation. The correlations were only possible after acceptance of a standard division into facies, which was applied to all the sections. This division is based on that of Reijers (1972) in whose paper each of the facies is described in detail. The standard division comprises: a back-reef facies with pack-, wacke- and mudstones (Reijers' facies b); a biostromal facies with pack- and grainstones (Reijers' facies c, partly); a biostromal facies with boundstones (Reijers' facies c, partly); a biohermal facies of boundstones (Reijers' facies d); a fore-reef facies with grain-, pack- and wackestones (Reijers' facies e and f) and a siliciclastic facies (Reijers' facies g). A further help for the lithological correlations were the biostratigraphical data obtained from conodonts.

Subdivision of the Portilla and the Candás Formation. - Several different local subdivisions have been proposed for both formations (Fig. 2) but it is preferable to use only one

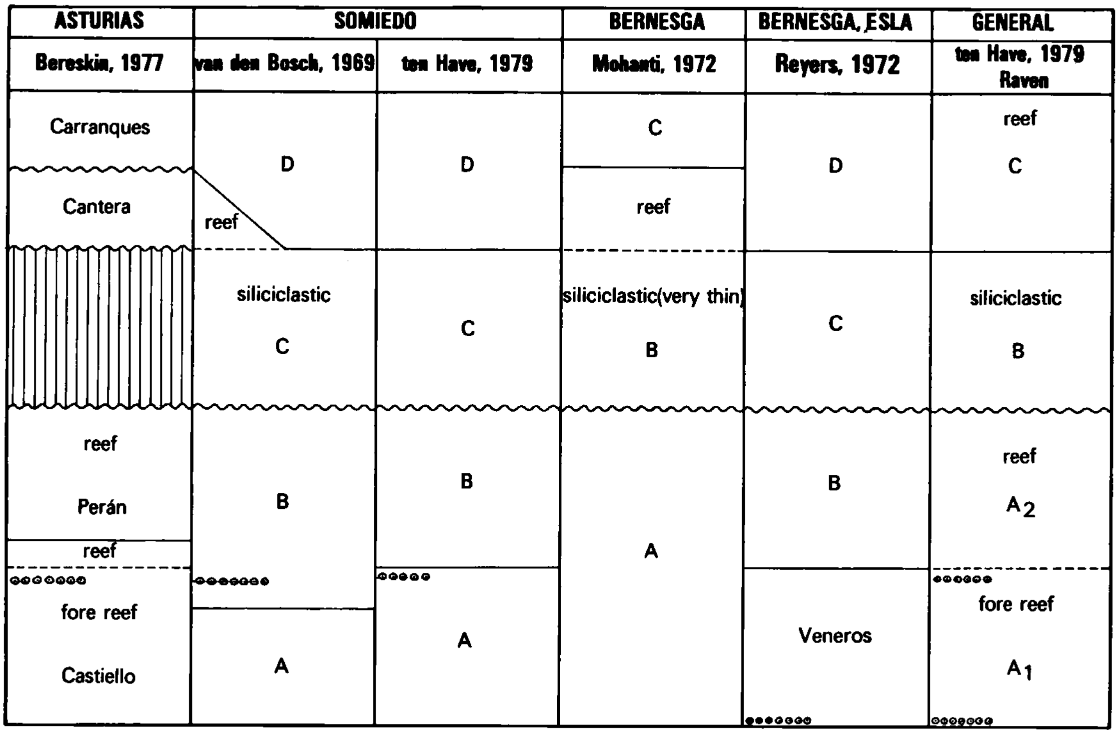

Fig. 2. Comparison of the different subdivisions in members of the Portilla and Candás. At the right side of the diagram the proposed informal subdivision (not to scale; ⊚⊚⊚ = oolite).

set of subdivisions for the whole region. The subdivision proposed by Bereskin (which is treated extensively in Sanchez de la Torre, 1977) was made for the Perlora syncline in Asturias but meant to facilitate correlation with other sections in Asturias and León. But the siliciclastic interval in the middle of the formation, present in nearly all other sections, is not present in that of Perlora. Further the contacts between the members are not defined, making a correlation with other sections difficult or doubtful. Against the subdivision proposed by Reijers (1972) there are other objections: due to his definitions of the members some of these interfinger, resulting in the occurrence of more lobes of one member in a single section. There is a strong opposition against such subdivisions (e.g. Wheeler & Mallory, 1953). Ten Have (1979) (Fig. 2) subdivided the Portilla Formation into three informal members which can be recognized immediately in the field: an upper and a lower biostromal limestone and a more siliciclastic middle member. These members are closely comparable to Reijers' members (Fig. 2) but are differently defined and therefore the objections against Reijers' subdivision do not exist for this subdivision. The same subdivision can also be used for the Candás Formation. This informal subdivision is accepted and used herein. For the description of the members I will refer often to section III in the Somiedo area (Encl. 1: Fig. 7).

2.2.2. *Member A*

The major boundary within the Portilla Formation is the lower surface of the first important siliciclastic interval. The sediments below this surface belong to member A, the sediments above it to members B and C but where oolitic grainstones are present immediately below the siliciclastics these also are included in member B. The erosive base of member B is an almost synchronous plane and therefore it was chosen as datum line in the correlations (Encl. 1: Figs. 2-7). The lowermost part of the Portilla Formation (unit A1) consists of cross-bedded, coarse-grained, detrital limestones (grainstones, packstones; section III on Encl. 1: Fig. 7). Generally, the limestones contain hard ferruginous streaks which accentuate the sedimentary structures. The limestone consists mainly of fragmented crinoids and bryozoans. Near the base and the top of the unit an oolite is present. The lower boundary with the Huergas Formation is not sharp: gradually more crinoid ossicles occur in the sands of the latter. In the entire basin reefs developed on top of these crinoidal and bryozoan limestones which had stabilized the substratum. This upper part of member A (unit A2) consists of biostromal and locally biohermal limestones: coarse to very coarse packstones, wackestones and locally boundstones. Occasionally crinoidal and bryozoan grainstones and packstones or fine-grained limestones (e.g. mudstones) are present. The transition from unit A1 into unit A2 may be gradual, the boundary is drawn at the top of the upper oolite in unit A1 or, where the oolite is not present, at the base of the first coral-rich or stromatoporoid-rich layer.

From the correlations (Encl. 1: Figs. 2-7) it becomes apparent that generally two phases of reef-building may be distinguished within unit A2, separated by fore-reef deposits. Only in the back-reef area near the Pardomino high the situation is more complicated: there as many as four phases are present, the middle two separated by an erosion surface (Encl. 1: Fig. 3) (Reijers & ten Have, 1983). I suppose that the number of phases of reef development could have been controlled by local epeirogenetic movements. After the interruption bioherms and biostromal boundstones re-appeared at the same places as before, indicating that these places were particularly favourable for reef-building organisms.

The limestone sedimentation generally began rather gradually: locally thin limestone layers are present in the uppermost part of the Huergas Formation. In the shallowest, innermost areas deposition of unit A1 started at the end of the Lower *varcus* Subzone (4.3; at Coallajú, Asturias and in the parautochthonous of Valdoré, Esla area). Elsewhere in the basin this sedimentation phase commenced at the beginning of the Middle *varcus* Subzone. The conodont samples prove that reef-building occurred during a relative short interval of time (Middle *varcus* Subzone; Fig. 27).

Because reef building was restricted to a short interval of time the facies ratio map for this member (Fig. 4) represents the facies distribution during the Middle *varcus* Subzone. For each section the ratios between the carbonate facies were plotted in a 100-percent triangle (Fig. 3; Krumbein & Sloss, 1963). From this plot it becomes apparent that most of the sections

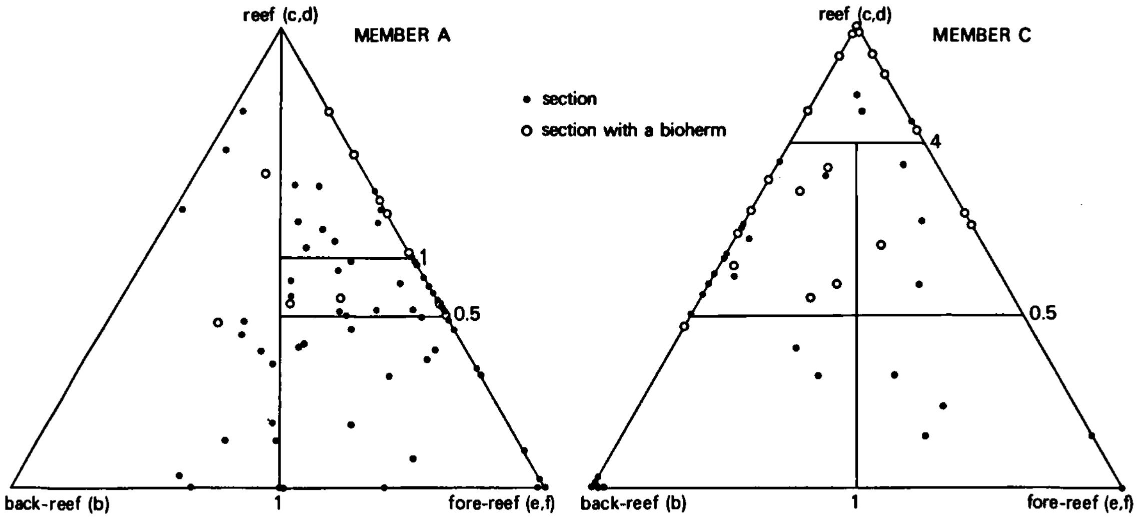

Fig. 3. Facies ratio triangles for members A and C of the Portilla Formation. The letters between brackets refer to the facies types distinguished by Reijers (1972).

occur at the fore-reef side of the triangle. At 12 sections boundstone build-ups are present which most probably are bioherms. Two palinspastical corrections were made for the facies ratio map: first the Esla nappe and the Valsurvio area were shifted southward to its original position and secondly the area south of the Sabero-Gordón line was shifted towards the west. Arguments for these displacements are given in 5.2. No other corrections were introduced in order to avoid a very uncertain basis map.

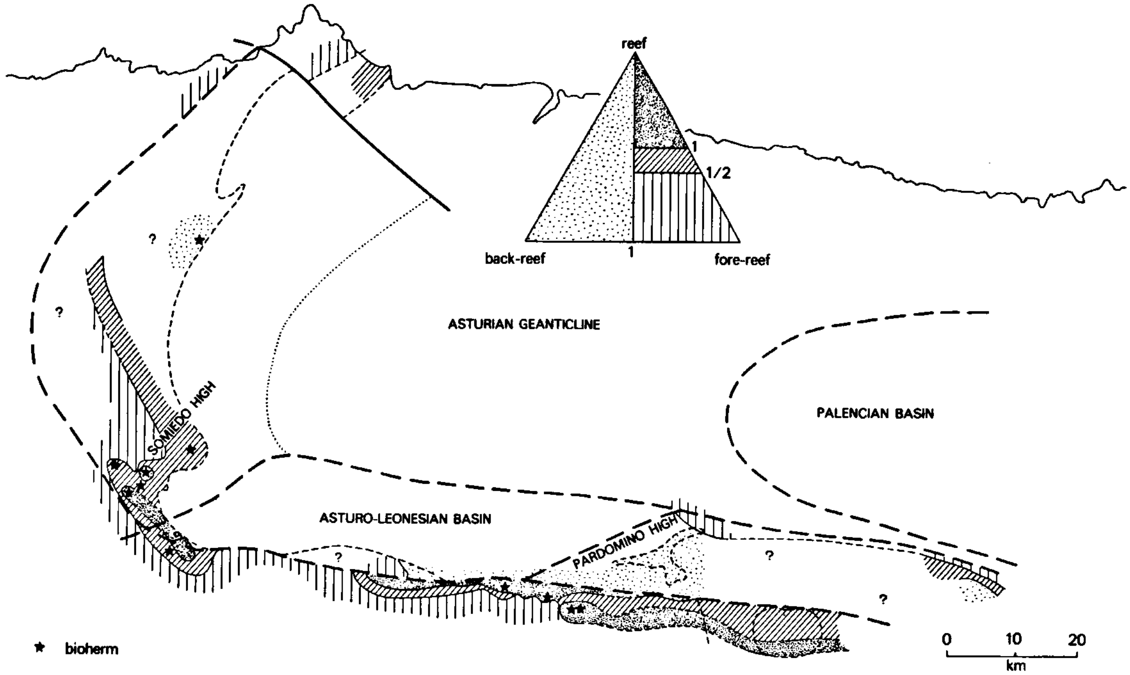

Fig. 4. Facies ratio map for member A of the Portilla Formation, based on the facies ratio triangle in Fig. 3 which is reproduced in reduced size in this figure. Different signatures indicate differences in facies ratio. Few palinspastic corrections were made (compare 5.3): the Esla nappe was shifted back to the south and the area south of the Sabero-Gordón line was shifted to the west. The position of the Valsurvio area is uncertain (5.2).

The facies ratio map (Fig. 4) demonstrates the presence of a narrow reef belt, mainly along the Intra-Asturo-Leonesian facies line. South of the reefs open marine fore-reef deposits are dominant; north of the reefs the back-reef facies prevails. There are two bays with fore-reef deposits: one in the Bernesga area, the other in the northern part of the Somiedo area. The bay in the Somiedo area was bordered in the south by a northeast-southwest oriented ridge which was surrounded by reefs (Encl. 1: Fig. 7). Ten Have (1979) interpreted it as a submarine ridge which he called the Somiedo high. Both in the extreme north of the Esla area and in the Ventanilla area deposits prevail which were interpreted as fore-reef deposits by Reijers (1972) and Wassink (1979) respectively (Encl. 1: Figs. 2 and 3). Fig. 4 demonstrates that in these areas along the León line boundstones occur, some of which may be interpreted as bioherms (A. Brouwer, Leiden, pers. comm.). Most probably there was a connection with the Palencian basin (Brouwer, 1962) where the shales of the Gustalapiedra Formation and nodular limestones of the Cardaño Member were deposited at the same time (3.2; Fig. 27). The limestones of member A in the Ventanilla and Valsurvio areas as well as in the north of the Esla area indeed contain much shale (Encl. 1: Figs. 2-3).

The thickest deposits occur at the extreme west of the Alba syncline (96 m in section PSA of Mohanti, 1972; compare Encl. 1: Fig. 6). Towards the southeast the thickness rapidly diminishes and near Sagüera the limestone disappears completely. The southermost limestone deposits are deformed by southward slumping (Pl. 1: Fig. 1; Encl. 1: Fig. 6). The axes of the slumps are parallel to the Intra-Asturo-Leonesian facies line. Reijers (1973) described these slumps but he failed to note that he studied only the plane parallel to the slump axes which lead to a wrong interpretation of the direction of slumping. Towards the east only a minor deformation occurs. Most probably south of the facies line the depth increased rapidly. Back-reef deposits occur only in few sections because most of these deposits in the inner zone of the basin were removed by later erosion.

2.2.3. Member B

Member B consists of sandstones, shales, limestones rich in siliciclastics, and thin intercalated limestone beds (section III, Encl. 1: Fig. 7 and Encl. 1: Fig. 3). The thickness of the member is highly variable, being thinnest in the Alba syncline and Asturias. In some sections siliciclastic beds are present within members A and C. Generally these are thin, small lenticles. Locally, however, such a bed may have a considerable thickness (e.g. at Luanco where thick sandstone beds occur within member C). In each section there is only one member B: the unit between both major biostromes. The boundary between members A and B is drawn between the upper biostromal layer and the lowermost siliciclastic layer or the thin oolithic grainstone at the base of this siliciclastic layer (if present). This is a sharp and locally erosive contact. At several localities the erosion cuts through compound corals (e.g. van der Baan, 1970; Mohanti, 1972). Transported skeletons of reefal organisms occur in the base of member B. Generally there are no indications for substantial erosion at the base of this member, but there are few exceptions, e.g. the Somiedo high (Encl. 1: Fig. 7).

Due to the difference in resistance to weathering, the middle member forms a depression between two prominent limestones. Since the siliciclastics generally are poorly exposed, rather few sedimentary structures were observed. In the Ventanilla and Valsurvio areas the member consists of (silty) shales with limestone lenticles. In these areas the top of the member is always eroded, being overlain by other formations (Encl. 1: Fig. 2; Wassink, 1979; Witte, 1980). In the Esla area the sequences belonging to the outer facies area have well-sorted, very fine to fine-grained quartz arenites which are parallel-laminated or bioturbated, containing horizontal and vertical burrows. Further greenish shales occur, containing brachiopods, bryozoans and ostracods. Unlike the sandstones these shales also occur to the north of the facies line (Encl. 1: Fig. 3; van der Baan, 1970; Reijers, 1972). Member B is thinner there where bioherms were present in member A. This indicates that these indeed rose above their environment.

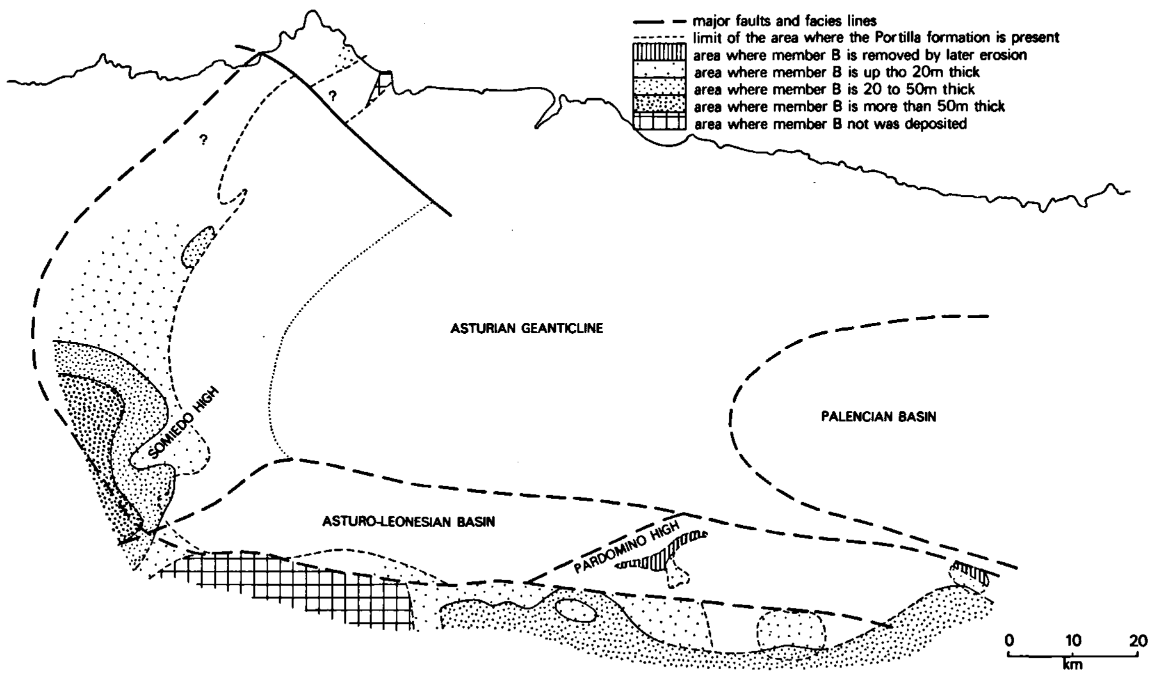

Fig. 5. Schematic representation of isoliths in member B of the Portilla Formation (compare with Fig. 4).

Member B is absent in a large part of the Bernesga area, it is only found in the Pedroso syncline and in the western extreme of the Alba syncline where there is a thin sandy intercalation between members A and C (Fig. 5; Encl. 1: Figs. 4-6). Where member B is absent it may be impossible to distinguish between members A and C. In the Somiedo area the siliciclastic deposits are much thicker (Fig. 5) except in a narrow northeast-soutwest oriented zone on the Somiedo high (2.2.2) where member B is thin (Encl. 1: Fig. 7). South of the facies line a thickness of over 60 m may be reached, thus comprising more than half of the formation. There the deposits consist almost exclusively of sandstones, quartz arenites as well as arkosic arenites and greywackes. Sedimentary structures include low-angle large-scale cross-bedding, shallow channels, parallel lamination and bioturbation (vertical burrows) (ten Have, 1979). In Asturias the member may be present as a thin sandstone or a shale bed. The greatest thickness is reached at Coallajú where 20 m of sandstone, limestone and shale are exposed. In all these areas except Asturias, oolites occur within member B. Up to three levels may occur within one section. Some of these limestone layers are preserved in a rather large area but most frequently they form only small lenticles with sharp lower and upper boundaries.

The sandy sediments of member B are interpreted as shallow-marine sands deposited mainly in intertidal and supratidal environments, partly as bars and barriers. The shape of the lithosomes and the sedimentary structures indicate that channels transported the sand from the source area and that part of the sand was further distributed along the coast by longshore drift. The thick deposits in the Somiedo area indicate that the major source of the sand was near that area. As the sand can not have originated from within the area itself, the source must be sought to the northeast or southwest. Ten Have (1979) supposes that the sand originated from the Narcea anticlinorium, but the facies of the calcareous intercalations in member B and the facies in members A and C indicate that depth increased towards the southwest. The sandstones are thicker in the southwest, partly compensating the more rapid subsidence in that area, and the source was somewhere on the Asturian geanticline. Another source was also at the geanticline but close to the Pardomino high. The absence of sandstones from the Esla parautochthonous (where shales were deposited, Encl. 1: Fig. 3) indicates that a channel discharged beyond this area. In the Bernesga area member B is absent from the main part of the area south of the Intra-Asturo-Leonesian facies line. Only at the eastern and western extremes a few m of crinoidal grainstones were deposited in a fore-reef environment, contemporaneously with member B. These are included in member C. The limestone lenticles and oolites within member B were deposited in areas where the supply of siliciclastic sediments failed temporarily. Carbonate deposition was followed by erosion (partly) removing these deposits. The repeated erosion explains the relatively limited thickness of member B which represents an interval of time much longer than that during which member A was deposited: the Upper *varcus* Subzone and a large part of the *hermanni-cristatus* Zone (Fig. 27).

2.2.4. Member C

Member C consists mainly of (locally silicified) thick-bedded, biostromal and biohermal limestones alternating with some crinoidal-bryozoan calcarenites, fine-grained pack-, wacke- and mudstones and siliciclastic layers (section III, Encl. 1: Fig. 7). This member is rich in fossils, especially thamnopores but also branching rugose corals, solitary corals and brachiopods. Locally the boundary between members B and C is sharp and conformable but elsewhere it must be drawn arbitrarily between the uppermost thick siliciclastic layer of the former and the lowermost thick biostromal layer of the latter. The boundary with the overlying Nocedo Formation is discussed in 2.2.4 and 2.3.1.

It is more difficult to make a palaeogeographical reconstruction for this member than for member A because: (1) member C represents a longer interval of time during which more changes in the facies distribution occurred. Member C was deposited from *hermanni-cristatus* Zone to Lower *asymmetricus* Zone but where member B is lacking it includes also the Upper *varcus* Zone. Thus member C is partly contemporaneous with member B but since only a few meters of grainstones were deposited in that interval this does not have much influence on the map. (2) Member C comprises

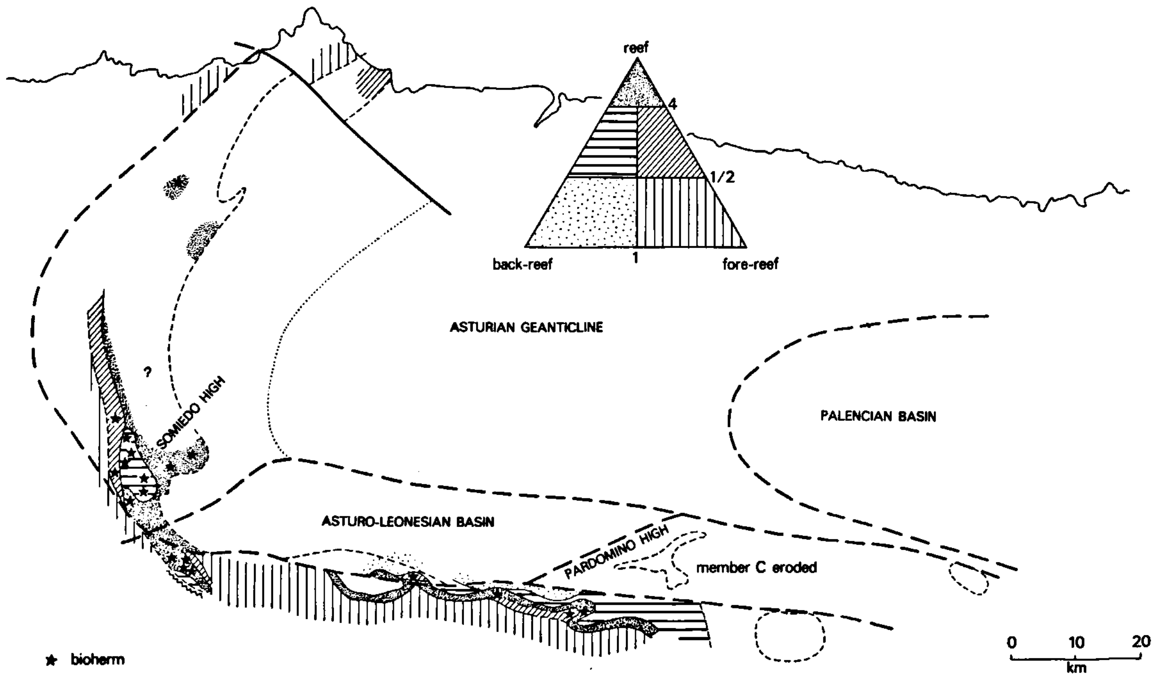

Fig. 6. Facies ratio map for member C of the Portilla Formation, based on the facies ratio triangle in Fig. 3 which is reproduced in reduced size in this figure. Different signatures indicate differences in facies ratio (compare Figs. 4 and 5).

relatively more reef deposits which contain few conodonts and thus the age of the sediments is not always precisely kown. In spite of these complications a facies ratio map has been drawn for member C (Fig. 6). Fig. 3 illustrates that these ratios changed considerably: in member A most sections were in the right-hand (fore-reef) half of the triangle with about as much reef as fore-reef facies; in member C most sections are in the left-hand (back-reef) half of the triangle, with more reef facies. This becomes even clearer when it is appreciated that from almost all the sections in the back-reef half of the plot for member A, the sediments of member C were removed by erosion or never deposited. Furthermore in contrast to the 12 sections which contain bio-hermal deposits in member A, 20 sections contain such sediments in member C. This is largely due to a further progradation of the reefs into the fore-reef environment which was evident from the beginning and continued during the deposition of member C.

In the eastern part of the Asturo-Leonesian basin the erosion is most pronounced (Fig. 6): both in the Ventanilla and Valsurvio areas member C is absent and in the Esla parautochthonous only a thin remnant is preserved (Encl. 1: Figs. 2 and 3). In the Esla nappe, the Pedroso syn-cline and the eastern part of the Alba syncline the unit is (partly) preserved, consisting mainly of biostromal limestones (Encl. 1: Figs. 3-5). In relation to member A the reef prograded slightly towards the south and the west (compare Figs. 4 and 6). Bioherms occurred more or less at the same localities which probably were particularly suited for reef growth. The back-reef facies now makes up an important part of many sections south of the facies line. This is in accordance with the progradation of the reefs. Farther west there was still a bay with fore-reef facies in the central part of the Bernesga area (Fig. 6). In this bay sedimentation was very slow as is evidenced by the thin deposits (about 15 m, Encl. 1: Fig. 5) representing the Upper *varcus* Subzone to Lowermost *asymmetricus* Zone (García Alcalde et al., 1979). The reefs probably built out into this area but this progradation was very slow. It is evidenced by the "large-scale cross-bedding" described by Mohanti (1972) from the western extreme of the Alba syncline (Pl. 1: Fig. 2). Between his sections PSJ and PSN there was a bioherm. The cross-bedding is present within the biohermal deposits and can be traced well into the back-reef de-posits southwest of it (Pl. 1: Fig. 3). This cross-bedding illustrates the progradation of the reef towards the northeast into the embayment. The erosion surfaces with thin karst within the back-reef area indicate the existence of a shallow ridge which emerged now and then (Mohanti, 1972). I do not agree with Reijers (1972) who interpretes the cross-bedding as formed by pro-gradation of the reef towards the nortwest and Mohanti's back-reef facies as a fore-reef facies. Although this would better fit in the palaeogeographical reconstruction the interpretation of Mohanti is better in accordance with the sedimentary evidence. Later on silty grainstones were deposited near Sagüera and silty biostromal boundstones near Mirantes: probably the southern part of the area had subsided.

The embayment which was present in the Somiedo area (member A) no longer existed when member C was deposited (compare Figs. 4 and 6). This may be due to the sedimentation of thick siliciclastic deposits in this area (Fig. 5). The reefs prograded towards the west. Besides that bioherms occurred at all the localities where bioherms were present in member A, they occurred also in the former embayment (the bioherms are only indicated where they are recognized within sections, they may be present too in between these sections). Within the reef belt there is a small area which alternately fell within the reef and the back-reef area. Farther north in Asturias reef growth was important too (Fig. 6). At Luanco (Asturias) the lower part of member C (212.5 to 427.5 m in the section measured by García López, 1976: Fig. 2) consists of an alternation of shales and reef limestones. I recognized seven limestone beds which vary in thickness between approximately 7 and 25 m. Generally the lower boundary of such a limestone bed is sharp; at the base of these limestone beds there are framestones composed of thick lamellar and spherical stromatoporoids. Among the stromatoporoids spherical and branching corals (both tabulate and rugose) occur, accompanied by bryozoans, solitary rugose corals, brachiopods, crinoids, etc. Towards the top of the limestone the diversity of the organisms decreases: first the lamellar and the spherical reef builders disappear. The upper part of each sequence consists mainly of branching corals. Gradually the amount of shale among these corals increases and via wackestones the limestone passes into mudstones and shales without microfossils (Fig. 7). Such a succession indicates a general increase in growth rate of the organisms and a decrease in energy of the environment. The gradual change in the fauna indicates that submergence was faster than the growth of the frame-builders. For such a situation Hoffmann & Narkiewicz (1977) predicted a reverse of the normal developmental pattern of reefs. The normal sequence (Walker & Alberstadt, 1975) is also the general sequence present in the Portilla (Reijers, 1980: Fig. 8), the Candás and in the Crémenes Limestone (2.3.3). With the decreasing energy less clay was winnowed away, the organisms could not keep pace with sedimentation and were suffocated in the clay. Most probably this happened when the sea bottom was below wave-base. A new sequence could start each time the supply of siliciclastics stopped. Maybe nearby faults acted as sediment trap.

Locally the upper part of member C is not preserved, e.g. in the Somiedo area at the Puerto del Somiedo (ten Have, 1979: sections XVI, VII; Encl. 1: Fig. 7), Lumajo and northwest of Quejo (ten Have, 1979, sections II and V; Encl. 1: Fig. 5). This is confirmed by conodont faunas present near the top of the Portilla. Just at the localities where the uppermost part of member C contains older faunas, member C (and the entire formation) is thin. This is ex-plained by slight uplift and erosion of the Somiedo high before (non-deposition) or after the deposition of this part of member C (erosion). Also in other areas erosion of the Portilla Formation was recognized although it is difficult to ascertain this in the field. At Coallajú (Asturias) member C is eroded and limestones of the Ermita Formation were deposited on top of it. Also in other sections in that area the Candás was eroded. In the Bernesga area, at Huergas de Gordón, the youngest Portilla deposits are from the upper part of the *hermanni-cristatus*

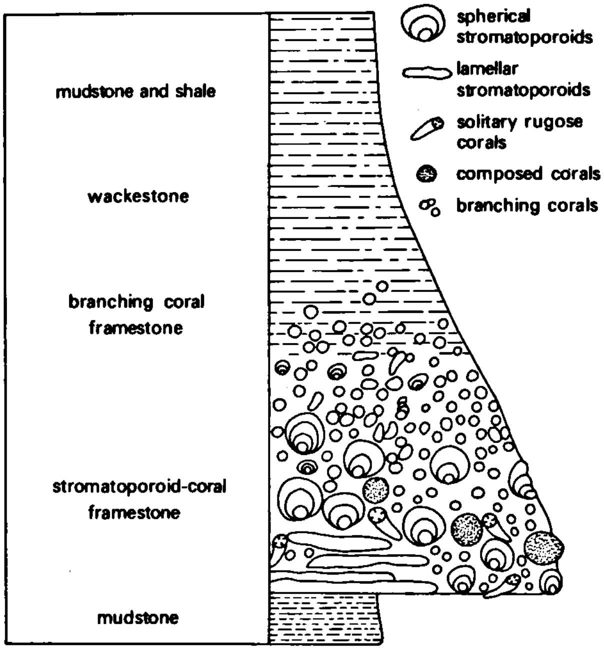

Fig. 7. Biostrome in the Candás Formation near Luanco (Asturias). The reefs developed when supply of siliciclastic sediments decreased and disappeared when depth and the deposition of clay increased. The shape of the stromatoporoids and corals indicates an increase in velocity of growth (not to scale).

Zone (García Alcalde et al., 1979) so that there is a small hiatus between Portilla and Nocedo. At Sagüera the uppermost part of the Portilla was deposited during the Lower *asymmetricus* Zone (Becker et al., 1979) and from there Buggisch et al. (1982) describe deep karst fissures indicating an interval of emergence.*) At Beberino (Bernesga area; van Staalduinen, 1973: fig. 8; Frankenfeld, 1981: fig. 28) and northwest of Cistierna (Esla area) the top of the Portilla is very irregular. There, the top of the formation was eroded by channels which were later filled with silty limestones which often were dolomitized afterwards (van Staalduinen, 1973; Sleumer, 1969; Frankenfeld, 1981). In the Esla parautochthonous and in the Valsurvio and Ventanilla areas the Portilla is affected much more by erosion which may reach into members B or A (Encl. 1: Figs. 2 and 3). In the northern part of the Valsurvio area there is a section where only 5 m of limestone are preserved (Witte, 1980). In the Ventanilla area locally karst fissures penetrate into the Portilla, being filled with sandstones of the Ermita Formation (N. Schelling, Leiden, pers. comm.). Thus before the deposition of the Nocedo Formation over a large area there was a regression. Erosion at some localities was accompanied by strong karst weathering but at other localities it left almost no trace.

2.3. NOCEDO FORMATION (FRASNIAN)

2.3.1. Introduction to the Upper Devonian deposits (Table 1)

Comte (1959: pp. 190-193) introduced the "Grès de Nocedo" with a type section near the village Nocedo in the Bernesga valley. The same section was described by Vilas Minondo (1971: pp. 138-142). In the type section 400 m of mainly sandy sediments are present (Fig. 8). On this sequence there are 100 m of shales, called "Schistes de Fueyo" (Comte, 1959: p. 193) (Fig. 8). Evers (1967), van den Bosch (1969) and Van Loevezijn & Raven (1983) included the Fueyo shales in the Nocedo Formation. Because of an important hiatus existing between Nocedo and Fueyo (2.4.4) herein, just as in Vilas Minondo (1971), the Nocedo and Fueyo are distinguished as separate formations.

The lower boundary of the Nocedo Formation was based on the first occurrence of the brachiopod *Apousiella bouchardi* (Comte, 1959: p. 241). In the International Stratigraphic Guide (Hedberg, 1976) it is recommended to define boundaries between lithostratigraphical units on lithological criteria. Therefore I propose to draw the boundary between Portilla and Nocedo above the uppermost limestone bed with a thickness of at least 10 cm.

Comte (1959: pp. 193-195) further described the "Grès de l'Ermitage" with its type section near the Ermita del Buen Suceso near Huergas de Gordón in the Bernesga valley (Fig. 8). This section lies south of the Intra-Asturo-Leonesian facies line. Sjerp (1967: p. 71) formally introduced the name Ermita Formation and described some members for the Isidro-Porma area. Van Adrichem Boogaert (1967: p. 159) proposed a new type section near Camplongo (also in the Bernesga valley but north of the facies line), where the Ermita is very thin. It is preferable to maintain the original type section because it is situated immediately overlying the type section of the Nocedo and Fueyo and because the Ermita is much thicker and more complete there. Since the formation has a distinctive development on either side of the facies line, the section near Camplongo may be used as a reference section for the inner area.

For the sections in which the boundary between Nocedo and Ermita can not be determined with certainty Brouwer (1962, 1968) proposed the name Aguasalio Formation. I prefer to draw an arbitrary boundary where necessary. Only in the Valsurvio area it is impossible to distinguish a

*) East of Sagüera the entire Portilla Formation is lacking (Pl.1: Fig. 7).

boundary in the thick, sandy Upper Devonian. For these deposits I use the name Camporredondo Formation (Koopmans, 1962). In order to get a uniform subdivision in all areas, herein the uppermost shales of the Valcovero Formation are included in the Camporredondo Formation.

Other names have also been used for the Upper Devonian deposits (Table 1) but none of these are used herein. Some of the names are discussed below. In his description of the Perán section Barrois (1882: p. 482) mentions unit 19: a red sandstone, 25 m thick, without fossils, lying regularly on the preceding limestones of the "Calcaire de Candás". He also correlates this unit with the "Grès de Cué", which comprise the thick quartzites in the Sierra de Cué, southeast of Llanes. Adaro & Junquera (1916) demonstrated that at least the major part of these deposits belong to the Ordovician and proposed the name "Arenisca del Naranco" for the sandstone on top of the Calcaire de Candás because, by error, they thought that this was the same sandstone as that outcropping at Monte Naranco which is known to be Middle Devonian. Barrois (1882: p. 495) used the name "Grès de Candás à *Gosseletia*" pre-empting the name proposed for the Upper Devonian sandstones: "Grès de Candás" (Comte, 1936) which should therefore not be used.

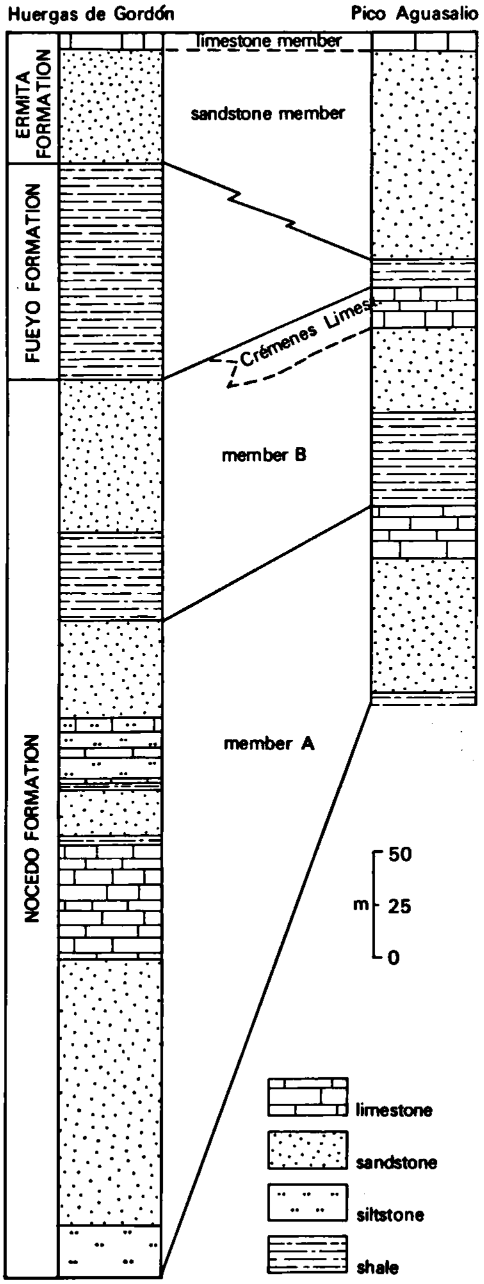

Fig. 8. Subdivision in lithological units of the Upper Devonian in two sections in the Asturo-Leonesian basin.

The "Piñeres-Sandstein" with its stratotype near Piñeres (northwest of Candás) was introduced by Radig (1961: p. 264). A description of this section was given in Radig (1958) but has never been published. The lower and upper limit of the stratotype are badly exposed but from correlations with the sections at Perán and Playa de Gargantera it becomes clear that these deposits were formed concurrently with the Nocedo Formation. Therefore the name Nocedo Formation, which has priority over the name introduced by Radig, is used for the late Frasnian siliciclastic deposits in Asturias.

Wagner et al. (1971) discovered conodont faunas at Aviados, north of Pola de Gordón and at Beberino, belonging to the *anchoralis-latus* Zone (herein part of these are included in other zones, 4.3) in the uppermost part of the Ermita Formation. Because of the supposed difference in age between the upper part of the Ermita limestone in most areas (early Tournaisian) and in these three sections (supposed to be early Viséan) they introduced the Baleas Formation with exactly the same facies as the underlying Ermita. Therefore the Baleas can not be distinguished from the Ermita on lithological criteria. Further research (2.5.6 and 4.3) proved that the hiatus between both limestones is less important than supposed by Wagner et al. and maybe even absent. Therefore there are no reasons to distinguish a separate Baleas Formation.

Marquínez (1978) introduced the name "Caliza de las Portillas" for the Famennian to lower Carboniferous limestone and sandstone in the Picos de Europa. The name Ermita Formation, however, has priority and is used herein. The "Caliza de Candamo" was defined in an unpublished thesisby Pello (1972) from a type section at San Román de Candamo (north of Grado in Asturias). Earlier it was named "Caliza Blanca" (Pello, 1968). None of these names was introduced formally and because both are equivalents of the Ermita Formation I prefer to use that name.

Subdivision of the formations. - South of the Intra-Asturo-Leonesian facies line generally three regressive sequences may be recognized (Fig. 8). Each of the sequences begins with shales, passing into sandstones and capped by a limestone.*) These three sequences can be recognized in a large part of the Asturo-Leonesian basin. The two lower sequences were designated as informal units by van Loevezijn & Raven (1983): units A and B respectively. These are herein indicated as members A and B. Van Loevezijn (1983) proposed formal names for these units: Gordón Member for the lower unit and Millar Member for the upper unit. In the Esla area one more unit is distinguished, a limestone in the top of member B. It was described by Westbroek (1964) as "Calcaire de Crémenes", herein referred to as Crémenes Limestone, with its type section at Pico Aguasalio, east of Crémenes (Fig. 13).

*) Locally the upper part of these sequences is eroded.

Sjerp (1967) distinguished two units in the Ermita Formation: the Valverde Ferruginous Sandstone Member and the Mampodre Limestone and Shale Member. These names may be used in the Isidro area. I prefer not to use the name Valverde Member outside this area because generally the sandstone is much thicker and only locally in the Asturo-Leonesian basin it is ferruginous. In this paper the names sandstone unit and limestone unit will be used informally.

2.3.2. Member A

2.3.2.1. Introduction. - This member is the part of the Nocedo Formation with the greatest horizontal distribution. It is present in a great deal of the Asturo-Leonesian basin, particularly the outer part, but is absent from the innermost part. The thickness of the member varies from 0 to 306 m at Huergas de Gordón (Bernesga area: Fig. 9; van Loevezijn, 1982).

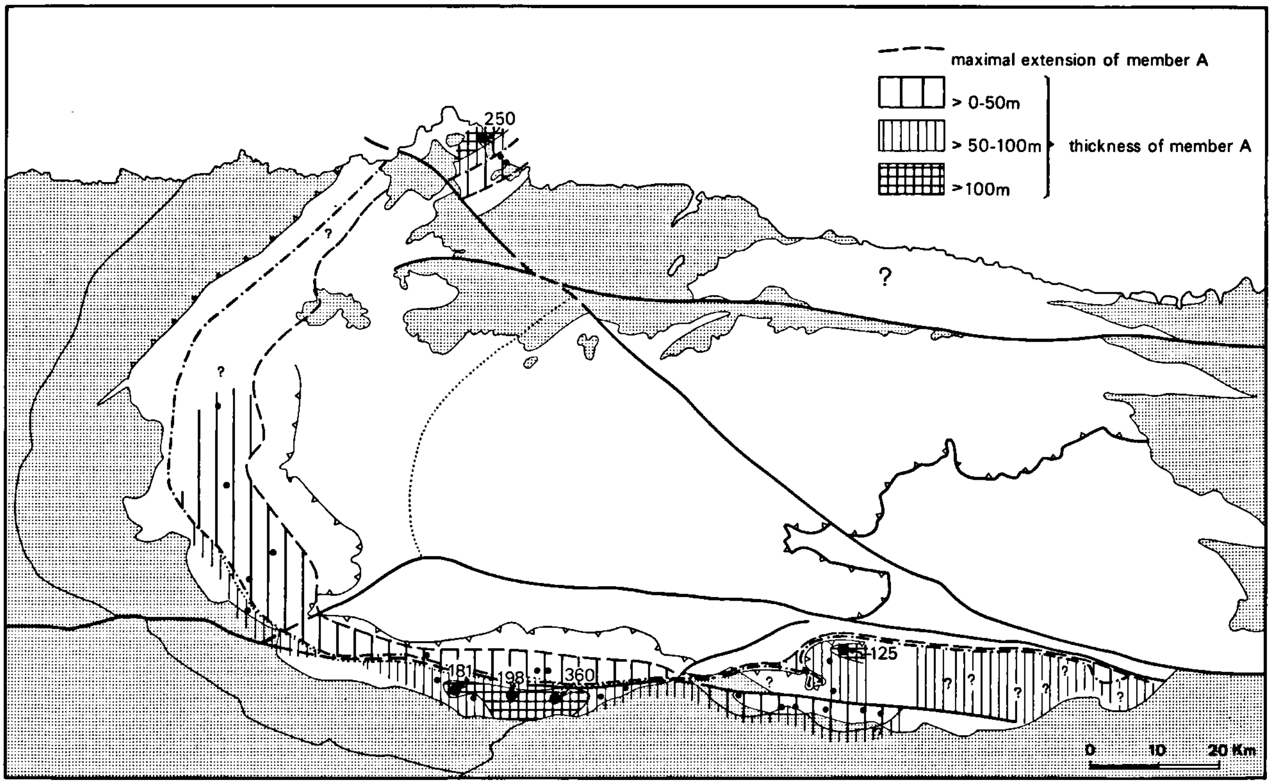

Fig. 9. Thickness and distribution of member A of the Nocedo Formation. It is not known whether deposits of this unit are present in the Coastal Ranges or not (2.5.2).

2.3.2.2. Sediment petrology. - Van Loevezijn & Raven (1983) distinguished five different facies (a to e) in the Upper Devonian deposits in the province of León. The same subdivision will be applied to this unit throughout the entire Cantabrian region.

The lowermost part of the member generally consists of bioturbated shales or siltstones (Fig. 10b). Facies e, which consists of alternating shales and sandstone beds, occurs only in a small part of the Bernesga area. In these deposits goniatites, small bivalves (*Buchiola*) and tentaculites occur. Facies d, which consists of completely bioturbated sandy shales and siltstones, occurs in almost all the sections, also in the Valsurvio and Ventanilla areas where the shales are dark grey. In these deposits bryozoans, bivalves, solitary rugose corals and crinoids are common.

Higher in the member the fine-grained sediments pass into mature quartz arenites, locally quartzites (facies c, Fig. 10b). These quartz arenites consist of quartz grains and a low percentage of matrix or cement formed by clay minerals, haematite and carbonate. Feldspar and other immature minerals are absent and heavy minerals occur only in extremely small quantities. The grains are moderately well to well-sorted and are coarser higher in the sequence. In the sandstones physical sedimentary structures are abundant. Large-scale current ripples and bars are most common. Fossils are scarce, the most common being crinoid ossicles.

The sandstones of facies c may alternate with or pass into cross-bedded crinoidal grainstones (facies b). The cross-bedding is formed by large-scale current ripples and small bars. The grainstones largely consist of grey crinoid ossicles. Further numerous fossil fragments occur which have a red colour due to haematite impregnation. The fragments were transported and redeposited after impregnation, often in thin laminae accentuating the cross-bedded structures. Besides crinoids also bryozoans and brachiopods are common. Locally other organisms such as solitary and compound corals (*Alveolites, Chaetetes*), stromatoporoids and gastropods (*Platyceras*) occur. Grey mudstones and wackestones may occur on, or alternating with, these grainstones. At Perán and Playa de Carranqués the limestone consists of red, sandy, cross-bedded crinoidal grainstones and a boundstone of branching rugose and tabulate corals. The uppermost part of the limestone consists of mudstone with calcite and fluorite veins, capped with a thin, red crinoidal grainstone. At Perán above this limestone there is a fine-grained red sandstone with fragments of *Hexagonaria* near the top. At other sections above the limestone locally an alternation of shales, sandstones and limestones occurs (facies a, known only from Huergas de Gordón, Bernesga area) or a sandstone (facies c). This is the general sequence present. Of course local variations occur, e.g. at Llombera and Matallana (Bernesga area) small limestone

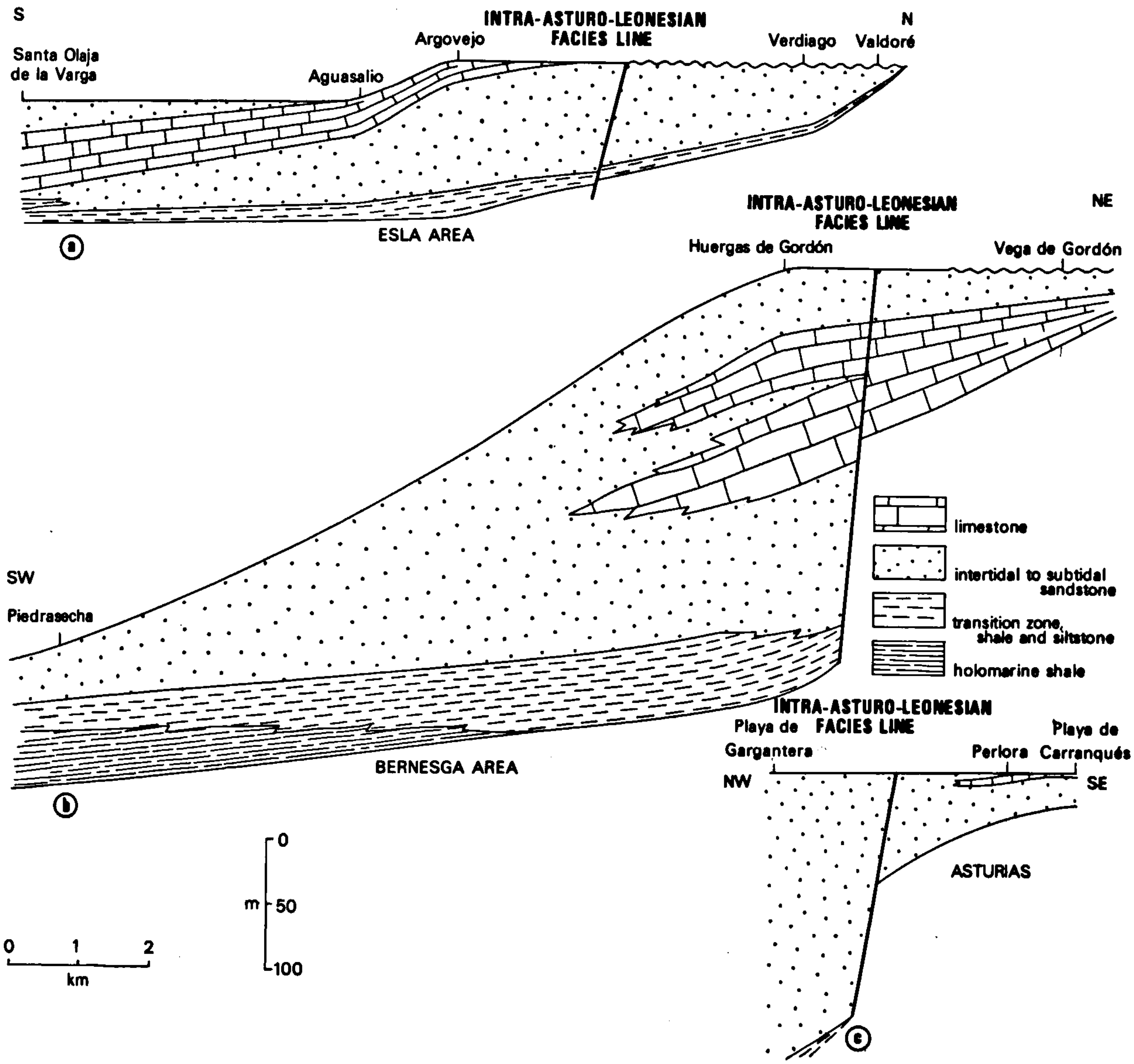

Fig. 10. Palinspastically reconstructed cross-sections through member A of the Nocedo Formation: (a) Esla area, (b) Bernesga area, (c) Asturias. The cross-sections are drawn at the same scale but the vertical scale is exaggerated.

lenticles are present in the lower part of this unit (Encl. 1: Fig. 8). The relative importance of the different facies is variable too. In general the facies mentioned for the lower part of the sequence (d and e) are the best developed in the outermost areas (in particular the Alba syncline) and those of the upper part of the sequence (a and b) in the inner areas. Facies c makes up the major part of this member.

Member A is thicker in the outer areas (Fig. 9). The cross-sections (Fig. 10) illustrate the difference between the various parts of the basin. In the Bernesga area a regressive sequence is developed, which is truncated in the north. The thickest deposits were formed just south of the facies line (compare Figs. 9 and 10). In the northern area a carbonate platform was present. In the Esla area the same relationship between the facies is present but there the influence of the facies line was less. The carbonate platform extended farther south. In Asturias the difference between the shallower platform with reef limestones and the thick barrier deposits (shallow water) in the northwest indicates that also in this area there was an active facies line and that the northwestern area was subsiding rapidly. Probably farther northwest the thickness of the deposits decreased, just as in the other areas.

The transition from a shallow carbonate platform towards a deeper area southeast of it, near Sagüera (Bernesga area) has been shown by van Loevezijn & Raven (1983: fig. 5). The difference in depth was compensated by more rapid sedimentation in the deeper areas, partly by mass-flows and limestone blocks originating from the carbonate platform of the Portilla Formation. The higher parts of member A have the same facies and thickness in both areas, indicating that the difference in depth had disappeared.

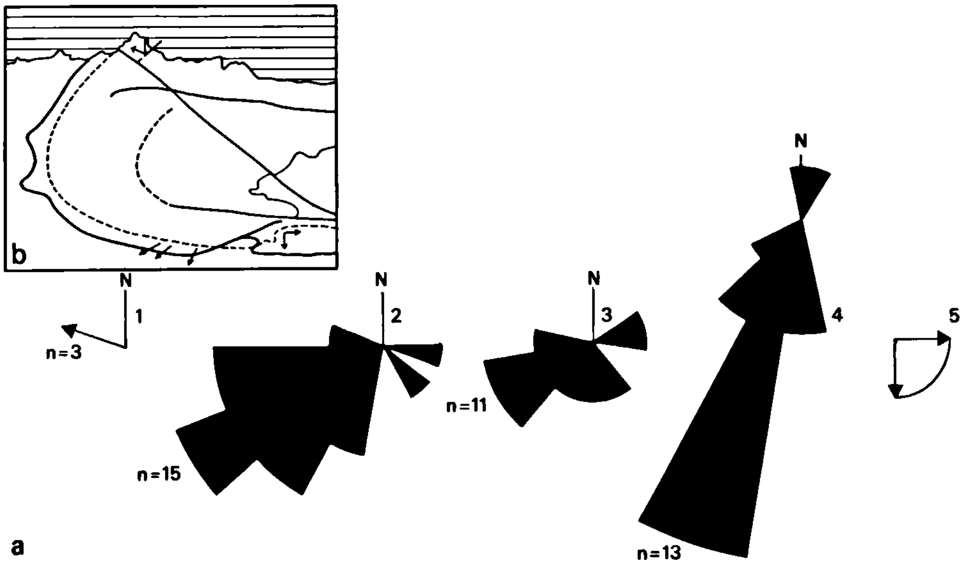

Fig. 11. Current directions measured in large-scale current ripples in member A of the Nocedo Formation: (a) Playa de Gargantera, (b) Sagüera de Coo (1970), (c) Portilla de Luna, van Loevezijn (1982), (d) Huergas de Gordón, (e) Esla area, Raven (1980a).

Measurements in large-scale current ripples in the Esla area generally indicate transport towards directions between south and east (Fig. 11a). In the Bernesga area at Huergas de Gordón the vector mean of the current directions measured in a sandstone bed just above the main limestone is towards N210°E (n=13; Pl. 2: Fig. 6). In the grainstones at Portilla de Luna van Loevezijn (1982) found a mean of about N240°E (n=11) and at Sagüera de Coo (1970) found a mean of about N250°E (n=15). At Playa de Gargantera (Asturias) only few ripples were measured with a mean of N285°E (n=3). At all these locations only few (generally smaller) ripples indicate currents in the opposite direction. Everywhere the currents were more or less perpendicular to the coast (Fig. 11b) indicating a dominant seaward current and sediment transport.

At many localities in the Asturian part of the Asturo-Leonesian basin member A consists of dark red sandstone (e.g. Playa de Gargantera (Pl. 2: Fig. 1), Perán, Playa de Carranqués, Cigüedres, north of Pola de Somiedo). Such red beds occur also dispersed within the other sections. These sands are well-sorted barrier deposits which owe their colour to a haematite matrix or cement. The occurrence of red beds may be caused by affluxes of brackish or fresh water (Franke & Paul, 1982).

2.3.2.3. Boundaries. - The lower boundary of the Nocedo Formation generally is sharp, especially in the area north of the facies line, but at some locations the transition is more gradual. In that case the boundary is drawn above the uppermost limestone bed with a thickness of at least 10 cm. In the eastern part of the Asturo-Leonesian basin the Portilla is very incomplete and there the erosion before deposition of the Nocedo was most pronounced (2.2.3). At Perán (Asturias) the boundary between Candás and Nocedo is sharp. On dark grey marls with stromatoporoids and corals in growth position lies a red sandstone. The boundary surface is full of large burrows (3 to 6 cm broad and meters long) indicating an interval of non-deposition. In the southeastern part of the Alba syncline the Nocedo rests upon the Santa Lucía Formation (Pl. 1: Fig. 7). Also the top of the member may be eroded. This is generally the case in the area north of the facies line where there is a large hiatus between this unit and the overlying Ermita Formation. South of the facies line there are generally no indications for erosion of the top.

2.3.2.4. Fossils and the age of the member. - The most detailed information on the age of the sediments was obtained from conodonts. A problem is that in most sections limestones occur only in the upper part of the unit. These were generally formed during the Lower and Middle *asymmetricus* Zone. The few samples with a different age are important for understanding the sedimentation pattern. The youngest sediments of the Portilla Formation were formed during the Lowermost or Lower *asymmetricus* Zone. The oldest sediments of the Nocedo are proved in sections with a calcareous base, e.g. the Llombera section (Fig. 12) where sediments of the lowermost part of the Lower *asymmetricus* Zone are preserved. At Olleros de Alba there is a large limestone lenticle. The older deposits are covered by Stephanian rocks but the base of the lenticle is at the boundary between the Lowermost and Lower *asymmetricus* Zone. At Matallana, however, the base of member A is in the Middle *asymmetricus* Zone (Encl. 1: Fig. 8). Before that moment the Pardomino high (including the area near Aviados and Matallana) emerged. Some of the sediments may have been deposited later than the Middle *asymmetricus* Zone, but only at Huergas de Gordón the age was proved (Upper *asymmetricus* Zone and *Ancyrognathus triangularis* Zone, García Alcalde et al., 1979).

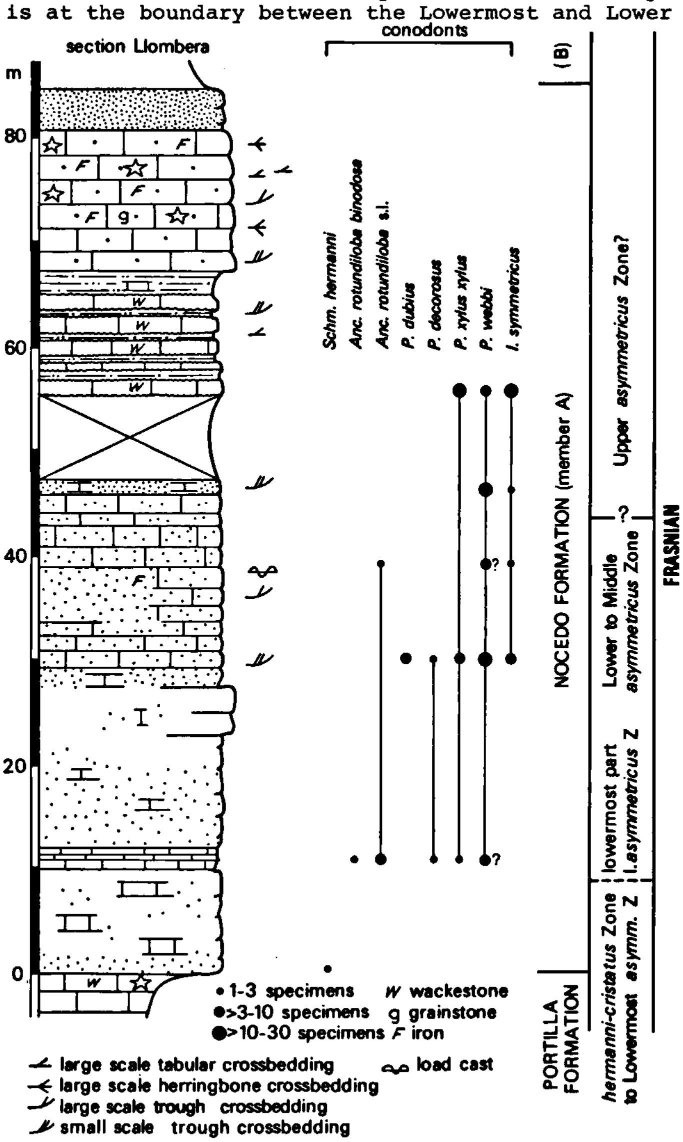

Fig. 12. Section through member A of the Nocedo Formation at Llombera (Bernesga area). Lithology after van Loevezijn (1982).

2.3.2.5. Interpretation of the sediments. - The absence of member A in the innermost areas results from non-deposition (the different lithosomes thin rapidly towards the north) and erosion (Figs. 10a and b). The high maturity of the sand is explained by the character of the sediment source and sorting by currents. The rapid decrease of the thickness towards the Asturian geanticline indicates that this area was subaerially exposed at that time. That was also the case with the Pardomino high. Therefore it is highly probable that the sands were provided by erosion of the mature sands of the Huergas, the San Pedro and the Barrios Formation. The longshore drift (see below) sorted the grains qua size and composition, building up sand bars of well-sorted quartz arenites. The poorly sorted grainstones not only contain many light carbonate grains but also many well-rounded grains of heavy minerals.

Van Loevezijn & Raven (1983) interprete the facies as follows: facies e, formed on the middle to outer shelf; facies d, deposited in a low-energy environment below wave-base; facies c, deposited in a near-coastal environment above wave-base; facies b, deposited in a shoal environment in shallow water, and facies a, deposited in a protected lagoon. These facies occur in a regressive sequence due to progradation of the coast. In the facies maps made by van Loevezijn & Raven (1983: fig. 6b) sands were formed around the Pardomino high and along the facies line with lagoonal deposits to the north and open marine deposits to the south.

Some of the sandy deposits of member A were deposited as bars and barriers, others are dominated by ebb-oriented structures with unimodal palaeocurrent patterns (Fig.

11; Pl. 2: Fig. 6) representing subtidal channels and ebb-deltas. A strong longshore drift prevented the formation of large deltas and the shifting subtidal channels eroded most of the barriers. The palaeocurrent patterns indicate the existence of dominant longshore drift from the east (Fig. 11). Only where both rate of subsidence and rate of sedimentation were high has a major part of a barrier been preserved. An example is the thick section (250 m) at Playa de Gargantera (Asturias) which was studied in detail by Sanchez de la Torre et al. (1976). Within member A they recognized several minor sedimentary cycles representing bars which developed (Pl. 2: Fig. 1), some of them emergent allowing the formation of dunes, some were eroded by migrating subtidal channels.

Subsidence was stronger in the outer area beyond the facies line but only in the near-coastal part compensated by increased sedimentation. Thus a mainly sandy wedge developed (Fig. 10) and farther offshore pelitic sediments form the major part of the sequence. The marked difference in thickness between either side of the facies line indicates that the southern area was a separate block which subsided more rapidly.

Where the supply of sand decreased, crinoidal bars or shoals formed on top of the barriers, locally forming a carbonate platform. The limestone deposition started in the south, but with decreasing supply of sand from the Pardomino high the platform gradually extended over the entire eastern part of the Bernesga area and the Esla area (Encl. 1: Fig. 8). Where stronger subsidence prevailed lagoonal deposits might form behind the barriers (e.g. Huergas de Gordón). Deposition of the sand continued in the Somiedo area and in Asturias. Eventually these sands, most probably provided by longshore currents, spread over the carbonate platform.

2.3.3. Member B

2.3.3.1. Introduction. - Member B occurs in the area south of the Intra-Asturo-Leonesian facies line in the province of León and is not known from the inner area. Although it is uncertain whether the second sequence present at the Asturian coast was deposited synchronously with member B, a description of these deposits is given in this paragraph.

2.3.3.2. Sediment petrology. - In the Esla area in member B there is a sequence of facies d to a. The member is thin and locally absent due to erosion. In the eastern part of the area shales B comprise a larger part of the section than in member A. A large part of the member is formed by facies c with the same sedimentary structures as in member A (2.3.2.2). The upper part of the sandstone may consist of dark red barrier sands with cross-bedded sets of up to a meter in thickness. In the eastern part of the area this sandstone is overlain by the Crémenes Limestone (Fig. 13). The lowermost part of this limestone consists of red crinoidal grainstones such as those present in member A of the Portilla Formation. Above these grainstones, however, with a sharp boundary there are biostromal boundstones. The lower part of the boundstones is composed of thin laminar stromatoporoids and corals in reddish marls or grainstones. At Robledo the coral fauna is varied; towards the north less corals occur, representing less genera, mainly branching rugose corals and few compound corals (*Hexagonaria*), together with crinoids (including some very large specimens), brachiopods, bryozoans, etc. Upwards in the sequence the amount of matrix and the variety in fossils decrease rapidly, leaving only stromatoporoids. Upwards this boundstone passes into a thick-bedded wackestone with brachiopods, ostracods, bryozoans, corals, etc., capped with a thin crinoidal grainstone. As in member A, on the limestone there is a thin layer of cross-bedded sandstone with much haematite (Raven, 1980a).

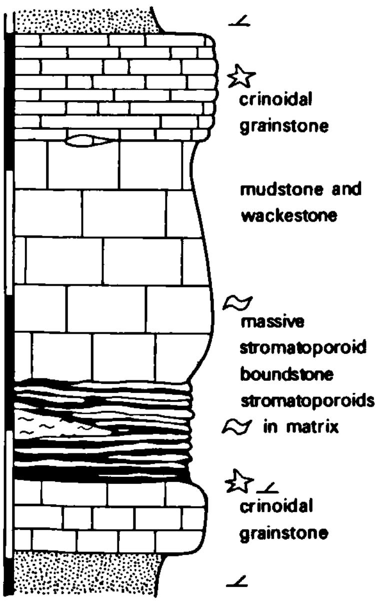

Fig. 13. Type-section of the Crémenes Limestone at Pico Aguasalio (Esla area). The scale bar indicates units of 5 m.

As in the Esla area, in the Bernesga and Somiedo areas member B contains much more fine-grained sediments than member A. Generally facies e makes up half or more of the sequence, passing via a thin layer of siltstones (facies d) upwards into sandstones and conglomerates (facies c). These coarser deposits are present in two fan-deltas, a large one near Barrios de Gordón (110 m thick deposits) and a small one near Puente de las Palomas. The fans have asymmetrical shapes. In the proximal fan lenticular orthoconglomerate units occur between sandstones. These orthoconglomerate units are graded (fining upward); cross-bedding is not present; the pebbles occur in a sand matrix. The majority of the pebbles are of vein quartz but pebbles of shale, sandstone and quartzite also occur. The largest cobble (quartzite, diameter 20 cm) was found at Barrios de Gordón; further away from the source grain size and bed thickness decrease rapidly (van Loevezijn, 1982; van Loevezijn & Raven, 1983).

At Playa de Gargantera (Asturias), with an erosive boundary above member A there is a sequence of about 67 m of grey bioturbated siltstones with some intercalated sandstone beds. These sandstone beds generally have a sharp lower boundary. Parallel lamination and few small-scale ripples were observed. The rest of the sequence is capped by Cretaceous rocks. Sanchez de la Torre et al. (1976) observed slumps and flute marks. In their opinion these deposits pass landward into sandstones and limestones.

Because the sandstones of member A pass gradually into the limestones I include the limestones in member A.

2.3.3.3. Boundaries. - Locally the lower boundary of member B is erosive (e.g. at Playa de Gargantera) but in most of the sections member B lies conformably upon member A. In the Esla area erosion occurred before deposition of the Fueyo Formation. This is evidenced by the marked variation in thickness of member B which is absent from some sections, and by the Crémenes Limestone near Robledo which appears in patches, interpreted as remnants of the eroded carbonate platform. The upper part of the section at Playa de Carranqués has suffered Mesozoic erosion.

2.3.3.4. Fossils and age of the member. - Only data on the age of the Crémenes Limestone are available. A more complete list of the fossils from that unit is given by Raven (1980a). Only the most important data will be discussed. The corals and stromatoporoids indicate that the limestone was deposited before the end of the Frasnian. The brachiopod fauna contains genera indicating a Frasnian (*Cupularostrum*, "*Camarotoechia*"), and a Famennian age (*Ptychomaletoechia*) (Sartenaer, 1968). A revision of this group is badly needed. The occurrence of specimens of the superfamily Atrypacea, however, clearly indicates the Frasnian: the Atrypacea became extinct at the end of the Frasnian (McLaren, 1970). One sample from the middle of the limestone at Robledo contained the tentaculite *Homoctenus ultimus ultimus* Zagora, 1964 (identified by J.P.S. Goeijenbier, Leiden) which is known from the Kellwasserkalk (boundary *Palmatolepis gigas* Zone and *Pa. triangularis* Zone). Further it contained the ostracods *Polyzygia neodevonica* (Matern, 1929), *Jenningsina* spec. indet. and *Microcheilinella* spec. indet. (identified by G. Becker, Frankfurt am Main). *P. neodevonica* occurs from late Givetian to the *Palmatolepis triangularis* Zone (Lethiers, 1974; Buggisch et al., 1978). The same sample contained conodonts: *Icriodus* cf. *symmetricus* and *Polygnathus xylus xylus* which are not known from deposits younger than the Frasnian. From other samples the conodont *I.* aff. *subterminis*, *Po. webbi*, *Belodella* spec. and *Coelocerodontus* spec. are known. Most of these species have a long vertical range but are not known from deposits younger than the Frasnian. The tentaculite suggests that the limestone was deposited near the *gigas* Zone-*triangularis* Zone transition.

2.3.3.5. Interpretation of the sediments. - The sediments in member B are more varied than those in member A. In the Esla area sedimentation was as in member A but now the limestone near the top of the member was formed by a biostrome. Since the Pardomino high only slightly emerged, mainly shales were deposited, only just east of the high with sandstones. Initially in the Bernesga and Somiedo areas similar deposits were formed but when the area north of the Intra-Asturo-Leonesian facies line became emerged fan-deltas were formed. The sediments of these fans were transported by mass-flows. Sediment with the same composition occurs in the Pliocene of the Pomarao region in Portugal and is interpreted as a sheetflood deposit which formed on a peneplain due to rare but heavy rains in an arid climate (van den Boogaard, 1967). The sediments may have formed north of the Intra-Asturo-Leonesian facies line when this area was emerged, from whence the sediment was transported towards the southern area where it formed the fan-deltas. The height of the fan near Barrios de Gordón indicates that both areas were separated by a fault. The longshore drift was strong enough to disturb the symmetry of the fans (current from east to west) but not to rework all the sediment for the formation of coastal barriers. In general member B has more shales and less sandstones and limestones than member A. A facies map was made by van Loevezijn & Raven (1983: fig. 6e).

2.4. FUEYO FORMATION (FAMENNIAN)

2.4.1. Introduction

For a general introduction to the lithological subdivision of the Upper Devonian the reader is referred to 2.3.1. The Fueyo Formation is present only south of the Intra-Asturo-Leonesian facies line (Fig. 14). In the Leonese part of the Asturo-Leonesian basin the formation is present everywhere but in the Asturian part of the basin the Upper Devonian is so badly exposed that there is no certainty whether the Fueyo is present or not. The thickness (800 m) of the Upper Devonian near Salas (Asturias) (Pello, 1972) indicates that there subsidence was very rapid and therefore the Fueyo may be present. The thickness varies from 16 m at Pico Aguasalio (Esla area) to 325 m at Piedrasecha (Bernesga area) (van Loevezijn, 1982). In the Valsurvio area the formation was not recognized and the decreasing thickness from the Bernesga area to the Esla area (Encl. 1: Fig. 9) indicates that it most probably pinched out here.

2.4.2. Sediment petrology

The Fueyo Formation is composed of shales, alternating with beds of well-sorted sheet siltstones and very fine to medium grained sandstones. The beds are less than one cm to several tens of cm thick. In the Esla area most of them are 5 to 30 cm thick, in the Bernesga area the thickness is more variable. The lower boundaries of the beds are sharp. In the Bernesga area the beds are graded (fining upwards) and the Bouma sequence (units A, C, D and E) has been observed. In most of the sand sheets the lower part is homogeneous and the upper part parallel-laminated. Load casts, ball-and-pillow structures, slumps and sole markings occur frequently. Some horizontal burrows of deposit-feeders are present in the upper part of the beds. At the type locality

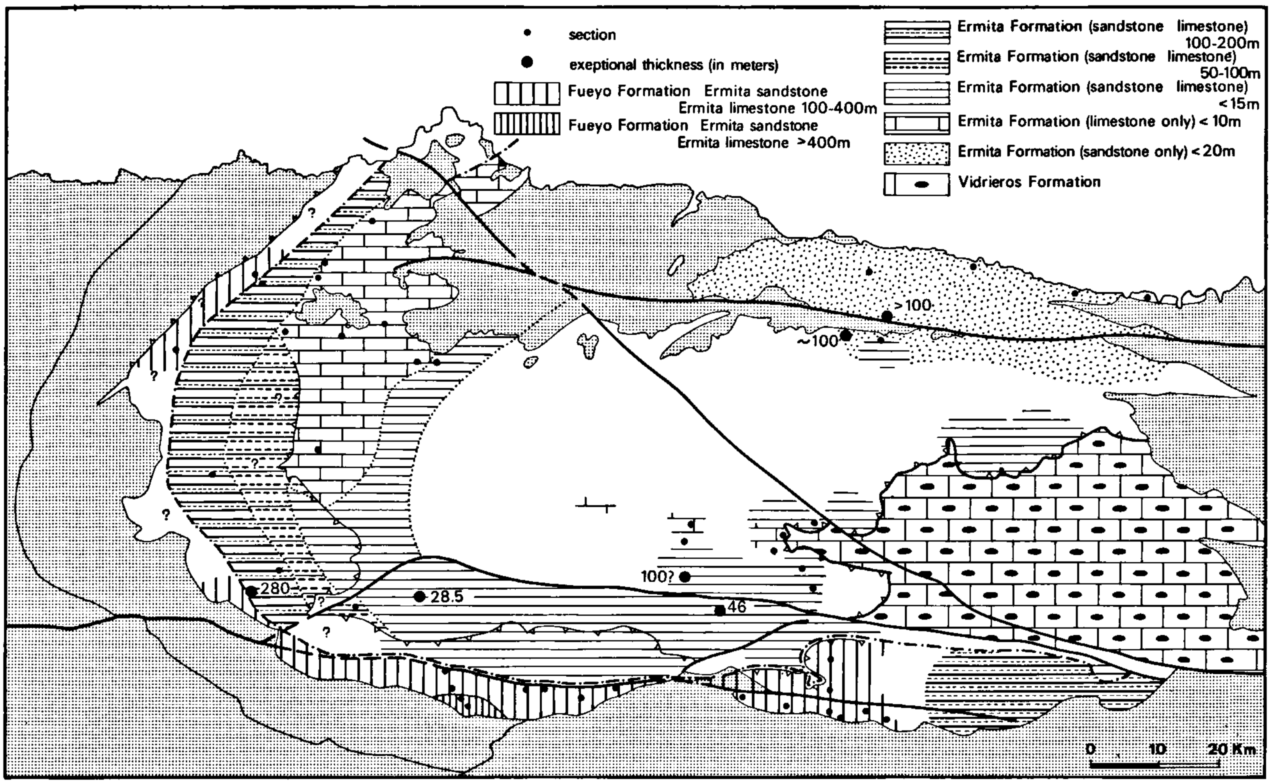

Fig. 14. Thickness and distribution of the upper Famennian deposits in the Cantabrian zone.

fossiliferous silt concretions occur within the shales. At Olleros de Alba numerous thin beds of silty to sandy limestones and calcareous siltstones and sandstones are present in the upper part of the formation, some of them deposited in channels. In the Esla area the formation is thinner and there the coarser beds fine upwards and (especially the upper part of the beds) are parallel-laminated. Usually the beds have a sharp lower surface which may be deformed by load casts, and some of the beds have a bioturbated upper surface where they pass into shale. These shales may be completely bioturbated. Upwards in the sequence the number of coarser grained beds increases, as do grain size and thickness of the individual beds. In both areas individual beds may be traced over tens of meters or more.

2.4.3. Boundaries

The lower boundary of the Fueyo Formation is sharp. It is drawn at the base of the shale. Almost everywhere it lies upon member B of the Nocedo Formation but locally in the Esla area that unit is absent and the formation lies upon member A (San Adrián and Cistierna). At Robledo de la Guzpeña the Crémenes Limestone in the top of the Nocedo is preserved in patches due to erosion before deposition of the Fueyo.

The upper boundary of the formation is chosen arbitrarily since upwards the grain size and the thickness of the coarser grained beds increase gradually. The boundary is drawn at the base of the first sandstone bed thicker than one meter, above which shales occur only sparsely. This sandstone bed may show large-scale cross-bedding, common in the Ermita Formation. At Nocedo de Bernesga the boundary between Fueyo and Ermita is a sharp surface which undulates due to load casting of the overlying sands (García Alcalde et al., 1979).

2.4.4. Fossils and age of the Formation

In the Bernesga area small bivalves and brachiopods are common. In nodules at the type locality Comte (1959) found bivalves (*Buchiola* and others), brachiopods, gastropods, goniatites and plant remains. In some of the calcareous beds at Olleros de Alba conodonts and numerous ostracods occur. From the Esla area only crinoid ossicles are known.

Since the underlying Crémenes Limestone was deposited during the latest Frasnian, sedimentation of the Fueyo may have started during the early Famennian or later. Upwards the Fueyo passes into shallow water sandstones (Ermita Formation) which were deposited during the late Famennian *costatus* Zone (Fig. 27). At Olleros de Alba the upper half of the Fueyo was deposited during the latest Famennian: sample OLL1 taken 48 m below the top of the formation contains a conodont fauna of the *costatus* Zone and most probably it belongs to the uppermost part of that zone (M. van den Boogaard, Leiden, pers. comm.; Klapper & Ziegler, 1979; the sample contains *Pandorinellina plumulus*). That sample also contains the conodont species *Palmatolepis gracilis* which lived in rather deep water and is not known from the Ermita, thus confirming the greater depth of deposition of the Fueyo in relation to the Ermita (Fig. 16). The uppermost limestone layer in the Fueyo at Sorribos de Alba contains "*Spathognathodus*" *bohlenanus* which occurs from the *marginifera* Zone to the *costatus* Zone. The only other data on the age of the Fueyo are those of brachiopods and bivalves identified by Comte (1959), but these fossils need a revision.

2.4.5. Interpretation of the sediments

The fine-grained deposits of the Fueyo were deposited on the shelf, commencing in the deeper shelf while succeeding levels were deposited in the transition zone. The coarse sediments were transported by low-density turbidity currents as is proved by the sedimentary structures. Sheet-like lithosomes with low-relief irregular undulations at their base and the dominance of parallel lamination (lower flow regime) are characteristic for the deposits of the Fueyo thus matching some of the characteristics of storm-induced turbidites which distinguish them from turbidites caused by a palaeoslope (Benton & Gray, 1981). On the other hand slumping indicates that at least some of the turbidity currents may have been caused by a palaeoslope. These deposits interfinger with and pass into the shallow-marine sandstones of the Ermita Formation due to progradation of the coast. Fig. 15 gives a correlation. The thickest deposits occur in the southeast of the Alba

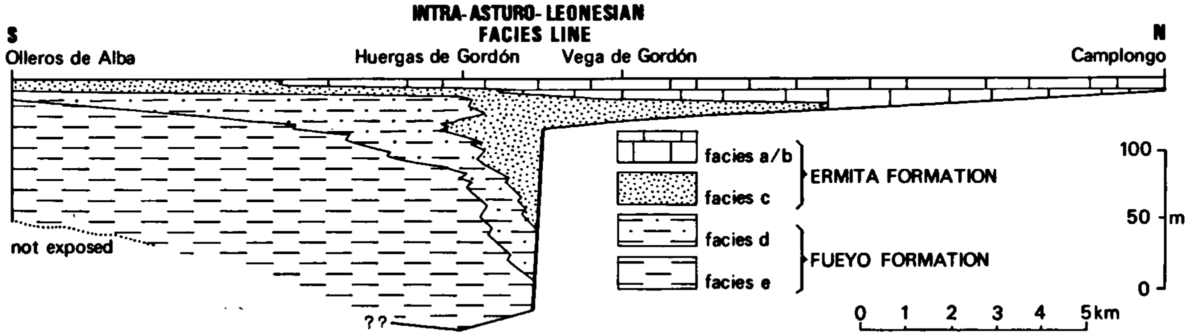

Fig. 15. Correlation of Fueyo and Ermita in the Bernesga area (not palinspastically corrected after data from van Loevezijn (1982) and myself.

syncline where mainly shales were deposited at the greatest water-depth (Encl. 1: Fig. 9). Arguments for the greater depth are the type of sediment (greyish shales), the sedimentary structures and the fossils (small bivalves as *Buchiola* and conodonts as *Palmatolepis*). Important source areas were in the northeast and northwest, near the Pardomino and Somiedo highs respectively.

In the Fueyo and Ermita a regressive sequence is observed (Fig. 15) which is closely comparable to similar sequences in the Nocedo Formation, the main difference being the larger horizontal distribution (compare Figs. 9 and 14). Therefore the Fueyo and Ermita are considered to have been deposited to a large extent contemporaneously. Where the age of the Ermita is known, no evidence of an age older than the *costatus* Zone has ever been found; therefore it is improbable that the deposits south of the Intra-Asturo-Leonesian facies line, which were formed after the same transgression, are much older. Indeed the sample from the Fueyo Formation indicates that also this formation was deposited during the *costatus* Zone (2.4.4). This implies that deposits representing almost the entire Famennian are lacking in the Asturo-Leonesian basin.

2.5. ERMITA FORMATION (FAMENNIAN AND TOURNAISIAN)

2.5.1. Introduction

For an introduction to the lithological subdivision of the Upper Devonian the reader is referred to 2.3.1. In the southern and western parts of the Asturo-Leonesian basin the thickness of the Ermita Formation may be a few hundred meters, but in the major part of the basin and on the Asturian geanticline the Ermita is absent or thin (0-50 m; Fig. 14). The Ermita interfingers with the Fueyo Formation (2.4.5). Where the Ermita is thick, the major part of the formation consists of sandstones and quartzites, often with a thin limestone in the upper part. Where the Ermita is thin, a greater part of the entire formation consists of limestone. Where the Ermita rests upon member A of the Nocedo Formation the boundary may be indeterminate, e.g. in the Valsurvio area where the name Camporredondo Formation is used (2.3.1).

2.5.2. Sediment petrology

In the outer zone and the outer half of the inner zone of the Asturo-Leonesian basin the Ermita generally is thick although it thins at the souternmost locations where the shales of the Fueyo are very thick. The composition of the sediment in the lower part of the formation is very constant except in the Bernesga area south of the Intra-Asturo-Leonesian facies line. It consists of very fine to fine white sandstone, well to very well sorted. Little or no matrix occurs and almost all the grains are quartz: heavy minerals are rare and not a single feldspar grain is present. Often these quartz arenites have been cemented during diagenesis to form orthoquartzites. In this fine sandstone large-scale tabulate cross-bedding (current ripples) is the commonest sedimentary structure, although large-scale trough-shaped cross-bedding, herringbone cross-bedding, parallel lamination, sand bars, bioturbation and wave ripples also occur. There are occasional thin horizons with coarser grains, usually as layers of moderately to badly sorted sandstone having grains up to granule size. Such layers occur in the basal part of broad and shallow channels (up to ten meters broad and up to one or two meters deep) and are particularly abundant in the upper part of the formation. Large-scale cross-lamination of megacurrent ripples and low-angle cross-bedding are preserved in the channels. Parallel laminated or bioturbated thin lenses of shale occur.

South of the Intra-Asturo-Leonesian facies line in the Bernesga area the sandstones commonly contain considerable amounts of argillaceous material and have a grey colour. In these quartz wackes sedimentary structures are less frequent than in the quartz arenites but large-scale and small-scale tabular cross-bedding is common (current ripples) as well as large-scale trough-shaped current ripples and horizontal and vertical burrows. Rare channels and slumps have been recorded. In the higher part of the quartz wackes a layer with chamosite ooids may occur

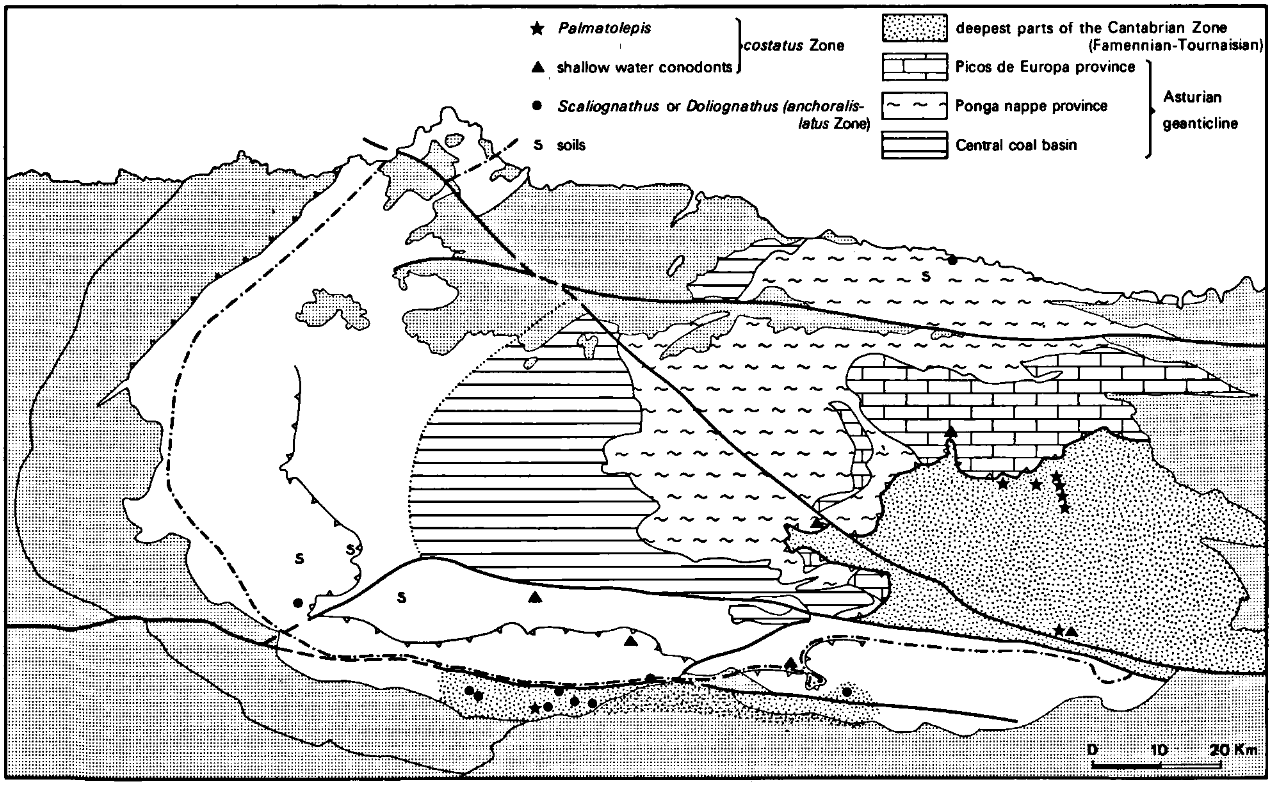

Fig. 16. Occurrence of some conodont species which are indicative for water depth (*Palmatolepis*, *Scaliognathus* and *Doliognathus* occur in deep water) and of soils in Upper Devonian deposits. Further the subdivision of the Asturian geanticline into structural provinces, as proposed by Julivert (1971) is given.

(Huergas de Gordón, Portilla de Luna, Sagüera) (van Loevezijn, 1982).

The siliciclastics may pass upward into limestones (e.g. Fig. 17). These are further described below. Where limestones are lacking and locally below the limestone the upper tens of centimeters of the sandstone are badly sorted, containing well-rounded quartz granules and pebbles, and glauconite (the occurrence of the latter mineral was confirmed by P. Seibert, Tübingen, pers. comm.). I will refer to this mineral as glaucony (sensu Odin & Matter, 1981). At Santa Olaja de la Varga (Esla area) it forms a greensand. West of the Intra-Asturo-Leonesian facies line only the sandy part of the formation is present (Pello, 1968).

Where the Ermita is thick usually a large part of the sandstones are very fine to medium, well-sorted sandstones with carbonate clasts and cement and with a dark colour caused by the presence of haematite cement. These may show a rhythmic alternation, the couplets being several cm thick (e.g. Puerto de la Cubilla, Caldás area; Páramo, Asturias). The sand may show abundant large-scale ripples. Where the formation is thin and near the erosive base poorly to very poorly sorted microconglomerates may occur, rich in carbonate clasts and with grains of fine silt to granule size, the granules consisting of quartz, quartzite or shale fragments. These granules were eroded from the formations on which the microconglomerate was deposited as is demonstrated by the bimodal distribution of grain size and the resemblance of the pebbles with the eroded underlying deposits. At Villanueva de la Tercia (Bernesga area) the limestone near the base of the Ermita contains pebbles with a maximum diameter of 8 cm (van Tongeren, 1975).

In the inner zone there are several locations with brown or red beds in the Ermita e.g. at La Cueta (Somiedo area) and San Emiliano (Caldas area) (Fig. 16) which van den Bosch (1969) and van Loevezijn (1982) interpreted as soils. According to these authors the red colour of the Ermita in the Caldas area is due to erosion and reworking of the soils. The soils consist of a leached zone of loose sand, above which there is a red bioturbated sandstone enriched in haematite and limonite, capped with a red shale layer. At San Emiliano eolian deposits also occur (van Loevezijn, 1982).

In the Coastal Ranges (Fig. 1) the Upper Devonian siliciclastic deposits lie upon sandstones and quartzites of the Ordovician Barrios Formation. The thickness is variable, ranging from a few centimeters to more than a hundred meters (Onís, Següenco) but is generally less than 20 m. The rocks are largely sandstones with microconglomerates or conglomerates developed occasionally at the base. Fragments of plutonic and volcanic rocks have been found in these deposits which have been attributed to sources known to occur in the Barrios Formation (Thibieroz, 1978; Martínez García, 1981). The exact age of the upper siliciclastics is not known with certainty but most authors agree from the sparse data available that they were deposited during the Late Devonian. In the whole Sierra Plana there is a layer of coal, several cm thick, containing plant remains with a diameter of 6 to 8 cm and more than a meter long (Hernández Sampelayo & Kindelan, 1950). Nearby, north of Onís, there is a thick quartzite below the Alba Formation and in a gallery a 100 m below the Alba Formation a late Devonian flora was found in the quartzite (Pello, 1972). According to Radig (1966) and Martínez García (1981) brachiopod faunas from Santiuste and Següenco indicate that the deposits were formed during the Frasnian but little is known about the brachiopods from the Upper Devonian sandstones in the Cantabrian zone and thus such conclusions remain uncertain. In this area generally the Ermita limestone is not present. Everywhere in the Cantabrian zone (except in the Palencian basin) there is a hiatus, comprising at least a large part of the Famennian but it may reach even into the Cambrian (Figs. 19, 27). This hiatus is largest on the Asturian geanticline. There, above the hiatus Ermita or younger deposits occur. At the southern part of the geanticline the few meters of Famennian deposits consist mainly of limestone, locally with a thin sandstone below it or intercalated in the limestone. The conodont faunas indicate deposition in a very shallow environment (4.4). Thus the few meters of sandstone which are present

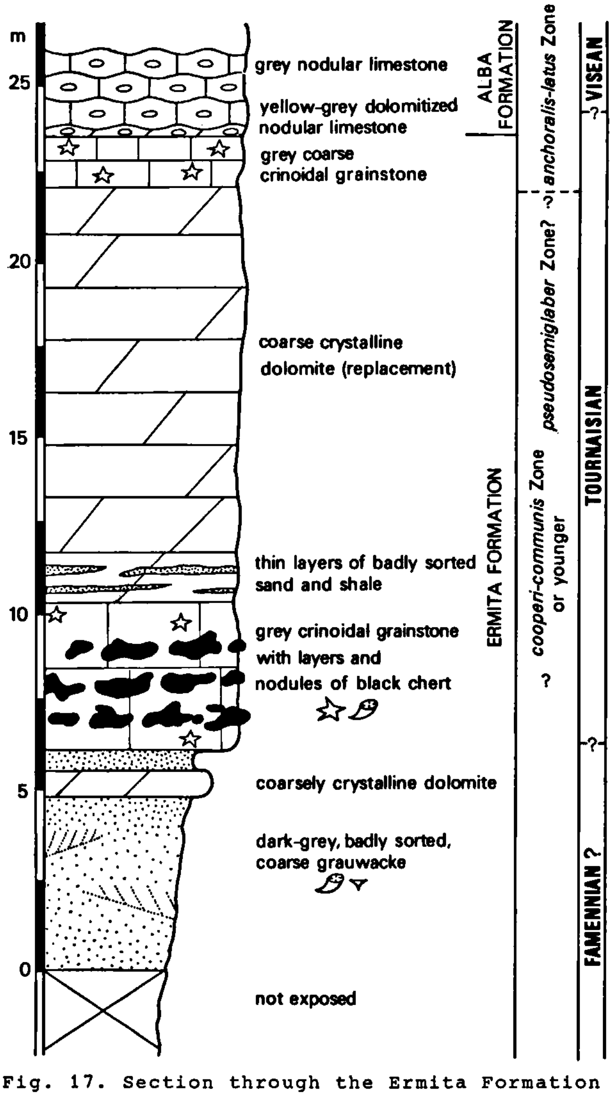

Fig. 17. Section through the Ermita Formation at Las Portillas (Picos de Europa). In reality this section is disturbed by numerous faults, this is a reconstruction.

in the Coastal Ranges (only thicker near Onís and Següenco, Fig. 14), locally containing some brachiopods or a thin coal layer, most probably were deposited in a still shallower environment. In such an environment crinoids probably did not occur in amounts large enough to form a limestone. I presume that these deposits are of late Famennian age, until good evidence shows that a part of them is older.

The limestone of the Ermita Formation usually consists of grey, sandy, coarse bioclastic grainstone with wavy bedding and locally with large-scale current ripples. Generally these grainstones consist of brecciated crinoid ossicles, brachiopods, bryozoans and some solitary corals and ostracods. Further iron-oxides (mainly haematite) occur, concentrated in thin irregular laminae which accentuate the irregularity of the surface of the beds. Some of these levels formed by diagenetic dissolution (stylolites), others by dissolution of the carbonate or non-deposition at the sea bottom (submarine hardgrounds). Very thin layers may be slightly oolitic (Sjerp, 1967: p. 72). Moreover less washed, finer grained packstones, wackestones and mudstones occur. These may be rich in fossil remains: bryozoans, brachiopods, ostracods, crinoids, solitary corals, goniatites, calcisphera, conodonts and fish remains. In the limestone some glaucony and anorganic phosphate may occur. At two localities irregular chert nodules occur: at Las Portillas (Picos de Europa area, Fig. 17; Pl. 2: Fig. 4) and northeast of Mirantes de Luna (Bernesga area, Fig. 18). Usually the grainstones are present in the alternations of sandstone and limestone and

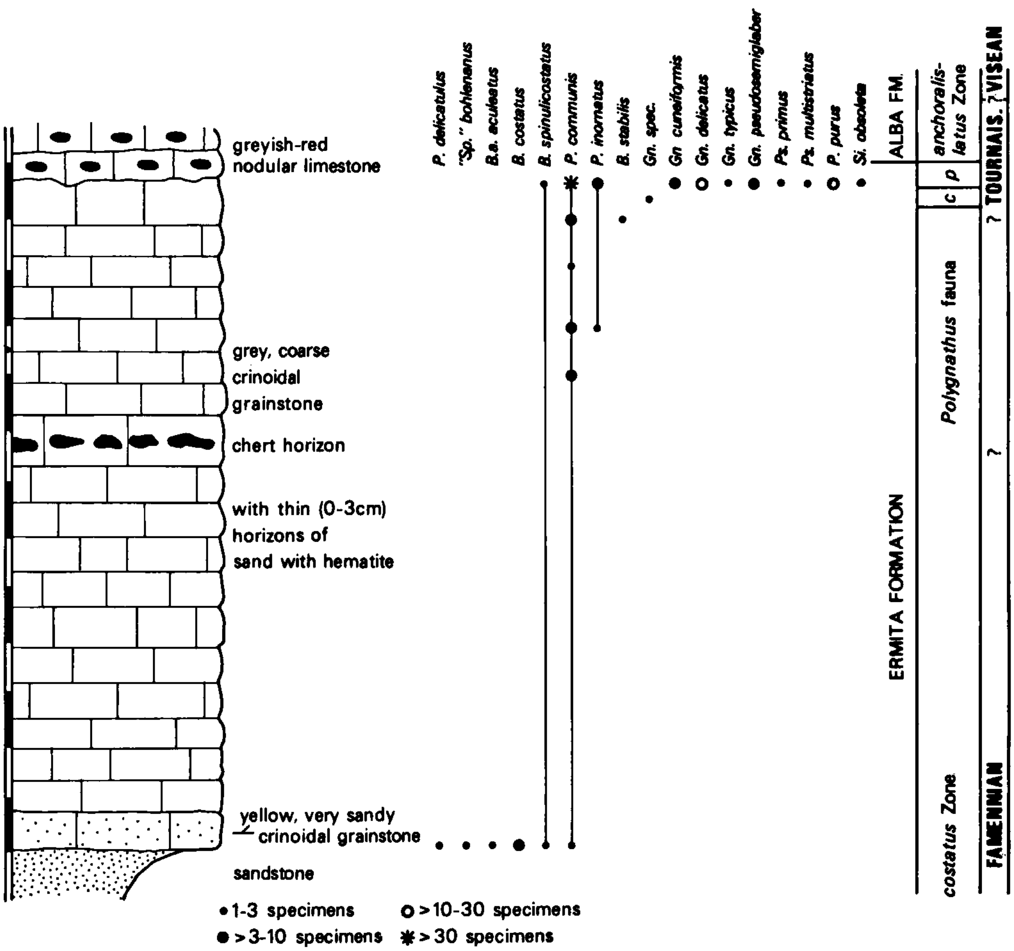

Fig. 18. Section through the upper part of the Ermita Formation northeast of Mirantes de Luna (Bernesga area); c = *cooperi-communis* Zone, p = *pseudosemiglaber* Zone.

in the basal part of the limestone. The higher part may consist of each of the limestone types, sometimes alternating. In a large part of Asturias only limestones occur (Fig. 14). There the Ermita limestone lies on the Candás Formation. In the innermost zone of the Asturo-Leonesian basin, where the Ermita lies on the siliciclastic Naranco Formation (= Huergas Formation) a thin sandstone may be present near the base of the limestone. On the Asturian geanticline these limestones often are dolomitized. This was observed in the Ponga nappe province (Fig. 16, e.g. La Uña, Oseja de Sajambre, Polvoredo) and in the Picos de Europa (Rosa de Lon, Las Portillas, Fig. 17).

2.5.3. Erosion at the base of the Ermita Formation

In the Cantabrian zone there is a hiatus below the Ermita, ranging from about 5 to 150 Ma (Fig. 19). In some sections the erosion is apparent from large-scale karst weathering (Caldas area) but at most of the localities only minor irregularities mark the disconformity. South of the

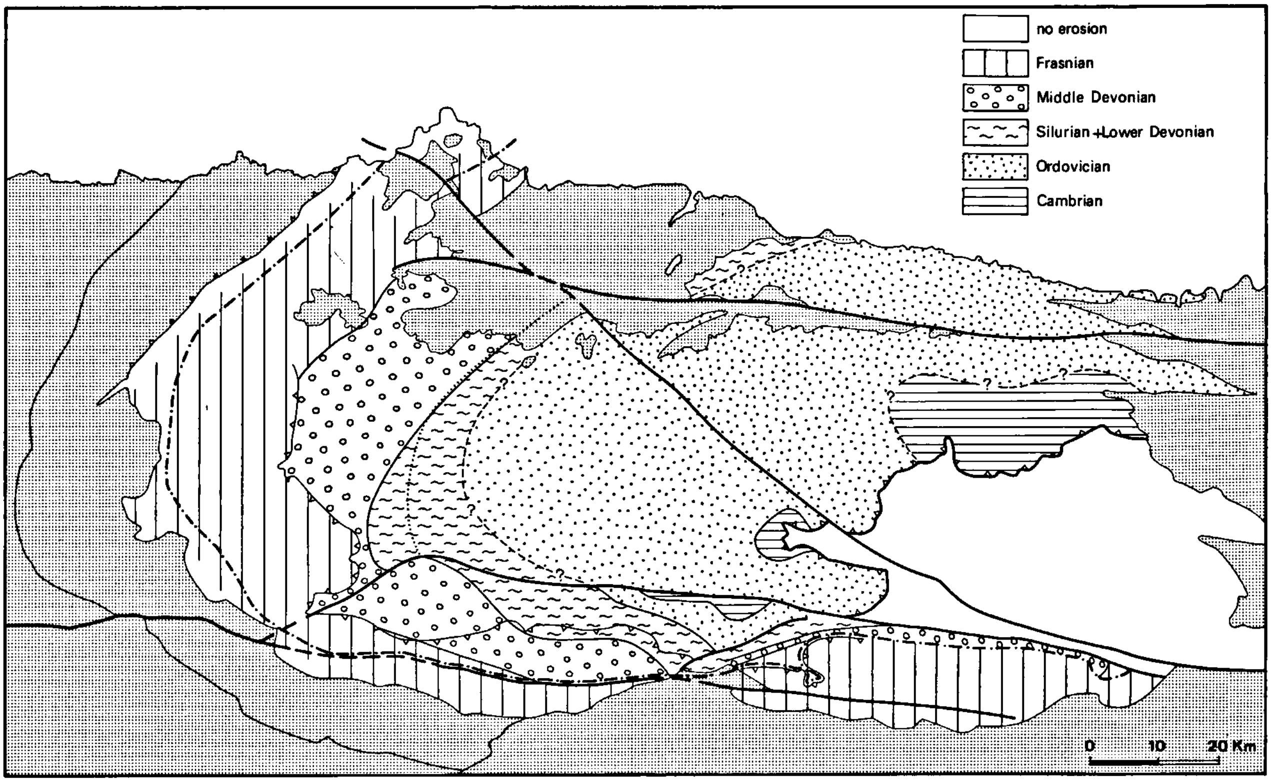

Fig. 19. Level to where the erosion during the Famennian reached.

Intra-Asturo-Leonesian facies line the underlying Fueyo passes gradually into the Ermita and the boundary is arbitrary.

The largest hiatus occurs on the Asturian geanticline where the formation rests upon Cambrian or Ordovician rocks (Fig. 19). In the Isidro and Picos de Europa areas the formation lies upon the Cambrian Láncara Formation (Sjerp, 1967; Martínez García, 1981). Elsewhere on the geanticline the Ermita rests upon the Barrios Formation; there the boundary is evident as a sharp undulating erosive surface with depressions of several to tens of centimeters, filled with cross-bedded sands (Pl. 2: Fig. 2).

In the Asturo-Leonesian basin the hiatus is usually much smaller. In many sections only a slightly undulating erosive surface marks the boundary. At San Martín de los Herreros (Ventanilla area, Fig. 20) carbonate pebbles from the underlying Portilla Formation occur in the lower part of the Ermita, the erosive surface cutting through coral colonies. Locally in the Ventanilla area the Ermita penetrates into the Portilla limestone via karst fissures (N. Schelling, Leiden, pers. comm.). At Fuentes de Sancena (Bernesga area) (Frankenfeld, 1981: figs. 42, 43) the Santa Lucía Formation is eroded: the karst surface undulates over several tens of centimeters forming broad depressions filled with quartzite of the Ermita. It is not clear where in the sections at Vega de Gordón and Beberino (Bernesga area) one has to draw the boundary between Nocedo and Ermita but some data indicate that only the uppermost few meters belong to the Ermita Formation. Near the base of this unit pebbles and cobbles of compound corals (*Hexagonaria*) occur. These almost certainly come from the Nocedo or older Devonian deposits so that they were probably

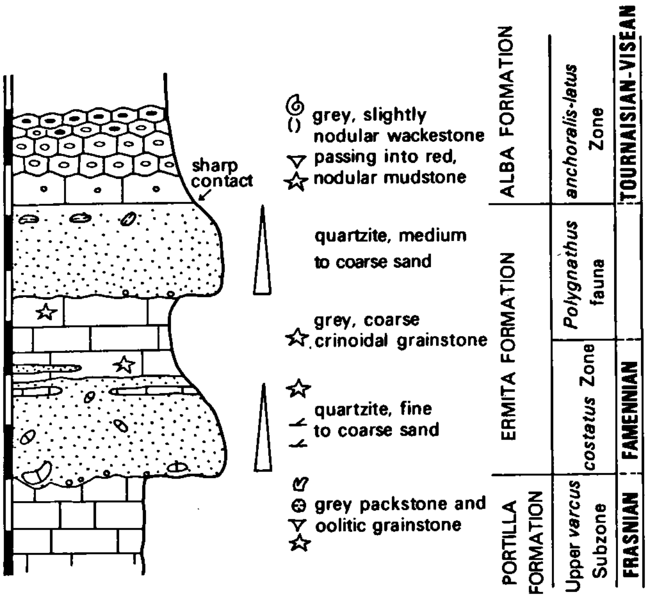

Fig. 20. Section through the Ermita Formation at San Martín de los Herreros (Ventanilla area).

being eroded during the deposition of the Ermita.

In the Caldas area (Bodón unit) the Ermita Formation lies upon the Santa Lucía Formation (the Caldas Formation of Smits, 1965) or on the La Vid Formation. The thickness of the underlying deposits diminishes from west to east: in the Sobía unit the Santa Lucía Formation has a thickness of 600 m below the Huergas Formation, at Puerto de la Cubilla the Santa Lucía measures 382 m, at Caldas de Luna 224 m (van den Bosch, 1969) and at Busdongo and Villanueva de la Tercia there is no Santa Lucía, the Ermita directly lying on the La Vid Formation. Smits (1965) calculated an angular unconformity of 0.5°, I estimate it at about 1.5°. This unconformity was caused by an increased uplift in the east (Pardomino high) and north (Asturian geanticline). The uplifted area was subaerially exposed, and karst weathering originated deep holes and fissures which later were filled with cross-bedded sands of the Ermita, which levelled off the area. Hence the thickness of the Ermita Formation is very variable and it has an irregular, patchy distribution. From southwest to northeast the thickness decreases from 85 m near Villavecino (Frankenfeld, 1981) to few meters near Caldas de Luna. The basal sandstone is succeeded by sandy crinoidal grainstones of the Ermita Formation and black shales of the Vegamián Formation, further levelling off the karstified area.

The irregular upper surface of the Santa Lucía Formation in the Caldas area is considered to be caused mainly by karst weathering, in agreement with Smits (1965), van den Bosch (1969) and van Staalduinen (1973). Arguments are: (1) the rapid decrease in thickness of the formation towards the east showing erosion of the formation, (2) the irregularity of the holes and fissures: broad and shallow depressions as well as deep irregular holes occurs next to each other. These structures may be several meters deep but some fissures reach to the base of the limestone, 58 m below the top (at Caldas de Luna). Some of the karst cavities collapsed during deposition of the Ermita Formation. Lithified blocks of sandstone and limestone slid into the depressions forming breccias (van Loevezijn, 1982). Locally (Caldas de Luna) sand was deposited so rapidly that cross-bedded sets were overfolded during sedimentation (van Loevezijn, 1982). Not all the irregularities in the upper part of the Santa Lucía Formation can be explained by karst weathering. Engeser et al. (1981) describe faults in the limestone with siliciclastic sediments penetrating between the separate blocks from below and above (P. 2: Fig. 3). Gravitation caused downsliding of the blocks and folding of the thinner strata. The fold axes are vergent towards the southwest.

It is important to know when the karst weathering and the faulting took place. Frankenfeld (1981: p. 33) argues that the upper part of the Caldas Formation was deposited during the Givetian and Frasnian because of the composition of the coral fauna. The thamnopores, which he mentions as typical for the Portilla Formation, occur also in the Santa Lucía Formation. De Coo (1974) considers the Caldas and Santa Lucía as the same formation. Data on the age of the upper part of the Santa Lucía from brachiopods, ostracods, conodonts and other fossil groups all indicate that in the inner zone deposition of the Santa Lucía terminated during the *Polygnathus costatus patulus* Zone (Couvenian 1 = late Emsian and early Eifelian) (García Alcalde et al., 1979; Michel, 1972; Huisman, 1981). Locally in the outer zone deposition may have terminated slightly later: in the southern limb of the Alba syncline deposition continued during the *Polygnathus costatus costatus* Zone (early Eifelian; Buggisch et al., 1982 and my own observations) but younger deposits are not known. The few condodont samples from the top of the formation in the Caldas area confirm this: they contain, amongst others *Icriodus corniger leptus* and *I. corniger retrodepressus* which are indicative for the *patulus* Zone (sections Puerto de la Cubilla, Huisman, 1981; 1 km east of Puerto de la Cubilla, Buggisch et al., 1982). Frankenfeld (1981) further supposes that the red ferruginous sandstones in the depressions in the Santa Lucía Formation in the Fuentes de Sancena area were deposited during the Frasnian. The brachiopod faunas on which he based his conclusion, however, have not been exhaustively studied. At Villanueva de la Tercia the crinoidal limestone below these ferruginous sandstones contains conodonts from the late Famennian *Bispathodus costatus* Zone. Therefore I suppose that the sandstones were deposited during the late Famennian, but as in the Coastal Ranges (2.5.2) further study of the macrofossils may modify this conclusion. Thus at some time during the interval between early Eifelian and late Famennian karst weathering must have occurred. The tectonic fissures were probably formed during the same interval but some data (Frankenfeld, 1981: p. 39) indicate that the movements continued during deposition of the ferruginous sandstones, thus during the late Famennian.

In Asturias the Ermita rests upon the Huergas, the Candás or the Nocedo Formation. Palinspastically reconstructed cross-sections (Evers, 1967: fig. 14) and maps of the level to which the erosion reached (Fig. 19) (Parga, 1969) demonstrate the angular unconformity below the Ermita. The age of the underlying sediments increases from the Asturo-Leonesian basin towards the Asturian geanticline. The absence of a basal conglomerate (locally a microconglomerate or conglomerate lenses may be present) is generally used as an argument for peneplainisation of the Asturian geanticline during the late Devonian (e.g. van Adrichem Boogaert, 1967).

2.5.4. *Erosion within the Ermita Formation*

In the sandstones as well as in the limestones of the Ermita Formation intraformational erosion is a common phenomenon, generally due to small-scale, local events. These may have been shifting channels and subaerial erosion. In the field it can be difficult to distinguish horizons of non-deposition (hardgrounds) and erosion surfaces; therefore only some better studied erosion surfaces are mentioned as examples.

At Nocedo de Bernesga (Bernesga area) the upper part of the formation consists of cross-bedded calcareous sandstone deposited during late Famennian and early Tournaisian. The upper part of the sandstone contains fragments of black shale from the Vegamián Formation (compare 2.6.4 and Encl. 1: Fig. 11). This layer and the 50 cm of crinoidal grainstone above it contain a varied conodont fauna of the *pseudosemiglaber* Zone.

There are also many erosive contacts between limestone and sandstone layers of the Ermita. An example is the section at San Martín de los Herreros (Ventanilla area, Fig. 20) where the uppermost erosive quartzite contains some pebbles of the underlying Ermita limestone.

2.5.5. *Erosion of the top of the Ermita Formation*

Usually the top of the formation is a sharp erosive surface upon which the Vegamián or the Alba Formation were deposited. In the southeast of the Alba syncline and in the northeast of the Leonides the Vegamián may show a basal glauconitic sandstone on top of an Ermita limestone, the boundary surface with dissolution features (karst) penetrating several cm into the limestone (at Vegamián, Evers, 1967: fig. 18; at Caldas de Luna, Peña Ubiña, etc.). The Alba Formation often starts with a grey or yellowish (locally nodular) fine-grained limestone or a red nodular limestone. Although these limestones have sharp boundaries with the Ermita, in the field it may be difficult to distinguish the formations: both may consist of yellowish grey limestone, and sharp boundaries may occur within the Ermita. The nodular mudstones and wackestones immediately below the red nodular limestone, containing a pelagic fauna, are considered as part of the Alba Formation.

Moreover it is difficult to establish the boundary in the neighbourhood of Beberino (Bernesga area). At Beberino itself the Ermita passes into the Alba via a number of hardgrounds which are partly enriched in haematite. Between the grey crinoidal grainstones of the Ermita and the red nodular limestones of the Alba reddish and greyish crinoidal wackestones occur which I provisionally assign to the Ermita Formation. Water movements above the sediment surface lead to breaking of the hardground and these fissures were filled by younger deposits (Fig. 21b). At Pico Aguasalio (Esla area; Fig. 21a) there is a surface between Ermita and Alba which was interpreted as a karst surface (van Adrichem Boogaert, 1967), a slump of unlithified limestones (Frankenfeld, 1981: p. 58), and a hardground with fissures which originated by water movements above the hardground (P. Seibert, Tübingen, pers. comm.). The shape of the fissures (Fig. 21a) favours the karst theory.

At most localities it is easy to distinguish the formations in the field, particularly where the Vegamián separates both formations or where fragments of the Ermita are present near the base of the Alba as at San Román de Candamo (Asturias) (Pello, 1972).

2.5.6. *Fossils and age of the formation*

Fossils are scarce in the sandstones: scattered crinoid ossicles are common but only locally a thin layer occurs with many fossils (brachiopods, crinoids, bivalves, gastropods and bryozoans), usually only as casts. The limestones contain a more varied fauna (2.5.2).

The conodonts from limestone and sandstone beds prove that the upper part of the formation was deposited from the late Famennian *costatus* Zone onwards. Sediments of this zone are present all over the Asturo-Leonesian basin and on part of the Asturian geanticline. The conodont faunas contain many species but are rather constant in composition in the Asturo-Leonesian basin and indicate deposition on the inner shelf, far from a sediment-supplying hinterland. If the supposition that the underlying Fueyo was deposited during the *costatus* Zone (2.4.4) is correct, the Ermita can not be older than that.

Within the basin and on the geanticline the *Polygnathus* fauna is also widespread. It is supposed to have been deposited in very shallow water: usually it is present above sediments deposited during the *costatus* Zone suggesting that part of these deposits may be younger. In a few samples from the Ermita on the geanticline shallow water faunas occur which may be attributed to the *costatus* Zone (4.3). Particularly a sample from Caín (Picos de Europa, collected by M. Hernán Goméz, Delft, sample HG42) contains a typical shallow-water fauna. This may indicate that during the late Famennian the Picos de Europa area was very shallow.

Only at locations where the Vegamián is lacking faunas of *cooperi-communis* Zone and *pseudosemiglaber* Zone may occur, mainly in the area between the Intra-Asturo-Leonesian facies line and the León line. A fauna from the *cooperi-communis* Zone is present in the greensand near the top of the Ermita at Santa Olaja de la Varga (Esla area). In the two sections where the *anchoralis-latus* Zone is mentioned from the Ermita Formation (Higgins & Wagner-Gentis,

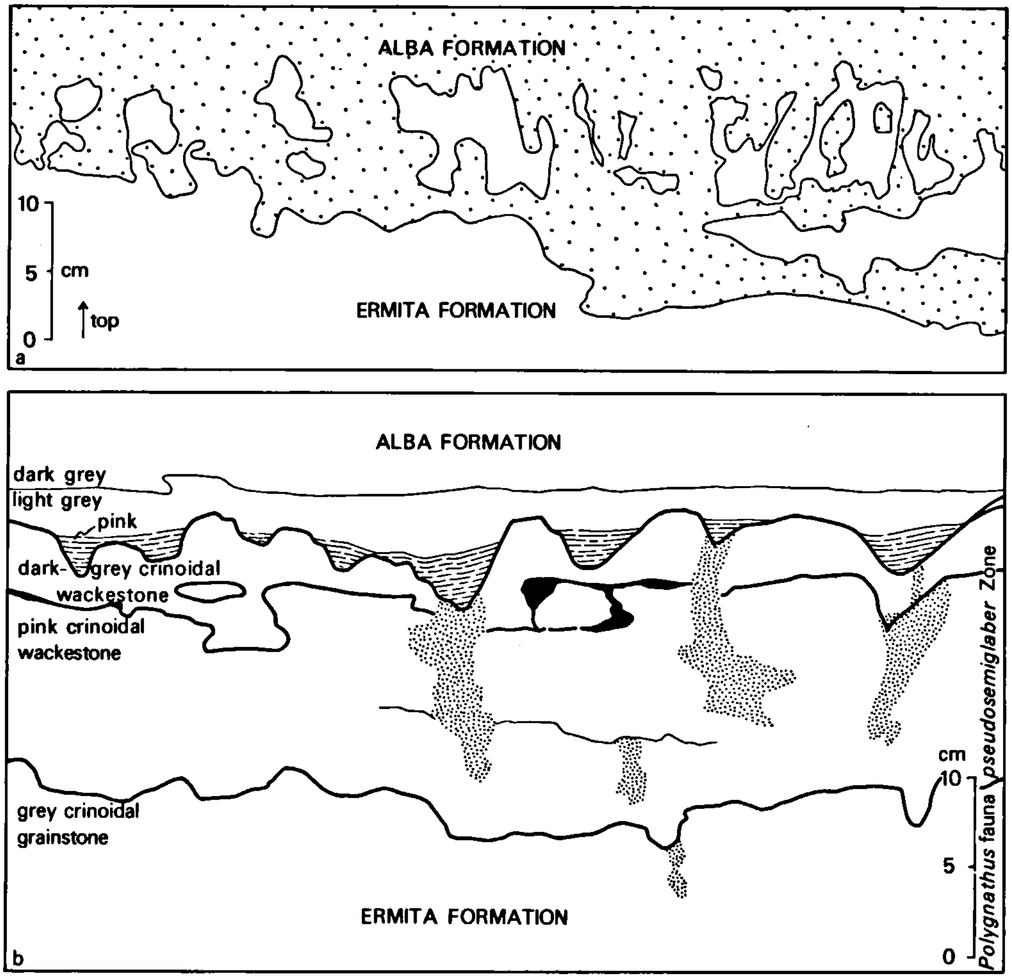

Fig. 21. Contacts between the Ermita Formation and the Alba Formation. (a) At Pico Aguasalio (Esla area) the Alba Formation appears to have been deposited over a karst surface in the top of the Ermita Formation. (b) At Beberino (Bernesga area) there is an interval of pink and grey crinoidal wackestones provisionally attributed to the Ermita Formation. Several hardgrounds are present, the upper one being the best developed (thick black lines indicate an iron (haematite) crust, the stippled parts contain dispersed haematite). All the deposits contain conodont faunas attributed to the *pseudosemiglaber* Zone, except the grey crinoidal grainstone of the Ermita Formation at Beberino which contains conodonts of the *Polygnathus* fauna.

1982) it occurs in a red crinoidal wackestone intermediate between Ermita and Alba (2.5.5). Only at Las Portillas and south of Caín (both Picos de Europa) a grey crinoidal grainstone of the Ermita Formation is known to contain a fauna from the *anchoralis-latus* Zone (Figs. 17 and 20). This grainstone occurs everywhere at the base of the lowermost nappe of the Picos de Europa. It indicates that the Picos de Europa were the shallowest part of the Cantabrian zone, although far from a sediment-supplying hinterland.

2.5.7. Interpretation of the sediments

The isopach maps (Fig. 14 and Encl. 1: Fig. 9) show that the Ermita Formation is the thickest south and west of the Intra-Asturo-Leonesian facies line. Away from this line the thickness decreases rapidly and is very small on the Asturian geanticline, on a large part of which (the Central Coal basin, Fig. 16) the formation is not known with certainty.

I postulate that the quartz arenites south of the facies line were deposited in a shallow, stable, high-energy environment. Channels shifted and constantly reworked the sediment resulting in an excellent sorting of grain size and a homogeneous composition of a mature sand. This was also favoured by the polycyclic nature of the sand originating from mature sandstones of older formations (Huergas, San Pedro and Barrios) exposed at topographical higher areas such as the Asturian geanticline and the Pardomino high. In between the channels, shoals were present where parallel-laminated sands and shales were deposited which were often bioturbated. The quartz wackes have been deposited in a slightly deeper environment than the quartz arenites and they pass laterally into the shales of the Fueyo Formation (2.4.5). The ratio and isopach map of Fueyo and Ermita (Encl. 1: Fig. 9) indicates the presence of a depocenter in the southwestern part of the Bernesga area and more stable areas to the east and west.

There was a remarkably large area in Asturias where only Ermita limestone occurs (Fig. 14). The prevailing conditions may be compared with those at the narrow ridge in the Bernesga area and with those in the Esla area north of the facies line, which during the late Tournaisian were stable, rather shallow areas (2.6.5). The limestones (crinoidal grainstones) were deposited above wave-base. At several places in the other part of the inner zone soils were present in the upper part of the Ermita sandstones suggesting intermittent conditions of emergence.

The Ermita was deposited after a transgression (*costatus* Zone) over the largely peneplained inner zone of the Asturo-Leonesian basin and the Asturian geanticline. Erosion most probably started on the geanticline during the early Devonian when the differentiation between the Asturo-Leonesian and Palencian basins began, and reached its maximum during the Famennian. Before the transgression began over limestones in the inner zone of the basin karst weathering occured. It may be that after the transgression the Asturian geanticline was only partly covered by a shallow sea. The depth of deposition in the outer zone of the Asturo-Leonesian basin was slightly greater, up to about 20 to 50 m (see also 2.6.5).

After the late Famennian transgression the coast gradually prograded so that Ermita sandstone extended over the Fueyo shales in the Esla and Bernesga areas. The presence of shallow-water faunas of the *Polygnathus* fauna above those of the *costatus* Zone indicate a shallowing in the entire area (4.4). The ooids and soils in the upper part of the Ermita also indicate a shallowing. Nowhere on the Asturian geanticline, nor in the basin early Tournaisian deposits are known with certainty. During the Tournaisian karst weathering and erosion reduced the thickness of the deposits. The karst penetrates only a few cm into the Ermita limestone, indicating that the area never was much above sea level. Sedimentation of the Ermita was resumed, together with that of the Vegamián Formation, during the late Tournaisian *cooperi-communis* Zone and locally continued into the *anchoralis-latus* Zone.

2.6. VEGAMIÁN FORMATION (TOURNAISIAN)

2.6.1. Introduction

Comte (1959) introduced the name "Couches de Vegamián" which he included in his "Griotte à *Goniatites crenistriata*. Brouwer & van Ginkel (1964) introduced the Sella Formation including both the "Couches de Vegamián" and the "Griotte de Puente de Alba". Rácz (1964) introduced the Getino Formation for Vegamián and part of the Alba. Van Ginkel (1965) considered the differences in lithology and fossil content important enough to distinguish the Vegamián and Alba as two formations. The stratotype for the Vegamián was chosen 1 km southsouthwest of the village Vegamián (Bernesga area) where 15 m of black and green shales with phosphatic nodules and chert are present (Comte, 1959). Evers (1967) foresaw the disappearance of the stratotype because of the construction of the Embalse del Porma (reservoir) and designated a new stratotype along the new main road near the present Mirador de Vegamián.

The Vegamián Formation is thin and has a rather constant lithology in the whole area where it occurs: the eastern half of the Cantabrian zone. Contrasting with the uniformity of the facies in such a large area is the patchy distribution of the formation. It occurs in the southeastern part of the Alba syncline, locally in the northeastern part of the Bernesga area, locally on the eastern part of the Asturian geanticline and in the Palencian basin. It is also present in the northern and southern parts of the Esla area (Fig. 22). At the other localities contemporaneously with the Vegamián limestones of the Ermita were deposited (2.5.6).

2.6.2. Sediment petrology

Characteristic for the Vegamián are black, laminated, cherty shales with frequent phosphate and markasite (pyrite) nodules. The Vegamián often has sharp lower and upper boundaries. Where the formation lies upon karst weathered limestones the lowermost part of it consists of a thin layer of dark grey or green sandstone with phosphate nodules and glaucony (sensu Odin & Matter, 1981), e.g. at Caldas de Luna. Where the formation lies upon siliciclastic sediments the lowermost part of it consists of siltstones and shales, often with yellow or grey colours, e.g. at Sorribos de Alba (Encl. 1: Fig. 10). The transitional layer is thicker near the Pardomino high (Frankenfeld, 1981: fig. 77). After each erosional phase such a transitional layer was deposited. The transition into the overlying Alba Formation may be gradual (e.g. at Vegamián, Encl. 1: Fig. 12) or slightly erosive (e.g. in the southeastern part of the Alba syncline, Encl. 1: Fig. 10). The black shales pass into red nodular limestones of the Alba Formation via olive green and red shales (Vegamián, Tolibia de Abajo, Bernesga area; Lois, Isidro area) or via a grey or yellow (nodular) limestone (Caldas de Luna, Caldas area). The phosphate nodules vary from less than 5 mm up to nodules with a maximum diameter of about 5 cm. The larger the nodules, the flatter they are. They are black but in thin section the colour is dark brown, being lighter away from the nucleus. The nodules have almost always a concentric layering but a few nodules show horizontal layering. They are composed of small dark brown pellets (faecal pellets?) of very fine to fine sand size. Between the pellets there is white (phosphatic?) cement, and radiolarians and small crystals of pyrite are numerous. Radiolarians also occur in the (phosphatic?) nucleus. In the Palencian basin the black shales are laminated, due to a difference in silt content. These shales cleave into slices. In the other areas the black shales generally are not laminated and have a blocky fracture.

In the southeast of the Alba syncline the formation occurs only south of a line from Rabanal de Fenar to Sagüera. The thickness varies from 0 m (Sagüera) to 5.2 m (Piedrasecha) (Encl. 1: Fig. 10). Most remarkable is the occurrence of a glauconitic sandstone layer with pebbles and sand grains of black shale, chert and phosphate (van Tongeren, 1975; Frankenfeld, 1981; van Loevezijn, 1982). The fragments of phosphate and some of the chert are rich in radiolarians. This layer is present in all the sections except at Sorribos de Alba. Below this layer the Vegamián consists of bright coloured (yellow and green) shales and sandstones which pass upward into black cherty shales. The sand layer with pebbles represents an erosional phase which removed the Vegamián from the area north of a line from Sagüera to Rabanal. At Nocedo de Bernesga the uppermost part of the sandstones of the Ermita Formation contains fragments of black shale from the Vegamián (2.5.4; Encl. 1: Fig. 11). This layer has exactly the same composition as the pebble layer within the Vegamián. Above it there is half a meter of crinoidal grainstones of the Ermita Formation. In the Pedroso syncline deposition of the limestone of the Ermita was interrupted, leading to the formation of hardgrounds (Fig. 21b and Encl. 1: Fig. 11). In the southeastern part of the Alba syncline after the erosional phase bright-coloured shales, black cherty shales and locally some limestone lenticles were deposited. Slight erosion occurred before deposition of the Alba Formation.

In the northeast of the Bernesga area the Vegamián is present in an area bordered by carbonate platforms and by the Asturian geanticline in the north (it is not known whether the western part of the geanticline was emerged or not). Geological maps give a wrong impression of the distribution of the formation in this area: van Staalduinen did not recognize the formation in the northern area, and Evers (1967) indicated the formation almost everywhere between Ermita and Alba, whereas in fact the distribution of the formation is very patchy (Fig. 22). In this area the sediments consist of cross-bedded, bright-coloured sands and shales, and (laminated) black cherty shales. The thickness of the deposits increases from a few tens of centimeters in the west to 12 m in the stratotype. In the eastern sections some erosional surfaces are present within the formation (Encl. 1: Fig. 12; Evers, 1967: fig. 19). In the Valdecastillo section between the black shales there is a layer of white shale containing macrofossils. This layer represents an interval of better circulation. Also in other sections such fossiliferous layers occur (Tolibia de Abajo, Genicera).

The formation is further present on a part of the Asturian geanticline (the Ponga nappe province, Fig. 18) and in the Palencian basin (Fig. 22). In these areas it consists of black shales with chert, phosphate nodules and finely dispersed pyrite. On the Asturian geanticline the thickness varies from 0 to a few meters, in the Palencian basin the formation is up to 50 m thick (south of Pico Gildar, van Adrichem Boogaert, 1967: fig. 29). In the area of the geanticline some erosional phases were recognized by Sjerp (1967) (see also Encl. 1: Fig. 12). In the Palencian basin, however, deposition was largely uninterrupted. Only van Veen (1965: p. 63) mentioned two lenses of quartzitic pebbles near Cardaño de Arriba. Van Adrichem Boogaert (1967: fig. 31) found slumped beds of quartzitic sandstone in the Gildar-Montó area.

In the souteastern part of the Bernesga area and in the southwest of the Esla area south of the Intra-Asturo-Leonesian facies line reddish brown and yellowish brown radiolarian cherts occur. In most of the sections (Llombera, Peredilla, San Adrián, La Ercina, Cistierna) red cherts are present above these brightly coloured cherts. No indications for the age are available but in the section at Matallana estación, on the reddish brown cherts there are red nodular limestones of the Alba Formation, deposited during the *anchoralis-latus* Zone and early Viséan thus proving that the cherts were deposited during the Tournaisian (van Tongeren, 1975). Indeed, near Cistierna (Esla area) these cherts and black shales with phosphate nodules occur next to each other (J.F. Savage, Utrecht, pers. comm.). Because the cherts pass vertically into comparable Viséan cherts of the Alba Formation they are considered as part of that formation.

2.6.3. Fossils and age of the formation

Macrofossils are rare in this formation, only a few sections containing such fossils are known: inarticulate brachiopods, bivalves, trilobites, goniatites and fish remains (Gandl, 1973). Further ostracods, conodonts and radiolarians are known from this formation. In a section near Valdecastillo I found a layer of white shale with ostracods, articulate brachiopods and bivalves. The brightly coloured siltstones and sandstones are often intensely bioturbated. Pelagic radiolarians with long spines are very common in the chert layers and the phosphate nodules, as well as locally in the shale.

The conodonts from the basal part of the formation in the Asturo-Leonesian basin and on the Asturian geanticline are all from the *cooperi-communis* Zone (4.3). In the Palencian basin deposition of the formation presumably started at the same time. Also in that area there is a hiatus below the Vegamián, and the youngest sediments of the Vidrieros were deposited during the late Famennian-early Tournaisian (*Protognathodus* fauna and *Polygnathus* fauna) (4.3). At Pico Mampodre (Isidro area) a fauna of the *pseudosemiglaber* Zone (4.3) is present within the Vegamián (sample SJ H6 of Sjerp, 1967). At Genicera (Bernesga area) *Doliognathus latus* and *Pseudopolygnathus pinnatus* occur in the black shales (collected by C.F. Winkler Prins, Leiden) proving the deposition of these shales during the *anchoralis-latus* Zone (4.3). In some areas where the Vegamián is not present deposition of the Alba Formation started during the *pseudosemiglaber*

Zone, elsewhere in the Cantabrian zone during the *anchoralis-latus* Zone. An exception is the Palencian basin where the base of the Alba Formation was deposited during the late Viséan (van Adrichem Boogaert, 1967). It is possible that in this basin the deposition of the Vegamián continued during a part of the Viséan. At the base of the formation there is an unconformity in the Barruelo area (Wagner & Wagner-Gentis, 1963) and at least locally in the Liébana area where the Vidrieros and Vegamián may be absent (Maas, 1974).

2.6.4. Erosional phases

Everywhere in the Cantabrian zone at the base of the Vegamián there is an unconformity. Van Adrichem Boogaert (1967: fig. 46) described a (disturbed) section near Cardaño de Arriba in the Palencian basin where he presumed that the Vidrieros passes gradually into the Vegamián. Van Veen (1965) and Lobato Astorga (1977: p. 47) studied the area in detail and concluded that there is an erosive disconformity between the Vegamián and the underlying formations. Erosion reached locally into the Murcia Formation, the 33 m thick Vidrieros Formation having been completely removed. A brecciated layer of nodular limestone is present in the base of the Vegamián at Enterrias (Liébana area, Maas, 1974: p. 384). Everywhere in the Cantabrian zone the base of the formation is a sharp, locally erosive, surface. In many sections where the Vegamián lies upon the limestone of the Ermita Formation subaerial dissolution (karst weathering) eroded part of the limestone. The irregular surface undulates over several centimeters but the hiatus may comprise most of the early Tournaisian. Generally glauconitic sands were deposited on this surface, locally with phosphatic nodules, except in the Palencian basin. This layer was deposited during the *cooperi-communis* Zone and may be present even where the Vegamián is lacking as at Santa Olaja de la Varga (Esla area) where a greensand was deposited during this zone. At many locations coarse sands with some glaucony occur in the top of the Ermita Formation. At Valdecastillo the Vegamián lies upon the Barrios quartzites, locally with a thin (0 to 20 m) yellowish sandstone of the Ermita in between. Below the Vegamián there are narrow fissures which may penetrate a meter into the underlying quartzites. The fissures are filled with manganese and iron oxides. The lower surface of the Vegamián is irregular because of the presence of depressions in the boundary surface. Below the Vegamián near Triollo (Cardaño area) comparable iron-manganese deposits occur in the Vidrieros Formation (Vermeulen, 1982). Vermeulen mentions slow sedimentation, or even slight erosion, an oxidizing environment and the absence of bioturbation as conditions for the formation of these deposits. Both the glaucony and the iron-manganese indicate that there was an interval of extremely slow deposition during the early Tournaisian.

The rate of deposition of the Vegamián was very low. Although some sections in the Palencian basin have up to 50 m of black shales (deposited during late Tournaisian and Viséan), usually only a few meters of sediment are present. The thickness of the formation was reduced by intraformational erosion and local erosion before deposition of the Alba. Encl. 1: Fig. 10 demonstrates that the area of the southeastern Alba syncline was slightly deformed before erosion took place. Because the formation is very thin, due to the deformation the influence of erosion varies over short distances. In the Palencian basin the boundary between Vegamián and Alba is sharp. Van Adrichem Boogaert (1967: p. 163) supposes that there was a break in the sedimentation: the lowermost part of the Alba in this area was deposited during the late Viséan and thus a hiatus may be present comprising part of the Viséan.

2.6.5. Interpretation of the sediments

The black shales of the Vegamián were deposited slowly in an euxinic environment. The oxygen deficiency is reflected in the black colour of the shales (sulphides) and resulted in a scarcity of benthonic fauna together with large numbers of pelagic radiolarians. Black shales are often associated with phosphatic nodules. Such phosphatic nodules are known from modern deposits at the upper continental slope and outer shelf near the coast of Peru-Chile and from a small area north of Walvis Bay (Namibia) where the nodules occur at the inner shelf at a depth of about 60 to 70 m and 16 to 19 km from the coast. In both areas the nodules occur only in narrow zones bordering an euxinic area. In both areas coastal upwelling promotes a high phytoplanctonic production (diatomaceous chert) and reduces to a minimum the inflow of siliciclastics. During summer a thermocline is present preventing mixing of the water (Mrs. M. Brongersma-Sanders, Leiden, pers. comm.). Black shales are supposed to be deposited below wave base in quiet, poorly aerated water and occur frequently at or close to sequences deposited during marine transgressions. Irregularities in the inundated surface or differential subsidence may result in the existence of deeper parts within the flooded area. Water tends to stagnate in these deeper parts so that black shales may be deposited (Hallam, 1981: pp. 95-101). The shallower parts of the flooded area may reach into oxygenated water where deposition of limestone (Ermita Formation) continues. If the sea level drops, erosion may occur and light-coloured coarser sediments may be deposited. When it rises again the water gradually stagnates again and deposition of black shales may be resumed. Thus the colours of the sediments reflect different conditions within the basin itself. Wetzel (1982) mentions all these kinds of sediments from upwelling regimes at the eastern margin of oceans. Thus I suppose that the Vegamián was deposited during a coastal upwelling interval after a transgression in the euxinic deeper parts of a carbonate platform and in euxinic parts of the deeper shelf which surrounded the platform. The radiolarian bearing sediments and the composition of the conodont faunas found in the southeast of the Alba syncline (Fig. 16) indicate that such a deeper sea was present in the West Asturian-Leonese zone. Although coastal upwelling occurred this sea was not necessarily oceanic:

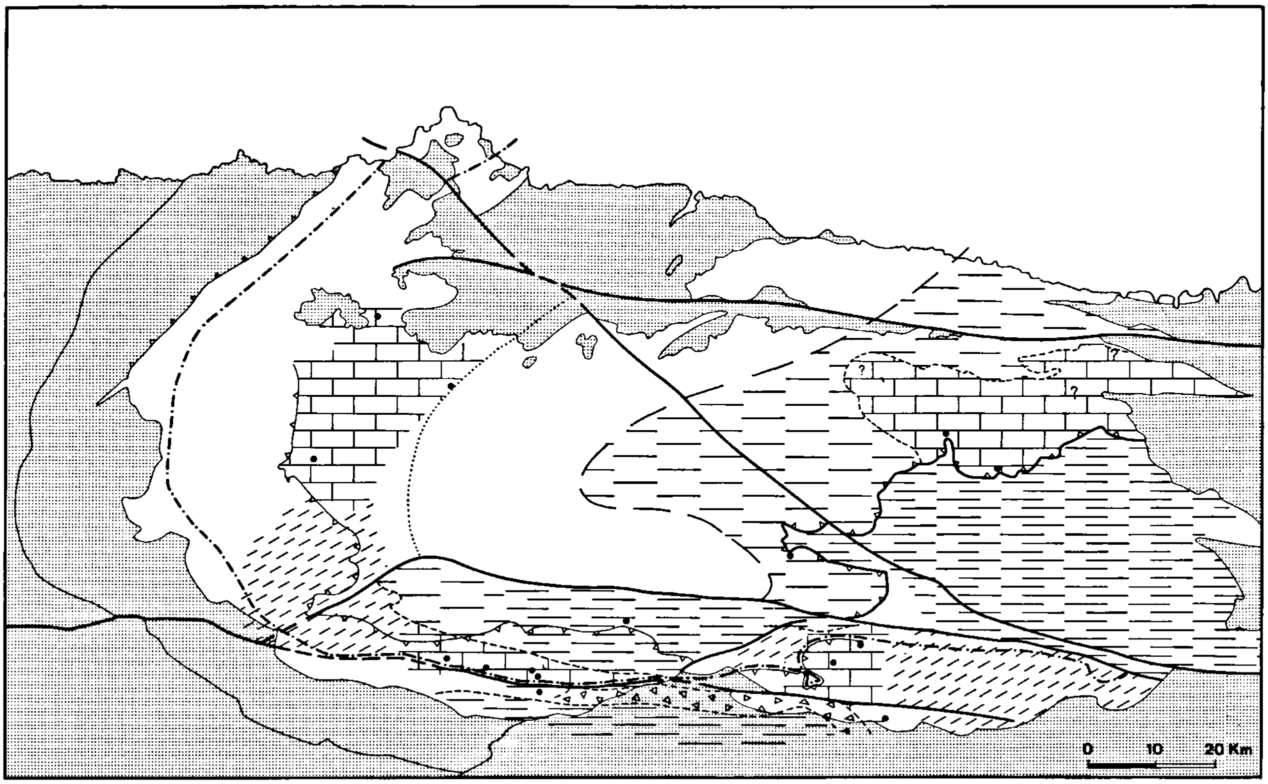

Fig. 22. Palaeogeographical reconstruction: lower part of the *pseudosemiglaber* Zone (late Tournaisian; for the legend see Fig. 23).

coastal upwelling occurs at the borders of the present oceans but the upwelled water may stream far onto the continental interior as occurred during the Permian (Phospharia Formation, U.S.A.; Hallam, 1981: p. 83) and Late Devonian (Ouachita basin, U.S.A.; Lowe, 1975).

Frankenfeld (1981) interpreted the upper part of the Ermita limestone and the pebble layer in the Vegamián Formation in the southeast of the Alba syncline as turbiditic deposits. The characteristic sequences for turbidites have not been observed so that I do not concur with this opinion. It does not seem possible that turbidites were deposited on topographical highs.

The deposits were formed in rather shallow water: although depths of at least 50 m are given for the deposition of phosphates and glaucony (Hallam, 1981; Odin & Matter, 1981) the maximum depth in the Asturo-Leonesian basin and on the Asturian geanticline is estimated to have been some tens of meters: the erosion surfaces indicate repeated subaerial exposure and the Ermita grainstones (deposited at the same time, 2.5) formed above wave-base. The Palencian basin, however, is supposed to have been slightly deeper: sedimentation was largely uninterrupted, and fauna, except pelagic radiolarians, is entirely lacking. I estimate the maximum depth between 60 and 100 meters.

The facies maps (Figs. 22-24) constructed with the new data presented in this section and the data from the conodonts, are completely different from those constructed by van Adrichem Boogaert (1967: fig. 65) and Frankenfeld (1981: fig. 74) in that these authors indicate erosion in large positive areas instead of the carbonate platforms. The Cantabrian zone (except the Palencian basin) was still a shallow area surrounded by deeper seas but the formation of deeper parts in which the Vegamián was deposited indicates that an inversion of the relief started during this age.

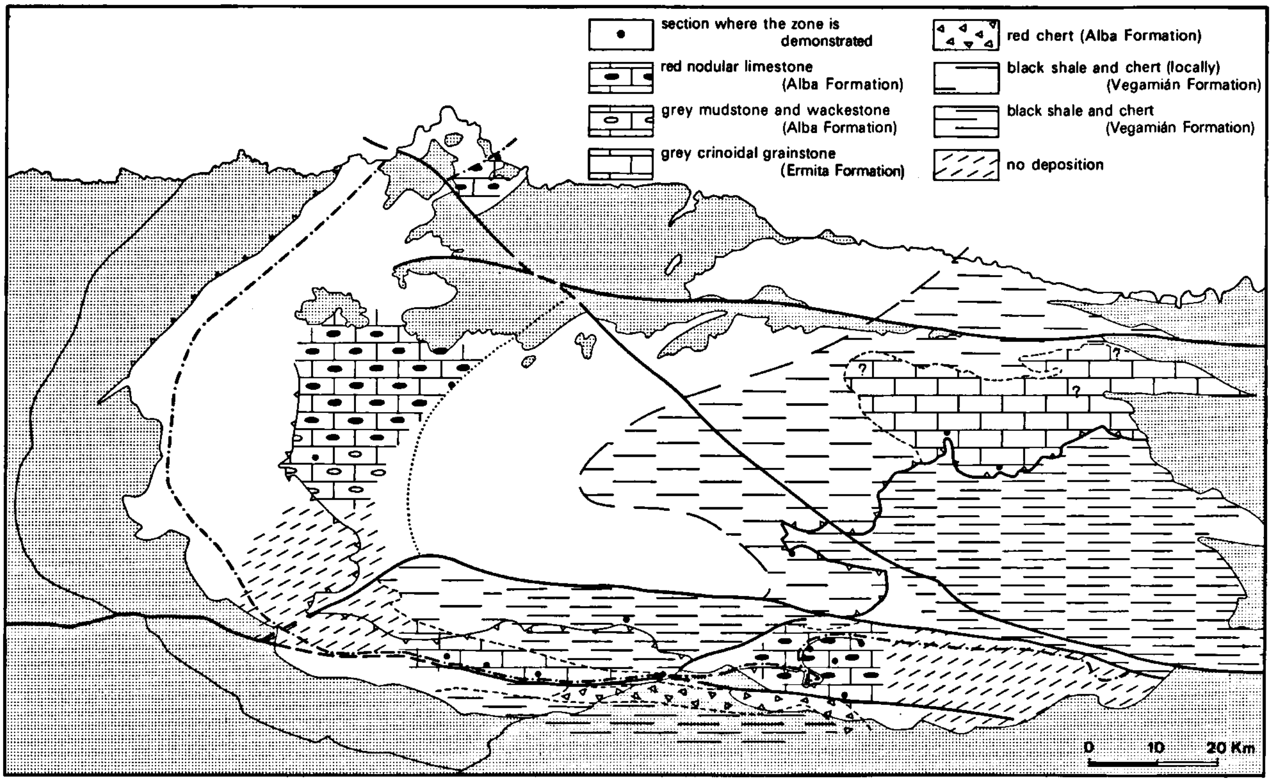

Fig. 23. Palaeogeographical reconstruction: upper part of the *pseudosemiglaber* Zone (late Tournaisian).

2.7. ALBA FORMATION (LOWER CARBONIFEROUS)

2.7.1. Introduction

Like the other formations discussed in this paper the Alba Formation has been renamed several times (Table 1). I will refer to it with the most commonly used name Alba Formation. Attention will be paid only to the lowermost part of the formation which may give some more information on the conditions in the region and on the length of the interval of time during which the Vegamián was deposited. Some problems concerning the base of the formation have been treated in 2.5.5.

Characteristic for the Alba Formation are red and grey nodular limestones but shales and bedded chert occur frequently. Detailed descriptions are given by Wagner et al. (1971, with references).

2.7.2. Erosion at the base of the formation

The Alba Formation lies disconformably upon the older Ermita and Vegamián (2.5.5 and 2.6.2). At many localities the contact between Alba and underlying formations is erosive, e.g. at San Román de Candamo (Asturias) where pebbles of white limestone from the Ermita occur near the base of the Alba (Pello, 1972). In the Liébana area erosion was stronger: locally the Vidrieros and Vegamián are absent, the Alba lying upon the Murcia Formation (Maas, 1974: p. 385). From the area east of Barruelo de Santullán (Palencia) there is even an angular unconformity below the Alba: the Devonian below it was folded before deposition of the nodular limestones (Wagner & Wagner-Gentis, 1963). In the southeast of the Alba syncline erosion occurred before deposition of the Alba (Encl. 1: Fig. 10). On the platforms (Esla area, Asturias) the transition from Ermita to Alba is generally sharp and red nodular limestones were deposited immediately afterwards. Only in areas close to the Vegamián basins a thin grey wackestone was

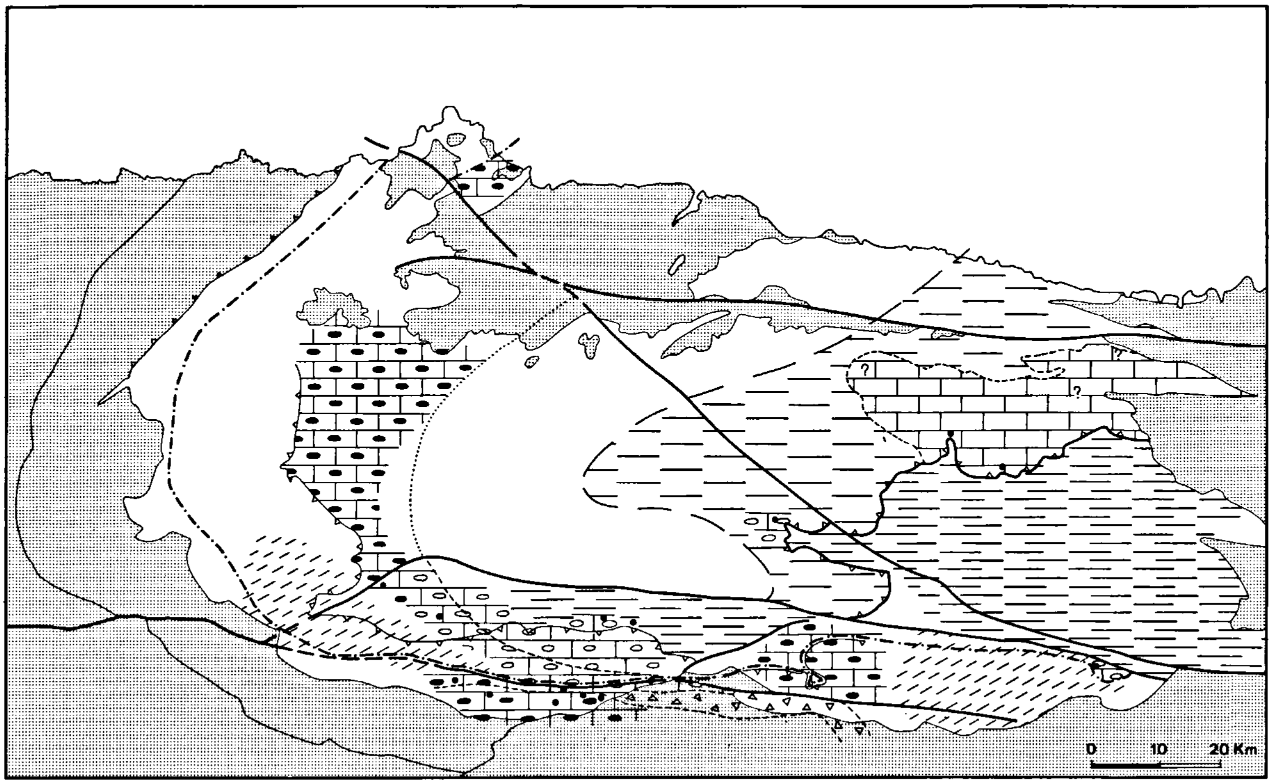

Fig. 24. Palaeogeographical reconstruction: lower part of the *anchoralis-latus* Zone (early Viseán; for the legend see Fig. 23).

initially deposited (e.g. Felechosa, Isidro area; San Martín de los Herreros, Ventanilla area; Santa Olaja de la Varga, Esla area; Aviados, Bernesga area; San Emiliano, Caldas area). In the deeper parts the transition from Vegamián black shales and chert to red nodular limestones is more gradual: via green shales, red shales, red shales with limestone nodules, red cherts, grey shales or grey wackestones.

2.7.3. Fossils and age of the lower part of the formation

Many different fossil groups are represented in the Alba Formation: benthonic as well as pelagic life forms, the pelagic fauna being most abundant. Cephalopods, ostracods, crinoids, echinoids, trilobites, brachiopods, bivalves, gastropods, bryozoans, foraminifera, solitary corals, radiolarians, conodonts, sponges and fish have been found. In the basal part of the formation the fossils may be partly dissolved by algae (van Tongeren, 1975) and bioturbation is common in the red nodular limestones.

Locally deposition of the Alba started in the Tournaisian e.g. the red cherts between Peredilla and Cistierna (2.6.2; Fig. 22). The oldest red nodular limestones are from the *pseudosemiglaber* Zone (late Tournaisian) and occur in the Esla area (Pico Aguasalio, Valdoré, Santa Olaja de la Varga) and in Asturias (Loredo: del Río & Menéndez Alvarez, 1978; Perlora) (Fig. 23). In the Bernesga area (Aviados, Beberino) and locally in Asturias (Entrago: Budinger & Kullman, 1964; Pello, 1972) grey wackestones and mudstones were deposited (Fig. 23). During the *anchoralis-latus* Zone deposition of red nodular limestone extended to the south of Asturias (Entrago, Puerto de la Cubilla) and further to the Bernesga area south of the Intra-Asturo-Leonesian facies line (Sagüera, Portilla de Luna, Piedrasecha, Olleros de Alba, Alcedo, Rabanal de Fenar) because the circulation in this area had improved (Fig. 24). At other locali ties grey wackestones and mudstones were deposited: along the facies line in the Bernesga area (Aviados, Nocedo de Bernesga), in the western part of the Leonides (Caldas de Luna), locally in the northwest of the Leonides (Venta de Getino: Higgins & Wagner-Gentis, 1982), at San Martín de los Herreros (Ventanilla area) and Riosol (Asturian geanticline) (van der Ark, 1982 and other data). Elsewhere deposition of black shales continued (Fig. 23); only at Las Portillas and south of Caín (both in the Picos de Europa) grey crinoidal grainstones continued to be deposited. In these areas deposition of the nodular limestones began later during the *anchoralis-latus* Zone. In the Palencian basin deposition of this limestone began in the late Viséan *bilineatus* Zone (van der Ark, 1982). The thicker sequence of Vegamián in the Palencian basin may be the result of prolonged deposition into Viséan time but there is no real necessity or proof of this.

2.7.4. *Interpretation of the sediments*

The origin of nodular limestone is still uncertain. It seems probable that nodular limestone may originate in various ways under different conditions. Different and opposed theories have been given by different authors, amongst others Gründel & Rösler (1963), Jenkyns (1974), Müller & Fabricius (1974) and Tucker (1974) but it is difficult to reconcile them. The limestones of the Alba Formation generally show the same characteristics as those described in the publications mentioned above. The nodules are supposed to have formed either during sedimentation or early diagenesis, at or near the sediment-water interface. Müller & Fabricius (1974) argue that the greatest part of the carbonate precipitated directly from the seawater as Mg-calcite and therefore special conditions are required: restricted circulation, high salinity and high temperature combined with a slow rate of deposition (estimated a few mm/10^3yr, Jenkyns, 1974). Deposition occurs in an oxygenated environment and part of the carbonate may be derived diagenetically from calcitic skeletons.

The present study may contribute to this problem since rather many data are available on the constitution of the depositional environment and on the history of sedimentation. During the late Tournaisian (*pseudosemiglaber* Zone) there was a carbonate platform in a part of the Asturo-Leonesian basin. In deeper areas with stagnant water and in an oxygen-depleted area in the southeast of the Alba syncline black shales were deposited (Vegamián Formation, 2.6; Figs. 22 and 25a). Due to a change in water circulation the oxygenation in these areas gradually ameliorated and other types of sediment were deposited. In the upper *pseudosemiglaber* Zone deposition of red nodular limestones started on the platforms in the Esla area and in Asturias

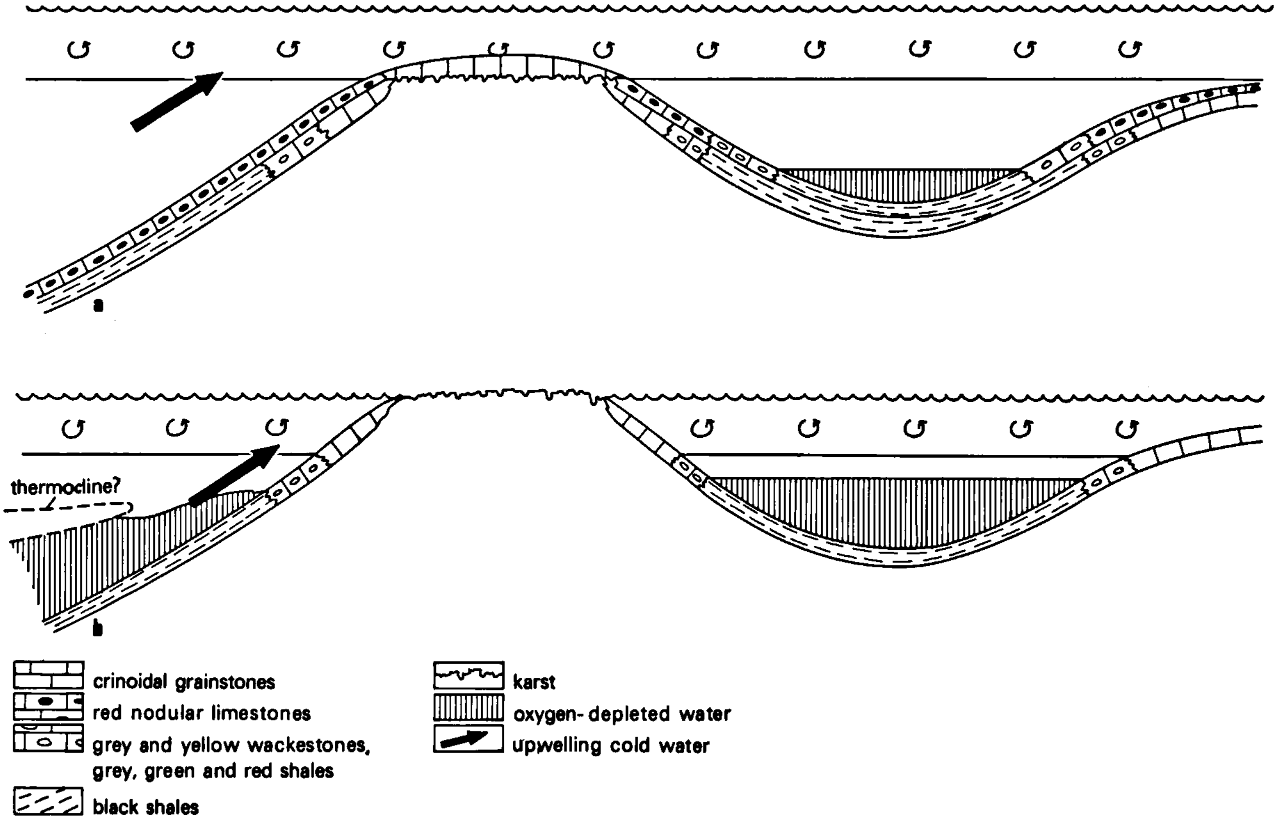

Fig. 25. Sketch of the relation between water-circulation and facies (not to scale; Tournaisian: Ermita, Vegamián and Alba Formation). (a) In the turbulent upper water layer between sealevel and wave-base crinoidal grainstones were deposited. In the lower water layer, which tended to stagnance, grey and yellow wackestones and grey, green and red shales were deposited. In the lowermost, oxygen-depleted, part of that layer black shales dominated. Phosphatic nodules formed at the border of the oxygen-depleted zone. Layers of radiolarian chert occurred in all the facies formed below wave-base. (b) Due to the rising sealevel circulation improved and sedimentation of red nodular limestones commenced extending over the other facies.

while black shales continued to be deposited in the deeper parts and grey mudstones and wackestones at their borders. At the same time red cherts were laid down along the southern slope of the carbonate platform and crinoidal grainstones in a part of the Pedroso syncline and in the Picos de Europa (Figs. 23 and 25b). Hence, Ermita, Vegamián and Alba lithosomes were deposited contemporaneously in different parts of the region. During the *anchoralis-latus* Zone circulation seems to have been restored first in the southeast of the Alba syncline and later also in the northeast of the Bernesga area and on the eastern Asturian geanticline: red (and grey) nodular limestones were then deposited all over the Cantabrian zone except in the Palencian basin (Fig. 24). Probably the thermocline (2.6.5) had disappeared and therefore circulation improved. I suppose that the crinoidal grainstones were formed above wave-base and the nodular limestones below it. The grey mudstones and wackestones with a predominant pelagic fauna formed in poorly oxygenated water near the borders of the basins, and the red mudstones and wackestones with abundant pelagic and benthonic faunas formed in well-oxygenated environments.*) The clay particles dispersed in the seawater settled out of the water in the quieter water below wave-base. The same clay formed the source of the black shales in anoxic environments. The red nodular limestones formed on the platforms but also in the deeper parts where these became better oxygenated. They never formed in the Palencian basin where only grey nodular limestones occur, indicating that that basin was still less oxygenated when deposition of nodular limestone started in that area. With these data I estimate that the red nodular limestones in Cantabria were deposited at depths between 10 and 50 m, in accordance with the occurrence of algae near the base of the formation (van Tongeren, 1975) and with the estimated depths of deposition of Ermita and Vegamián. The grey nodular limestones were deposited at the same depths and deeper. Probably red nodular limestones may also form in much deeper environments where the oxygenation of the environment is sufficient, but great depths of deposition must be excluded for Cantabria. Thus it seems that these nodular limestones formed below wave-base in areas with slow sedimentation. In well-oxygenated environments red limestones formed and in less-oxygenated environments grey and yellow limestones.

During the Tournaisian and early Viséan the Picos de Europa were the shallowest part of the Cantabrian zone: there deposition of crinoidal grainstones continued even during part of the *anchoralis-latus* Zone. Later during the Viséan the Picos de Europa continued to be a very shallow area (P. Seibert, Tübingen, pers. comm.). The Ponga Nappe province was slightly deeper. The main part of the Asturo-Leonesian basin was a shallow platform. The occurrence of *Scaliognathus anchoralis* and *Doliognathus latus* in the southeast of the Alba syncline (Fig. 16; 4.4) indicates that this area was the deepest part of the Cantabrian zone with a good water circulation.

3. PALENCIAN BASIN

3.1. INTRODUCTION

In the Palencian basin (Fig. 1) the Devonian and Lower Carboniferous are remarkably constant throughout, and much less varied than in the Asturo-Leonesian basin. In the main, deposits consist of shales and limestones. In this basin the Devonian is about 500 m thick (van Veen, 1965) against about 1200 m at Pico Aguasalio in the Esla area (Rupke, 1965). In contrast to the varied fauna of benthonic organisms in the Asturo-Leonesian basin, the faunas of the Palencian basin consist mainly of pelagic organisms and are poor in individuals of tentaculites, conodonts and cephalopods (Brouwer, 1968; Kullmann, 1968). The most detailed description is given by van Veen (1965) who introduced the formation names. A description of the Vegamián and the Alba Formation is given in 2.6 and 2.7.

3.2. GUSTALAPIEDRA FORMATION AND CARDAÑO MEMBER

Van Veen (1965) introduced the Gustalapiedra Formation with its stratotype along the Arroyo de Gustalapiedra, northwest of Cardaño de Arriba. The formation consists of shales and nodular limestones. Van Veen (1965) also introduced the Cardaño Formation, a nodular limestone with its stratotype north of Cardaño de Arriba. Ambrose (1972) writes that in the valley of the Río Carrión about 50 m of shales are present between the Cardaño and the Murcia Formation, and because the nodular limestone grades laterally into shales he includes the Cardaño as a member in the Gustalapiedra Formation. The shales between the Cardaño and the Murcia Formation may then be included in the Gustalapiedra Formation.

Conodonts from the Gustalapiedra Formation and the Cardaño Member were studied by Budinger (in Budinger & Kullmann, 1965), van Adrichem Boogaert (1967), Mouravieff (in Lobato Astorga, 1977) and myself. There are few data from the shaly part of the Gustalapiedra Formation below the Cardaño Member. Most samples are from the middle and upper part of these shales and were deposited during the *varcus* Zone (van Adrichem Boogaert, 1967; Lobato Astorga, 1977; see 4.3). A sample taken 10 m below the Cardaño (along the Río Carrión, northeast of Triollo) contained a conodont fauna of the middle *Polygnathus costatus costatus* Zone (early Eifelian). From the Cardaño Member more samples were studied, many of them with faunas

*) Lime mud may be precipitated from seawater by bacteria (C. Monty, Liège, pers. comm.).

rich in species but others, especially from the lower and middle part, are barren. The samples from the lower part belong to the *varcus* Zone except at Cardaño de Arriba where the oldest fauna is from the *hermanni-cristatus* Zone (van Adrichem Boogaert, 1967; Lobato Astorga, 1977 and new data). Van Adrichem Boogaert's sample CU3, taken near the base of the Cardaño in the Gildar-Montó area, belongs to the Middle *varcus* Subzone. The upper part of the member was deposited during the Upper *gigas* Zone but Lobato Astorga (1977: p. 33) mentions a sample taken 2 m below the top (in the stratotype) which contains a mixed fauna of conodonts from the Lower and Upper *Palmatolepis triangularis* Zone and it is possible that also at other parts of the basin the member contains an interval deposited after the *gigas* Zone. In the Arauz-Polentinos area the upper part of the Cardaño was deposited during the *varcus* Zone (van Adrichem Boogaert, 1967, conodonts; Goeijenbier, 1981, tentaculites). There 50 m of shales were deposited during the Frasnian. The great differences in age of the top of the Cardaño favour Ambrose's opinion that the Cardaño at best can be included into the Gustalapiedra Formation (compare Fig. 27). Then the total thickness of the formation is about 150 m in the section northwest of Cardaño de Arriba.

3.3. MURCIA FORMATION

3.3.1. Introduction

The Murcia Formation was proposed by van Veen (1965), with a stratotype at the western side of Pico Murcia, northwest of Cardaño de Arriba. Ambrose (1972) considers this name as a junior synonym of the Moradillo Sandstone introduced by Wagner (1955). Wagner & Wagner-Gentis (1963) introduced the Moradillo Sandstone Formation but without giving a good description. The proposed stratotype near Barruelo de Santullán is badly exposed and Ambrose (1972) even suggested using that of the Murcia Formation. Here I shall use the well established name and the well described stratotype of the Murcia Formation.

3.3.2. Sediment petrology

In the stratotype near Pico Murcia and north of Cardaño de Arriba the shales in the upper part of the Gustalapiedra Formation grade into the sandstones of the Murcia Formation. These are dark grey, well-sorted, fine to medium grained quartzitic sandstones. Locally they are grey or brownish and often real quartzites occur (van Veen, 1965; Ambrose, 1972). The average thickness of the beds is 50 cm but individual beds may be from one grain to two meters thick. Upwards these sandstones pass into the nodular limestones of the Vidrieros Formation. In the section near Cardaño de Arriba the lower part of the formation is formed by an alternation of shales, siltstones and sandstones. The upper part consists mainly of quartzitic sandstones and quartzites. In the sandstone load casts, flute marks, small-scale current ripples, parallel, wavey and convolute lamination and grading occur. Grading occurs in two types: as a general decrease in grain size from base to top of the bed and by an increase of sorting, the very coarse sand and granule being restricted to the lower part of some of the beds. In a detailed section measured near Cardaño de Arriba (Fig. 26) Bouma intervals were recognized. In the shales parallel lamination is the common structure. Van Veen (1965) mentions worm tracks and Kunst (1981) found vertical burrows (escape burrows?) in the upper part of beds north of Santibañez de Resoba. The formation is 60 to 200 m thick (van Veen, 1965): 107 m in the stratotype at Pico Murcia (Frikken, 1978) and thicker in the east (van Veen, 1965). In the southern part of the Cardaño area, along the Cardaño line (south of Cardaño and near Vidrieros) locally coarse sandstones with cross-bedding, load casts, flute marks and groove marks were found: the sole marks in east-west and southeast-northwest directions (Frikken, 1978).

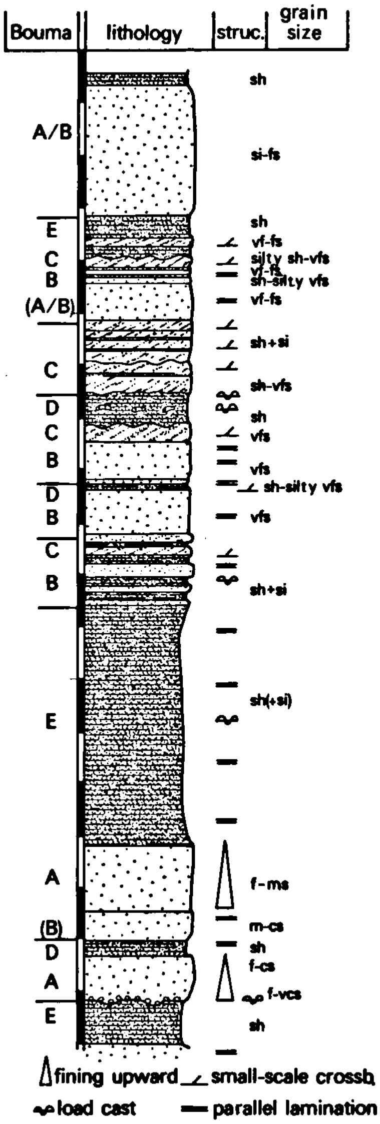

Fig. 26. Detailed section of a fine-grained interval in the lower part of the Murcia Formation north of Cardaño de Arriba. In the left column the Bouma intervals of turbidite sequences are indicated, the next column gives the scale (each bar is 10 cm), the grain size is indicated as follows: sh = shale, si = silt, vfs = very fine sand, fs = fine sand, ms = medium sand, cs = coarse sand, vcs = very coarse sand.

In the Gildar-Montó area the lower part of the formation consists of very fine grained and thin sandstone beds which have a constant thickness over a considerable area and are separated by shales. Towards the middle part the grains become coarser (even up to quartz pebbles with a diameter of 5 cm), the beds are graded and shales are lacking. In the upper part the grain size and the thickness of the beds diminish again (Frankenfeld, 1981). The formation is about 100 m thick (van Adrichem Boogaert, 1967: fig. 24).

In the Liébana area the formation consists of a light-coloured quartzite with sandstone beds of 0.2 to 1 m thick, alternating with thin beds of fine black shales. Frequently the base of the sandstone beds is loaded and also flute marks and groove marks occur (Maas, 1974: p. 384). The formation is 40 to 60 m thick (van Adrichem Boogaert, 1967: p. 162).

3.3.3. Age of the formation

Data from the adjacent formations suggest that deposition of the Murcia started after the Upper *gigas* Zone and at least locally only after the *Palmatolepis triangularis* Zone (Fig. 27). The oldest limestones of the overlying Vidrieros Formation are from the *marginifera* Zone. Thus the Murcia Formation was deposited at least during the *crepida* Zone and the *rhomboidea* Zone (compare 4.3). Maas (1974: p. 384) suggests that in the Liébana area the Murcia and Vidrieros interfinger with each other which would imply the continuation of the deposition of the Murcia later during the Famennian but such an interfingering was not found in the other, better exposed areas.

3.3.4. Interpretation of the sediments

The data from the different areas indicate that the coarse sediments were deposited from high-density turbidity currents. The detailed section (Fig. 26), measured in the predominantly shaly lower part of the formation, shows some thin layers and laminae of the distal part of small turbidites but generally these sandstones are thicker, up to tens of centimeters thick beds, mainly of medium sand. The scarce data available suggest that there were two major sources: one to the west of the Gildar-Montó area (where the coarsest sediments occur, the grain size diminishing towards the east) and one to the southeast of the Cardaño area (grain size and thickness of the lithosome diminish towards the west and northwest and sedimentary structures indicate transport in that direction). The formation is thin in the Liébana area where Maas suggested interfingering with the Vidrieros. Therefore the siliciclastic material appears to have been derived from the Asturian geanticline and from an area south of the Cardaño line. Also during the Silurian and early Devonian siliciclastic sediments in the southeast of the Palencian basin were derived from a land area south of the Cardaño line (Krans, 1982). Krans mentions transport directions from between south and east, just as I found for the Murcia.

3.4. VIDRIEROS FORMATION

Van Veen (1965) introduced the Vidrieros Formation with its stratotype near Pico Murcia. Kullmann (1960) had proposed the name "Montó-Schichten" for an equivalent sequence but the limits were not clearly defined (Ambrose, 1972). Another early synonym is the Verbios Formation, introduced by Wagner & Wagner-Gentis (1963). Ambrose prefers the name Vidrieros Formation because of the inadequate exposure of the proposed stratotype of the Verbios. The Vidrieros Formation consists of nodular limestones and shales (Pl. 2: Fig. 5). Conodont faunas older than the *marginifera* Zone were not found. Only the sample Ca3 from east of Santibañez de Resoba may be slightly older: it contains a fauna from somewhere in the interval Upper *crepida* to *marginifera* Zone. The youngest samples are found above the deposits of the *costatus* Zone and contain the *Protognathodus* fauna and *Polygnathus* fauna (late Famennian-early Tournaisian; 4.3; Triollo, Santibañez de Resoba).

3.5. FOSSILS

Faunas in the Palencian basin typically consist of small specimens (juvenile, dwarfed?) or aberrant forms. In the Abadía and Lebanza Formations a rich benthonic fauna is present but in the younger formations only a few benthonic macrofossils occur from a limited number of species.

The corals are represented by solitary rugose corals which lack dissepiments (Kullmann, 1968) as is characteristic for corals living in unfavourable environments. Many of the species are endemic for the Palencian basin. Kullmann (1968) concluded that the cephalopods are only represented by few species which have a cosmopolitan distribution. The dacryoconarids (tentaculites), which locally may compose up to 60 % of the limestones, remained extremely small (J.P.S. Goeijenbier, Leiden, pers. comm.). The conodonts are well preserved as are the numerous accessory elements. Although some samples contain a large number of specimens and species compared to other condensed sequences, the faunas are on the whole relatively poor. Many of the conodonts are small, which indicates that the Palencian basin was generally a very quiet environment, rather unfavourable for the benthonic fauna and even for pelagic animals. A comparison of the conodont faunas (4.4) indicates that the Palencian basin initially was a shallow area but from the *hermanni-cristatus* Zone onwards it deepened, becoming deeper than the Asturo-Leonesian basin.

3.6. INTERPRETATION

The oldest sediments known from the Palencian basin are late Silurian. Then the Asturo-Leonesian and Palencian basins were already differentiated (Ambrose, 1972; Krans, 1982). The difference increased during the Devonian, together with the rising of the Asturian geanticline and the zone along the Cardaño line in Palencia. The differences between the basins gradually increased during the Devonian. Instead of the biostromal limestones and the sandstones of the Asturo-Leonesian basin, in the Palencian basin deposition consisted mainly of nodular limestones and shales. The nodular limestones of the Abadía, the Cardaño, the Vidrieros and the Alba Formation are very much alike, so much that they usually can be identified only with help of a diagnostic fauna. There are only two aberrant types of sediment: the Murcia sandstones and the Vegamián black shales and cherts. The siliciclastics of the Murcia Formation were derived from shallower areas and transported towards the deeper parts of the basin by turbidity currents. Unlike the sheet-like turbiditic storm deposits of the Fueyo Formation in the Asturo-Leonesian basin, those of the Murcia most probably were deposited as turbidite fans. These were not caused by changes within the basin itself: when the supply of sand terminated limestones were deposited once more. This occurred also at other moments: the sedimentation of nodular limestones (Polentinos Member) in the Abadía Formation was also interrupted several times by turbidity currents: graded sand beds occur near the Abadía de Lebanza (J.F. Savage, Utrecht, pers. comm.). Generally such coarser sediments did not reach the Palencian basin while erosion at the Asturian geanticline is reflected in most of the Devonian in the Asturo-Leonesian basin. Maps showing the level of erosion below the upper Famennian deposits (Fig. 19 and Parga, 1969: fig. 1) indicate most pronounced erosion on the eastern part of the Asturian geanticline, thus suggesting a tilting of the central block towards the Asturo-Leonesian basin and a steep slope towards the Palencian basin. Therefore most of the erosion products were transported towards the Asturo-Leonesian basin. It was only during the early Famennian, an episode of emergence of the Asturian geanticline and the Asturo-Leonesian basin, that any clastic detritus reached the Palencian basin.

The conditions in the basin changed during the early Tournaisian as is well demonstrated by a section at Santibañez de Resoba (studied by C.F. Winkler Prins and M. van den Boogaard, Leiden, pers. comm.). Above nodular limestones of the Vidrieros Formation which contain a fauna of the *costatus* Zone with a constant composition there are a thin shale and a laminated limestone with the *Protognathodus* fauna and *Polygnathus* fauna. Above these limestones there are black shales of the Vegamián which were deposited under anoxic conditions in stagnant water (2.6.5). Unfortunately part of the sequence below the Vegamián may have been removed by erosion (2.6.4). Also below the Alba Formation there is a hiatus caused by erosion (2.7.2). The grey colour of the limestones of the Alba Formation in this basin indicates that during the late Viséan the circulation in the basin was not very good.

The lithological and palaeontological data suggest that during the Devonian the Palencian basin formed part of an outer shelf to continental slope.

4. CONODONT FAUNAS

4.1. INTRODUCTION

The limestones of the Portilla and the Nocedo were deposited in and near reef in which conodonts are scarce. After study of the samples taken by others and those taken by myself during the first field summer I chose to sample what had become evident as most promising rocks: the fore-reef facies (4.4) in a large number of sections. This contrasts the approach of García López (1972, 1976) who sampled extensively a few sections. Sandstones were not sampled as it is improbable that these contain many conodonts (4.4). As the rate of sedimentation during late Famennian and early Carboniferous was low, more intensive sampling was required in this interval, particularly in the upper part of the Ermita Formation. For practical reasons samples were mainly taken from limestones. Generally samples of 2 to 3 kg were taken, some of these were split into segments which were treated separately.

Samples were broken into small pieces and then dissolved in the laboratory by being kept 24 hours in diluted formic acid (10 %) followed by another 24 hours in fresh acid. The residue was then sieved (meshes of 2.0 and 0.1 mm). The intermediate residue (0.1-2.0 mm) was dried, after which the heavy fraction was separated off with the help of bromoform. Lastly the magnetic fraction was separated off with a magnetic separator. The conodonts were picked out under a binocular microscope.

About 1200 samples were processed, a third of which contained no identifiable conodonts, and only half the remainder yielded determinations suitable for zonal definition. Most species were photographed during examination with an S.E.M. (gold coating). Photographs of the most important species are reproduced at Plates 3-6. Only the problematic and interesting species will be discussed hereafter (4.2). Further the biozonation (4.3) and palaeoecology (4.4) will be discussed. A separate communication on the observations of the Colour Alteration Index (C.A.I., Epstein et al., 1977) is in preparation.

4.2. SYSTEMATICS

Bispathodus spinulicostatus (E.R. Branson, 1934)
Ziegler (in: Ziegler, 1975) mentions *B. spinulicostatus* as a rare form from the late Famennian, more commonly occurring during the early Tournaisian. Johnston & Higgins (1981) mention the species from younger deposits: the late Tournaisian where it occurs with *Ps. multistriatus*. In Cantabria it occurs from *costatus* Zone to the lower part of the *anchoralis-latus* Zone, being most common in the *pseudosemiglaber* Zone (Fig. 28). Specimens such as the one figured by Johnston & Higgins (1981, pl. 5: fig. 7) as *B. spinulicostatus* were identified as *Ps. dentilineatus (= primus)* by Rhodes et al. (1969, e.g. pl. 5: fig. 12). I consider such specimens to be *B. spinulicostatus*. In *B. spinulicostatus* at the left side of the anterior part of the platform the denticles are much weaker developed than in *Ps. primus* or even absent. It is not known whether there is a hiatus in the sequences between the occurrences of *B. spinulicostatus* in the *costatus* Zone and the *cooperi-communis* Zone in Cantabria. Johnston & Higgins (1981) suggest that the younger form evolved during the early Tournaisian and probably is a homeomorph of *B. spinulicostatus* s.s. because they found the forms in two levels separated by a long interval in which it was not present.

Clydagnathus gilwernensis Rhodes, Austin & Druce, 1969 (Pl. 6: Fig. 13)
This species was first recognized by M. van den Boogaard (Leiden, pers. comm.) in a sample from the Ermita Formation in the Picos de Europa (sample HG42, coll. RGM 248954). Later on it was also found in a sample from the Ermita at Valdoré (in van Adrichem Boogaert's sample IIf166, coll. RGM 125143), from the Vidrieros Formation at Santibañez de Resoba (Palencian basin, M. van den Boogaard, pers. comm.) and in a sample from an olistostrome southwest of Fuente Dé (Palencian basin, a mixed fauna, coll. RGM 248955). Sandberg & Ziegler (1979) argue that this species evolved from *Clyd. cavusformis* Rhodes, Austin & Druce, 1969 during the Early Carboniferous, probably within the *sulcata* Zone. In the section at Santibañez de Resoba, however, the species occurs in the Vidrieros Formation below faunas from the *costatus* Zone (M. van den Boogaard, pers. comm.). The sample from the Picos de Europa also proves that the species existed already during the Devonian: there it occurs with *Pelekysgnathus inclinatus* Thomas, 1949 which is known from *velifer* Zone through *praesulcata* Zone (Sandberg & Ziegler, 1979) thus being restricted to the Famennian (compare Figs. 27, 28).

Gnathodus pseudosemiglaber Thompson & Fellows, 1970 (Pl. 6: Figs. 2,4)
In the past the classification of the species belonging to this genus has been problematic but since the appearance of a classification which is not based on one area (Lane et al., 1980) good determinations are possible. Because the authors of older publications on the Cantabrian Mountains used other criteria for the subdivision of the genus older identifications are often in error. In consequence a species as *G. pseudosemiglaber* was identified by Higgins (1974; as *G. texanus pseudosemiglaber*) as occurring in low numbers in the *anchoralis-latus* Zone (*anchoralis* Zone) and younger zones while *G. semiglaber* (Bischoff, 1957) and *G. typicus* Cooper, 1939 (identified as *G. antetexanus* Rexroad & Scott, 1964) were described as very common. *G. pseudosemiglaber*, however, is a very common species in the Cantabrian Mountains making up the majority of *Gnathodus* specimens.

The species occurs well before the *anchoralis-latus* Zone in faunas containing, amongst others, *Siphonodella* species and other conodonts which are not known from the *anchoralis-latus* Zone (Fig. 28). The abundance of the species and the occurrence in Cantabria well before it is known in other areas may indicate that the species originated in Cantabria. In several samples from the *pseudosemiglaber* Zone (4.3) and the base of the *anchoralis-latus* Zone specimens occur with a row of nodes at both sides of the posterior part of the carina thus suggesting a close relation to *G. cuneiformis* Mehl & Thomas, 1947 (Pl. 6: Figs. 1, 3). Both lateral rows of nodes of *G. cuneiformis* get fused with the carina thus forming a thick carina and the other nodes fuse into a parapet and a pseudoparapet (Pl. 6: Figs. 2,4). Lane et al. (1980) place both species in different groups and suppose that *G. pseudosemiglaber* evolved from *G. semiglaber*. The groups originated during the *isosticha*-Upper *crenulata* Zone in which there still occurred forms intermediate between most species (Lane et al., 1980: fig. 1), also *G. pseudosemiglaber* originated during that zone, therefore it is very difficult to ascertain how the species evolved from one another.

Icriodus eslaensis van Adrichem Boogaert, 1967 (Pl. 4: Figs. 7, 9-13)
There is much confusion about the status of this taxon. Several authors consider the taxon synonymous with *I. brevis* Stauffer, 1940 (Klapper in: Ziegler, 1975) or *I. obliquimarginatus* Bischoff & Ziegler, 1957 (Seddon, 1970). On the other hand Bultynck (1975) introduced a new subspecies, later raised to species: *I. latecarinatus* (Brice et al., 1979; Bultynck & Hollard, 1980). With Bultynck (1972, 1975) I believe that *I. eslaensis* is a taxon which can be distinguished very well from *I. obliquimarginatus* while the relation with *I. brevis* remains doubtful. The distinction between *I. eslaensis* and *I. latecarinatus* can not be easily made in the samples from the Cantabrian Mountains since most of the samples are small, the preservation of many conodonts is bad and the variation is large. I agree with Bultynck (1975) that the samples from younger deposits (from higher in the Middle *varcus* Subzone into the *asymmetricus* Zone) generally show a greater variation but I include all the specimens in *I. eslaensis* s.l.

Icriodus aff. *subterminis* Youngquist, 1947 (Pl. 4: Figs. 1-5)
The axis is straight, platform margins are biconvex in juvenile specimens and subparallel in adults (Pl. 4: Figs. 1-5). The median row consists of 5 to 6 discrete denticles which are round in outline; only the anterior denticles are smaller and laterally compressed. The five denticles of the lateral rows are round or elongate transversely. Because they may extend beyond the platform the conodont may be rather broad. The denticles are alternate. The blade consists of two or three plump denticles which may be partly fused. Posteriorward these denticles increase in size but because the blade may be curved downward the denticles not necessarily extend above those of the platform. Aborally the basal cavity may be moderately to very broad, with a distinct spur. Phylogenetically late forms (*gigas* Zone) may have an extremely large basal cavity. Remarks: this form closely resembles the conodonts referred to as *I.* aff. *subterminis* in Orchard (1978), only the variation in the Cantabrian material is greater. The taxon may be confused with *I. symmetricus* (Pl. 4: Figs. 8, 14-16). The latter species has a more slender platform, a larger number of lateral teeth and a less broad basal cavity without a spur.

Palmatolepis gracilis Branson & Mehl, 1934 (Pl. 4: Fig. 24)
This species occurs commonly in the Vidrieros Formation (Palencian basin) and few specimens are known from the Fueyo Formation (sample OLL1, Olleros de Alba, coll. RGM). One specimen (sample Ca6, south of Pico Murcia, Cardaño area; Pl. 4: Fig. 24) has an outer lobe with a secondary carina as is typical for other species of this genus but not for *Pa. gracilis*. A specimen illustrated by van den Boogaard & Schermerhorn (1975, pl. 4: fig. 1) shows the same feature. It is considered to be an aberrant form.

Polygnathus communis Branson & Mehl, 1934 and *P. purus* Voges, 1959
As van Adrichem Boogaert (1967: p. 184) already has remarked, specimens of *P. communis* occur, in which the characteristic excavation behind the basal cavity is absent. Van Adrichem Boogaert indicated these specimens as *P.* cf. *communis communis*. Also Voges (1959: p. 290) found such specimens. But most of the samples contain only specimens with the excavation. In specimens lacking the excavation, particularly in juvenile specimens, it may be difficult to separate *P. communis* from *P. purus*. In both species the platform has a crenulated surface (Pl. 5: Figs. 6-8). There are specimens which form a series representing a gradual transition from *P. purus* to *P. communis carina*. In the Cantabrian Mountains *P. communis* occurs in a large interval and in large numbers. On the contrary *P. communis carina* occurs in few samples and in low numbers. Also between these taxa there is a gradual transition. Only specimens with two clear riblets at the anterior side of the platform are considered as *P. communis carina* (Pl. 5: Fig. 8). Due to the long vertical range and the broad horizontal distribution of *P. communis* it is very variable in shape and ornamentation. In this respect it is closely comparable with *P. xylus*. Both species probably lived in the surface layer of the sea. There is only a short interval around the Frasnian-Famennian boundary from which none of the species is mentioned. Both species have a long lived form (*P. communis communis* and *P. xylus xylus*) and a short lived form with serrated anterior platform edges (*P. xylus ensensis* and *P. communis dentatus* Druce, 1969). *P. communis* probably took the niche of *P. xylus* when the latter had become extinct.

Polygnathus dubius Hinde, 1879 (Pl. 3: Fig. 9)
Due to the variability of *P. dubius* it may be difficult to distinguish it from *P. decorosus* Stauffer, 1938, especially in the Cantabrian Mountains were both species may occur within a single sample. Generally the free blade is broken, so that the platforms are comparable. The platform of *P. dubius* is broad, the outer platform wider and rounder than the inner platform. The platform reaches its maximum width half-way or at the posterior half (Pl. 3: Fig. 9). *P. decorosus* has a narrower, more symmetric platform which reaches its maximum width in the anterior half of the platform (Pl. 3: Figs. 7-8). The basal pit in *P. decorosus* is at or near the junction of platform and free blade but in *P. dubius* at about a quarter of the platform length farther towards the posterior end.

According to Klapper & Johnson (1980) the vertical distribution of *P. dubius* would be restricted to *hermanni-cristatus* Zone and Lowermost *asymmetricus* Zone. Klapper & Ziegler (1979) indicate the occurrence of *P. decorosus* in Lower *asymmetricus* Zone to *gigas* Zone. In several samples from the Cantabrian Mountains, however, both species occur together. The occurrence of *P. dubius* with *A. gigas* (amongst others in samples N43 and N46) proves that the species occurs in the Middle *asymmetricus* Zone. This is in accordance with findings by Orchard (1976) who mentions the species from the Lower *asymmetricus* Zone and García Alcalde et al. (1979) who mention the species with *A. gigas*.

Polygnathus inornatus E.R. Branson, 1934, *P. spicatus* E.R. Branson, 1934 and *P. longiposticus* Branson & Mehl, 1934
The classification of these species is rather difficult. The specimens of which the margins are upturned and the right margin is distinctly higher than the left one are considered as *P. inornatus* (Pl. 5: Figs. 10, 12). The specimens without this characteristic may be divided into two groups: one with a broad basal pit which is indistinctly outlined and enclosed in a broad, flat area having a curved carina and blade, the geniculation points being opposite, referred to as *P. spicatus* (Pl. 5: Fig. 11, only known from the Palencian basin) and the other with a large basal cavity and generally a slightly curved carina with a higher denticle at the posterior end, referred to as *P. longiposticus*.

Polygnathus linguiformis Hinde, 1879 (Pl. 3: Figs. 11-16)
This is a very common species in the Portilla Formation, particularly in the lower part (member A). Almost all the specimens belong to the nominal subspecies (Pl. 3: Figs. 11, 12, 16) but some specimens from the Middle *varcus* Subzone are different. Most of the different specimens are juveniles with a straight outer margin with six to eight distinct nodes which merge into rather strong transverse ridges. Some of these juveniles have a high flange-like development of the margin with very sharp saw-tooth-like nodules. Adults of this form were not found and the specimens are referred to as *P. linguiformis* forma A. Most of the juveniles have flatter nodes and lack the high margin (Pl. 3: Fig. 14). They evolve into adults with low margins, almost parallel to the carina which consists of five to ten distinct or coalesced nodes. At the posterior side the outer margin is sharply curved inward. Adult specimens have a distinct tongue with (in some specimens transversely compressed) nodes or (rarely) transverse ridges, this tongue is curved steeply downward (Pl. 3: Fig. 15). These specimens are considered to be *P. linguiformis mucronatus* Wittekindt, 1966.

Polygnathus cf. "*P. norrisi*" Uyeno, 1967 (Pl. 5: Fig. 1)
A few specimens of this "species" were found in the Vidrieros Formation (e.g. sample Ca4 of the *marginifera* Zone, Pl. 5: Fig. 1). Comparable conodonts are known from the Frasnian of Canada, U.S.A. and Morocco (Klapper & Johnson, 1980) and the Famennian of Germany, Austria and Australia (Druce, 1976). Although the structure of the ornamentation is conformable, the shape of the platform and blade is very variable. This indicates that it is a pathological differentiation as was already supposed by Müller (1969). This is very probable since a comparable brush-like ornamentation occurs also in other genera (Müller, 1969; Druce, 1976; M. van den Boogaard, Leiden, pers. comm.). Helms (1959, pl. 5: fig. 8) figures a specimen of *P. diversus* Helms with such an aberrant ornamentation. The occurrence of such aberrant forms appears to be restricted to certain intervals but in my opinion it is not a good choice to base the division of the *dengleri* Zone on the first occurrence of this "species" (Klapper & Johnaon, 1980).

Polygnathus xylus Stauffer, 1940 (Pl. 3: Figs. 5, 6)
This is one of the most common species in the Portilla and the Nocedo Formation. It has a very long vertical range: from below the *varcus* Zone to the *gigas* Zone. This upper limit is deduced from samples in the Crémenes Limestone (2.3.3.4) and is in accordance with the data in Druce (1976, p. 204). Two subspecies are distinguished: *P. xylus xylus* and *P. xylus ensensis* Ziegler, Klapper & Johnson, 1976.

P. xylus ensensis is a little variable form with serrations just posterior of the geniculation points. Only the phylogenetically late form occurs (*varcus* Zone), which, in accordance with the description by Ziegler et al. (1976) has few serrations: one to three at the inner side and none or one at the outer side of the platform (Pl. 3: Fig. 5).

The phylogenetically early specimens of the nominate subspecies (*varcus* Zone) show little variation and are in accordance with the description in Klapper et al. (1970) (Pl. 3: Fig. 6). The late specimens (*asymmetricus* Zone to *gigas* Zone), however, show much more variation. Juvenile specimens agree well with the early forms but the adult specimens may have an irregular platform: wider and rounder at the outer side and curved inward at the anterior third of the platform. The lateral margins may have some nodes which may coalesce into a nearly smooth ridge.

Polygnathus webbi Stauffer, 1938 (Pl. 4: Figs. 19-20)
This is the most common conodont in the faunas from the *asymmetricus* Zone and *gigas* Zone. It shows a great variation, from a strongly ornamented platform with cross ridges or a strong ridge on the lateral margins (Pl. 4: Fig. 20) to a nearly smooth platform (Pl. 4: Fig. 19). Also the width of the platform is variable, but always the anterior platform margin is higher on the right side than on the left. The adcarinal troughs are rather deep at the anterior and shallow at the posterior part of the platform.

Polygnathus spec. A (Pl. 5: Fig. 2)
In the samples from the *costatus* Zone and the *Polygnathus* fauna in the Asturo-Leonesian basin and the Asturian geanticline, conodonts occur which have a pseudokeel at the aboral side, thus belonging to the group of *P. symmetricus* E.R. Branson, 1934, *P. spicatus* E.R. Branson, 1934 and *P. mehli* Thompson, 1967. It was not possible to attribute it to one of these species. It differs from them in having geniculation points which are not opposite, the one at the inner side of the platform declining steeply. The pseudokeel is broader than in *P. spicatus*. At the anterior side the platform narrows. The maximum width of the platform is just behind the middle. Towards the anterior side the platform narrows but the anterior ends are bent outwards. The platform margins are upturned, the outer one more than the inner one (Pl. 5: Fig. 2). The specimens resemble those figured as *P. lacinatus* Huddle, 1934 in Rhodes et al. (1969, pl. 11: figs. 1-7, 11) which occur in much younger deposits (with *P. bischoffi* and *Gnathodus*). Herein I will refer to it as *P. spec.* A.

Polygnathus spec. (Pl. 3: Fig. 10)
One of the specimens of the genus *Polygnathus* has an outer lobe as most of the species of the genus *Palmatolepis* (Pl. 3: Fig. 10). A central node, however, is not present. The shape of the basal pit is characteristic for *Polygnathus*. The specimen has a great overall resemblance

to *P. xylus*.

Pseudopolygnathus marginatus (Branson & Mehl, 1934)
This species occurs in small numbers in samples from the Ermita Formation. It may be distinguished from the other species of the genus by the triangular platform which is strongly rounded at the anterior side but has a posterior side which is tapering and curved downwards. At the anterior side there are two strong ribs, perpendicular to the carina. The rest of the platform is covered by weaker ribs. At the posterior end of the carina there are some higher denticles. The free blade has higher pointed denticles.

Pseudopolygnathus spec. A (Pl. 5: Figs. 2, 3)
This species has a slender, asymmetrical platform which is tapering at the posterior end. The platform margins have a row of nodes, terminating well before the posterior end and at the outer side extending farther towards the anterior end than at the inner side. The outer platform margin is curved upwards. The carina is nearly straight. At the aboral side there is a very large basal cavity which extends even beyond the platform margins. I found this species in two samples of the Vidrieros Formation in the Palencian basin: sample Vil3 and Vil5, both from the *costatus* Zone in the Liébana area. Van Adrichem Boogaert (1967, pl. 3: fig. 7) also found this species in the Palencian basin and illustrated a juvenile specimen, identified as *Pseudopolygnathus fusiformis* Branson & Mehl, 1934.

Siphonodella
Many of the specimens from the upper Tournaisian belonging to this genus are damaged so that the anterior and posterior parts of the platform occur separately in the samples. That obstructs the identification of the specimens. Most specimens belong to *S. obsoleta* Hass, 1959 and *S. isosticha* (Cooper, 1939) or intermediate forms and few specimens belong to *S. cooperi* Hass, 1959. Some of the specimens have an extremely long and narrow platform. The identification of the specimens from the lower Tournaisian deposits is also difficult.

"*Spathognathodus*" *bohlenanus* Helms, 1959 (Pl. 6: Fig. 7)
In the deposits of the *costatus* Zone from the Asturo-Leonesian basin and the Asturian geanticline, conodonts which are in accordance with the specimens illustrated as *Spathognathodus bohlenanus* in Ziegler (1962) are rather common. These occur also, in low numbers, in the Palencian basin. These conodonts differ from the original description in having higher denticles at the posterior side of the element. Sandberg & Ziegler (1979) mention "*Sp.*" *bohlenanus* only from deposits older than the *costatus* Zone. In a specimen from the Vidrieros Formation the denticles, particularly at the posterior part of the unit, bend towards the posterior end. Furthermore the denticles are not separated but touch each other (Pl. 6: Fig. 7). This specimen has a laterally curved platform, turned inwards more sharply in its posterior third. It is referred to as "*Sp.*" spec. A aff. *bohlenanus*.

4.3. CONODONT BIOZONATION

The interval discussed in this paper is divided into conodont zones which may be used all over the world. Not all the zones were recognized because siliciclastic sedimentation was important in Cantabria, and in these sediments conodont faunas are scarce (4.4). Therefore the interval between *gigas* Zone and *costatus* Zone is not recognized in the Asturo-Leonesian basin, and the interval between *Ps. triangularis* Zone and *marginifera* Zone is not recognized in the Palencian basin. In addition the common development of rapidly deposited reef limestones as well as the intervention of hiatuses within the sequence restricts the evidence of the conodont zoning (Fig. 27). Although conodont faunas may provide information for detailed correlations, the smallest units are subzones which lasted about 0.5 to 1 Ma (*). The variations in age, for example in the boundary planes between members, often could not be proved because the differences in age are too small for the solving capacity of conodont biozones. For the Tournaisian deposits a regional zonation is used. In this section I will discuss the conodont zones which were recognized in Cantabria. Fig. 27 gives a correlation of the deposits, based on conodonts.

The *varcus* Zone was subdivided by Ziegler et al. (1976). In the Asturo-Leonesian basin the *varcus* Zone comprises the upper part of the Huergas Formation and the lower part of the Portilla Formation. The Lower *varcus* Subzone is recognized by the joined occurrence of *Eognathodus bipennatus* and *Polygnathus timorensis* or *P. varcus*. Only the upper part of the subzone is present locally in the basal deposits of the Portilla: at Coallajú (Asturias) and in the parautochthonous of the Esla area. In section 24 (Reijers, 1972) north of Cistierna (Esla area) *Eo. bipennatus* occurs with *P. linguiformis mucronatus* which is indicative for the lowermost part of the Middle *varcus* Subzone. The Middle *varcus* Subzone is recognized by the occurrence of *P. ansatus*, *P. linguiformis mucronatus* or *P. linguiformis* subsp. A with *Icriodus obliquimarginatus*, *P. pseudofoliatus* or *P. xylus ensensis*. The subzone is present in the rest of the deposits of member A of the Portilla Formation. Generally the subzone will be easily recognized in the basal part of the unit and with more difficulties in the upper part of the unit since the reef deposits present in that part of the unit contain

(*) = (the Late Devonian probably lasted 10-15 Ma and is divided into 30 subdivisions; Ziegler, 1978)

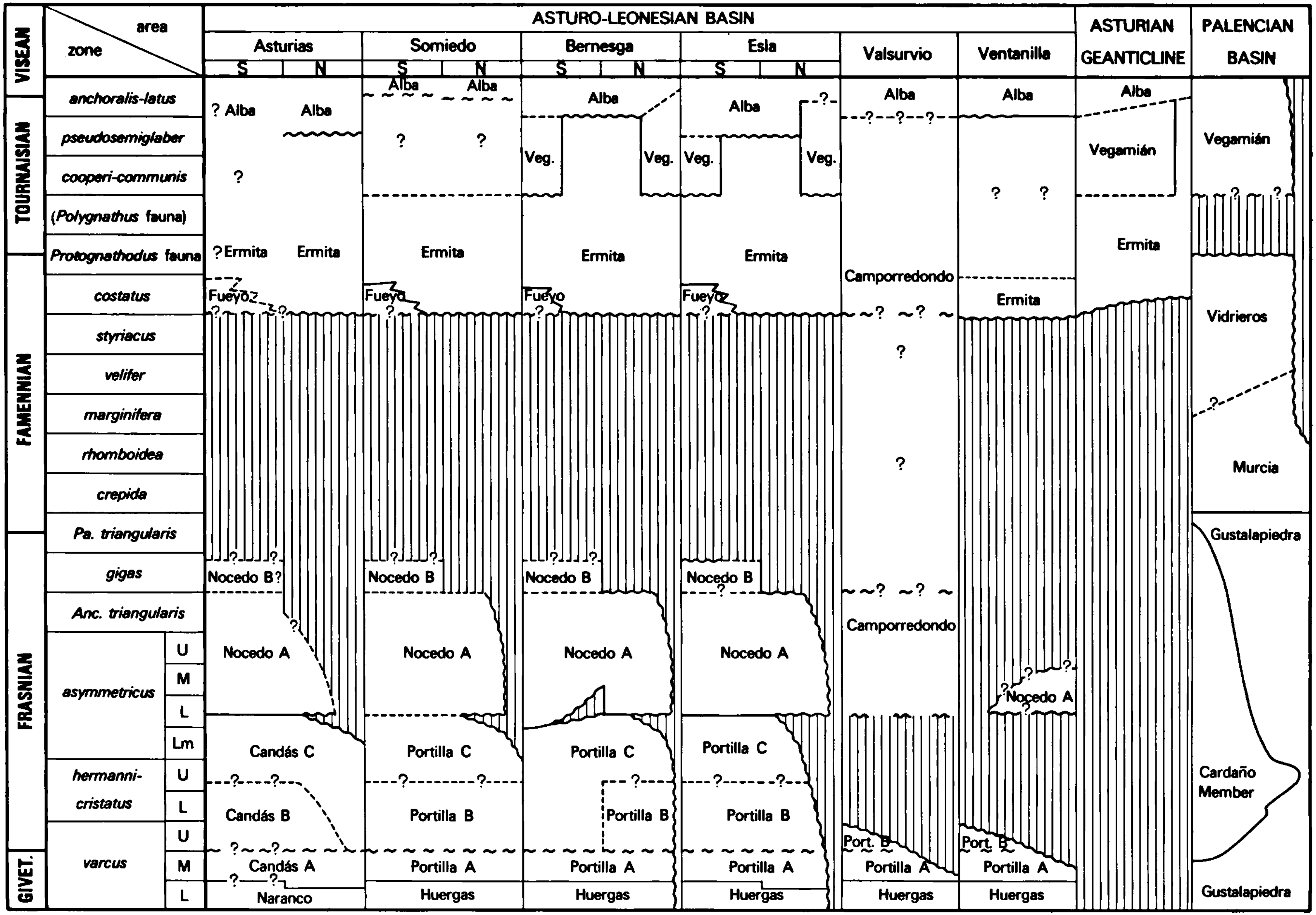

Fig. 27. Biostratigraphic correlation chart for Middle Devonian to Lower Carboniferous deposits in the Cantabrian Mountains. The vertical scale of the three lower zones is exaggerated.

impoverished faunas. The Upper *varcus* Subzone may be recognized by the occurrence of *I. symmetricus* or *P. latifossatus* with *P. varcus* or *Latericriodus latericrescens*. Generally it is not possible to recognize this zone with certainty since siliciclastic deposition was predominant during this subzone, containing faunas with few species, mainly *L. l. lateriscrescens*. That subspecies lived longer and was more common in the U.S.A. than in Germany (Ziegler et al., 1976) thus indicating an intermediary position for Spain. In the Palencian basin the *varcus* Zone was demonstrated already by van Adrichem Boogaert (1967). It was recognized in the middle and upper part of the Gustalapiedra Formation below the base of the Cardaño and in the lower part of the Cardaño Member. Van Adrichem Boogaert's sample CU3 from his section CU I (Gildar-Montó area) contains amongst other *P. linguiformis mucronatus* and *P. xylus ensensis* and thus is attributed to the Middle *varcus* Subzone.

The *hermanni-cristatus* Zone is subdivided into two subzones (Ziegler, 1966, 1971). Since most of the species which characterize this zone are rare or absent from the Asturo-Leonesian basin it was probably not always recognized where it is truly present. The zone is most easily recognized by the joint occurrence of *P. linguiformis* and *P. dubius*. *P. dubius* was only recently mentioned from locations in Europe (Becker et al., 1979; García Alcalde et al., 1979; Morzadec & Weyant, 1982), yet in the Asturo-Leonesian basin it is a common species. In a few sections the zone is represented in slowly deposited fore-reef sediments, e.g. at Huergas de Gordón (Bernesga area); there faunas may be present containing more of the species which prefer deeper water (such as schmidtognathids). The upper part of member B and a large part of member C of the Portilla Formation were deposited during this zone. In the Palencian basin the zone is represented in few samples from the lower part of the Cardaño Member.

The *asymmetricus* Zone is subdivided into four parts (Ziegler, 1971). It is not possible to recognize the Lowermost *asymmetricus* Zone with certainty but the samples containg *P. webbi* without any *Ancyrodella rotundiloba* may belong to this zone. In many of such samples *P. decorosus* is a common species which perhaps existed even before the Lower *asymmetricus* Zone. Some of the deposits of member C of the Portilla Formation and of member A of the Nocedo Formation may have formed during this subzone. The Lower and Middle *asymmetricus* Zone generally could not be distinguished separately: only in samples where *Anc. rotundiloba* occurs with *Anc. gigas* or *Anc. lobata* could the Middle *asymmetricus* Zone be separated. I have recognized the Upper *asymmetricus* Zone in none of my samples but García Alcalde et al. (1979) mention this subzone from Huergas de Gordón. They also mention the *Ancyrognathus triangularis* Zone from the same locality. The *asymmetricus* Zone and *triangularis* Zone were recognized in deposits belonging to member A of the Nocedo Formation.

The *gigas* Zone has not been recognized with certainty but the conodonts and other fossils indicate that the Crémenes Limestone of the Nocedo Formation was deposited during this zone (2.3.3.4).

In the Palencian basin the *asymmetricus* Zone to *Palmatolepis triangularis* Zone are represented within the sediments of the Cardaño Member. The presence of deposits formed during the latter zone is concluded from information in Lobato Astorga (1977) who mentions a sample containing a fauna with *Anc. curvata*, *Pa. gigas*, *Pa. minuta* and *Pa. delicatula*, amongst others.

The oldest fauna from the Vidrieros Formation may be the one found E of Santibañez de Resoba (Cardaño area, sample Ca3): besides *Icriodus* it contains *Palmatolepis glabra pectinata* which is known from Upper *crepida* Zone through Upper *marginifera* Zone (Ziegler, 1977).

Conodont zones of late Famennian and Tournaisian

Many different subdivisions have been made for this interval: of the late Famennian by Ziegler (1962) and of the Tournaisian by Bischoff (1957), Voges (1959) and Meischner (1970) but recently the subdivision of the Tournaisian was replaced by one proposed by Sandberg et al. (1978) and Lane et al. (1980), based on the genera *Siphonodella* and *Gnathodus*, which is followed here. By the means quoted above it is possible to make a very fine subdivision of the total elapsed time, however due to limitations of the sequences and faunas such a detailed subdivision is not possible in Cantabria. *Palmatolepis* (important for the Famennian deposits) occurs nearly only in the Palencian basin, *Siphonodella* is rather rare and most specimens are broken, many faunas contain only species with a very long range and many small hiatuses interrupt the condensed sequences of this time. Higgins (1974) and Higgins & Wagner-Gentis (1982) proposed a subdivision into *costatus* Zone, *Protognathodus* fauna, *Siphonodella cooperi-Polygnathus communis* Zone and *anchoralis* Zone. The revision of the genus *Gnathodus* (Lane et al., 1980) induces a change in this division. I will propose an adapted division, based on samples from the Asturo-Leonesian basin and the Asturian geanticline (Figs. 28 and 29).

costatus Zone: The lower boundary of the *costatus* Zone (Ziegler, 1962) is defined with the first occurrence of *Bispathodus costatus*. This Zone is easily recognized in the whole Cantabrian zone because the species after which it was named is very common. Species which do not occur above this zone are among others "*Spathognathodus*" *bohlenanus*, "*Sp.*" *inornatus*, "*Sp.*" *strigosus*, *Bispathodus ultimus*, *Bi. ziegleri* and *Polygnathus delicatulus*.

Polygnathus fauna: Above the *costatus* Zone generally there is an interval with faunas very poor in species, only the long ranging *Polygnathus communis*, *P. inornatus* and species of *Pseudopolygnathus* and *Bispathodus* being numerous. Samples with this fauna are not very use-

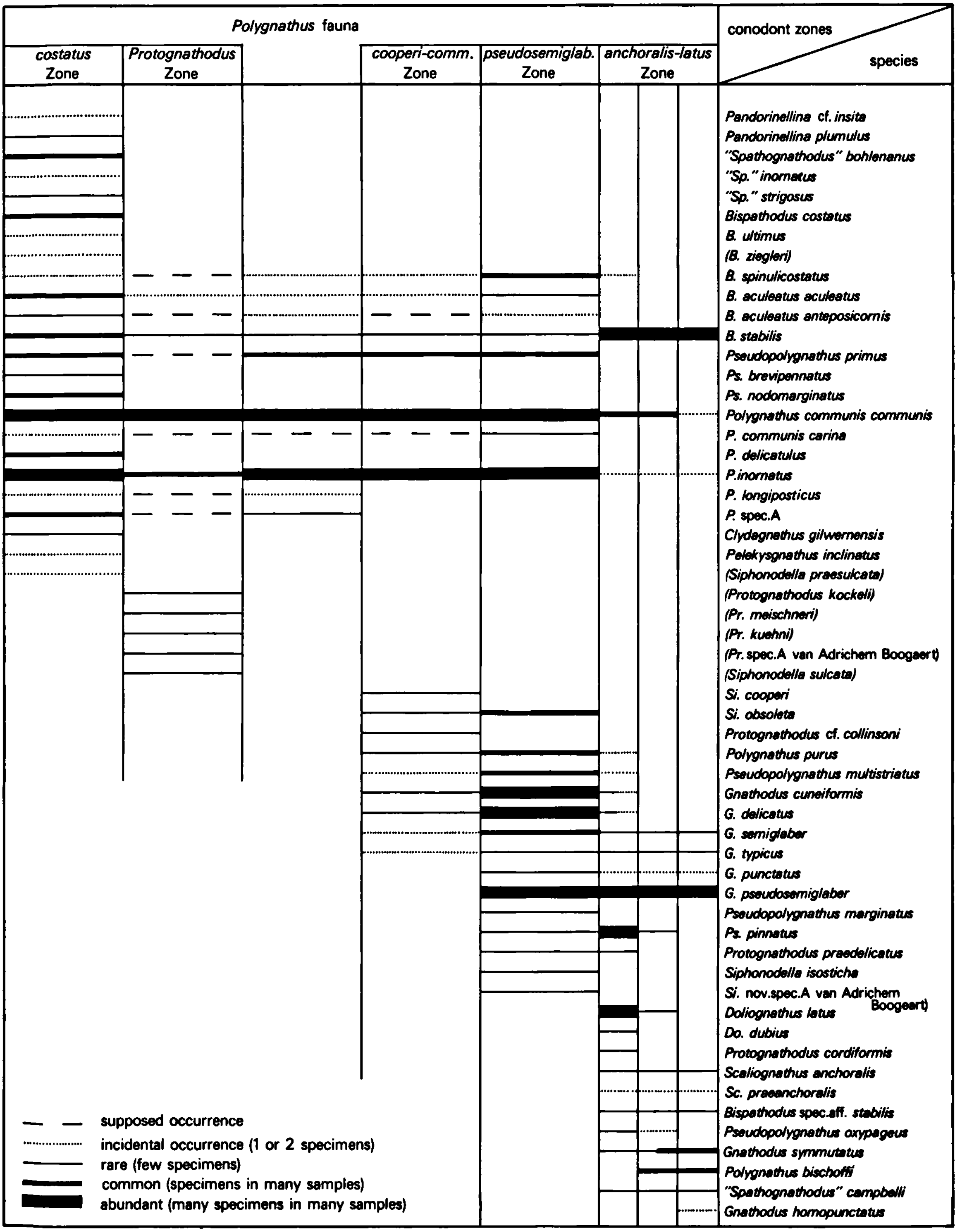

Fig. 28. Vertical distribution of conodont species in the upper Famennian to lower Viséan deposits of the Asturo-Leonesian basin and the Asturian geanticline. Between brackets the species which I did not find during the present investigation but which are mentioned in literature.

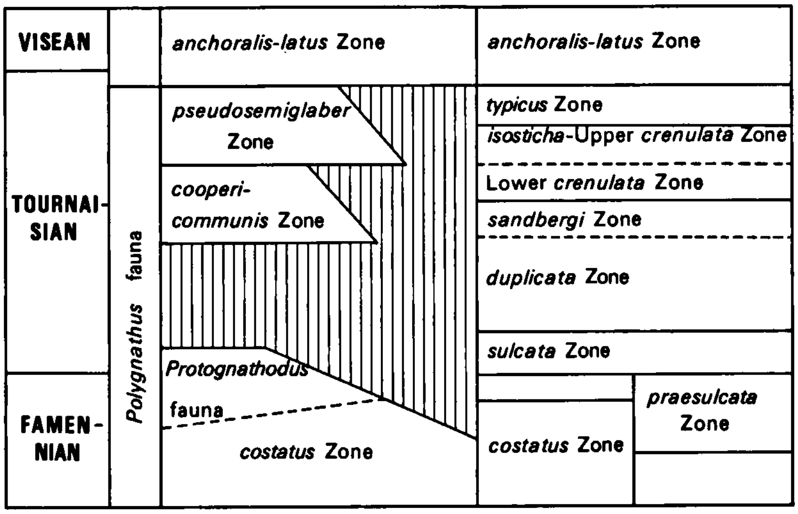

Fig. 29. Local conodont zonation proposed for the Cantabrian Mountains (modified after Higgins, 1971) compared with the standard zones proposed in Sandberg et al. (1978) and Lane et al. (1980).

ful in stratigraphy: they may be impoverished faunas from any one of the zones of late Famennian and Tournaisian. Usually these faunas occur in sequences between levels containing conodonts of the *costatus* Zone and those with conodont faunas of the late Tournaisian. They will be referred to with the informal name *Polygnathus* fauna.

Protognathodus fauna: Few samples containing many specimens of *P. communis* also contain species of the genus *Protognathodus*: *Pr. kockeli*, *Pr. meischneri*, *Pr. kuehni* or *Pr.* spec. A van Adrichem Boogaert, 1967. These are referred to as *Protognathodus* fauna, an informal name proposed by Higgins & Wagner-Gentis (1982). Samples with this fauna are known from Santiago de las Villas, Olleros de Alba (Higgins & Wagner-Gentis, 1982), Triollo (van Adrichem Boogaert, 1967) and Santibañez de Resoba (M. van den Boogaard, Leiden, pers. comm.).

Above this interval there are faunas rich in species, containing many species of *Gnathodus*. In the samples from the Ermita, the Vegamián and the lower part of the Alba Formation three different associations of conodonts may occur. The oldest association contains a fauna still dominated by *P. communis* and *P. inornatus* but accompanied by some species of *Gnathodus* and *Siphonodella*. The middle association is dominated by a large number of different *Gnathodus* species and further contains a large variety of *Polygnathus*, *Siphonodella*, *Pseudopolygnathus* and *Bispathodus*. The youngest association contains few species of *Gnathodus* but is dominated by *Gn. pseudosemiglaber* accompanied among others by *Doliognathus latus* and *Scaliognathus anchoralis*.

cooperi-communis Zone: The older association belongs to the *Siphonodella cooperi-Polygnathus communis* Zone of Higgins (1974). It is dominated by *Polygnathus communis* and *P. inornatus* but characterized by few specimens of *Gnathodus delicatus*, *Gn. cuneiformis*, *Gn. typicus*, *Si. obsoleta* and *Si. cooperi*. The name of the zone proposed by Higgins is unfortunate as *Si. cooperi* is a very rare species and *P. communis* is very common in all the faunas of Famennian and Tournaisian in the Cantabrian Mountains. However, I consider the number of samples attributed with certainty to this zone too low to justify the introduction of a new name. The upper boundary is redefined herein at the first occurrence of *Gn. pseudosemiglaber*. Many of the samples which Higgins placed in this zone most probably belong to the *pseudosemiglaber* Zone: many specimens of *Gnathodus* identified by Higgins as *semiglaber*, *cuneiformis* and *texanus* according to the description in Lane et al. (1980) belong to *Gn. pseudosemiglaber*. This is the case with Higgins, 1962, pl. 3: fig. 30; 1971, pl. 5: fig. 12 and Higgins & Wagner-Gentis, 1982, pl. 34: figs. 4, 6 and 14. There is also a difference of opinion as to the identification of other *Gnathodus* species. Therefore the samples with such doubtful identifications have not been used for construction of the facies maps. The *cooperi-communis* Zone occurs in the Ermita Formation at Pereda (Asturias) (Pello, 1972), Villafeliz and Peña Ubiña (Caldas area) (Frankenfeld, 1981), Aviados, Pola de Gordón (?) (Higgins, 1971) and Cabornera (Bernesga area) and Aguasalio and Santa Olaja de la Varga (Esla area). Further it is known from the Vegamián Formation at Santiago de las Villas (Bernesga area) (Higgins & Wagner-Gentis, 1982), Caldas de Luna and Puerto de la Cubilla (Caldas area) (Frankenfeld, 1981), Mampodre (Isidro area) (van Adrichem Boogaert, 1967) and Covadonga (Coastal Ranges) (Frankenfeld, 1981). Some of the samples attributed to this zone may in fact be impoverished faunas of the *pseudosemiglaber* Zone.

pseudosemiglaber Zone: The middle association contains *Gn. pseudosemiglaber* which is mentioned only from the upper *anchoralis-latus* Zone and younger faunas (Lane et al., 1980). The association in which it occurs, however, is clearly older as is demonstrated by the occurrence of *Siphonodella*, *Ps. marginatus*, *Ps. primus* and *P. purus*, among others, while species like *Doliognathus latus* and *Scaliognathus anchoralis* are lacking. Intermediate forms between *Gn. cuneiformis* and *Gn. pseudosemiglaber* are known only from this association which may indicate that the species *Gn. pseudosemiglaber* evolved within the Cantabrian area. In younger zones it is extremely common, dominating many of the faunas. This middle association is considered as intermediate between the *cooperi-communis* Zone and the *anchoralis-latus* Zone. A new local zone is proposed: the *Gnathodus pseudosemiglaber* Zone. The lower boundary is the first occurrence of *Gn. pseudosemiglaber*, the upper boundary the first occurrence of *Sc. anchoralis* or *Do. latus*. The zone is characterized by rich and varied faunas dominated by various species of the genus *Gnathodus* and containing species of *Polygnathus, Siphonodella, Pseudopolygnathus* and *Bispathodus*. Restricted to this zone are *Ps. marginatus, Siphonodella isosticha* and *Siph.* spec. A van Adrichem Boogaert, 1967. The zone is known from the Ermita Formation at Loredo (del Río & Menéndez Alvarez, 1978) and Entrago (Asturias), Geras, Mirantes de Luna and Nocedo de Bernesga (Bernesga area), Remolina, Aguasalio, Robledo (Esla area) and Las Portillas (Picos de Europa). It is also known from the Vegamián Formation at Puerto de la Cubilla (Caldas area) and Mampodre (Isidro area) and the lower part of the Alba Formation at Entrago (Budinger & Kullmann, 1964; Pello, 1972), Loredo (del Río & Menéndez Alvarez, 1978) and Perlora (Asturias), Aviados (Bernesga area) and Aguasalio, Valdoré and Santa Olaja de la Varga (Esla area).

anchoralis-latus Zone: I follow Lane et al. (1980) in replacing the *anchoralis* Zone by the *anchoralis-latus* Zone. It is known from the lower part of the Alba Formation in the Asturo-Leonesian basin and on the Asturian geanticline and also from the Vegamián Formation at Genicera (Bernesga area) and from the Ermita Formation at Las Portillas and south of Caín (both Picos de Europa). With help of *Ps. pinnatus* and *Pi. bischoffi* the zone may be subdivided into a lower part with abundant *Ps. pinnatus*, a middle part with both species and an upper part with *P. bischoffi* (van der Ark, 1982).

It is always possible that smaller or larger hiatuses may be present between the parts of sections assigned to different zones. In several sections sedimentation seems to have continued uninterrupted during late Famennian and Tournaisian, whereas in others discontinuities are apparent within the Ermita and the Vegamián or at the base of the Vegamián and the Alba Formation (e.g. Encl. 1: Fig. 11).

4.4. CONODONT PALAEOECOLOGY

During the last decade conodont specialists have become increasingly aware of the influence of the environment on the occurrence of conodont species (e.g. Barnes, 1976). Although the scarce knowledge on the mode of living of the conodont animal hampers the drawing of conclusions, some major trends are generally recognized and the preference of genera and species for certain environments has gradually become better known. Probably the discovery of the conodont animal (Briggs et al., 1983) will stimulate further ecological studies. In this section I shall combine data from the literature with data from the present study in order to detect major changes in the environments and for a comparison of the Asturo-Leonesian and Palencian basins.

The samples from the Santa Lucía Formation in the Asturo-Leonesian basin (Huisman, 1981) and of the Abadía Formation in the Palencian basin are dominated by *Icriodus* and simple cones (*Belodella, Coelocerodontus* and *Acodus* s.l.), indicating deposition in very shallow water. In both the Asturo-Leonesian and the Palencian basin comparable shallow water faunas, dominated by *Icriodus* and *Polygnathus*, are present in the deposits formed during the *varcus* Zone. A statistical test (Raven, 1980b) applied to the conodont samples from the Portilla Formation in the Somiedo area (*varcus* Zone to Lower *asymmetricus* Zone) demonstrated the unequal distribution of the conodonts over the different facies (Fig. 30). The largest number of conodonts occurs in the fore-reef environment. In contrast simple cones (*Belodella, Coelocerodontus*) occur mainly in the reef environment and relatively few in the fore-reef and back-reef facies (Fig. 30). *Latericriodus* and *Polygnathus* species with a broad platform (here *Latericriodus latericrescens* and *Polygnathus linguiformis*) are almost completely restricted to the fore-reef environment with a preference for the agitated environment in which crinoidal grainstones were deposited. *Icriodus* and *Polygnathus* species with a narrow platform (the *varcus* group sensu Klapper et al., 1970) occur in small numbers in the back-reef and reef environments and show a preference for the fore-reef packstones. This type of information may be used for the comparison of environments deposited during a certain conodont zone. One must be very careful, however, in applying this information to the comparison of environments of different ages because the relative abundance of the conodont genera depends not only on the environment but may also be influenced by the extinction of species.

Within each of the four described units of the Portilla Formation (2.2) *Icriodus* makes up about 50 % of the conodonts whereas the relative abundance of the other conodonts is rather variable. *Polygnathus* specimens with a broad platform are extremely scarce in members B and C since *P. linguiformis* was already nearly extinct and *P. webbi* evolved later. *Polygnathus* species with a narrow platform become more abundant in the upper part of the formation, mainly due to the newly evolved species *P. dubius* and *P. decososus*. *Latericriodus latericrescens* is abundant in member B but not present in member C because in Cantabria the genus became extinct in the *varcus* Zone.

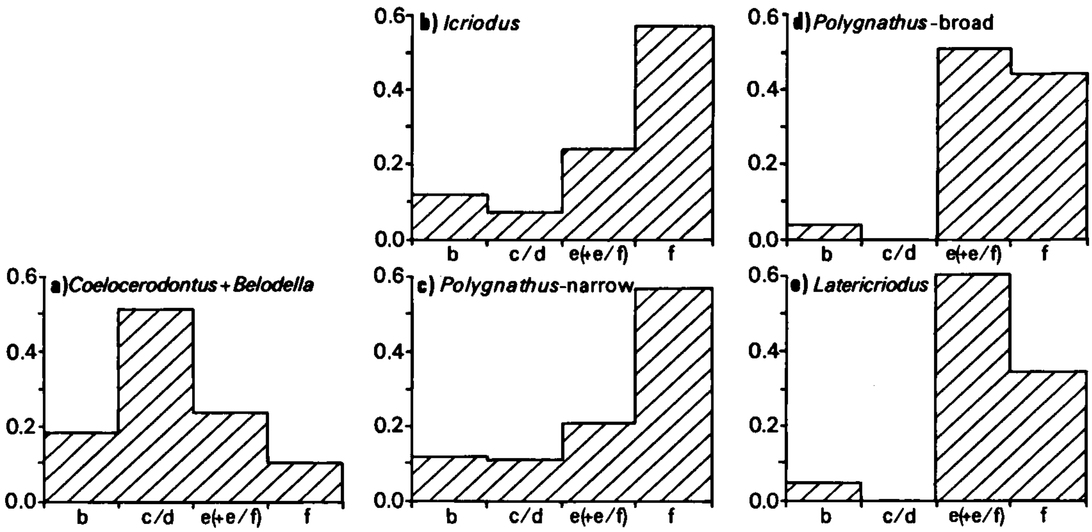

Fig. 30. Relative histograms for the different conodont taxa with relation to the distinguished facies (Portilla Formation, Somiedo area; b = back-reef facies, c/d = reef facies, e = fore-reef grainstones, f = fore-reef packstones and wackestones; based on 57 conodont-bearing samples with 536 platform elements; after Raven, 1980b).

Conodonts are most abundant in the lower part of member A (unit A1) with a mean of 17 P-elements per sample (average weight 2-3 kg) (Table 2). They are much scarcer in unit A2 and in member C, with a mean of 2 to 5 P-elements per sample (Table 2). In member B the occurrence of siliciclastics may have had a negative influence on the occurrence of conodonts. In unit A2 and in member C conodonts are scarce despite the fact that the greater part of the samples were taken from fore-reef deposits. Probably the presence of coral-stromatoporoid reefs also influenced the distribution of conodonts close to the reefs outside the reef belt proper.

During the *hermanni-cristatus* Zone faunas poor in species developed in the siliciclastic and reef areas of the Asturo-Leonesian basin while in the Palencian basin many more species

	Portilla								Nocedo			
	unit A1		unit A2		member B		member C		member A		member B	
	n	%	n	%	n	%	n	%	n	%	n	%
Polygnathus (broad platform)	313	17.5	61	18.2	1	2.0	15	2.8	546	62.5	45	60.0
Polygnathus (narrow platform)	395	22.0	75	20.1	13	25.5	196	36.8				
Latericriodus	75	4.2	24	6.4	12	23.5	-	-	-	-	-	-
Icriodus	951	53.0	187	50.1	19	37.2	267	50.2	226	25.9	28	37.3
simple cones	76	4.2	19	5.1	6	11.8	50	9.3	1	0.1	2	2.6
Eognathodus	11	0.1	-	-	-	-	-	-	-	-	-	-
Ancyrodella	-	-	-	-	-	-	1	0.2	96	11.0	-	-
others	-	-	-	-	-	-	2	0.4	4	0.5	-	-
total	1793	101.0	269	99.9	51	100.0	532	99.7	873	100.0	75	99.9
number of samples	108		76		24		124		60		17	
number of conodonts per sample	16.6		4.9		2.1		4.3		14.6		4.4	
'conodont pearls'	577		269		-		551		-		-	
number of taxa (minimum)	18		16		8		17		17		6	

Table 2. Comparison between the faunas from the members of the Portilla Formation and the Nocedo Formation; n = number of platform elements % = percentage of the fauna for each genus.

occurred, among which *Polygnathus* species with a broad platform such as *cristatus*, *ordinatus*, *pennatus* and *limitaris* (van Adrichem Boogaert, 1967) which are not known from the Asturo-Leonesian basin. These conodonts indicate that the Palencian basin was slightly deeper than the Asturo-Leonesian basin while being separated in some way from that basin.

Polygnathus and *Icriodus* were the main faunal components in the Asturo-Leonesian basin during the *asymmetricus* Zone, with a small proportion of *Ancyrodella* (Table 2). Most samples were taken from the carbonate platform and lagoon deposits (facies a and b in van Loevezijn & Raven, 1983; 2.3.2.1). The differences in depth and turbulence between these environments were so small that more or less the same conodont faunas could develop, only the preservation differs: large and transported conodonts are found in the crinoidal shoal deposits and smaller, better preserved conodonts in the lagoonal deposits. The abundance of the conodonts (with a mean of 15 P-elements per samples) is comparable to that of the fore-reef deposits of unit A1 from the Portilla Formation.

The faunas from the *gigas* Zone, Asturo-Leonesian basin, contain small numbers of species and individuals as is typical for the reef deposits from which the samples were taken (a mean of 4 P-elements per sample, Table 2).

In the Palencian basin the faunas became more abundant in numbers and species and even new genera such as *Palmatolepis* and *Mesotaxis* occur, indicating a greater depth of deposition (Mouravieff & Bouckaert, 1973) than in the Asturo-Leonesian basin. Due to the emergence of the Asturo-Leonesian basin during the main part of the Famennian, conodont faunas are available only from the uppermost Famennian deposits. In the Palencian basin sedimentation continued, the faunas from that basin contain many *Polygnathus* and *Palmatolepis* species and low numbers of *Icriodus* and "*Spathognathodus*", thus indicating deposition on the outer shelf to continental slope (compare Table 3).

A number of models of biofacies for the late Devonian have been proposed: one for the *styriacus* Zone in the U.S.A. (Sandberg, 1976; Sandberg & Ziegler, 1979) and another one for the Belgian Famennian (Dreesen & Thorez, 1980). The faunas from the *costatus* Zone in the Asturo-Leonesian basin consist mainly of *Polygnathus* and *Bispathodus*. Such faunas do not occur in the model of Sandberg but the composition of the fauna is close to those regarded as typical for the continental shelf to continental rise and slope. In Dreesen & Thorez's model, however, the comparable fauna is regarded as typical for a shallow subtidal environment (between mean low water level and the wave-base). Sandberg developed his model on an area where the facies zones and thus the ecological zones are broad while Dreesen & Thorez's model is based on an area with a high siliciclastic influx and with very narrow facies and ecological zones. In deposits from a shallow subtidal crinoidal shoal such as those in the Ermita Formation, Dreesen & Thorez predict a polygnathid-icriodid fauna whereas the deposits of the Ermita contain a polygnathid-bispathodid fauna. This may be explained by the relatively flat sea-bottom in Cantabria during the late Famennian, on which the crinoidal shoals were developed, forming a shallow environment but far from a sediment-supplying hinterland. It can be taken to indicate that the composition of conodont faunas depends not only on depth but also on the distance from the coast and the supply of siliciclastics.

	P. basin		A-L. basin	
	n	%	n	%
Palmatolepis	75	18.2	-	-
Bispathodus stabilis	68	15.1	36	3.9
Bispathodus (other species)	162	39.4	216	23.6
Pseudopolygnathus	16	3.9	62	6.8
Polygnathus communis	47	11.4	182	19.9
Polygnathus (other species)	9	2.2	335	36.6
Icriodus	1	0.2	-	-
Pandorinellina	-	-	5	0.5
Protognathodus	2	0.5	-	-
'*Spathognathodus*'	31	7.5	77	8.4
total	411	98.4	913	99.7
number of samples	10		35	
number of conodonts per sample	41.1		26.1	
number of taxa (minimum)	21		22	

Table 3. Comparison between the faunas from the *costatus* Zone in the Palencian basin and the Asturo-Leonesian basin; n = number of platform elements, % = percentage of the fauna. Ten samples from the Vidrieros Formation (Palencian basin) were counted (Ca6, Vi2, Vi4, Vi10, Vi13, Vi15, Vi17, Vi18, Vi21 and Vi22) and 35 samples from the Ermita Formation (Asturo-Leonesian basin) (E5, E8, E11, E13, E14, E19, E20, E22, E25, E26, E30, E32, E33, E38, E39, E40, N52, N58, N69, N79, OS1, OS2, U0, U1a, RI8, RI10, PE1a, PE1b, PE2, SAL1, IIf221 and AG4).

In the late Famennian (*costatus* Zone) deposits of both basins, "*Spathognathodus*" is common and not indicative for a particular environment (Table 3). *Polygnathus* occurs in much higher numbers in the Asturo-Leonesian basin and on the Asturian geanticline. *Polygnathus communis communis* occurs in all the environments but is more numerous in the shallower parts, and *P. communis carina* does not occur at all in the Palencian basin. *Bispathodus* is common in all the areas but much more numerous in the Palencian basin. *Palmatolepis* occurs only in the Palencian basin and in one sample from the southern extreme of the Asturo-Leonesian basin: in the Alba syncline beyond the carbonate platform. Low numbers of *Pandorinellina, Clydagnathus, Pelekysgnathus* and *Icriodus* occur in some samples. Generally these genera occur in shallow water. Few specimens of these genera occur also in samples with *Palmatolepis*, thus indicating that there the species lived in the surface layer of the water (photic zone) with *Palmatolepis* deeper below. It is also possible that some of these conodonts were transported into deeper water than that in which they lived. On the Asturian geanticline the faunas consist mainly of *Polygnathus communis* (about 50 %) and other polygnathids (30 %). In some of these faunas (from La Uña, Caín) *Pelekysgnathus, Clydagnathus, Pandorinellina* cf. *insita, Pand. plumulus* and *Bispathodus aculeatus anteposicornis* occur which are unique to this area and a few sampled localities in the innermost zone of the Asturo-Leonesian basin (Villanueva de la Tercia, Rodillazo, Valdoré) and the border of the Palencian basin (Fig. 16). These and lithological data indicate that after the late Famennian transgression the Asturian geanticline was still the shallowest area within the Cantabrian zone.

Faunas which may represent deepter-water conditions are known only from Santiago de las Villas, Olleros de Alba (both southeastern part of the Alba syncline) (Higgins & Wagner-Gentis, 1982), Triollo and Santibañez de Resoba (both Cardaño area, Palencian basin) (van Adrichem Boogaert, 1967; M. van den Boogaard, Leiden, pers. comm.). These have been described by Higgins & Wagner-Gentis (1982) as *Protognathodus* fauna and are characterized by the occurrence of *Protognathodus* species with *Polygnathus, Bispathodus* and *Pseudopolygnathus*. It may be present in the uppermost part of the Vidrieros Formation or the Ermita Formation.

Many samples contain few species, comprising *Polygnathus* with a few *Pseudopolygnathus* and *Bispathodus*, the conodonts showing signs of transport. All the species preserved in this *Polygnathus* fauna (4.3) have a long vertical range. I suppose that it lived in very shallow water, during the late Famennian and the Tournaisian.

	co-co. Zo.		*pseud.* Zo.	
	n	%	n	%
Siphonodella	4	2.7	80	5.7
Gnathodus	16	11.0	615	43.9
Bispathodus	14	9.6	47	3.3
Pseudopolygnathus	10	6.8	68	4.8
Polygnathus communis and *purus*	76	52.0	424	30.3
Polygnathus	24	16.4	166	11.8
Protognathodus	2	1.4	1	0.1
total	146	98.9	1401	99.9
number of samples	4		9	
number of conodonts per sample	36.5		155.7	
number of taxa (minumum)	16		23	

Table 4. Comparison between the faunas from the *cooperi-communis* Zone and the *pseudosemiglaber* Zone; n = number of platform elements, % = percentage of the fauna. Four samples from Ermita and Vegamián were counted for the left-hand column (AG5, SOo, IIf223 and 64MA3), 9 samples from Ermita and Vegamián were counted for the right-hand column (AG2, LPA4a, MIR6, NOC1, NO-C, PC1, REM2, V9, SJ/H6).

In the *cooperi-communis* Zone the variety of species increases with *Gnathodus* and *Siphonodella* species occurring in the faunas (Table 4). There genera lived in rather deep water, indicating an increase in depth in the Asturo-Leonesian basin and on the eastern side of the Asturian geanticline, where these faunas have been found.

The slightly younger *pseudosemiglaber* Zone is represented by varied faunas containing not only many species but also large numbers of individuals of each species and large conodonts too (Table 4). Many of these conodonts are species with broad platforms. The genus *Gnathodus* in particular is very abundant but *Polygnathus, Pseudopolygnathus, Siphonodella* and *Bispathodus* are also important parts of the fauna. This fauna lived in deeper water than those from the previous zones, presumably below wave-base. It is found in the Ermita, the Vegamián and the Alba Formation.

The *anchoralis-latus* Zone is represented by equally abundant faunas containing *Gnathodus, Pseudopolygnathus, Polygnathus, Bispathodus, Scaliognathus, Doliognathus* and others. All these faunas are considered to have lived in rather deep water, at least below wave-base. *Scaliognathus* and *Doliognathus* are regarded to have lived farther seaward than *Gnathodus* and *Pseudopolygnathus* but their occurrence also extended landward onto the carbonate platform (Lane et al., 1980). Only in the southernmost part of the Asturo-Leonesian basin do these genera make up a distinct part of the fauna (Fig. 18), thus indicating that there the greatest depth was reached of the further very shallow Cantabrian zone.

Adaptive radiation occurred within certain genera during the late Famennian (*costatus* Zone; in the genus *Bispathodus*) and the late Tournaisian to early Viséan (*cooperi-communis* Zone to *anchoralis-latus* Zone: in the genera *Gnathodus, Doliognathus, Scaliognathus*), not only in Cantabria but on a global scale. These intervals coincide with the major transgressions in Cantabria, which were probably caused by eustatic sea level rises. Fåhraeus (1976) concluded that adaptive radiation of conodontophorids occurred during the increase of shelf areas (transgressions) but the sequences in the area studied by me are too incomplete to draw definite conclusions.

Since conodonts have a high specific gravity (apatite: S.G. 3.1 - 3.2) they behave as heavy minerals during transportation. In addition their various shapes may lead to hydrodynamic sorting. The thicker and larger elements tend to withstand transportation and often are concentrated in lag deposits. The thinner and smaller blades and bars are transported more easily and may get lost by breaking. Therefore in high-energy shallow-water environments such as the Asturo-Leonesian basin usually the thick and large elements prevail while the thinner and smaller elements form a minority although there are more of the latter in conodont apparatuses. In a low-energy environment such as the Palencian basin both types of elements are found together, both being rather well preserved.

Heavy minerals are very scarce in the well-sorted sands of the Nocedo Formation and consequently it is unlikely that many conodonts occur in these sands. In some of the limestones heavy minerals are abundant and then it is often found that conodonts also occur in large quantities, for example near the top of the Ermita Formation. One would expect that in such a case many of the conodonts would be reworked but although the conodonts may have been transported generally there are no indications that vertical mixing occurred. However, in the abundant faunas from the *pseudosemiglaber* Zone we find species whose known vertical ranges do not overlap. The composition of the faunas from the *pseudosemiglaber* Zone is very constant in the entire Asturo-Leonesian basin and on the geanticline so that it seems most probable that these species lived together. Hence the vertical distribution of several species, which has admittedly been poorly defined up to now, must be extended (Fig. 28). This conclusion is further supported by the occurrence of some of the species even in the faunas from the *anchoralis-latus* Zone. A factor of uncertainty in these samples is caused by the nature of the rocks: the deposits of the *pseudosemiglaber* Zone and *anchoralis-latus* Zone form condensed sequences so that in samples taken at the boundary of the zones faunas may be mixed but such samples are very rare.

In many conodont samples from limestones of the Portilla and the Nocedo small apatite spheres (with a mean diameter of about 0.3 mm) occur. Such spheres have been attributed to several groups of organisms; Leuteritz et al. (1972), however, regard them as inorganic precipitates. Glenister et al. (1976) consider them to be conodont pearls. The spheres are not known from conodont samples of the Santa Lucía Formation, the Fueyo, Ermita, Vegamián, Alba or any of the deposits in the Palencian basin. Thus the spheres appear to be restricted to certain shallow water environments. In molluscs pearls form when mantle epithelium comes into the connective tissue of the mantle, which may be caused by alien particles such as sand grains or by parasites. The first is more probable to occur in shallow, turbulent water. In the Portilla Formation the spheres occur in all the units except member B (Table 2), indicating that prevailing siliciclastic sedimentation does not favour the formation of the spheres. In members A and C the spheres are common in the back-reef (b) and reef (c), as well as in the fore-reef facies (e/f). Only in the biohermal deposits (facies d) the spheres occur less frequently, as do conodonts, probably because both were winnowed from the boundstones. The spheres occur in grainstones, wackestones and boundstones but most frequently in packstones. The number of spheres per sample is very variable: in most of the samples only few spheres occur (0 to 10) but in some the spheres are very abundant (50 to 350). In the samples with the largest numbers of spheres, conodonts often are scarce or absent. This may be due to mechanical sorting or may indicate that the spheres were not formed by the conodont animal. The colour of the spheres is variable: even within one sample, milk white, pale yellow, light brown, dark brown and reddish spheres may occur although the conodonts in the sample have only one colour. In some samples the spheres are so translucent that the nucleus is visible from the outside. It is formed by silt-size sediment particles which occur also separately in the heavy fraction of the dissolved limestone samples.

5. SYNTHESIS

5.1. THE CAUSE OF THE CYCLICITY IN THE DEVONIAN DEPOSITS OF THE ASTURO-LEONESIAN BASIN

From the description of the Devonian deposits it is apparent that large sedimentary cycles are present in the Devonian of the Asturo-Leonesian basin. Such cycles were recorded from the Santa Lucía Formation, the Portilla Formation (Reijers, 1980) and the Nocedo Formation (Raven, 1980a; van Loevezijn & Raven, 1983). Reijers (1980) has recognized a regressive sequence followed by a transgressive sequence in both the Santa Lucía and the Portilla Formations. Because my correlations based on biostratigraphical data differ essentially from those accepted by Reijers the vertical sequences have had to be re-interpreted. Within the Portilla and the Nocedo the major cycles of sedimentation are made up of regressive sequences of shale, sandstone and limestone which formed due to progradation of the coast. Comparable sequences are present in the older Santa Lucía Formation. Each cycle has its own characteristics and within each cycle lateral variations occur. Nevertheless they are all of more or less the same magnitude, comprising three to four conodont subzones (Fig. 27). This indicates that the alternation of biostromal limestones and siliciclastics was a process which occurred repeatedly. Frankenfeld (1981), however, explains the disappearance of the Portilla reefs with unique circumstances: the reefs disappeared due to an extension of the "Asturisches Sandfeld" and could not re-appear due to increased water depth. This increased water depth would be due to a major Frasnian transgression. Evidence for the major Frasnian transgression was obtained from erroneous correlations (Frankenfeld, 1981, correlates amongst others the Crêmenes Limestone with the Portilla Formation). Further, although a depth of many tens of meters would be required to obstruct coral growth there are no indications for depths of more than a few tens of meters at most. Moreover, the re-appearance of biostromal limestones in members A and B of the Nocedo Formation (Frasnian) proves that the environment had not changed as much as Frankenfeld supposed. The reefs of the upper part of the Portilla Formation disappeared due to the supply of siliciclastics, just as had happened with the reefs in member A of the Portilla. It is important to know what caused the cyclicity.

In case of continuous uplifting of the Asturian geanticline a eustatic sea-level rise would result in a reduction of the source area, erosion would decrease and limestone deposition would prevail. After termination of the sea-level rise gradually more of the source area would emerge, erosion would increase again and lead to renewed siliciclastic supply and progradation of the coast. In such eustatically controlled cycles the limestone would be formed after a transgression and thus the major boundary between the cycles would be that between the sandstone and the limestone. In the Portilla and the Nocedo, however, this boundary never is a break in the sedimentation: the boundary is between the limestone and the shale.

Intermittent uplifting of the Asturian geanticline, however, may explain the origin of the cycles. After each major uplift shales were deposited in the subsided marginal areas, sands in the intermediate zones while erosion would prevail in the uplifted central area (Fig. 31). When erosion decreased a carbonate platform could form on the shallow sandy shelf.

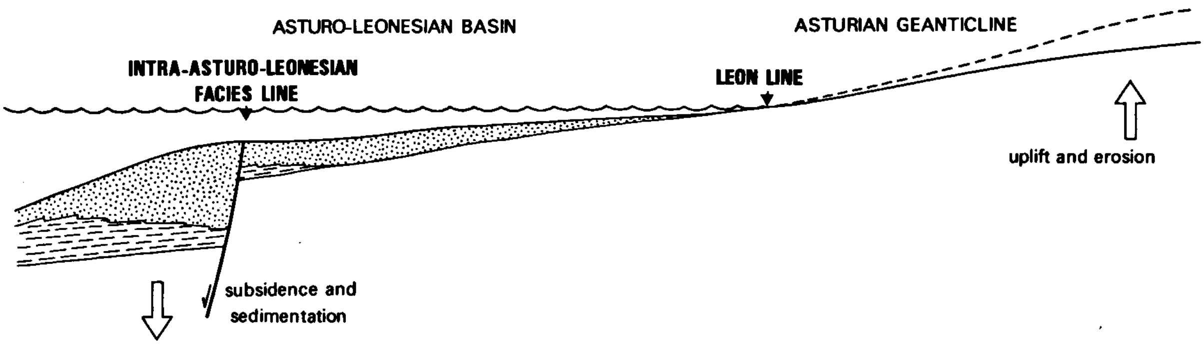

Fig. 31. Epeirogenetic movements caused the cyclicity in the Portilla and the Nocedo: uplifting of the Asturian geanticline resulted in erosion and the deposition of a coarsening upward sequence in the Asturo-Leonesian basin.

A new uplift would start a new cycle. Since the Asturo-Leonesian basin was a narrow inner shelf bordering the Asturian geanticline these cycles formed simultaneously in the entire basin. Local structures such as the Intra-Asturo-Leonesian facies line, the Pardomino high and others only slightly affected the cycles. Slight epeirogenetic movements could have brought about local repetitions within the sequences, e.g. member A of the Portilla Formation in the parautochthonous of Valdoré (Esla area; Encl. 1: Fig. 3). The result of this process is deep erosion on the geanticline and rather complete sequences in the marginal seas.

Although sea-level changes did not determine the cyclicity itself, the height of the sea level controlled sedimentation: during a high sea-level stand a larger part of the Asturian geanticline would be flooded so that erosion would be weaker and carbonates would be deposited, while a low sea level would lead to deposition of predominant siliciclastics. Such changes may explain the predominant carbonate deposition during the late Givetian and early Frasnian (Portilla Formation) and the prevailing siliciclastics in the Frasnian (Nocedo Formation). This is evidenced by the well-dated base of the Portilla Formation: in the Middle *varcus* Subzone, when sea level was at a high stand, reef growth occurred in shallow, tropical seas all over the world (Klapper & Johnson, 1980).

5.2. SOME IMPLICATIONS FOR THE STRUCTURAL GEOLOGY

5.2.1. The influence of faults on sedimentation

In the Bernesga area the Sabero-Gordón line actually coincides with the Intra-Asturo-Leonesian facies line but farther west they deviate from each other. Near Aviados the facies line and the Sabero-Gordón line pass through the narrow anticline north of the Alba syncline, just as elsewhere in the Bernesga area. East of the Bernesga area the facies line separates the Esla nappe from the parautochthonous of Valdoré and the Valsurvio area from the Ventanilla area (Fig. 1). Towards the west the Sabero-Gordón line appears to merge into the Villablino fault which may be traced far into the West Asturian-Leonese zone (Julivert, 1981:

fig. 2) as does the León line (Savage, 1979). The facies line, however, follows the Asturian arc. From the Abelgas syncline (van den Bosch, 1969) onwards it may be traced through the middle of the Palomas syncline (Somiedo area), passing west of the Saliencia syncline, through the middle of the Villazón syncline towards the northeast where it is cut by the Cardaño line. In the Candás area the facies line passes through the syncline in between of Luanco and Piñeres, herein called Luanco syncline (Fig. 1). Thus in the west the facies line passes through synclines but where it coincides with the Sabero-Gordón line it passes through a narrow anticline. The course of the line was deduced from differences in thickness and facies of the Devonian deposits on each side of it. Particularly for the siliciclastic Upper Devonian there is a substantial difference in thickness: e.g. in the Villazón syncline the thickness is 800 m at Salas in the western limb (Pello, 1972) and only 250 m at Cornellana in the eastern limb (Pello, 1968); in the Luanco syncline member A of the Nocedo Formation measures 242 m at the northwestern limb at Luanco and only 76 m in the southeastern limb at Pineres (Radig, 1961). Large differences in thickness occur also in the Candás Formation, e.g. west of the facies line 665 m thick at Luanco (García López, 1976) and east of it less than 300 m. Julivert et al. (1973) mention 500-600 m thick deposits of shales and limestones with a scarce fauna in the extreme west and 200 m thick reefal limestones in the east of the area southwest of Avilés (compare Pl. 1: Fig. 6 with Pl. 1: Figs. 1-4 and Pl. 2: Fig. 7).

The Intra-Asturo-Leonesian facies line already influenced the deposition of the late Silurian-early Devonian San Pedro Formation (Krans, 1982) and maintained its influence during Devonian and Carboniferous: e.g. during the Namurian a carbonate platform was present north of the line and south of it entomozoan shales were deposited in a deeper water environment (Heward & Reading, 1980). Also during the Devonian the facies line separated the inner area from the outer area forming a step in the shelf along which reefs were present when carbonate deposition prevailed (e.g. Portilla Formation, 2.2, Reijers, 1980) and coastal barriers when siliciclastic deposition was superior (e.g. San Pedro Formation, Krans, 1982; Nocedo Formation, 2.3; van Loevezijn & Raven, 1983: fig. 6). The combination of a high rate of subsidence and rapid sedimentation in the outer area resulted in the formation of clastic wedges just south of the facies line. These might be shallow-water siliciclastics as in member A of the Nocedo Formation at Luanco and Huergas de Gordón (2.3.2, Fig. 10) or finer grained deeper water siliciclastics as in the Fueyo Formation in the Bernesga area (2.4.2, Fig. 15). North of the facies line thinner deposits were formed in shallower environments, e.g. lagoonal deposits in member A of the Nocedo Formation (van Loevezijn & Raven, 1983: fig. 6b). The differences in thickness may be very large, also in sections which were very close. Only an active normal fault separating the two blocks may explain all these features, just as Krans (1982) described for the late Silurian and early Devonian and as Rupke (1965) and Al-Mehaidi (1972) described for the late Devonian in the Esla area. Subsidence and uplifting might vary along the fault as is demonstrated by the presence of fan-deltas with conglomerates along the fault escarpment in the Bernesga area while reefs were present in the Esla area (late Frasnian, member B of the Nocedo Formation; van Loevezijn & Raven, 1983: fig. 6e). If the facies line was formed by a normal fault the actual curvation of the facies line in Asturias must be of secondary origin. Indeed Ries et al. (1980) found palaeomagnetic evidence that a substantial part of the curvation of the Asturian arc is of secondary origin and originated by folding during the Late Carboniferous.

Sedimentation during Devonian and Carboniferous was also influenced by other faults. Particularly the distribution of Tournaisian deposits (Figs. 22-24) indicates that faults separated the Bernesga area from the Esla and Somiedo areas and subdivided the Bernesga area in a northern and a southern block. In the southern block the thickness of the Tournaisian increases rapidly away from the fault. In the northern block thick deposits of a deeper facies are present in the northeast, thin deposits of a shallow facies in the south, and deposits are lacking in the southwestern part of the block, thus indicating a tilting of the block towards the northeast (Figs. 22-24).

Parallel to the east-west striking part of the Intra-Asturo-Leonesian facies line wrench faulting occurred during the Stephanian. A sinistral displacement along the Sabero-Gordón line was proposed by Al-Mehaidi (1972) and Bastida et al. (1976). The latter authors estimated the displacement at 15 to 20 km. Also along the León line a wrench fault developed. Probably these features are all due to the same fundamental structure in the basement.

5.2.2. The Esla nappe and the horizontal displacement along the Sabero-Gordón line

The Esla nappe was first recognized by de Sitter (1959) and later studied in more detail by Rupke (1965), Bastida et al. (1976), Raven (1980a) and Arboleya (1981). These authors agree that the Esla nappe was originally south of the Sabero-Gordón line and later shifted towards the NNE. This is confirmed by a comparison of the Upper Devonian in the Esla and Bernesga areas. In the Bernesga area the maximal subsidence was reached in the area south of the facies line which was parallel to but south of the Sabero-Gordón line (see above and 1.4). Isopachs and isogrades of sand-shale ratio run approximately parallel to the facies line. Towards the north the strata become thinner or disappear, partly due to stronger erosion during the late Devonian. North of the facies line the Ermita Formation is 5 to 20 m thick and south of it at least some tens of meters. Member B of the Nocedo Formation and the Fueyo Formation do not occur north of the facies line. Although the sediment supply from the Pardomino high slightly altered the pattern of isopachs and isogrades I imagine that the same configuration as sketched above for the area west of the Porma fault was present in the Esla area too. That would imply that the facies line, and thus also the Esla nappe, were originally south of the Sabero-Gordón line. Since the facies line was formed by an active fault (see above) it is most probable that in the Esla area the line was at more or less the same distance from the Sabero-Gordón line as in the Bernesga area.

When the different units of the nappe (Aguasalio syncline, Felechas syncline and Peña Corada unit) are shifted towards the south the relative position of these units still is not in accordance with the various facies in each of the units. In particular the isopach map for member A of the Nocedo Formation (Fig. 9) indicates that additional palinspastic correction is necessary. Bastida et al. (1976) offered an acceptable solution: that, after the northward overthrusting displacement of the Esla nappe, a horizontal displacement along the Sabero-Gordón line shifted the Peña Corada unit towards the east. They gave some arguments based on the continuity of geological structures and estimated the displacement at 15 to 20 km. In this reconstruction the Esla nappe is a broad (about 40 km) nappe with the Peña Corada unit in the west and the Aguasalio syncline in the east. The eastern part of it was displaced over minimally 14 km (Arboleya, 1981) but for the western part (which actually is still south of the Sabero-Gordón line) the displacement may have been less. All the lithological data are in accordance with such a reconstruction. When the Peña Corada unit was displaced towards the east, of course also the other parts of the South Cantabrian block (Krans, 1982), the Alba syncline and the southern limb of the Abelgas syncline, must have been displaced. Thus, in a palinspastic reconstruction the southern limb of the Abelgas syncline is very close to the Somiedo area (Figs. 4-6) which would explain the great similarity between the sections of the Portilla Formation in both areas (Encl. 1: Fig. 5).

5.2.3. The Valsurvio and Ventanilla areas

The Valsurvio dome and the San Martín thrust sheet (Ventanilla area) are situated east of the Esla nappe (Fig. 1) which raises the question whether these areas are in autochthonous or allochthonous position. The deposits are not very different from those in the Esla area but folding lead to the dome-shape of the Valsurvio area and metamorphism has given the sediments a different appearance. For the palinspastic reconstruction the facies distribution in several formations may be an important argument. In the Valsurvio area the Upper Devonian deposits are very thick. Koopmans (1962) estimated the thickness at about 400 m. Witte (1980) measured a section of 1010 m thick but that extraordinary thickness must be due to faults. It is not possible to give a more precise estimate of the thickness because the sequence is disturbed by faults and because it was not possible to distinguish the lithological units of the Upper Devonian which were recognized in other areas. But certainly the thickness of these deposits is considerable which may be taken to indicate that they were deposited south of the Intra-Asturo-Leonesian facies line and thus probably south of the Sabero-Gordón line. In an exposure of the Camporredondo Formation farther east (southeast of Ventanilla) the thickness was estimated at 150 m (N. Schelling, Leiden, pers. comm.) and therefore also this locality is supposed to have been situated south of the facies line. In the San Martín thrust sheet between Ventanilla dn Rebanal de las Llantas erosion occurred before deposition of the Ermita Formation, through which only locally a thin shale of the Nocedo Formation was preserved. Thus the Ventanilla area closely resembles the parautochthonous of Valdoré (Esla area; 2.2.2). The stratigraphical data indicate that the Ventanilla area (the San Martín thrust sheet) is in a parautochthonous position and the Valsurvio area (the Valsurvio dome) in an allochthonous position. Since arguments against or in favour of this theory may be obtained from the structures, detailed structural research is recommended.

5.3. GEOLOGICAL HISTORY

In the south and west the Asturo-Leonesian inner shelf was bounded by the Intra-Asturo-Leonesian facies line and passed into the deeper shelf of the West Asturian-Leonese zone. At the other side the Asturo-Leonesian shelf generally passed into the Asturian geanticline. The easternmost part of the shelf, however, passed into the outer shelf (and continental slope?) of the Palencian basin which may have been connected to the Pyrenean trough (1.4).

In the Asturo-Leonesian basin during early Givetian time siliciclastic sedimentation was predominant. Later during the Givetian and Frasnian, carbonate platforms occasionally developed on the shallow inner shelf. The carbonate deposition was repeatedly interrupted by phases of siliciclastic sedimentation when the Asturian geanticline emerged and was being eroded. Longshore drift towards the west generally distributed the sand all along the facies line, resulting in the formation of coastal barriers with protected lagoons north of it. Due to the continued uplift during the late Frasnian the Asturian geanticline extended so far that the inner zone of the Asturo-Leonesian basin was uplifted and peneplained; locally along the facies line fan-deltas developed. Elsewhere on the shelf deposition was predominantly of clay. In the Palencian basin sediment supply was much lower: during the early Givetian clay deposition prevailed but when carbonate deposition started in the Asturo-Leonesian basin, in the Palencian basin limestones were deposited. During the early Famennian the Asturian geanticline extended so far that the entire Asturo-Leonesian basin was emergent: there nothing was laid down until the late Famennian. At the other hand, in the Palencian basin the eroded siliciclastics were transported by turbidity currents and deposited, thus interrupting the sedimentation of nodular limestones. There are no indications for a change of depth in that basin. Gradually the erosion on the geanticline and in the Asturo-Leonesian basin decreased and again nodular limestones were deposited in the Palencian basin.

Due to an important transgression during the late Famennian once again the sea spread over the peneplained areas of Asturo-Leonesian basin and Asturian geanticline. In the major part of the area a thin deposit of sandstone and limestone was formed; in the deeper, more rapidly subsiding area of the outer area of the Asturo-Leonesian basin thicker sandstones and shales with storm beds were deposited. The coast prograded gradually and during the early Tournaisian large areas probably emerged once again. Epeirogenetic movements created a new relief with a low on the eastern side of the Asturian geanticline and in the northeast of the Bernesga area and with stable higher areas: the Esla and Asturian platforms and a narrow strip along the facies line in the Bernesga area. Thus the reversal of topography started, which is a general characteristic for the Lower Carboniferous in the Variscides (Brouwer, 1978). In the West Asturian-Leonese zone there was a first phase of deformation during the same interval (Dvořak et al., 1977).

After a new transgression during the late Tournaisian coastal upwelling of cold nutrient-rich water south of the Cantabrian zone caused a thermocline to be established, so that water stagnated in the southeast of the Alba syncline as well as in the deeper parts of the platform and the Palencian basin. In these basins black shales with phosphate nodules and radiolarian cherts were deposited. There were probably many small isolated basins judging from the discontinuous nature of the outcrop of the Vegamián and Ermita but exact details of their geometry and constitution can not be established. Due to epeirogenetic movements deposition was interrupted repeatedly by erosion. Since the Asturo-Leonesian basin and the Asturian geanticline still formed a shallow area surrounded by deeper seas, condensed sequences were deposited. With the gradual rise of sea level most of the area came below wave-base so that nodular limestones were deposited in the shallower, well-oxygenated areas. Gradually circulation increased and one by one also in the deeper areas deposition of nodular limestones commenced.

Sedimentation in the Cantabrian zone was greatly influenced by epeirogenetic movements and synsedimentary faults. The Intra-Asturo-Leonesian facies line was formed by a synsedimentary normal fault which during the Devonian and Carboniferous lead to differences in thickness and facies between the areas at both sides of it but differences also occurred laterally along the facies line. The Pardomino high with the Porma fault and the Somiedo high with the León line marked the boundaries between the three parts of the inner shelf of which the Bernesga block was the most mobile part.

The nature of the sediments appear to have been profoundly influenced by the nature of the circulation of the sea water along a coast as well as the depths of water maintained during sedimentation. In the Asturo-Leonesian basin there was an important change from the Givetian and Frasnian coral-stromatoporoid reefs, generally occurring at the eastern side of continents (Schuhmacher, 1976), to upwelling of cold water during the Early Carboniferous, generally occurring at the western side of continents (Schuhmacher, 1976), with an intermediate phase of emergence. The upwelling started during the Tournaisian, maybe after a first phase of deformation in the West-Asturian-Leonese zoneand the beginning of the reversal of topography. This may indicate that the start of the upwelling was related to these events which may have been caused by tectonic plate interactions. The order in which the circulation improved in the Vegamián basins and the initial occurrence of black shales on the shallowest part of the shelf in front of the carbonate platform (southeastern part of the Alba syncline) indicate that the cold water entered the area via the West Asturian-Leonese zone.

REFERENCES

Adaro, L. & Junquera, G., 1916. Hierros de Asturias. Mem. Inst. geol. min. España, Criaderos de hierro de España, 2, 610 pp.

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

Al-Mehaidi, H.M., 1972. The geology of the devonian rocks between Robledo and Boñar. Léon province, Spain. Institute for Aerial Survey and Earth Sciences, Delft. Internal report.

Ambrose, T., 1972. The stratigraphy and structure of the pre-Carboniferous rocks north-west of Cervera de Pisuerga, Cantabrian Mountains, Spain. University of Sheffield, unpublished Ph.D. Thesis, 200 pp.

Arboleya, M.L., 1981. La estructura del manto del Esla (Cordillera Cantábrica, León). Bol. geol. min., 92, pp. 19-40.

Ark, P. van der, 1982.Conodonten, ouderdom en afzettingsmilieu van de basis van de Alba-Formatie in het Cantabrisch Gebergte (Spanje). Leiden University, Dept. of Stratigraphy and Palaeontology Internal report, 68 pp.

Baan, D. van der, 1970. Stratigrafie van de Portilla formatie in het oostelijke Esla gebied. Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 38 pp.

Barnes, C.R., ed., 1976. Conodont Paleoecology. Spec. Pap. geol. Assoc. Canada, 15, 324 pp.

Barrois, C., 1882. Recherches sur les terrains anciens des Asturies et de la Galice. Mém. Soc. géol. Nord, 2/1, 630 pp.

Bastida, F., Marcos, A., Arboleya, M.L. & Méndez, I., 1976. La unidad de Peña Corada y su relación con el manto del Esla (Zona Cantábrica, NW de España). Brev. geol. Asturica, 20, pp. 49-55.

Beard, D.C. & Weyl, P.K., 1973. Influence of texture on porosity and permeability of unconsolidated sand. Bull. Amer. Assoc. Petroleum Geol., 57, pp. 349-369.

Becker, G., Frankenfeld, H. & Schulze, R., 1979. Neue Daten zum Riffsterben im Oberdevon des Kantabrischen Gebirges (N-Spanien). Clausth. geol. Abh.,

30 (Festschrift Reinhard Schönenberg), pp. 19-33.
Benton, M.J. & Gray, D.I., 1981. Lower Silurian distal shelf storm-induced turbidites in the Welsh Borders: sediments, tool marks and trace fossils. Jour. geol. Soc. (London), 138, pp. 675-694.
Bischoff, G., 1957. Die Conodonten-Stratigraphie des rhenoherzynischen Unterkarbons mit Berücksichtigung der *Wocklumeria*-Stufe und der Devon/Karbon-Grenze. Abh. hess. Landesamt Bodenforsch., 19, 64 pp.
Bosch, W.J. van den, 1969. Geology of the Luna-Sil region, Cantabrian Mountains (NW Spain). Leidse Geol. Meded., 44, pp. 137-225.
Boogaard, M. van den, 1967. Geology of the Pomarão region (southern Portugal). Amsterdam University, Ph.D. Thesis, 113 pp.
- & Schermerhorn, L.J.G., 1975. Conodont faunas from Portugal and southwestern Spain, part 2. A Famennian conodont fauna at Cabezas del Pasto. Scripta geol., 28, 36 pp.
Brice, D., Bultynck, P., Deunff, J., Loboziak, S. & Streel, M., 1979. Données biostratigraphiques nouvelles sur le Givétien et le Frasnien de Ferques. Ann. Soc. géol. Nord, 98, pp. 325-344.
Briggs, D.E.G., Clarkson, E.N.K. & Aldridge, R.J., 1983. The conodont animal. Lethaia, 16, pp. 1-14.
Brouwer, A., 1962. Deux faciès dans le Dévonien des Montagnes cantabriques méridionales. Brev. geol. Asturica, 8. pp. 3-10.
-, 1968. Devonian of the Cantabrian Mountains, northwestern Spain. In: Oswald, D.H. (ed.). Int. Symp. Devonian System (Calgary, 1967), II, pp. 37-45.
-, 1978. Die europäischen Variszíden als Teilstücke des nordatlantischen Paläozoikums. Zeitschr. dtsch. geol. Ges., 129, pp. 557-563.
-, 1982. The Variscan Cantabrian zone (Iberian peninsula): a Palaeozoic platform. N. Jb. Geol. Paläont. Abh., 163, pp. 148-152.
- & Ginkel, A.C. van, 1964. La succession carbonifère dans la partie méridionale des montagnes cantabriques. C.R. 5me Congr. int. Stratigr. géol. Carb. (Paris, 1963), pp. 307-319.
Budinger, P. & Kullmann, J., 1974. Zur Frage von Sedimentationsunterbrechungen im Goniatiten- und Conodontenführenden Oberdevon und Karbon des Kantabrischen Gebirges (Nordspanien). N. Jb. Geol. Paläont. Mh., 1964, pp. 414-429.
Buggisch, W., Meiburg, P. & Schumann, D., 1982. Facies, Paleogeography and Intra-Devonian stratigraphic gaps of the Asturo-Leonese Basin (Cantabrian Mts./ Spain). N. Jb. Geol. Paläont. Abh., 163, pp. 212-230.
-, Raben, A. & Hühner, G., 1978. Biostratigraphische Parallelisierung und Faziesvergleich von oberdevonischen Becken- und Schwellen-Profilen E Dillenburg (Conodonten- und Ostracoden-Chronologie, Oberdevon I-V, Dillmulde, Rheinisches Schiefergebirge). Geol. Jb. Hessen, 106, pp. 53-115.
Bultynck, P., 1972. Middle Devonian *Icriodus* Assemblages (Conodonta). Geologica et Palaeontologica, 6, pp. 71-86.
-, 1975. Conodontes de la Formation de Fromelennes, du Givetien de l'Ardenne Franco-Belge. Bull. Inst. royal Sci. nat. Belg., 50/10, 30 pp.
- & Hollard, 1980. Distribution comparée de conodontes et goniatites dévoniens des plaines du Dra, du Ma'der et du Tafilalt (Maroc). Aardkundige Meded., 1, pp. 7-73.
Carls, P., 1982. Das kantabrische Devon und der Ibero-armorikanische Bogen aus keltiberischer Sicht. N. Jb. Geol. Paläont. Abh., 163, pp. 183-187.
Comte, P., 1936. Le Dévonien moyen et supérieur du León. C.R. Acad. Sci. (Paris), 202, pp. 1198-1200.
-, 1959. Recherches sur les terrains anciens de la Cordillère Cantabrique. Mem. Inst. geol. min. España, 60, 440 pp.
Coo, J.C.M. de, 1970. Faciësveranderingen in de Albasynclinaal (Verslag van een sedimentologisch onderzoek naar faciësveranderingen in midden- en bovendevonische afzettingen in de Albasynclinaal, Cantabrisch Gebergte, Spanje). Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 18 pp.
-, 1974. Lithostratigraphy of the Devonian Santa Lucía Limestones in León, Spain. Leiden University, Ph. D. Thesis, 87 pp.
Dreesen, R. & Thorez, J., 1980. Sedimentary environments, conodont biofacies and paleoecology of the Belgian Famennian (Upper Devonian). An approach. Ann. Soc. géol. Belg., 103, pp. 97-110.
Druce, E.C., 1976. Conodont biostratigraphy of the Upper Devonian reef complexes of the Canning Basin, Western Australia. Bull. Bur. miner. Resour. Geol. Geophys., 158/1, 303 pp.
Dunham, R.J., 1962. Classification of carbonate rocks according to depositional texture. In: Ham, W.E. (ed.). Classification of carbonate rocks. Mem. Amer. Assoc. Petroleum Geol., 1, pp. 108-122.
Dvořak, J., Mirouse, R., Paproth, E., Pelhate, A., Ramsbottom, W.H.C. & Wagner, R.H., 1977. Relations entre la sédimentation Eodevono-Carbonifère et la tectonique Varisque en Europe Centrale et Occidentale. In: La Chaîne Varisque d'Europe moyenne et occidentale. Colloq. int. CNRS (Rennes, 1974), 243, pp. 241-273.
Embry, A.F. & Klovan, E.J., 1972. Absolute water depth limits of Late Devonian paleoecological zones. Geol. Runschau, 61, pp. 672-686.
Engeser, T., Kullmann, J. & Schönenberg, R., 1981. Zerreiss-Strukturen (pull-apart-structures) im Rahmen der tiefoberdevonischen Tektonik des Kantabrischen Variszikums (Nordspanien). Zeitschr. dtsch. geol. Ges., 132, pp. 181-189.
Epstein, A.G., Epstein, J.B. & Harris, L.D., 1977. Conodont color alteration - an index to organic metamorphism. Prof. Paper U.S. geol. Survey, 995, 27 pp.
Evers, H.J., 1976. Geology of the Leonides between the rivers Bernesga and Porma (Cantabrian Mountains, NW Spain). Leidse Geol. Meded., 41, pp. 83-151.
Fåhraeus, L.E., 1976. Conodontophorid ecology and evolution related to global tectonics. In: Barnes, C.R. (ed.). Conodont paleoecology. Spec. Paper geol. Assoc. Canada, 15, pp. 11-26.
Franke, W. & Paul, J., 1982. Über den Ursprung der Rotfärbung in Sedimentgesteinen aus der Bohrung Schwarzbachtal 1. Senckenb. Lethaea, 63, pp. 285-292.
Frankenfeld, H., 1981. Krustenbewegungen und Faziesentwicklung im Kantabrischen Gebirge (Nordspanien) von Ende der Devonriffe (Givet/Frasne) bis zum Tournai. Clausth. geol. Abh., 39, 91 pp.
Frikken, H.W., 1978. De stratigrafie en sedimentologie van het Midden en Boven-Devoon in het gebied rond Cardaño de Arriba, provincie Palencia, N.W. Spanje. Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 18 pp.
Gandl, J. 1973. Die Karbon Triboliten des Kantabrischen Gebirges (NW-Spanien), 1: Die Trilobiten der Vegamián-Schichten (Ober-Tournai). Senckenb. Lethaea, 54, pp. 21-63.
García Alcalde, J.L., Arbizu, M.A., García López, S. & Méndez Bedía, I. (eds.), 1979. Guidebook of the field trip. Meeting of the international subcommission on Devonian stratigraphy (Spain, 1979). Oviedo (Universidad de Oviedo), 41 pp.
García López, S., 1972. Analisis del contenido en conodontes de un corte de la Caliza de Candás en la costa Asturiana. Brev. geol. Asturica, 16, pp. 49-53.

-, 1976. La Caliza de Candás en la zona de Luanco (Asturias) y su contenido en conodontos. Trab. Geol. Univ. Oviedo, 8, pp. 173-185.

Ginkel, A.C. van, 1965. Spanish Carboniferous fusulinids and theis significance for correlation purposes. Leidse Geol. Meded., 34, pp. 175-225.

Glenister, B.F., Klapper, G., & Chauff, K.M., 1976. Conodont pearls? Science, 193, pp. 571-573.

Goeijenbier, J.P.S., 1981. Tentaculieten uit het Onder- en Midden-Devoon van het Palentijnse Bekken (Cantabrisch Gebergte, Spanje). Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 107 pp.

Gründel, J. & Rösler, H.J., 1963. Zur Entstehung der oberdevonischen Kalkknollengesteine Thüringens. Geologie, 12, pp. 1009-1038.

Hallam, A., 1981. Facies interpretation and the stratigraphic record. Freeman, Oxford, 241 pp.

Have, A.H.M. ten, 1979. Stratigrafie en faciës van de devonische Portilla kalksteen formatie in het Somiedo-Luna gebied, Cantabrisch Gebergte, N.W. Spanje. Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 120 pp.

Hedberg, H.D. (ed.), 1976. International stratigraphic guide. Wiley, New York, 200 pp.

Helms, J., 1959. Conodonten aus dem Saalfelder Oberdevon (Thüringen). Geologie, 8. pp. 634-677.

Hernández Sampelayo, P. & Kindelan, J.A., 1950. Mapa geológico de España E. 1 : 50.000. Memoria explicativa de la hoja No. 32 Llanes (Asturias). Coullaut, Madrid, 109 pp.

Heward, A.P. & Reading, H.G., 1980. Deposits associated with a Hercynian to late Hercynian continental strike-slip system, Cantabrian Mountains, northern Spain. In: Ballance, P.F. & Reading, H.G. (eds.). Sedimentation in oblique-slip mobile zones. Spec. Publ. int. Assoc. Sedimentol., 4, pp. 105-126.

Higgins, A.C., 1962. Conodonts from the griotte limestone of northwest Spain. Notas Comunic. Inst. geol. min. España, 65, pp. 5-22.

-, 1971. Conodont biostratigraphy of the Late Devonian-Early Caboniferous rocks of the South Central Cantabrian Cordillera. Trab. Geol. Univ. Oviedo, 3, p. 179-192.

-, 1974. Conodont zonation of the lower Carboniferous of Spain and Portugal. Belgian geol. Surv. int. Symp. on Belgian micropalaeont. limits (Namur, 1974), 4, 17 pp.

- & Wagner-Gentis, C.H.T., 1982. Conodonts, goniatites and the biostratigraphy of the earlier Carboniferous from the Cantabrian Mountains, Spain. Palaeontology, 25, pp. 313-350.

Hoffmann, A. & Narkiewicz, M., 1977. Development pattern of Lower to Middle Paleozoic banks and reefs. N. Jb. Geol. Paläont. Mh., 1977, pp. 272-283.

Huisman, J., 1981. Conodonten uit de Santa Lucía-Formatie (Cantabrisch Gebergte, NW Spanje). Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 77 pp.

Jenkyns, H.C., 1974. Origin of red nodular limestones (Ammonitico Rosso, Knollenkalke) in the Mediterranean Jurassic: a diagenetic model In: Hsü, K.J. & Jenkyns, H.C. (eds.). Pelagic sediments on land and under the sea. Spec. Publ. int. Assoc. Sedimentol., 1, pp. 249-272.

Johnston, I.S. & Higgins, A.C., 1981. Conodont faunas from the lower Carboniferous rocks at Hook Head, County Wexford. Jour. Earth Sci. royal Dubl. Soc., 4, pp. 83-96.

Julivert, M., 1971. Décollement tectonics in the Hercynian cordillera of northwest Spain. Amer. Jour. Sci., 270, pp. 1-29.

-, 1981. A cross-section through the northern part of the Iberian Massif. In: Zwart, H.J. & Dornsiepen, U.F. (eds.). The Variscan Orogen in Europe. Geol. en Mijnb., 60, pp. 107-128.

-, Fontbote, J.M., Ribeiro, A. & Conde, L.E., 1974. Memoria explicativa del mapa tectónico de la Península Ibérica y Baleares E. 1 : 1.000.000. Madrid (Inst. geol. min. España), 113 pp.

-, Truyols, J., Marcos, A. & Arboleya, M.C., 1973. Mapa geológico de España E. 1 : 50.000. Memoria explicativa de la hoja No. 13 Avilés. Inst. geol. min. España, Madrid, 30 pp.

Klapper, G. & Johnson, J.G., 1980. Endemism and dispersal of Devonian conodonts. Jour. Paleont., 54, pp. 400-455.

-, Philip, G.M. & Jackson, J.H., 1970. Revision of the *Polygnathus varcus* Group (Conodonta, Middle Devonian). N. Jb. Geol. Paläont. Mh., 1970, pp. 650-667.

- & Ziegler, W., 1979. Devonian conodont biostratigraphy. In: House, M.R., Scrutton, C.T. & Bassett, M.C. (eds.). The Devonian System. Spec. Paper Palaeont., 23, pp. 199-224.

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

Krans, T.F., 1982. Block-movements and sedimentation in the Upper Silurian and Lower Devonian of the Cantabrian Mountains (NW Spain). N. Jb. Geol. Paläont. Abh., 163, pp. 163-172.

Krumbein, W.C. & Sloss, L.L., 1963. Stratigraphy and sedimentation, 2nd ed., Freeman, San Francisco, London, 260 pp.

Kullmann, J., 1960. Die Ammonoidea des Devon im Kantabrischen Gebirge (Nordspanien). Akad. Wiss. Lit. Mainz, Abh. Math. Naturw. Kl., 2, pp. 1-105.

-, 1968. Associations of rugose corals and cephalopods in the Devonian of the Cantabrian Mountains (Northern Spain). In: Oswald, D.H. (ed.). Int. Symp. Devonian System (Calgary, 1967), II, pp. 771-776.

- & Schönenberg, R., 1975. Geodynamische und paläökologische Entwicklung im Kantabrischen Variszikum (Nordspanien). N. Jb. Geol. Paläont. Mh., 1975, pp. 151-166.

- & -, 1978. Facies differentation caused by wrench deformation along a deep-seated fault system (León line, Cantabrian Mountains, North Spain). Tectonophysics, 48, pp. T15-T22.

- & -, 1982. Geodynamik und Palökologie im Kantabrischen Variszikum. N. Jb. Geol. Paläont. Abh., 1963, pp. 143-148.

Kunst, F., 1981. De sedimentaire en strukturele ontwikkeling in het gebied tussen Cervera de Pisuerga en Triollo, Cantabrisch Gebergte. Prov. de Palencia, Spanje. Utrecht University, Dept. of Structural Geology. Internal report, 70 pp.

Lane, H.R., Sandberg, C.A. & Ziegler, W., 1980. Taxonomy and phylogeny of some lower Carboniferous conodonts and preliminary standard post-*Siphonodella* zonation. Geologica et Palaeontologica, 14, pp. 117-164.

Lethiers, F., 1974. Ostracodes du passage Frasnien-Famennien de Senzeilles (Ardenne). Palaeontographica, (A) 147, pp. 39-69.

Leuteritz, K., Pietzner, H., Vahl, J. & Ziegler, W., 1972. Aufbau, Zusammensetzung und Entstehung von Calciumphosphat-Sphären in paläozoischen Kalken. Geologica et Palaeontologica, 6 pp. 111-137.

Lobato Astorga, L., 1977. Geología de los valles altos de los ríos Esla, Yuso, Carrión y Deva (NE. de León, NO. de Palencia, SO de Santander). Institución Fray Bernardino de Sahagun, León, 192 pp.

Loevezijn, G.B.S. van, 1982. Stratigrafie en facies van de Nocedo Formatie en de Ermita Formatie, Cantabrisch Gebergte, NW Spanje. Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 72 pp.

-, 1983. Upper Devonian block movements and sedimentation in the Asturo-Leonese basin (Cantabrian Mountains, Spain). Leidse Geol. Meded., 52, pp. 185-192.

- & Raven, J.G.M., 1983. The Upper Devonian deposits in the northern part of León (Cantabrian Mountains, northwestern Spain). Leidse Geol. Meded., 52, pp. 179-183.

Lotze, F., 1945. Zur Gliederung der Varisziden der Iberischen Meseta. Geotekt. Forsch., 6, pp. 78-82.

Lowe, D.R., 1975. Regional controls on silica sedimentation in the Ouachita system. Bull. geol. Soc. Amer., 86, pp. 1123-1127.

Maas, K., 1974. The geology of Liébana, Cantabrian Mountains, Spain: deposition and deformation in a flysch area. Leidse Geol. Meded., 49, pp. 379-465.

Marquínez, J.L., 1978. Estudio geologico del sector SE de los Picos de Europa (Cordillera Cantábrica, NW de España). Trab. Geol. Univ. Oviedo, 10, pp. 295-316.

Martínez García, E., 1981. El Paleozóico de la Zona Cantábrica Oriental (Noroeste de España). Trab. Geol. Univ. Oviedo, 11, pp. 95-127.

Matte, P., 1968. La structure de la virgation hercynienne de Galice (Espagne). Géol. alp., 44, pp. 157-280.

McLaren, D.J., 1970. Presidential address: time, life and boundaries. Jour. Paleont., 44, pp. 801-815.

Meischner, D., 1970. Conodonten-chronologie des deutschen Karbons. C.R. 6me Congr. int. Stratigr. géol. Carb. (Sheffield, 1967), III, pp. 1169-1180.

Michel, M.P. 1972. *Polyzygia* Guerich (Ostracoda) in the Devonian of Asturias and León (Spain). Leidse Geol. Meded., 48, pp. 207-273.

Mohanti, M., 1972. The Portilla Formation (Middle Devonian) of the Alba syncline, Cantabrian Mountains, prov. León, northwestern Spain: carbonate facies and rhynchonellid palaeontology. Leidse Geol. Meded., 48, pp. 135-205.

Morzadec, P. & Weyant, M., 1982. Lithologie et conodontes, de l'Emsien au Famennien, dans la rade de Brest (Massif Armoricain). Geologica et Paleontologica, 15, pp. 27-46.

Mouravieff, A.N. & Bouckaert, J., 1973. Mailleux's (1940) *Manticoceras intumescens* (Beyrich) in the Belgian Frasnian with respect to the conodont zonation. Geologica et Paleontologica, 7, pp. 93-95.

Müller, J. & Fabricius, F., 1974. Magnesian-calcite nodules in the Ionian deep sea: an actualistic model for the formation of some nodular limestones. In: Hsü, K.J. & Jenkyns, H.C. (eds.). Pelagic sedimentation on land and under the sea. Spec. Publ. int. Assoc. Sedimentol., 1, pp. 235-248.

Müller, M.J., 1969. Bürstenbildung bei Conodonten. Paläont. Zeitschr., 43, pp. 64-71.

Odin, G.S. & Matter, A., 1981. De glauconiarum origine. Sedimentology, 28, pp. 611-641.

Orchard, M.J., 1978. The conodont biostratigraphy of the Devonian Plymouth Limestone, South Devon. Palaeontology, 21, pp. 907-955.

Parga, J.R., 1969. Consideraciones sobre la erosión Fameniense en la Cordillera Cantábrica. Brev. geol. Asturica, 13, pp. 46-49.

Pello, J., 1968. Sobre la existencia de Devónico superior en la región central de Asturias y los problemas que plantea el contacto Devónico-Carbonífero al W. de Oviedo. Brev. geol. Asturica, 12, pp. 11-16.

-, 1972. Estudio geológico de la región central de Asturias. Oviedo University, Dept. of Geology. Unpublished Ph. D. Thesis, 198 pp.

Pettijohn, F.J., 1957. Sedimentary rocks, 2nd ed. Harper & Bros, New York, 718 pp.

Rácz, L., 1964. Carboniferous calcareous algae and their associations in the San Emiliano and Lois-Ciguera Formations (prov. León, NW Spain). Leidse Geol. Meded., 31, pp. 11-112.

Radig, F., 1958. Stratigraphie und Tektonik der asturischen Küste zwischen San Esteban de Pravia und dem Cabo de Torres (Prov. Oviedo, Nordspanien). Münster University, Dept. of Geology. Unpublished Ph. D. Thesis, 136 pp.

-, 1961. Zur Stratigraphie des Devons in Asturien (Nordspanien). Geol. Rundschau, 51, pp. 249-267.

-, 1966. Eine Oberdevon-Fauna aus dem östlichen Asturien (Spanien) und die Schichtlücke unter den Knollenkalken des Visé. Zeitschr. dtsch. geol. Ges., 115, pp. 515-523.

Raven, J.G.M., 1980a. De bovendevonische afzettingen in het Esla-gebied (Cantabrisch Gebergte, Spanje): sedimentatie en tektoniek. Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 72 pp.

-, 1980b. Conodonten uit het Givetien en Frasnien van het Asturo-Leonese Bekken (Cantabrisch Gebergte, Spanje). Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 77 pp.

Reijers, T.J.A., 1972. Facies and diagenesis of the Devonian Portilla Limestone Formation between the river Esla and the Embalse de la Luna, Cantabrian Mountains, Spain. Leidse Geol. Meded., 47, pp. 163-249.

-, 1973. Stratigraphy, sedimentology and paleogeography of Eifelian, Givetian and Frasnian strata between the river Porma and the Embalse de la Luna, Cantabrian Mountains, Spain. Geol. en Mijnb., 52, pp. 115-124.

-, 1974. Hinge movements influencing deposition during the Upper Devonian in the Esla area of the Cantabrian Mountains, Spain. Geol. en Mijnb., 53, pp. 13-21.

-, 1980. Sedimentary mechanisms in Spanish Devonian carbonates. Geol. en Mijnb., 59, pp. 87-96.

- & Have, A.H.M. ten, 1983. Ooid zonation as indication for environmental conditions in a Givetian-Frasnian carbonate shelf-slope transition. In: Peryt, T.M. (ed.), Coated grains. Springer, Berlin.

Reineck, H.-E. & Singh, I.B., 1973. Depositional sedimentary environments. Springer, Berlin, 439 pp.

Rhodes, F.H.T., Austin, R.L. & Druce, E.C., 1969. British Avonian (Carboniferous) conodont faunas, and their value in local and inter-continental correlation. Bull. brit. Mus. nat. Hist. Geol., Suppl. 5, 513 pp.

Ries, A.C., Richardson, A. & Shackleton, R.M., 1980. Rotation of the Iberian arc: palaeomagnetic results from north Spain. Earth and planet. Sci. Lett., 50, pp. 301-310.

Río, P. del & Menéndez Alvarez, J.R., 1978. Estudio lito y bioestratigrafico de la Caliza de Candamo (Loredo, zona central de Asturias). Trab. Geol. Univ. Oviedo, 10, pp. 379-387.

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

Sanchez de la Torre, L., 1977. Guia de las sesiones de campo. Formaciones detriticas y carbonatadas del Devónico medio y superior de la Cordillera Cantábrica. VIII congreso nacional de sedimentología (Oviedo-León, julio 1977).

-, Colmonero, J.R. & Suárez de Centi, C., 1976. Serie de la arenisca del Devónico superior en la playa de Gargantera. Trab. Geol. Univ. Oviedo, 8, pp. 187-201.

Sandberg, C.A., 1976. Conodont biofacies of late Devonian *Polygnathus styriacus* Zone in western

United States. In: Barnes, C.R. (ed.). Conodont paleoecology. Spec. Pap. geol. Assoc. Canada, 15 pp. 171-186.

- & Ziegler, W., 1979. Taxonomy and biofacies of important conodonts of late Devonian *styriacus*-Zone, United States and Germany. Geologica et Palaeontologica, 13, pp. 173-212.

-, -, Leuteritz, K. & Brill, S.M., 1978. Phylogeny, speciation, and zonation of *Siphonodella* (Conodonta, Upper Devonian and Lower Carboniferous). Newsl. Stratigr., 7, pp. 102-120.

Sartenaer, P.J.M.J., 1968. Famennian rhynchonellid brachiopod genera as a tool for correlation. In: Oswald, D.H. (ed.). Int. Symp. Devonian System (Calgary, 1967), II, pp. 1043-1060.

Savage, J.F., 1979. The Hercynian orogeny in the Cantabrian Mountains, N. Spain. Krystalinikum, 14 (1978), pp. 91-108.

-, 1981. Geotectonic cross sections through the Cantabrian Mountains, Northern Spain. In: Zwart, H.J. & Dornsiepen, U.F. (eds.). The Variscan Orogen in Europe. Geol. en Mijnb., 60, pp. 3-5.

- & Boschma, D., 1980. Geological maps of the southern Cantabrian Mountains (Spain). Leidse Geol. Meded., 50, pp. 75-114, encls.

Schuhmacher, H., 1976. Korallenriffe. Ihre Verbreitung, Tierwelt und Oekologie. München, Bern, Wien (B.L.V.), 275 pp.

Seddon, G., 1970. Frasnian conodonts from the Sadler Ridge-Bugle Gap area, Canning Basin, Western Australia. Jour. Soc. Western Australia, 52, pp. 21-30.

Sitter, L.U. de, 1959. The Rio Esla nappe in the zone of León of the Asturian Cantabric Mountain chain. Notas Comunic. Inst. geol. min. Espana, 56, pp. 3-24.

-, 1962. The structure of the southern slope of the Cantabrian Mountains; explication of a geolocial map with sections, scale 1 : 100.000. Leidse Geol. Meded., 26, pp. 255-264.

Sjerp, N., 1967. The geology of the San Isidro-Porma area (Cantabrian Mountains, Spain). Leidse Geol. Meded., 38, pp. 55-128.

Sleumer, B.H.G., 1969. Devonian stromatoporoids of the Cantabrian Mountains (Spain). Leidse Geol. Meded., 44, pp. 1-136.

Smits, B.J., 1965. The Caldas Formation, a new Devonian Unit in León, Spain. Leidse Geol. Meded., 31, pp. 179-187.

Staalduinen, C.J. van, 1973. Geology of the area between the Luna and Torío Rivers, Southern Cantabrian Mountains, NW Spain. Leidse Geol. Meded., 49, pp. 167-205.

Thibieroz, J., 1978. Transgresión de la "Caliza Griotte" (Viseense) encima de un zocalo acido (Pico Primiello, concejo de Amieva, Asturias). Brev. geol. Asturica, 12, pp. 62-64.

Tongeren, P.C.H. van, 1975. Stratigrafie en paleogeografie van het Boven Devoon en het Onder Karboon van het Cantabrisch Gebergte (Spanje). Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 43 pp.

Tucker, M.E., 1974. Sedimentology of Palaeozoic pelagic limestones: the Devonian Griotte (southern France) and the Cephalopodenkalk (Germany). In: Hsü, K.J. & Jenkyns, H.C. (eds.). Pelagic sediments on land and under the sea. Spec. Publ. int. Assoc. Sedimentol., 1, pp. 71-92.

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

Vermeulen, J.J., 1982. De geologie van het gebied rond Triollo, provincie Palencia, NW Spanje. Utrecht University, Dept. of Structural Geology. Internal report, 48 pp.

Vilas Minondo, L., 1971. El Paleozoico Inferior y Medio de la Cordillera Cantábrica entre los ríos Porma y Bernesga (León). Mem. Inst. geol. min. España, 80, 169 pp.

Villena, J., Pendón, J.G., Pardo, G. & Meléndez, A., 1981. Rasgos sedimentológicos del Carbonífero inferior de Montalbán (provincia de Teruel). Acta geol. hisp., 14 Homenatge a Lluís Solé i Sabarís (1979), pp. 208-212.

Voges, A., 1959. Conodonten aus dem Unterkarbon I und II (Gattendorfia- und Pericyclus-Stufe) des Sauerlandes. Paläont. Zeitschr., 33, pp. 266-314.

Wagner, R.H., 1955. Rasgos estratigráfico-tectónicos del Paleozóico Superior de Barruelo (Palencia). Estud. geol., 11, pp. 145-202.

- & Martínez García, E., 1974. The relation between geosynclinal folding phases and foreland movements in northwest Spain. Stud. geol. (Salamanca), 7, pp. 131-158.

- & Wagner-Gentis, L.H.T., 1963. Summary of the stratigraphy of Upper Palaeozoic rocks in NE Palencia, Spain. Proc. k. nederl. Akad. Wetensch., (B) 66, pp. 149-163.

-, Winkler Prins, C.F. & Riding, R.E., 1971. Lithostratigraphic units of the lower part of the Carboniferous in northern León, Spain. In: Wagner, R.H. (ed.). The Carboniferous of Northwest Spain. Trab. Geol. Univ. Oviedo, 4, pp. 603-663.

Walker, K.R. & Alberstadt, L.P., 1975. Ecological succession as an aspect of structure in fossil communities. Paleobiology, 1, pp. 238-257.

Wassink, J.C.L., 1979. De Rivera-Formatie tussen Ventanilla en San Martín (Palencia). Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 38 pp.

Westbroek, P., 1964. Systématique et importance stratigraphique des rhynchonelles du Calcaire de Crémenes (Dévonien Supérieur, province de León, Espagne). Leidse Geol. Meded., 30, pp. 243-252.

Wetzel, A., 1982. Cyclic and dyscyclic black shale formation. In: Einsele, G. & Seilacher, A. (eds.). Cyclic and event stratification. Springer, Berlin, pp. 431-455.

Wheeler, H.E. & Mallory, V.S., 1953. Designation of stratigraphic units. Bull. amer. Assoc. Petroleum Geol., 234, pp. 2361-2365.

Witte, J.M., 1980. De lithostratigrafie en de faciës van het Devoon van de Valsurvio dome, Palencia, Spanje. Leiden University, Dept. of Stratigraphy and Palaeontology. Internal report, 133 pp.

Ziegler, W., 1962. Phylogenetische Entwicklung stratigraphisch wichtiger Conodonten-Gattungen in der Manticoceras-Stufe (Oberdevon, Deutschland). N. Jb. Geol. Paläont. Abh., 114, pp. 142-168.

-, 1966. Zum höchsten Mitteldevon an der Nordflanke des Ebbesattels. Fortschr. Geol. Rheinl. Westfal., 9, pp. 519-538.

-, 1971. Conodont biostratigraphy of the european Devonian. In: Sweet, W.C. & Bergstrom. S.M. (eds.). Symposium on conodont biostratigraphy. Mem. geol. Soc. Amer., 127, pp. 227-284.

-, ed., 1975. Catalogue of Conodonts, vol. II. Schweizerbart, Stuttgart, 404 pp.

-, ed., 1977. Catalogue of Conodonts, vol. III, Schweizerbart, Stuttgart, 574 pp.

-, 1978. Devonian. In: Papers geol. time scale Symp., 25th int. geol. Congr. (Sydney, 1976). Stud. geol. amer. Assoc. Petroleum Geol., 6, pp. 337-339.

-, Klapper, G. & Johnson, J.G., 1976. Redefinition and subdivision of the *varcus*-Zone (Conodonts, Middle-?Upper Devonian) in Europe and North America. Geologica et Palaeontologica, 10, pp. 109-140.