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FACIES AND DIAGENESIS OF THE DEVONIAN PORTILLA LIMESTONE FORMATION BETWEEN THE RIVER ESLA AND THE EMBALSE DE LA LUNA, CANTABRIAN MOUNTAINS, SPAIN

BY

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ABSTRACT

In the central part of the Cantabrian Mountains, between the artificial lake in the river Luna in the west and the river Esla in the east, outcrops of the Portilla Limestone Formation were investigated. A fairly uniform development could be observed in four structurally different areas. Six lithostratigraphic units: four members and two beds, were recognized and correlations were established on the basis of lithostratigraphic characteristics. In these units six facies types could be distinguished. Lithological samples, characteristic of the facies types, served as a basis for the investigation of the diagenetic products and processes, especially the early diagenetic ones. Mainly in view of differences in lithofacies, and of the presence of biogenetic upgrowths related with the Sabero-Gordón line, this structural line is supposed to have acted already during deposition of the Portilla Limestone sediments. Arguments are presented in favour of an autochthonous origin of the Peña Corada Unit, earlier recognized as an allochthonous unit. A number of selected stratigraphic sections were systematically investigated on the faunal content, and a Givetian-Frasnian age could preliminarily be deduced for the Portilla Limestone Formation. Near Matallana only the base is possibly Eifelian; Frasnian fauna was observed nowhere.
The field observations, the additional laboratory observations and the data collected during investigation of the diagenetic processes and products served as a basis for a regional sedimentation model (Fig. 19). Isopach-lithofacies maps (Figs. 15, 16, 17, 18) illustrate the basin-fill history.
After a period in which mainly siliciclastic material was supplied by a hinterland situated in the north and in the vicinity of the Porma fault, carbonate sedimentation began. In the area east of the Porma fault, deposition of carbonate material started in the north and progressively shifted towards the south; in the area west of the Porma fault deposition of carbonate material progressed from the Matallana section on towards the west and the east. Both west and east of the Porma fault a vast shallow platform occurred, with prolific crinoid-bryozoan growth. This resulted in deposition of encrinic/encrinal grainstones and packstones, interrupted in some places by siliciclastic sediments. In the area east of the Porma fault a slope was present on which in Member B patchy reefs of the biostromal type developed. In the area west of the Porma fault enormous accumulations of bioclastic (coral) material (banks) were deposited. Bioturbated mudstones, wackestones and packstones show a relation to these biostromal and bank sediments. In the area east of the Porma fault this pattern was suddenly interrupted by supply of siliciclastic material, presumably coming from the north and the northwest. During the deposition of this siliciclastic material the flooded reefs could partly re-establish themselves, but were flooded afresh by a second pulse of siliciclastic material. In the area west of the Porma fault the siliciclastic material is interbedded and admixed in the carbonate sediments. Only in the extreme west is pure siliciclastic material present. The last phase of deposition of the Portilla Limestone sediments resulted in the development of bioherms and biostromes with associated sediments such as slightly dolomitized mudstones, and packstones containing pellets, ostracods, gastropods and fairly large quantities of bioturbation structures. The bioherms occurred mainly in places where reefs had previously developed. Due to the Upper Devonian erosion nothing can be said with certainty concerning the upper part of the sequence in the area north of Valdoré in the autochthonous area, and in the Pedroso syncline west of the Bernesga valley.

SAMENVATTING

In het centrale gedeelte van het Cantabrisch gebergte, tussen het stuwmeer in de Luna in het westen en de rivier de Esla in het oosten, werden ontsluitingen van de Portilla Kalksteen Formatie bestudeerd. In de vier structureel wezenlijk verschillende gebieden kon een vrij uniforme ontwikkeling worden waargenomen. Zes lithostratigrafische eenheden, nl. vier 'members' en twee 'beds' konden worden onderscheiden en voornamelijk op lithostratigrafische kenmerken werden correlaties vastgesteld. In deze eenheden werden zes hoofd-faciëstypen onderscheiden. Lithologische monsters welke kenmerkend zijn voor de onderscheiden faciës typen, dienden als uitgangspunt voor het onderzoek van voornamelijk de vroeg-diagenetische producten en processen. Voornamelijk op grond van verschillen in lithofaciës, en van de aanwezigheid van biogene bouwwerken, welke een relatie vertonen met de Sabero-Gordón lijn, wordt verondersteld dat deze structurele lijn reeds actief was gedurende de sedimentatie van de Portilla kalksteen sedimenten. Er worden argumenten naar voren gebracht, welke pleiten voor een autochtone oorsprong van de 'Peña Corada Unit', welke eerder als allochtoon werd onderscheiden. Enkele uitgekozen stratigrafische profielen werden systematisch onderzocht op de erin voorkomende fauna, en (voorlopig) kon een Givetien tot Frasnien ouderdom worden afgeleid voor de Portilla Kalksteen Formatie. Alleen bij Matallana is de basis mogelijkerwijze Eifelien, en hier is nergens Frasnien fauna waargenomen. De veldwaarnemingen, de aanvullende laboratoriumwaarnemingen en de gegevens welke werden verzameld tijdens het onderzoek van de diagenetische processen en producten dienden als basis voor een regionaal sedimentatie model (Fig. 19). De isopachen-lithofaciës kaarten (Figs. 15, 16, 17 en 18) geven een beeld van de geschiedenis van het opvullen van het bekken.
Na een periode waarin overwegend siliciclastisch materiaal werd aangevoerd uit een achterland wat gelegen is in het noorden en in het gebied van de Porma breuk, begon de afzetting van kalk. In het gebied ten oosten van de Porma breuk begon de kalkafzetting in het noorden en verschoof ze naar het zuiden, in het gebied ten westen van de Porma breuk ging de afzetting van kalk in westelijke richting en in oostelijke richting vanuit de omgeving van Matallana. Zowel ten westen als ten oosten van de Porma breuk kwam op een uitgestrekte, ondiepe vlakte een overvloedige groei voor van crinoiden en van bryozoën. Dit resulteerde in afzetting van de encrinische tot encrinitische 'grainstone' of 'packstone', welke afzetting op enkele plaatsen onderbroken is door siliciclastische sedimenten. In het gebied ten oosten van de Porma breuk was een helling aanwezig waarop in Member B 'patchy reefs' van het biostromaire type tot ontwikkeling kwamen. In het gebied ten westen van de Porma breuk werden enorme opeenstapelingen van bioclastisch (voornamelijk koraal) materiaal (banken) afgezet. Bioturbate 'mudstones', 'wackestones' en 'packstones' zetten zich af in relatie met deze biostromaire en bank sedimenten.

In het gebied ten oosten van de Porma breuk werd dit patroon plotseling onderbroken door aanvoer van siliciclastisch materiaal, waarschijnlijk uit het noorden en het noordwesten. Gedurende de afzetting van dit siliciclastische materiaal konden de overspoelde riffen zich deels herstellen, maar een tweede stoot van siliciclastisch materiaal overspoelde ze weer. In het gebied ten westen van de Porma breuk komt het siliciclastische materiaal gelaagd voor tussen kalksteenlagen, en is het hiermee vermengd. Slechts in het uiterste westen van het bestudeerde gebied komen weer pure siliciclastische sedimenten voor. De laatste fase van de afzetting van de Portilla Kalksteen sedimenten leidde tot de ontwikkelingen van biohermen en biostromen met geassociëerde sedimenten zoals licht gedolomitiseerde 'mudstones', en 'packstones' welke pellets, ostracoden, gastropoden, alsmede tamelijk veel bioturbate structuren bevatten. Deze biohermen komen voornamelijk voor op plaatsen waar eerder al riffen tot ontwikkeling waren gekomen. Ten gevolge van de boven-devonische erosie kan niets met zekerheid worden gezegd van het bovenste deel van de opeenvolging in het gebied ten noorden van Valdoré in het autochtoon, en in de Pedroso syncline, ten westen van het dal van de Bernesga.

SUMARIO

En la parte central de la cordillera Cantábrica, entre el embalse de la Luna al oeste y el Río Esla al este, aflora la Formación de las calizas de la Portilla, objeto de este estudio. Su antigüedad probablemente se remonta al Givetiense-Frasniense. En las cuatro regiones estructurales, esencialmente diferentes, se muestra un desarrollo bien uniforme. Cuatro paquetes y dos capas se distinguen y las correlaciones se basan especialmente en características litoestratigráficas. Por medio de una investigación detallada del terreno, confirmada por trabajos en el laboratorio, se han distinguido seis tipos primarios de facies. Fundado principalmente en diferencias de litofacies y en la presencia de 'construcciones biogenéticas' que presentan cierta relación con la línea Sabero-Gordon, se supone que aquella línea estructural ya era activa durante la sedimentación de las calizas de la Portilla. Se presentan argumentos que abogan por una origen autóctona de la Unidad de la Peña Corada, anterioremente distinguida como alóctona. Algunos cortes estratigráficos fueron sometidos a examen sistemático para averiguar que clase de fauna contenían lo que permitió deducir una antigüedad Givetiense-Frasniense. Solamente cerca de Matallana, el lado inferior tenga posiblemente una antigüedad Eifeliense. Aquí no se ha observado fauna de la Frasniense. Muestras litológicas representando tipos diferenciados de facies sirven de punto de partida para una investigación de los productos y procesos diagenéticos, especialmente los tempranos. Estos productos y procesos diagenéticos tempranos fueron investigados por separado en cada una de las cuatro regiones estructurales investigadas y se ha procurado relacionarlos con los tipos de facies reconocidos. Las observaciones sobre el terreno, las experiencias complementarias en el laboratorio y los datos coleccionados durante la investigación de los procesos y productos diagenéticos podrían ser agrupados en un modelo representativo de la sedimentación regional en el que la sucesión vertical que se muestra en partes de los perfiles está representada lateralmente (Fig. 19). Mapas con isopacas y litofacies de los cuatro paquetes (Figs. 15–18) representan los sucesivos ciclos sedimentarios. Después de un periodo de sedimentaciones siliciclásticas acarreadas por los afloramientos terrestres situados en el norte y en los alrededores de la falla de Porma, la sedimentación de las calizas comenzó cerca de la línea de León y cerca de Matallana. Al este de la falla de Porma la sedimentación de las calizas se desviaba hacia el sur; al oeste la sedimentación avanzaba hacia el oeste y el este. Sobre una plataforma extensa y poco profunda, tanto al oeste de la falla de Porma como al este, tuvo lugar un crecimiento excesivo de crinoideos y bryozoos dando por resultado 'grainstones' y 'packstones' de los tipos 'encrinal' y 'encrinite' los cuales están interpolados en sedimentos siliciclásticos. Al este de la falla de Porma se presenta una pendiente sobre la cual en el paquete B se desarrollaron 'patchy reefs' del tipo biostromo. Al oeste de la falla de Porma, enormes acumulaciones de material bioclástico (principalmente corales), fueron acumulados en bancos. Además se depositaron en relación con aquellos sedimentos biostromos y bancos 'mudstones', 'wackestones' y 'packstones' con osificios de litófagos. Al este de la falla de Porma, este modelo fue repentinamente interrumpido por transporte de material siliciclástico, probablemente del norte y del noroeste. Durante la acumulación de este material siliciclastico que anesó, los arrecifes que habían formado la techumbre podrían restablecerse parcialmente, pero una segunda avenida de material siliciclástico, los anegaba otra vez. Al oeste de la falla de Porma, el sedimento siliciclástico se presenta interealada en capas de calizas y mezclado con ellas. Solamente en el extremo oeste del terreno estudiado, los sedimentos siliciclásticos puros reaparecen. La última fase de acumulación de las calizas de la Formación de la Portilla ocasionó el desarrollo de biohermos y biostromos con sedimentos asociados, tales como 'mudstones' y 'packstones' ligeramente dolomitizados y conteniendo 'pellets', 'ostracodos' y 'gastropodos' así como razonablemente grandes cantidades de estructuras 'bioturbatas'. Los biohermos se presentan especialmente en aquellos lugares donde anteriormente se desarrollaron arrecifes. A consecuencia de la erosión del Devónico Superior nada se puede decir con seguridad de lo que sucedió en la parte superior en el terreno al norte de Valdoré en el área autóctona y en el sinclinal de Pedroso al oeste del valle del Río Bernesga.

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CHAPTER I

GENERAL REMARKS

AREA AND SCOPE OF STUDY

In the early fifties members of the Geological Institute of the University of Leiden began a detailed geological investigation in the southern part of the Cantabrian Mountains in the provinces of Palencia and Léon in northwestern Spain. The investigation can broadly be divided into a detailed mapping project and into a stratigraphic and palaeontologic study. In the framework of the latter attention recently has been focused on the regional, the sedimentological and the environmental description and interpretation of the succession. This paper is a contribution towards that end. Based on a study of facies and diagenesis an attempt is made to give an environmental interpretation and a depositional history of the Devonian Portilla Limestone Formation between the River Esla in the east and the artificial lake in the river Luna in the west. The field work for this study was carried out between 1968 and 1971. The area studied is subdivided into four sub-areas separated from each other by structural lines (Fig. 1). These sub-areas and the number of stratigraphic sections selected for this study, are:

(1) the Peña Corada area (P.C. on Fig. 1) of Rupke, 1965 (15 sections),
(2) the Esla autochthonous area (E.A. on Fig. 1) of Rupke, 1965 (8 sections),
(3) the Alba syncline area (A.S. on Fig. 1) of Evers, 1967 (5 sections),
(4) the Pedroso syncline area (P.C. on Fig. 1) of van Staalduinen, in prep. (4 sections).

AN OUTLINE OF GEOLOGY

On a slowly subsiding Cryptozoic basement Palaeozoic sediments were deposited, and the first important tectonic movement during the Palaeozoic is the tilting and subsequent erosion of the northern part of the Leonides (de Sitter, 1959, 1962) just prior to the Famennian. Large-scale folding and thrusting (especially in the Bernesga-Esla area) took place shortly before or during the early Westphalian (Sudetic phase). The Asturian phase is held responsible for refolding of the thrust structures previously formed. Westphalian sediments rest unconformably upon Lower Palaeozoic rocks; Stephanian, coal bearing, sediments lie unconformably both on Westphalian and on older Palaeozoic rocks. The Alpine orogeny placed the sediments in their present position. Some units in the Esla area have been interpreted by Rupke (1965) as thrust sheets. One of these is the Esla Nappe.* This has consequences for the present study because, according to this interpretation, outcrops on the Peña Corada Unit are out of depositional position. Although in the present study the Peña Corada area is regarded as autochthonous (cf. Chapter VI) reference will be made to the 'Peña Corada area' and to the 'Esla autochthonous area' in order to distinguish between these two areas already recognized and named as such by

* The Esla Nappe consists of the Peña Corada Unit, the Felechas syncline and the Agua Salio syncline. The distance of travel of the nappe structure, as calculated by Rupke (1965, p. 57), is 16 km. This distance even becomes 25 km, if we follow Rupke's calculation, but use the more recent data concerning the thicknesses of the various formations in the points considered by him.

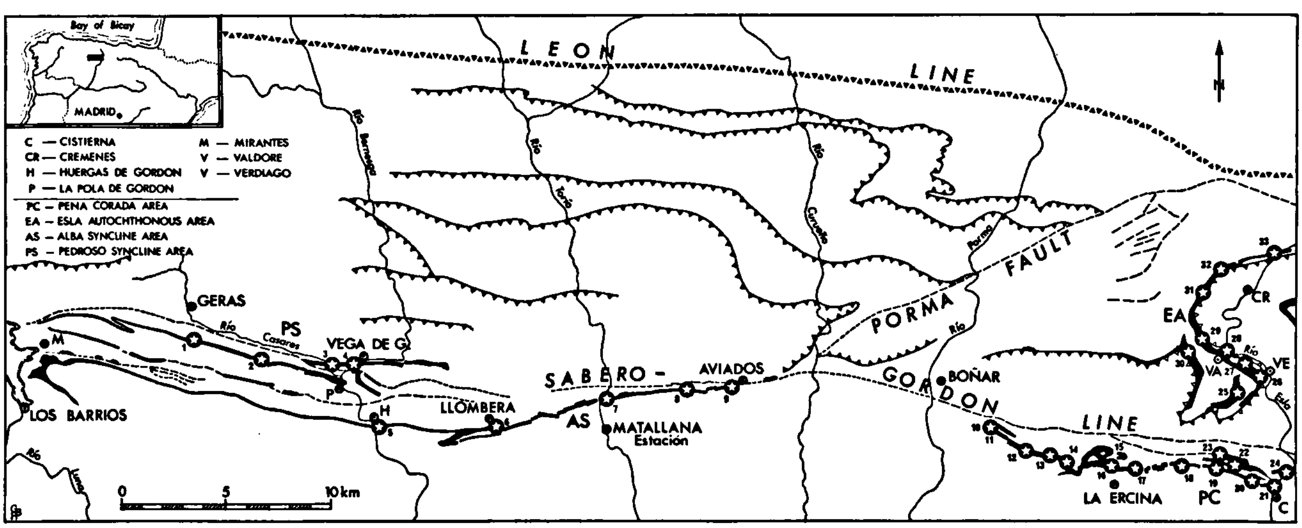

Fig. 1. Outcrop and index map of area studied.

Rupke (1965, p. 45–48). In the Porma-Bernesga area Evers (1967, p. 127) recognized the broad and intricately folded Alba and Corada synclinoria which, partly covered by younger sediments, may be regarded as one WNW-ESE striking synclinorium essentially south of the Sabero-Gordón line. Van Staalduinen (in prep.) mapped the continuation of these stuctures towards the west. Here the Alba syncline and the Pedroso syncline together form the Alba synclinorium. Still further to the west, van den Bosch (1969, p. 192) mapped the Abelgas syncline. The outcrops of the Portilla Limestone Formation in the Alba syncline west of the Bernesga River are subjected to detailed investigation by Mohanti (in prep.). North of this syncline the Sabero-Gordón line forms the southern boundary of the fourth area recognized in the present study, the Pedroso syncline. The Sabero-Gordón line separates an area of progressively higher uplift and subsequent erosion in the northern Leonides from an area of continued subsidence and rapid sedimentation in the southern sub-basins (Evers, 1967, p. 83). Rupke (1965, p. 38, 39) discusses the character of this line as a facies boundary. Another line which divides the Devonian basin into two sub-basins, and which acts partly as a facies boundary during deposition of the Portilla Limestone sediments, is the Porma fault.

METHOD OF APPROACH AND TECHNIQUES USED

The framework of the present study and the way in which the scope of this work, presenting an environmental picture of the Portilla Limestone Formation, was approached, is shown on Fig. 2.

Out of the great variety of techniques applicable to a study of carbonate rocks (cf. Wolf et al., 1967b) those, pictured in Fig. 3 were selected and used. Because in total approximately 1200 samples were examined and described, it was necessary for reasons of time to select 25 % of these samples, being the most representatives, for detailed examination. For checking the visual estimates out of these 164 samples were point-counted.

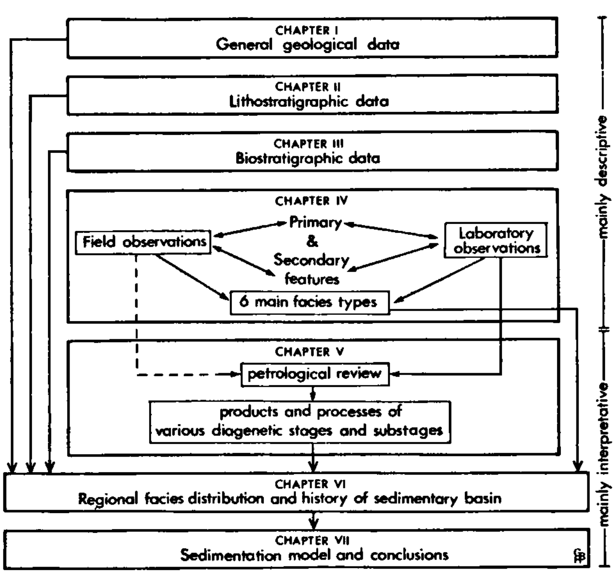

Fig. 2. Framework of present study.

REMARKS ON TERMINOLOGY AND CLASSIFICATION

In order to avoid misunderstanding of terminology and classification a summary is given of the terms and ideas used in this study. Most detrital limestones are characterized by the types and the relative quantities of textural components (Leighton & Pendexter, 1962, p. 36). Four components are dominant, viz. grains, lime mud, cement and pores.

Grains. – These are discrete particles, capable of forming a rock framework. The arbitrary lower size is 0.03 mm. Various categories of grains have been defined. There are four main groups.

(1) Detrital grains. The general term for carbonate sediments, as proposed by Wolf (1965b), is 'limeclast' or 'detrital carbonate grain'. Among these 'extraclasts' are obviously derived from pre-existing rocks. 'Intraclast' * (Folk, 1959, 1962) is a term used for carbonate fragments formed within the basin. Lumps (composite grains, often with superficial irregularities) and composite ooids and/or composite pellets (the grapestones of Imbrie & Purdy, 1962) can also be regarded as intraclasts sensu Wolf.

(2) Skeletal grains. These can be subdivided into fragmented and non-fragmented, skeletal grains. A subdivision according to the types of fossils is often possible.

(3) Pellets. This term is often used for all micritic grains lacking discernible internal structure and being ovoid or subrounded in shape.

(4) Ooids.** In this group it is possible to make a detailed subdivision.

(4a) Ooids (Sanders & Friedman, 1967), or true ooids, have clear layers of oriented crystals with a concentric or a radial structure.

(4b) Pseudooids (Carozzi, 1957) are grains without a clear coating of concentric or radial crystals. They may resemble pellets.

(4c) Superficial ooids, if only one layer is present.

(4d) Grains similar to, but larger than ooids and less regular in form. They are commonly crenulated, and rocks composed of these grains are known as pisolites.

(4e) Grains (nuclei) encrusted by Algae or by Foraminifera.

(4f) Axiolitic grains (Bissell & Chilingar, 1967, p. 96). These are radial-cylindric grains.

Lime mud and micrite. – Various definitions exist (cf. Folk, 1959, 1965; Leighton & Pendexter, 1962; Feray et al., 1962; Dunham, 1962). Bissell & Chilingar (1967) use

* For a genetic discussion of intraclasts in the Portilla Limestone Formation, see Chapter V.

** Oolite is used for rocks mainly composed of ooids.

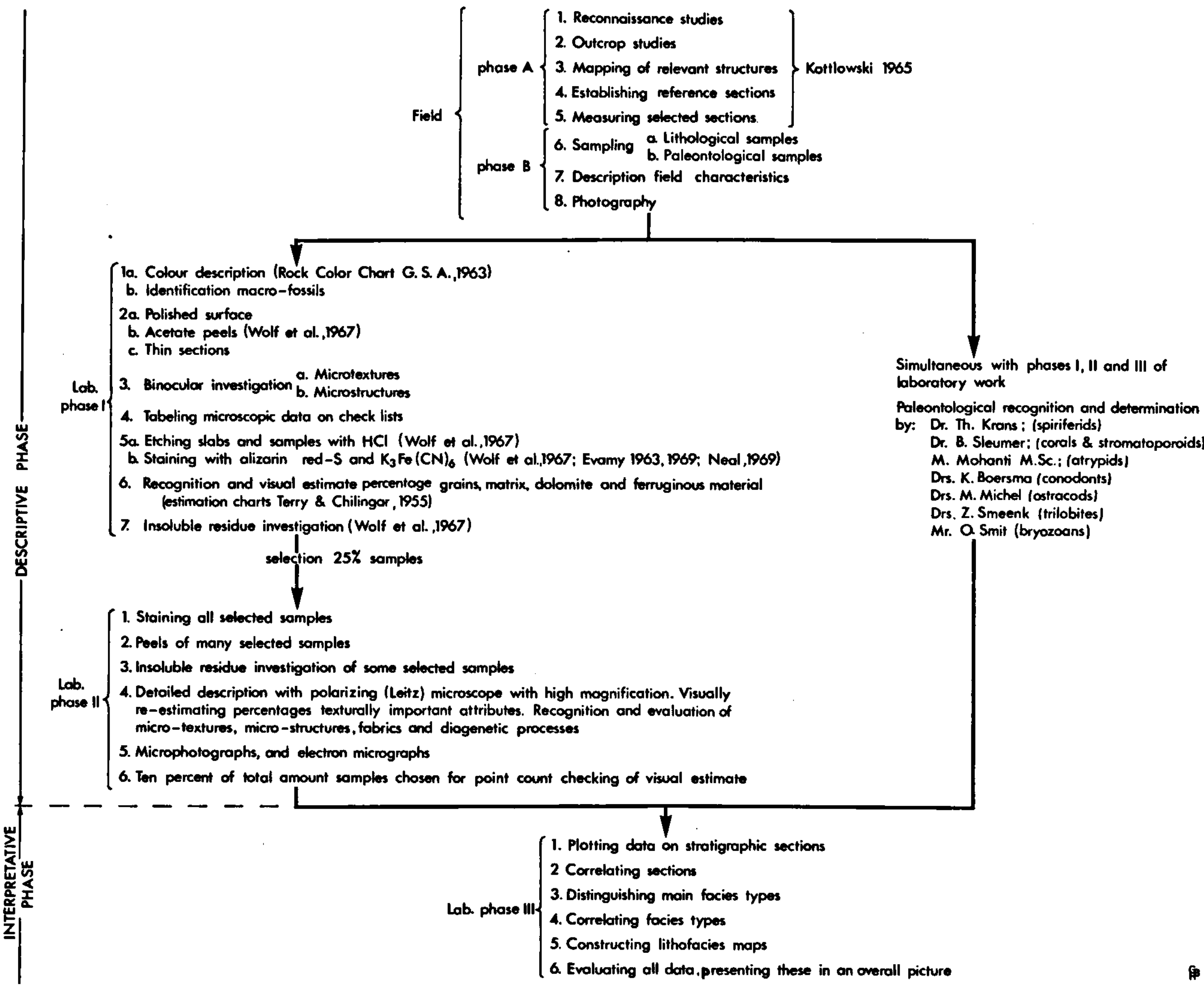

Fig. 3. Flow chart of techniques used in present study.

micrite for material, either crystalline or finely grained, which is 0.05 mm or smaller in diameter. It can be lime mud or its indurated equivalent. This 'boundary of convenience' has been followed in the field and in this sense the term micrite is used in the descriptive part of this work. In Chapter V time and mode of genesis is discussed and greater refinement has been made, whereby Folk (1965) has generally been followed. Matrix is used for interstitial material made up of grains or crystals greater than 0.05 mm in which sedimentary particles are embedded.

Cement. – According to its origin cement can be divided into ortho-sparite and pseudo-sparite (Bissell & Chilingar, 1967, p. 166). It may occur in many morphological shapes clearly recognized by Bathurst (1958) and extensively discussed by Folk (1965). Sparite will be used in the descriptive part of this work to designate any transparent or translucent calcite crystal larger than 0.05 mm. In Chapter V time and mode of genesis is discussed, and a more refined terminology will be used.

Dimensional terms such as rudaceous, arenaceous and lutaceous (Grabau, 1904, 1913) are used in a general way. During the microscopic and the binocular inspection these terms, where applied, were redefined following Bissell's and Chilingar's (1967) modification of Wentworth's grade scale. Authigenic components are classified according to the crystallinity scale on their table IV (p. 102, 103).
With regard to sorting, in the present study use is made of Folk's suggestion '...it would probably be better to define sorting by an arbitrary descriptive series of ranked adjectives, rather than setting up two distinct classes of 'well sorted' and 'poorly sorted'...' (Folk, 1962, p. 80).

The non detrital sediments (the skeletal structures like bioherms and biostromes) are not primarily defined by the textural components, but by their present day shape. Cumings' (1933, p. 333, 334) definitions that will be used are: '...Bioherm... a reef, bank or mound; for reef like, moundlike or lenslike or otherwise circumscribed structures of strictly organic origin, embedded in rocks of different lithology...'; '...Biostromes... purely bedded structures, such as shell beds, crinoid beds, coral beds, etc. consisting of and built mainly by sedentary organisms, and not swelling into moundlike or lenslike forms... which means a layer or bed...'.

With regard to the origin, skeletal limestones can be defined as reefs or banks (Nelson et al., 1962, p. 234). The ecological potential of the organism responsible for forming the deposits is of prime importance. '...Reef... a skeletal limestone deposit formed by organisms possessing the ecological potential to erect a rigid, wave resistant, topographic structure...'. '...Bank... a skeletal limestone deposit formed by organisms which do not have the ecological potential to erect a rigid, wave resistant structure...'.

The final choice of a textural classification based on the parameters described is difficult. Leighton and Pendexter (1962) suggested that the determination of the grain/micrite ratio is valuable in a textural classification of limestones. Their classification is primarily a descriptive one, but has some genetic implications and environmental significance in that the textural bases of the classification are strongly influenced by depositional processes. With a grain/micrite ratio a numerical value is assigned which denotes the gradation between two extremes. A high number refers to a small quantity of original mud, indicating a highly turbulent area, a low number indicates the contrary. This idea is followed in the present study. Dunham's classification is according to depositional texture. He stresses the importance of the presence or absence of mud, the abundance of mud and/or grains, and finally the presence of signs indicating binding. His classification is easy to use in the field, and provides a firm basis for the field observations of this study.

Remarks in accordance with the work of Leighton and Pendexter (1962) are often additional and were included during the second phase of the study, in the laboratory.

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CHAPTER II

LITHOSTRATIGRAPHY

The stratigraphy of the Palaeozoic in the Leonides is essentially based on the reconnaissance work of Comte (1959). His units are nowadays accepted, chiefly unaltered or with slight modifications, as formations in the sense of the 'Code of the Stratigraphic Nomenclature' (1961), (the 'Code', hereafter) and of the 'Statement of principles of stratigraphic classifications and terminology' (1960), (the 'Statement' hereafter).

DEFINITION OF THE PORTILLA LIMESTONE FORMATION

Among the units described by Comte are the 'Calcaires de la Portilla' of which the section '...tout au long de la rive droite de l'arroyo de la Portilla...' (1959, p. 199) in the valley of the river Torío evidently served as the type section. In the Peña Corada area and in the Esla autochthonous area Comte described some more sections. In the vicinity of Valdoré immediately above the Portilla Limestone Formation he distinguished the 'Calcaires de Valdoré'. Van Adrichem Boogaert (1967) redefined the Portilla Limestone Formation by including this Valdoré Limestone. Lithostratigraphically there are no objections against this procedure.

From data that have become available from this study and from careful mapping by Evers (1967), it is evident that the type section is incomplete (Enclosure IV). Moreover, the development in the type section cannot be regarded as representative for other areas investigated. Therefore the type section is supplemented with reference sections (Art. 13, a and i of the 'Code' and remark B 4 of the 'Statement'). For that purpose sections 14 (Veneros) and 24 were chosen. The geographical positions of the type section and the reference sections are: type section (Matallana): section 7: longitude, 1°50'18" W of Madrid, latitude, 42°51'0" N; section 14 (Veneros): longitude, 1°34'0" W of Madrid, latitude, 42°49'20" N; section 24: longitude, 1°26'15" W of Madrid, latitude, 42°49'18" N.*

TYPE SECTION AND REFERENCE SECTIONS

Comte described the lithological aspect of the type section in the following way (1959, p. 199, 200).

'...Les calcaires de la Portilla peuvent aussi être analysés tout au long de la rive droite de l'arroyo de la Portilla. Ils sont formés de 20 a 30 mètres de calcaires gris clairs en bancs irréguliers, assez riches en Polypiers et en Brachiopodes, séparés par des feuillets schistogréseux et schistomarneux...'. '...Les calcaires qui viennent ensuite sont disposés en gros bancs ou semi-massifs, ils renferments au début de nombreux Polypiers ramifiés et plus haut surtout des Polypiers massifs. La puissance de ces calcaires est d'environ 40 mètres. Leur pendage ainsi que celui des précédents est d'environ 70° secteur nord...'.

This description can now be supplemented by the following information (cf. Enclosure IV). The lower boundary of the succession is covered by Carboniferous sediments. The lowest lithological unit exposed (regarded as a member in this study) is Member B (12 m). Bedding is medium to thick. Colour is reddish grey with, especially in the uppermost part, a reddish green hue. Cross-beddings and grading features were observed. Shale is interbedded as partings. Platy tabulate corals, massive tabulate corals and compound rugose corals are present in great quantities. The boundary with the next lithological unit, regarded as a member, viz. Member C (29 m) is gradual. This member consists almost entirely of fragments of *Thamnopora* sp. and *Coenites* sp. Argillaceous material is interbedded and present as thin laminae. It is shaly and silty red material of the same kind as is present in the nose of the Alba syncline (Mr. M. Mohanti, pers. comm., 1970, and pers. obs.). Sedimentary features like wedging-out, gullies and oriented deposition of the rod-like *Thamnopora* sp. are common. Rhombohedral dolomite crystals are very commonly present. This unit grades into the next one, Member D (11 m). Bedding in this member is distinct. Graded bedding, burrowing (especially in the upper part), cross-bedding (especially in the upper 4 m), nodules and shale partings are present. Colour is light grey. Among the fossils especially the platy tabulate corals, the stromatoporoids and the massive tabulate corals as well as the compound rugose corals are present. The Nocedo Formation overlies this member, and the boundary shows slightly erosional features.

The reference section at Veneros has not previously been described. The main lithological characteristics are given below. The lower boundary of the lowermost member is the first limestone bed following the shales and the siltstones of the Huergas Formation. Thickness of this member is 4.50 m. The first 60 cm are composed of ooid-rich wavy, platy and yellowish weathered calcarenite. Bedding is 10–20 cm thick. The fresh colour is light grey, with some orange brown hematite dots. The oolite is often rather strongly recrystallized and along fractures hematitic material has intruded. Patchy secondary dolomite can be an important constituant.

* Longitude has been given west of Madrid because geological maps prepared in the Geological Institute in Leiden are based on topographical maps of the Instituto Geografico y Catastral de Madrid, that are designed in this way. Longitude of Madrid is 3°40'30" E with respect to Greenwich.

The upper part of the member is a regular, slightly wavy, yellowish weathered, medium calcarenite with crinoids, brachiopods and bryozoans. Cross-bedding is clearly visible. Erosion features were observed especially in the upper part. The contact with the overlying Member B is rather sharp. Bedding phenomena, fossil content and weathering colour change abruptly. Bedding in Member B is thicker; colour is less yellow, and coral layers, having a nodular aspect due to resistency differences, are intercalated. The depositional texture is coarse packstone, in some places wackestone and boundstone. The general aspect is biostromal and in some places small bioherms can be seen, together with slope phenomena such as small-scale slumping and pressure phenomena such as pressure solution. Member C (22 m) is poorly exposed. Only in a few places it is visible that shaly material is present. A few nodular and marly limestone lenses are present between this siliciclastic material with fragments of *Actinostroma stellulatum* and of auloporoid corals. Member D (50.50 m) contains in the lowermost part grainstones. Two nodular biostromal layers are situated at the transition into the upper part of Member D. The upper part is a regular alternation on the one hand of marly, shaly and dolomitic limestone, on the other hand of pure limestone and dolomitic limestone. Bedding thickness is 20–40 cm. Rock type is wackestone to mudstone. In this sequence nodular cherty beds are very clearly visible, and can be used as marker beds. They contain Phillipsastraeidae, *Thamnopora* sp. and *Alveolites* sp. In the uppermost parts of this section burrowing becomes important. The boundary between this member and the Nocedo Formation is sharp.

The reference section 24 has been described by Comte (1959, p. 214, 215). His description of the lithology is as follows.

'...Après avoir franchi une bande de Houllier qui prolonge vers l'Est le bassin de Sabero, on traverse, en suivant le cours de l'Esla, les Calcaires de Santa Lucia précédents, les Grès et Schistes de Huergas ou *Spirifer elegans* est commun et enfin les Calcaires de la Portilla...'. '...On observe ensuite la succession:

1. Calcaires gris en bancs minces, irréguliers, séparés par des feuillets marneux .... 8 m
les premiers lits sont schistoides
2. Calcaire gris massif à Polypiers .... 12 m
3. Calcaires marneux gris foncé .... 8 m
Ces calcaires sont très friables et se débitent en feuillets ou petits parallélépipèdes
4. Calcaires gris en bancs épais ou semi- massifs à Polypiers .... 12 m
5. Calcschistes grossiers noirâtres et grès marneux gris ou jaunâtres .... 4 m
Ces lits sont très friables et altérables
6. Calcaires marneux schistoides gris .... 2 m
7. Calcaires gris foncé en bancs épais ou semi-massifs à Polypiers .... 10 m
8. Calcaires massifs gris clair à Polypiers ..... 20 m
9. Grès calcarifères ou marneux, et calcaires en bancs irréguliers .... 30 m
A part quelques zones cette assise est relativement peu décalcifiée en surface. Les bancs sont en général minces sauf près du sommet...'.

'...1, 2, 3, 4 représentent les Calcaires de la Portilla; 5, 6, 7, 8, 9 représentent du Frasnien sous un faciès a prédominance calcaire comme autour de Valdoré...'.

This description can be supplemented by the following new observations. The lowermost lithological unit, regarded as a member, viz. Member A (14 m) begins where the first limestone bank can be traced over a great distance. Transition from the Huergas Formation into the Portilla Limestone is gradual. The uppermost levels of the Huergas Formation contain limestone lenses, coaly inclusions and plants. Branching bryozoans are also present. Member A is predominantly a grainstone with large quantities of crinoidal and bryozoan debris. At 1 m above the base we find the ooid bed (60 cm). Colour of the sediments of this member is brownish yellow and in some levels cross-bedding, gullies and slightly irregular bedding planes occur. Member B is predominantly biostromal (14 m). The lower part consists of packstone; the upper part of boundstone. The upper part is extremely rich in corals and has a nodular appearance. Some cross-bedding is present; in some levels bedding is very thick. Member C (23 m) is partly exposed. A minor thrust fault is present in this section and reduces the total thickness slightly. The middle part of this sequence is nodular and irregularly layered; here we find grainstones and packstones. The upper part is partly a boundstone, with a vast number of corals of extraordinarily large dimensions. Colour is yellowish brown to grey. Member D (40 m) is a packstone in the lower layers, a boundstone in the upper part. Especially the upper 20 m are very thickly layered. In the middle part of this member a clearly distinguishable biostromal deposit is present. The upper part is a biohermal deposit with all the characteristics, e.g. screes, pressure solution, sliding. This upper part is extremely rich in stromatoporoids, tabulate and rugose corals. In some places the corals are silicified.

DEFINITION OF LITHOSTRATIGRAPHIC UNITS IN THE PORTILLA LIMESTONE FORMATION

The earlier workers in the Cantabrian Mountains sometimes divided the Portilla Limestone Formation in different ways. In the Peña Corada area four different units can be seen and the fourfold subdivision as proposed by Rupke (1965, p. 22) is accepted with slight differences in this study. With some difficulty the same units can be recognized in the Esla autochthonous area. Here the subdivision as proposed by Rupke is also accepted in broad outlines. In the north flank of the Alba syncline, east of the Bernesga river, the various units as distinguished by Evers (1967, p. 94), are arranged into four distinguishable units. In the Pedroso syncline mainly three units are

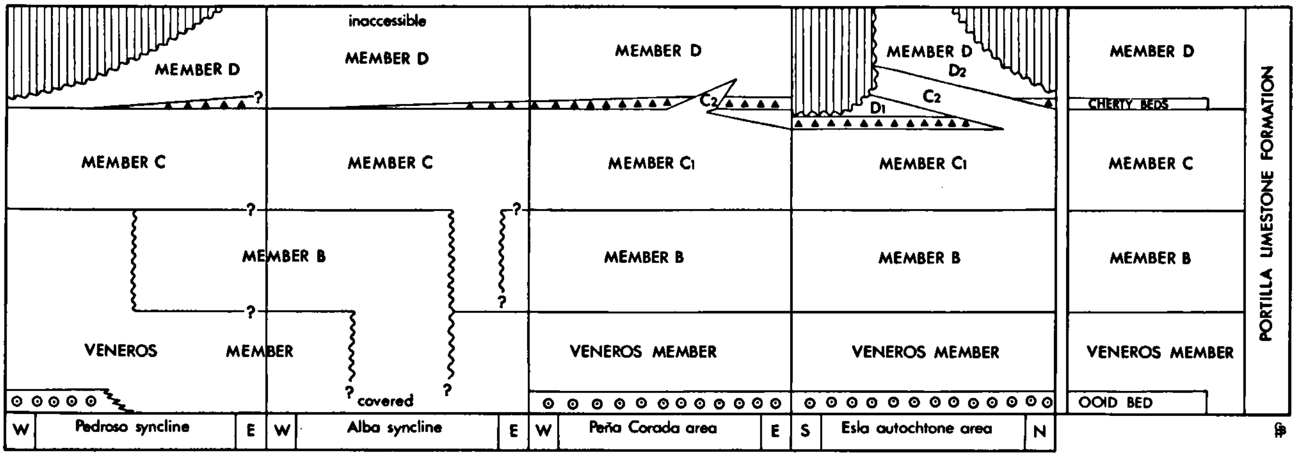


Table 1. Lithostratigraphic correlation of recognized units.

clearly visible. In Table 1 the proposed subdivision of the Portilla Limestone Formation is presented in a diagrammatic way. The correlation of the units distinguished is established in Enclosures I–IV. The following discussion of the lithostratigraphic units is based on field observations. In order to get a more detailed idea of the composition of, and of the differences between the various members and areas, field estimates on quantities of components have been made. These were subsequently chequed by point counting. These point count results are presented in 100 % circular diagrams (Figs. 4–7).

Member A (The Veneros Member)

Member A is the lowermost member of the Portilla Limestone Formation. It includes the ooid bed, commonly formed by the first layers. In a few places it does not occur, due to non-deposition or to tectonic complications. In the type section (7) e.g., the lowermost part of the section is covered by Carboniferous sediments; in section 21 non-deposition is postulated. The specific characteristics of Member A for each section are pictured in Enclosures I–IV, and Fig. 4 shows the average composition. This member consists of a rather

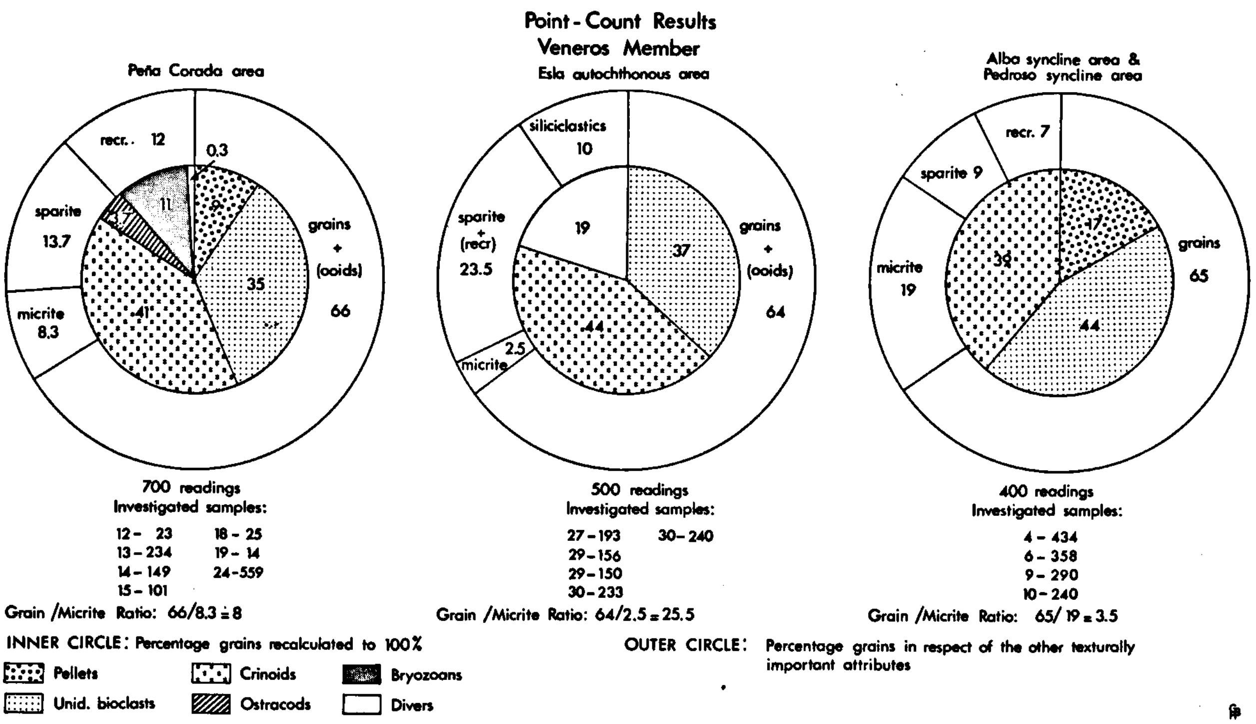


Fig. 4. Point-count results of selected samples from the Veneros Member.

pure, in many places thoroughly recrystallized medium, sometimes coarse calcarenite, containing crinoidal and bryozoan fragments. Only in a few places (e.g. in sections 4, 5, 6, 10, 22 and 28) are significant quantities of siliciclastics admixed and/or interbedded. In the north flank of the Alba syncline, east of the Bernesga River, the character of the lowermost member of the Portilla Limestone Formation changes completely in comparison with the other areas. In the easternmost sections crinoidal-bryozoan grainstones and packstones similar to those discussed, are present. More to the west, however, especially in the sections near Llombera and Huergas de Gordón, much siliciclastic material and many branching, platy and massive corals give a more biostromal aspect to the lower member (like the aspect of Member B which will be dealt with below). In general, however, Member A is fairly regularly developed, and shows many characteristics of the blanket deposits of Krumbein & Sloss' 'Blanket Carbonate Associations' (1963, p. 524, 525).

Member B

Member B is the second member in succession of the Portilla Limestone Formation in the two reference sections. In the type section it is the first exposed member as already has been pointed out. In section 21, too, it is the lowermost exposed member. Non-deposition of Member A is here postulated. With the exception of sections 1, 2 and 10, this member is present everywhere in the area studied. In section 10 Member B laterally grades into Member A as can be ascertained by closely studying the field and the lithological aspects. In the Pedroso syncline, the upper part of Member A seems to be laterally equivalent with Member B elsewhere. Transition from Member A into Member B is generally gradual. The boundary is drawn at the base of the first nodular, coral-rich layers (Enclosures I–IV). In sections 17, 26 and 29 in the Peña Corada area and in the Esla autochthonous area, Member B is composed of two or even three tongues, interfingering with Member A. In the entire Alba syncline area, east of the Bernesga River, a close interrelation with Member A is manifest. It is difficult to say what happens in section 25, since erosion has worn away much of this succession. In the Esla autochthonous area sections 28 and 29 show interfering especially with Member C.

In general Member B is biostromal. This is especially conspicuous in the Peña Corada area. Many tabulate and rugose corals, branching bryozoans and stromatoporoids, often in growth position, are the main rock-forming elements. We are here dealing with coarse to very coarse packstones, wackestones and locally even with boundstones. In three sections, 14, 18 and 20, small bioherms have been observed. Appreciable amounts of siliciclastic material are present in the sections 12, 29 and 30. In the Alba syncline area, especially in the vicinity of Aviados a biostromal layer, typical of this member, is intercalated in crinoid-bryozoan calcarenites, typical of Member A. In the Pedroso syncline Member B is not present as a traceable element in the field. After detailed study a locally important lens with biostromal aspects became distinguishable. The average composition of this member for the different areas is pictured in Fig. 5.

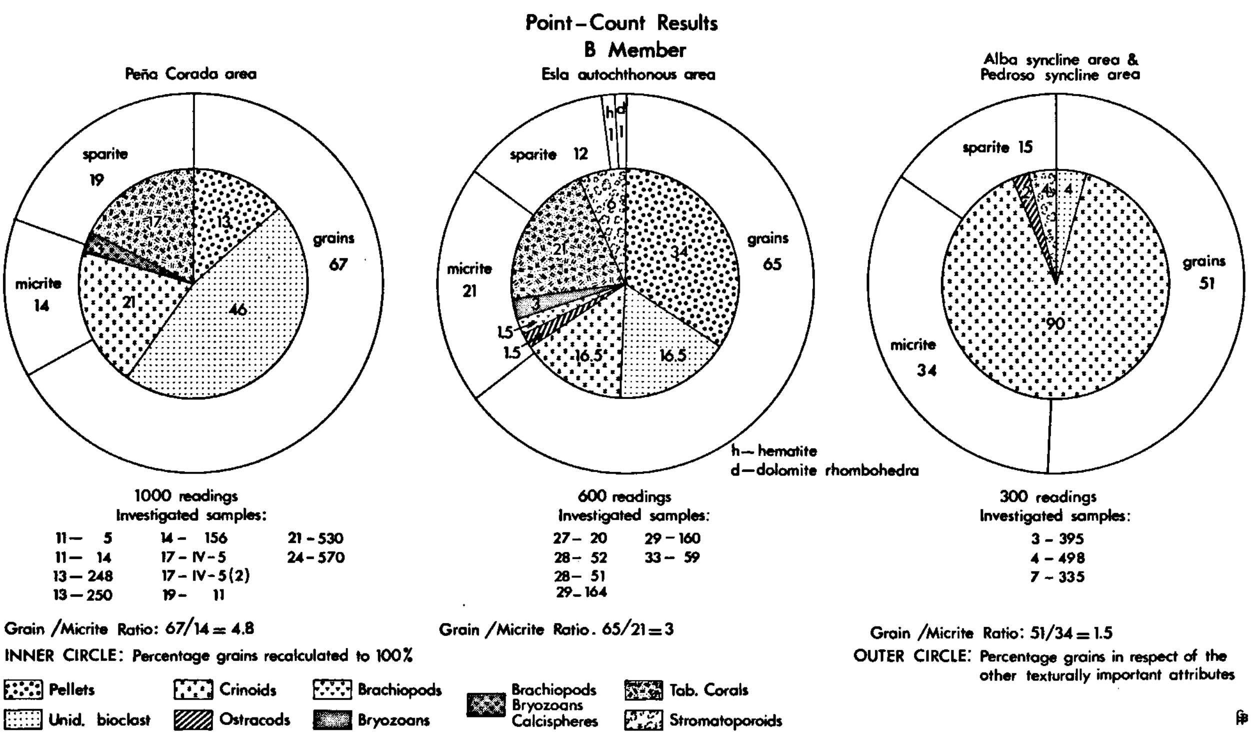


Fig. 5. Point-count results of selected samples from Member B.

Member C

The third member of the Portilla Limestone Formation is Member C. It is everywhere present except in section 21, where Member D rests conformably on Member B. The boundary between Members B and C is sharp in the Peña Corada area and in the Pedroso syncline. Elsewhere a gradual transition exists. The contact is conformable everywhere. Where transition is gradual, the lower boundary is drawn at the base of the first siliciclastic layer. Only in sections 31 and 5 is the lower boundary rather arbitrary. In a few places erosion features occur. In the eastern part of the Peña Corada area as well as in the northern part of the Esla autochthonous area Member C consists of two tongues, that seem to connect.

In general we find in Member C a siliciclastic and a limestone interval, both possessing different aspects in the four areas. Due to the generally poor exposure especially of the siliciclastic part, this member was difficult to study. In the sections near Aviados and Matallana Member C, present as a limestone, shows a biostromal and locally a bank-like aspect. Considerable quantities of branching and platy tabulate corals were accumulated. This differs importantly from the general aspect of this member elsewhere. In the vicinity of the Bernesga River the siliciclastic material predominates. It is here interbedded between rather thin biostromal deposits. In the western part of the Pedroso syncline Member C is purely siliciclastic and consists of quartz sandstone and locally even of quartzite. This member is traceable into the 'Mirantes Limestones' (the limestones, northeast of Mirantes near the Embalse de la Luna), into the exposures near Mallo and into the Babía area. In the Peña Corada area and in the Esla autochthonous area Member C consists mainly of siliciclastics, only interrupted by relatively unimportant limestone lenses. The average composition of the limestone part of this member is pictured in Fig. 6.

Member D

In the Peña Corada area and in the Pedroso syncline the lower boundaries of Member D are sharp and conformable contacts (Enclosures I, II and IV). In the sections east of 17, Member D is bifurcated into tongues D_1 and D_2. Both tongues have sharp lower boundaries (Enclosure II). The D_1 tongue in the Esla autochthonous area pinches out in northerly direction. The lower boundaries here as well as in the Alba syncline, are often arbitrary (Enclosures III and IV). In the whole area studied Member D is almost everywhere conformably overlain by the Nocedo Formation with a fairly sharp boundary. Only in sections 4, 5 and 16 (Enclosures I, IV) the transition is gradual. In the northern part of the Esla autochthonous area (Enclosure III) and in sections 1, 2 and 3 (Enclosure IV) the uppermost part of the succession is eroded and the upper Devonian sediments are present with an unconformable contact, which cuts progressively deeper into the Devonian sequence towards the north. In the Esla autochthonous area the Lancara

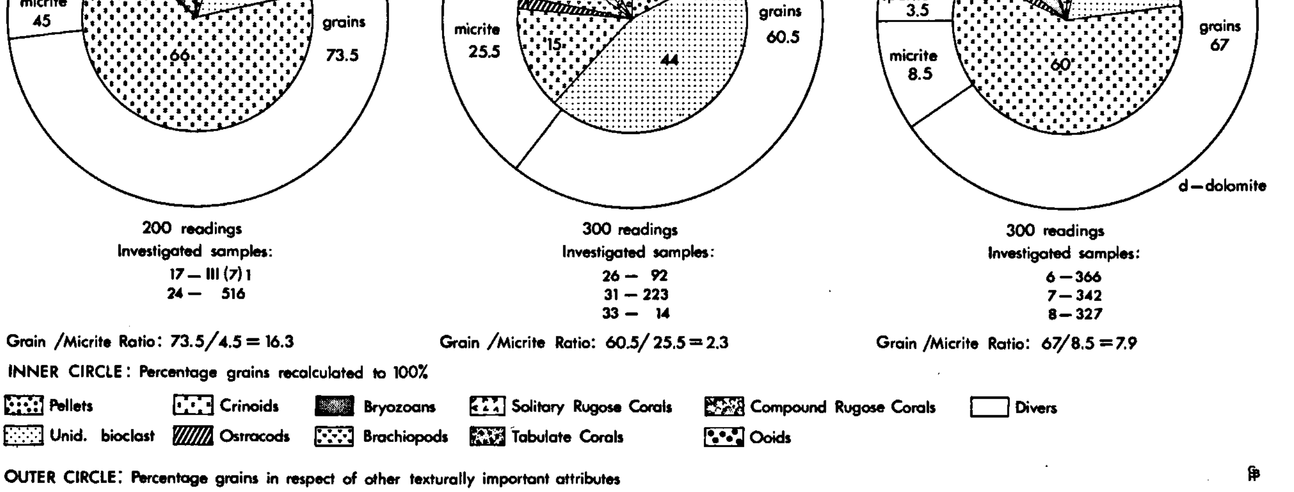


Fig. 6. Point-count results of selected samples from Member C.

Formation (middle Lower Cambrium; Lotze & Sdzuy, 1961) is tectonically superimposed upon the plane of the angular unconformity. Here Cambrian dolomites are in direct contact with Devonian sediments.

In the Peña Corada area, Member D is a rather intricate complex of high energy biostromal and locally biohermal sediments, laterally alternating with sediments deposited in quiet areas. The biostromal and biohermal deposits are mainly coarse packstones (sometimes wackestones) and boundstones; the sediments in the quiet areas are mainly fine-grained packstones, wackestones and mudstones. In these deposits, silicification is a common feature. In the Esla autochthonous area Member D is mainly composed of sediments deposited under very quiet conditions. Here we find wackestones and mudstones, and at some places in these mudstones indications of intertidal to infratidal deposition in the form of bird's eye limestones have been noted. In a few places crinoidal-bryozoan wackestones and grainstones are present, indicating slightly higher turbulence conditions. In the northern flank of the Alba syncline in the east of the area Member D consists of crinoidal-bryozoan calcarenites, similar to those of Member A. In the type section (7) Member D consists mainly of massive corals, often bounded together in a biostromal to biohermal unit. In the section near Llombera and more to the west the uppermost parts of the sections are inaccessible, but in the valley of the Bernesga River we find very clear biostromal developments with abundances of corals in a great variety, sometimes separated by siliciclastic layers.

In the Pedroso syncline Member D consists of biostromal to biohermal deposits. In the extreme west the biostromal aspect is predominant; in the vicinity of Vega de Gordón the section is extraordinarily thick, and grainstones and packstones with the same aspect as in the Veneros Member are intercalated. The average composition of this member for the different areas is pictured in Fig. 7.

Apart from the four member that are distinctly visible and can easily be traced in the field, some smaller units are discernible, and sometimes serve as important markers. The most important ones are the following beds.

Ooid bed in Member A (The Veneros Member)
The thickness of the ooid bed in Member A varies between 0.60 m and 1.50 m (Enclosures I–IV). It has a yellowish-grey weathering colour and is intercalated in the crinoidal-bryozoan calcarenites of Member A. The distribution of this bed over almost the entire Peña Corada area, almost the entire Esla autochthonous area and a large part of the Pedroso syncline is striking. Non deposition, however, occurs in sections 33, 31, 28, 23, 21 and 13, and in the entire Alba syncline (Mr. M. Mohanti, pers. comm., 1970). These ooid beds, too, have been observed in the 'Mirantes limestones' and in the section near Mallo. Den Hengst (1970) and van den Bosch (1969) reported the presence of the same ooid beds in the Babía area. These beds are everywhere present in the same stratigraphic position, and act

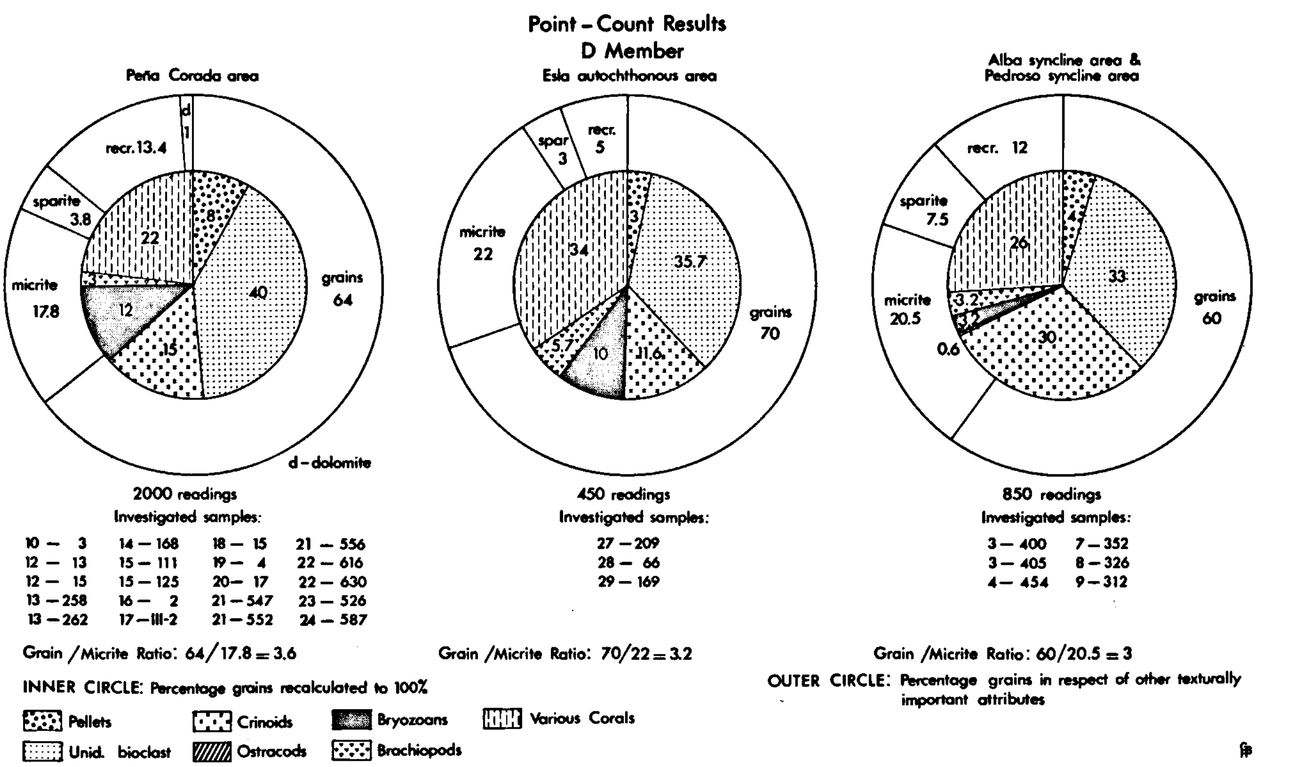


Fig. 7. Point-count results of selected samples from Member D.

generally as a transition between the Huergas shales and siltstones (a gradual boundary) and the encrinic and arenitic beds of Member A (a sharp boundary). In a few cases (in sections 1, 2, 24, 26, 27 and 29) these ooid beds are present at only a few metres from the first limestone beds of the sequence. In sections 12 and 19 the lowermost parts of the sequences are cut out by faulting with the result that the ooid beds, if deposited there, where not found.

The nodular cherty beds
These units are present in the Peña Corada area, especially in Member D. They are distinctly visible in areas where sediments were deposited under quiet conditions. Laterally from these areas we find biostromal and/or biohermal deposits. The nodular cherty beds are here traceable by means of silicified corals. In the Esla autochthonous area the nodular cherty beds are also present, but in a much more irregular manner. In the north flank of the Alba syncline, east of the Bernesga River, chert is only locally present in smaller quantities. In the Pedroso syncline we only find chert in the vicinity of Vega de Gordón.

ESTABLISHMENT OF ONE FORMAL MEMBER AND ONE KEY BED

The previously described Member A has a wide regional extension. With only a few exceptions the development is uniform. It is therefore logical to give it the rank of a formal member (art. 13 of the 'Code'). This member cannot be defined in the type section, due to incompleteness of the basal layers in part of the Alba syncline. It will therefore be defined in the reference section at Veneros. The name Veneros Member is hereby proposed (art. 10 f of the 'Code' and remark B 50 of the 'Statement').

The ooid bed in the Veneros Member is also of wide regional extent. In some areas it is always present at the boundary between the Huergas Formation and the Portilla Limestone Formation. There it serves as a key or a marker bed (art. 8 b of the 'Code'). For the same reasons as for the Veneros Member this key bed, too, will be defined in the reference section at Veneros. 'Ooid bed in the Veneros Member' is proposed as a name. The characteristics of both formal units proposed here have already been described (p. 171) and this description serves as the definition.

THE UNDERLYING AND OVERLYING SEDIMENTS

The sediments underlying the Portilla Limestone Formation belong to the Huergas Formation (Comte, 1959, p. 188). The type locality is direct below the measured Portilla Limestone section 5, in the village of Huergas de Gordón, in the valley of the Bernesga River. The lithology as described by different authors varies but there is agreement on a general division into a lower part, consisting mainly of shales and siltstones, with corals, trilobites and brachiopods, a middle part, consisting of weathered sandstones and quartzites often showing cross-beddings stressed by weathering, and an upper part, being a complex of decalcified sandstones, sometimes intercalated with shales and containing many brachiopods in some levels. Red coquinoid levels, rich in hematite, and quartz-iron sandstones are also reported (van den Bosch, 1969; Evers, 1967). Burrowing is locally very common. In the vicinity of Cistierna plants were observed in the upper layers by the present author. These observations support van den Bosch's interpretations, that in the lower part the Huergas sediments were generally deposited under slightly reducing conditions. Subsequently conditions sometimes changed, and even could become oxidizing. Beach deposits are frequently found. The general sediment association of the Huergas Formation according to van den Bosch is typical of an unstable shelf. The transition from the Huergas sediments into the Portilla sediments is abrupt in some places, but normally gradual (Enclosures I–IV).

The type section of the overlying Nocedo unit (Comte, 1959, p. 190–194) is situated near the village of Nocedo in the valley of the Bernesga River. There is again a rather good uniformity to be noted all over the area investigated, with regard to the composition of the Nocedo sediments. In general the sediments deposited in the lower part of the Nocedo Formation are fine-grained sandstones and sometimes even siltstones, whereas higher in the sequence the grain size increases. In a few places breccia-like sediments (possibly San Pedro ferruginous pebbles) are intercalated (Evers, 1967, p. 96; Rupke, 1965, p. 24). In the Nocedo quartzites and sandstones, limestones occur in the Agua Salio syncline and in the Bernesga valley, but they are only of local significance. The coarsening upwards of the sediments towards the top of the formation is thought to be connected with epeirogenetic uplift of the area (Rupke, 1965; Evers, 1967).

CHAPTER III

DATING OF THE STRATIGRAPHIC SEQUENCE BY MEANS OF BRACHIOPODS, CONODONTS AND MISCELLANEOUS TAXA

Recently different opinions have been expressed regarding the chronostratigraphic correlations of the Portilla Limestone Formation [Struve & Mohanti, 1970; Mohanti, in prep. versus Comte (1959), van Adrichem Boogaert (1967) and Dr. Th. F. Krans and Mr. K. Th. Boersma (pers. comm., 1970)]. In order to establish a chronostratigraphic framework investigation in a systematic way of sections 5, 7, 14, 28 and 33 was carried out. Additional information was gathered from sections 4, 9, 30 and 32. Identified faunal components are plotted in Enclosures I–IV.

BRACHIOPODS

a. *New material* (identifications Dr. Th. F. Krans and M. Mohanti MSc., Leiden)
(1) Eifelian.
Section 7 (base of the sequence): *Euryspirifer supraspeciosus* most probably indicates an Eifelian age.

(2) Givetian.
Section 33, sample 1 (2 m from base): *Cyrtina heteroclita*.
Section 33, sample 2 (4 m from base): *Spinocyrtia* cf. *alatiformis*.
Section 32, sample 22 (base Portilla Limestone Formation): *Spinocyrtia ascendens*.

This association possibly indicates a (Lower to Middle) Givetian age.
Section 33, samples 6–12 (8 m–22 m): *Spinocyrtia plicatula*.
Section 33, sample 18 (33 m from base): *Spinocyrtia* sp.

This faunal association was dated as (Upper?) Givetian.
Section 28, samples 55/56 (36 m from base): *Spinocyrtia plicatula*.
Section 28, sample 61 (48 m from base): *Spinocyrtia ascendens*.
Section 28, sample 63 (54 m from base): *Mucrospirifer bouchardi*.

This association indicates a Givetian age, but the latter sample possibly already indicates a Frasnian age.
Section 27, sample 192a (6 m from base): *Spinocyrtia plicatula*.

This most probably indicates a Givetian age.
Section 24, between samples 577–580 (42 m–50 m): *Spinocyrtia* sp., *Cyrtina* sp., and *Tingella* sp.

This association indicates a Middle Devonian age, possibly Givetian.
Section 14, sample 170 (60 m from base): *Tenticospirifer* sp.

This could indicate (Upper) Givetian to (Lower)Frasnian age. The noted range of this species is up to sample 173 (70 m from base).
Section 7, samples 342–356 (occurring from 15 m from the base to the top): *Spinatrypa (Isospinatrypa)* cf. *wotanica* and some collected comparable species probably suggest a (Lower) Givetian age.
Section 5, samples 374, 375 (6 m from base): *Mucrospirifer* cf. *bouchardi*.

Here the association with conodonts (cf. discussion hereafter) leaves no other possibility than (Upper) Givetian age, although elsewhere it occurs in the Frasnian.

(3) Frasnian.
Section 32, sample 32 (24 m from base): *Mucrospirifer bouchardi*.
This possibly indicates a Frasnian age.
Section 14, samples 170–182 (top section 14): *Tenticospirifer* sp., *Cyrtospirifer* cf. *stolbovi*, *Guerichella* cf. *pseudomullifida*, *Emanuella* sp., *Mucrospirifer bouchardi*, *Cyrtina heteroclita*, *Minatothyris* sp., and *Cyrtospirifer* sp.

These species indicate a Frasnian age, in which possibly a subdivision can be made into Middle and Lower Frasnian.
Section 4, sample 452a (51 m from base): *Cyrtina* sp. might indicate a Frasnian age.

b. *Earlier investigators*
Comte (1959) described Portilla Limestone sediments from sections 5, 7, 10, 24, 28 and 33 (numbers of sections in present study). After an analysis of the Devonian of the Cantabrian Mountains, and after comparison with other regions in Europe, the lower and middle part of the Portilla Limestone Formation were dated by him as (Upper) Givetian, on the basis of *Cupressocrinus orassus, Stropheodonta nobilis, Spirifer mediotextus, Spirifer berberinensis*. The upper part of the Portilla Limestone sediments was dated as (Lower) Frasnian (Assise de Fromelennes), on the basis of *Spirifer tenticulum*. The 'Calcaires de Valdoré' were dated as (Middle) Frasnian on the basis of *Spirifer bouchardi* (cf. hereafter) but were regarded as a separate unit. Van Adrichem Boogaert (1967) included these 'Calcaires de Valdoré' in the Portilla Limestone Formation, and this is maintained in the present study.

Struve & Mohanti (1970) recorded a Middle Devonian (Eifelian) atrypid fauna from the lower to middle part of the Portilla Limestone Formation in the nose of the Alba syncline. On the basis of a third species of *Spinatrypa (Invertrypa)*, viz. *Spinatrypa (Invertrypa)* *cantabrica*, a presumed time marker of late Eifelian could be established. Above this time marker, *Spinatrypa (Spinatrypa)* and *Spinatrypa (Isospinatrypa)* sp. cf. *wotanica* are definitely Givetian species. In addition, these authors state that only a few atrypids could indicate Frasnian age. An example is *Spinatrypa (Exatrypa)* sp. They therefore conclude that '...it is highly improbable that beds of Frasnian age are represented in the Portilla Formation of the Alba syncline...' (p. 164).

CONODONTS

a. *New material* (identifications Mr. K. Th. Boersma, Leiden)
(1) Upper Emsian.
Section 5, samples 374/375 (6 m from base): *Polygnathus foveolatus*.

This species is only known from the Upper Emsian (Mr. K. Th. Boersma, pers. comm., 1971) and is present in a sample in which further (Upper) Givetian species have been found. Here we are definitely dealing with a mixed fauna.

(2a) Givetian.
Section 24, sample 573 (33 m from base): *Icriodes expansus* and *Spathognadotus* sp. indet., together indicate Givetian age.
Section 5, samples 379/380 (14 m from base): *Icriodus latericrescens latericrescens* and *Belodus triangularis* together indicate a Givetian age.

(2b) Varca zone sensu van Adrichem Boogaert (1967).
Section 33, sample 1 (base): *Icriodus* cf. *I. latericrescens latericrescens, Icriodus expansus, Icriodus eslaensis, Icriodus linguiformis linguiformis* and *Polygnathus varca* together indicate an (Upper) Givetian age.
Section 24, sample 558 (transition Huergas Formation-Portilla Limestone Formation): *Polygnathus linguiformis linguiformis, Polygnathus linguiformis mucronata, Polygnathus varca* and *Icriodus eslaensis*.

This fauna indicates an (Upper) Givetian age.
Section 5, samples 374/375 (6 m from base): *Polygnathus linguiformis, Polygnathus varca* and *Icriodus eslaensis* is an association indicating an (Upper) Givetian age.

b. *Earlier investigators*
Among other areas, van Adrichem Boogaert (1967) investigated the Río Esla area (p. 135–138) and for some locations dealt with the conodont faunas found in the top and in the base of the Portilla Limestone Formation. These locations roughly correspond with sections 26, 28, 31, 32 and 33 in the present study. On the basis of *Icriodus* sp. indet., *I. cymbiformis, I. alternatus* and *Polygnathus decorosa*, the top of the Portilla Limestone Formation is placed in the Middle Frasnian.

On the basis of *Icriodus* sp. indet., *I. curvatus, I. nodosus, I. latericrescens latericrescens, I. expansus, I. eslaensis, I. cymbiformis, Polygnathus linguiformis linguiformis* and *Polygnathus varca*, the base of the Portilla Limestone Formation is placed in the varca zone sensu van Adrichem Boogaert, which corresponds with (Middle to Upper) Givetian. According to remarks by van Adrichem Boogaert (p. 157, 181) the distribution of icriodid conodonts is facies dependent.

MISCELLANEOUS TAXA

Trilobites (identifications: Mr. Z. Smeenk, Leiden)
Very little material indicating definite ages has as yet been collected. The only species that can so far be dated with certainty is *Neocalmonia (Heliopyge) iberia*, that occurs in section 14, sample 173 on to the top (80 m–102 m). According to Haas (1970) it indicates lower Frasnian. In the uppermost ten metres of the same section, *Phacops (Phacops)* n.sp. (cf. Smeenk, in prep.), *Phacops (Phacops)* sp., *Scutellum* sp., *Otarion* sp. and *Proetus (Proetus)* sp. do occur. *Neocalmonia (Heliopyge) iberica* also occurs in section 5 in sample 378 (12 m from base).

Ostracods (identifications: Mr. M. Michel, Leiden)
In the top of section 14 *Polyzygia neodevonica* occurs, and indicates a Frasnian age. Some species belonging to the family of Primiopsidae were found in section 24, sample 558 (transition Huergas Formation-Portilla Limestone Formation), which, according to Mr. M. Michel (pers. comm., 1971), indicate a Lower Devonian (Emsian) to a Middle Devonian (Eifelian?) age.

CONCLUSIONS

Two time lines can be drawn with some reserve, viz. the Eifelian-Givetian boundary and the Givetian-Frasnian boundary. The former is present in section 7 (?); the latter was found in sections 5, 14 (?), 24 (?) and 28 (?) (cf. Table 2). This is confirmed by some datings of trilobites and ostracods.

A general conclusion is that the base of the Portilla Limestone Formation is older in the north of the Esla autochthonous area, and younger in the south. In the basal part of section 33 the brachiopods are more or less of the same age as those in the base of section 7. The conodonts, however, suggest a (Middle) Givetian age for the base of section 33.

The ages derived by Struve & Mohanti (1970) for the basal part of the Portilla Limestone Formation in the westernmost part of the Alba syncline are more or less in accordance with those of section 7 (Matallana). In general, however, the sections in the areas investigated in the present study are younger than those described by Struve & Mohanti (1970). The only conclusion possible is that sedimentation started earlier in the west, and shifted progressively towards the east. This will be further elaborated in Chapter VI.

We must be aware, however, of the difficulties that

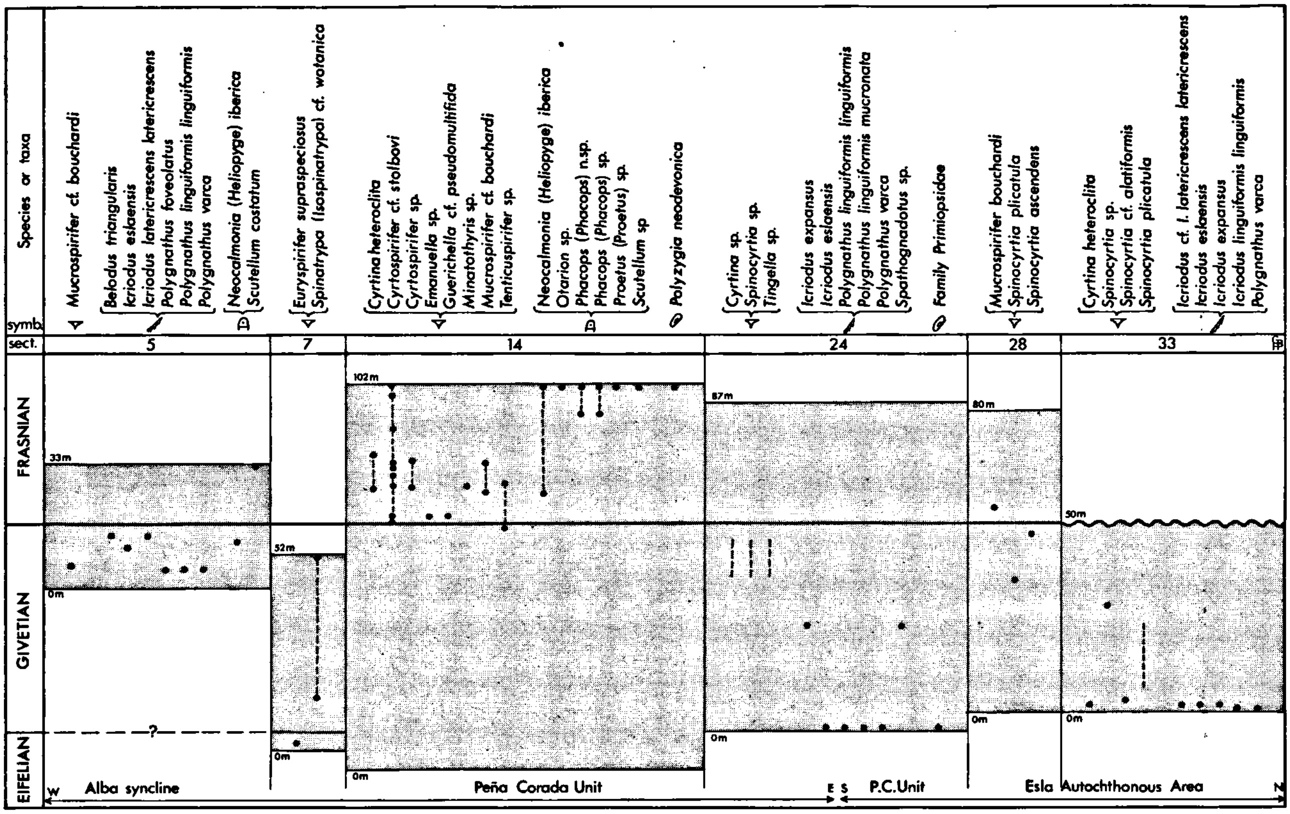


Table 2. Biostratigraphic correlations of some selected stratigraphic sections.

arise when comparing one group of faunal components with another. Discrepancies like those confronted with when comparing conodonts with brachiopods can also play a part in the differences in dating between Struve & Mohanti (1970, atrypid brachipods) and Dr. Th. F. Krans and Mr. K. Th. Boersma (spiriferid brachiopods and conodonts).

Another difficulty arises with *Mucrospirifer bouchardi*. It occurs in the basal part of section 5, in the upper part of section 14 and in the middle part of section 28. In the first section the association with conodonts strongly suggests an (Upper) Givetian age although in sections 14 and 28 a Frasnian age is suggested. Taking *Spirifer bouchardi* (correct name *Mucrospirifer bouchardi*) as a guide fossil, Comte (1959) derived a (Middle) Frasnian age for the 'Calcaires de Valdoré'. It may be preliminarily concluded either that the diagnosis of *Mucrospirifer bouchardi* has to be revised, or that the range of *Mucrospirifer bouchardi* is in need of revision (Dr. Th. F. Krans, pers. comm., 1971).

The top of the complete sections in the Portilla Limestone Formation lies mainly in the Frasnian (sections 5, 14, 24 and 28). Section 7, however, yields a Givetian age for the upper part. A Frasnian age can possibly be found in the uppermost part after close investigation. The enormous quantities of corals reflect an environment that presumably was not favourable for brachiopods and/or conodonts as is confirmed by the scanty presence of the former. Sections 28–33 are incomplete due to the Upper Devonian erosion. Nowhere has Frasnian been recovered, so that we are led to presume that the erosion cut out the sequence up to the top of the Givetian.

One difficulty remains to be discussed, viz. the mixed faunas present in the basal part of the Portilla Limestone Formation in sections 5 and 24. In the former section *Polygnathus foveolatus* (characteristic of the Upper Emsian, according to Mr. K. Th. Boersma) occurs together with an (Upper) Givetian conodont fauna. In the latter section an ostracod species of the family of the Primiopsidae also strongly suggests an Emsian age, although here a possibly Eifelian age has to be borne in mind. Similarities between both findings are that the faunal elements are present:

(1) in the detrital basal part of the Portilla Limestone Formation.

(2) South of the Sabero-Gordón zone.

(3) Not far from possible source areas (viz. the Santa Lucia Formation).*

Possible source areas are the Sta. Lucia exposures north of section 5 and the Sta. Lucia exposures [mapped as Portilla Limestone Formation by Rupke, 1965, but in reality Sta. Lucia, as described by Reijers (1969, p. 43),

* The Upper part of the La Vid Formation and the lower part of the Santa Lucia Formation are of Emsian age.

on the basis of *Actinostroma papilosum* (determination Dr. B. H. B. Sleumer, 1969)] NW of Cistierna and south of Saelices. In combination with conglomeratic remnants of the San Pedro Formation in the Crémenes Limestone and in the Ermitage Formation (Rupke 1965, p. 24, 27, 38) in the Esla area, and in the calcareous lower member of the Nocedo Formation (Evers, 1967, p. 96, 102) in the Alba syncline area, this mixed fauna with Sta. Lucia elements in the basal part of the Portilla Limestone Formation may indicate tilting and erosion north of the Sabero-Gordón line as early as during the (Upper) Givetian.

CHAPTER IV

FIELD AND LABORATORY OBSERVATIONS

Within lithostratigraphic units discussed in Chapter II seven different facies can be separated. These represent '...areally restricted parts where environmental conditions, causing the facies differentiation (were) stable for a sufficient length of time...' (Laporte, 1967, p. 77). These facies are units which, from the here given description, should be recognizable in the field by subsequent workers. Arbitrary boundaries sometimes were inevitable, especially where contacts between the facies are gradational, or where exposure is not complete. The positions of these facies in the lithologic columns are indicated on Enclosures I–IV. The facies laterally display a mosaic pattern which will be discussed in Chapter VI.

FIELD OBSERVATIONS

The ooids
The beds containing ooids are slightly wavy, medium layered, yellowish grey (5 Y 8/1) to moderate orange pink (5 YR 8/4) weathered, impure calcarenites (Pl. I–1). These calcarenites are locally impure because of the significant amounts of siliciclastics that may be admixed. Streaks and stylolites, containing much ferruginous material, are present. The contacts between the layers can be slightly dolomitized. Cross-bedding is very frequent. This is extremely prominent in the vicinity of Valdoré and Velilla. Here, too, gullies and grading features are well preserved (Pl. I–2). A few small-scale slumpings were observed. Normal ooids, superficial ooids and, in a few cases, pseudooids are recognizable. The ooids vary in size between 0.5 and 1.5 mm. In the field bimodality can be recognized. Nuclei of the ooids are siliciclastic grains, bioclastic fragments and in a few cases there are no nuclei (pseudooids). Macroscopically visible fossils are bryozoans, crinoids, fragments of brachiopods and pelecypods.

The pelletoidal, ostracodal, calcispheral, gastropodal and brachiopodal mudstone, wackestone and packstone: Facies b
Colour of the beds of this facies grades from medium grey (N 5) to medium dark grey (N 4). Sometimes yellowish grey (5 Y 8/1) and reddish greyish pink (5 R 8/2) colours occur. Bedding is obvious and regular to slightly wavy. Locally the appearance of the layers is slightly nodular as a result of irregular weathering and of stylolites present. A regular alternation can occur of massive, fairly pure limestone beds, with layers containing large quantities of marly and shaly material, in which fossil fragments are concentrated. Gullies and pinchings-out over short distances are common sedimentary structures. Some graded beds are present, and cross-bedding does occur. In the north of the Esla autochthonous area minor slumps occur. The presence of bird's eye limestone in sections 31,27 and 26 is noteworthy. Nodular chert is typical for this facies. It is concentrated in certain beds. The nodules can reach lengths of up to 50 cm. In some sections (e.g. section 14) six different nodular cherty beds were observed. Texturally the rock is a mudstone, a wackestone or a packstone. The grains are mainly bioclastic. Fragments of *Styliolina* sp., calcispheres, gastropods, ostracods, branching tabulate and compound rugose corals (Pl. IV–3), as well as significant amounts of unidentifiable bioclastic fragments are present. The coral fragments often show signs of transport. Especially the activity of scavengers brings about mingling of the sediment resulting in a strongly reworked, fairly uniform deposit. In a few cases, however, entire bioclastic elements are present in the fine interstitial material. In those cases the sediment has a wackestone-like appearance although, when using Dunham's classification in a strict sense, it is a packstone. Dolomitization can be significant. Plate IX–1 illustrates the gross characteristics of this facies.

The coral, bryozoan, packstone-boundstone with argillaceous interstitial material: Facies c
Thickness of the layers mainly vary between 10 and 100 cm. Bedding is moderately to very irregularly developed and wavy (Pl. IV–8). In the Alba syncline area bedding sometimes is obscured by recrystallization. Contacts between the layers are often stylolitic. The aspect of the layers can be nodular when a great amount of fossil fragments, e.g. branching and platy tabulate corals, massive rugose corals and stromatoporoids, bryozoans and brachiopods are present. In the Esla autochthonous area fossils are often extremely large. In section 23 platy tabulate corals were measured up to 240 cm in width (Pl. IV–6). Remarkable amounts of zaphrentoid rugose corals occur in beds containing much siliciclastic material (Pl. IV–5) in the north of the Esla autochthonous area. In the Alba syncline fossils are abundant, though in comparison with the other areas only a few different species occur (Pl. IV–7). In sections 5 and 6 gastropods, ostracods and trilobites are present. They seem to be restricted to this area. Bituminous layers, containing minor amounts of greyish-black (N 2) to black (N 1) splintery shaly material, are sometimes interbedded. Pyrite is present in some localities (Pl. IV–5). In general colour of the beds of this facies is brownish grey (5 YR 4/1) to dark grey (N 3) and in the Alba syncline moderate red (5 R 4/6), with pale olive (10 Y 6/2) streaks. In section 10, 14 and 23 small slumps were seen. Loadcasting structures (Pl. IV–8) occur. In section 15

boundstone-lentils of small dimensions in Member C (Pl. IX–2) are comparable to the sediments in Member B of section 18 (Pl. IV–4). In the Alba syncline, cobble-sized intraclasts together with fossils constitute the framework of the packstone-boundstone texture of the rock. Very fine-grained bioclastic and shaly material is interstitially present. Cross-bedding, pinching-out and interfingering layers are fairly common in some intervals (Pl. IV–4). Especially in the Alba syncline rod-like bioclastic fragments are often deposited in an oriented way. Ooids occur in sections 12 and 19. In these sections and in the Alba syncline dolomitization is often macroscopically visible. In some localities of the Esla autochthonous area, and from Llombera towards the Bernesga River facies c occurs with mixed aspects (Pl. IV–3).

The coral, stromatoporal, boundstone-packstone with a biohermal aspect: Facies d

This facies is only recognized in Members B and D, and mainly in the Peña Corada area. The bioherms in Member D are generally thicker and better developed than those in Member B. Bedding is very thick to indistinguishable. Colour is very light grey (N 8) to white (N 9) with, in some levels, pale yellowish points and streaks, due to hematite-coated dolomite rhombohedral crystals. In many places, pale reddish brown (10 R 5/4) macroscopically recognizable dolomite rhombohedra were observed more or less parallel to the supposed bedding. The percentage of dolomite increases towards the top of these sediments. Recrystallization often obscures the primary textures and structures. Where recrystallization is not too strong packstones and boundstones can be recognized. Faunal elements are platy, massive and branching tabulate corals, compound and massive rugose corals, massive and platy stromatoporoids. Less important are branching and fan-shaped bryozoans, crinoids and some ostracods. This facies occurs in dome-like structures. In the field these domes are stressed by exfoliation (Pl. X–5). The lateral extension is seldom more than 200 m. The average dip that can be measured is 5° . After studying the isopach map (Fig. 16) it is noticed that in some sections (e.g. section 18) where this facies occurs, increased thickness is present. This may suggest vertical upgrowth. The previously described features justify the conclusion that we are dealing with real bioherms (Fig. 8).

Beds underlying the bioherms may have a wavy appearance, possibly due to the biohermal weight (Pl. X–6). Stylolites are common in and between these beds. These sediments are extremely fossiliferous. Fossils often cause a nodular aspect. The material between the fossils grades from calcirudites to calcilutites. Mainly sand-sized fragments are present. Lateral, with respect to bioherms, we find screes, composed of nodular, vaguely distinguishable beds, dipping 5° . These screes partly 'envelope' the bioherms sensu stricto (Pl. X–5, and Fig. 8). The faunal content does not differ importantly from that in the bioherm. Silicification of fossil fragments, however, is far more common in these enveloping screes than in the bioherms sensu stricto. In the screes a remarkable regular alternation is visible of layers with fossils, either in growth position or accumulated more or less in situ, with layers in which bioclastic fragments occur, showing clear signs of transport and, in a few cases, of high turbulence. Large compound coral fragments may be present in a reversed position (cf. Sleumer, 1969, p. 14). The total thickness of the sediments forming the scree increases whereas the sediments composing the bioherm sensu stricto decrease in thickness (Pl. X–7).

The encrinal (encrinite) bryozoan grainstone: Facies e

Bedding is well developed. The contacts between the beds are slightly wavy. Cross-bedding is a very common structure. Weathering colour is light brown grey (5 Y 6/1) to yellowish grey (5 Y 8/1) in many places. Sometimes it grades from medium grey (N 5) into medium dark grey (N 4). The most important field characteristics are reproduced in Plates I–1, I–2 and II–1. Macroscopically visible fossil fragments are enormous quantities of crinoid ossicles and branching and fan-shaped bryozoans. Some brachiopods are also present. Post-depositional effects have in many places obscured the fossil content and the sedimentary structures. In those places bedding is often very thick and massive. Macro-

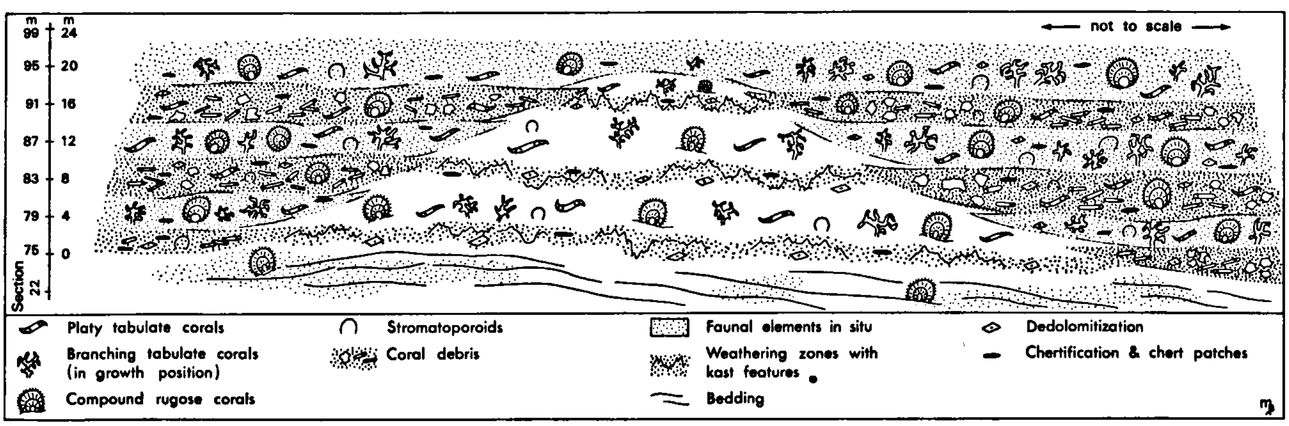


Fig. 8. Diagrammatic representation of a bioherm in section 22, Member D.

stylolites roughly indicate the original position of the layers (Pl. II–1). A grain-supported texture is common in this facies, and we are generally dealing with grainstones.

The encrinal, bryozoan, coral, brachiopodal packstone and wackestone, with marly intercalations: Facies f
Bedding is irregular and possesses a wavy appearance, and there are often important resistance differences, produced by 'knoll'-shaped massive corals, solitary zaphrentoid corals, brachiopods, fan-shaped and branching bryozoans and various fossil fragments bedded in soft interstitial material. Sometimes the aspect of the beds can even be breccious (Pl. V–1) due to the enormous quantities of bioclastic debris. Individual beds are seldom thicker than 10 cm. The fragments grade from arenite to rudite size. Sedimentary structures, such as cross-bedding, gullies, grading and small slumpings, are common. Colour of the sediments grades from yellowish grey (5 Y 8/1) to greyish yellow green (5 GY 7/2) and in some places we find brownish grey (5 YR 4/1) intercalations of marly material (Pl. II–2) but also pinkish grey (5 YR 8/1), recrystallized, purer limestone lenticles (Pl. II–3). Bioturbation and pelleting is generally common. Only in section 33, where this facies regularly alternates with facies g, is the amount of pellets significantly diminished, Here the large size of the solitary rugose zaphrentoid corals (up to 15 cm) is remarkable (Pl. II–2). The general characteristics of this facies are illustrated in Plates II–2, II–3, II–4. In the field sediments belonging to this facies can easily be classified as packstones and in a few cases as grainstones and as wackestones. Small pieces of the samples incidentally show binding structures, and consequently the texture in those spots is that of boundstone. These structures are related to the corals present in those places. Locally we find thin biostromal deposits.

The bioclastic packstone and wackestone, admixed with significant quantities of siliciclastics: Facies g
Due to the often poor exposure of this facies, little can be said about the characteristics. In general it consists of significant amounts of sand-sized, silt-sized or shaly siliciclastics. In the Pedroso syncline the sandstone layers contain almost only quartz. Burrowing is here fairly common. In the vicinity of Valdoré they are rather pure sandy and silty, with, intercalated, limestone lenticles containing crinoids, bryozoans, brachiopods, calcispheres and tentaculites. Cross-bedded limestone layers are present (Pl. IX–3). These contain ooids. Siliciclastics too occur as interbeddings or admixtures in limestones, producing argillaceous limestones and in a few extreme cases lime-siltstones (Plates II–5, II–6). The sandy and silty material in limestones often accentuate sedimentary structures such as cross-beddings, gullies and grading (Pl. II–7). Zaphrentoid and compound rugose corals and some branching, some massive tabulate corals are typical components of the argillaceous limestone part of this facies. Branching and fan-shaped bryozoans are also present. In many cases, the fan-shaped bryozoans (Pl. II–8) trapped the fine-grained sediment and produced very calm micro-environments.

LABORATORY OBSERVATIONS

The facies distinguished and discussed in the previous part of this chapter are subject of microscopic investigations the results of which will presently be reported.

The ooids
The percentages of ooids, according to point-counting, vary between 17 vol. % and 54 vol. % in the ooid beds. The average (1000 readings in five samples) is 29.3 vol. %. Ooids are normal or superficial. In the structure of both types concentric and radial arrangements of the crystals are present. The ooids can be spherical, rodlike (axiolitic) and sometimes irregular, due to replacings or budlike protuberances (Pl. I–3). In a few cases only a partial micritic envelope has developed, possibly as a result of primary sheltering effects. This happens, for instance, in some places in the northern part of the autochthonous area (Pl. I–4). The nuclei, mainly bioclastic fragments, sometimes show grain diminution phenomena (degrading neomorphism, Folk, 1965). In nuclei of crinoidal fragments selective hematitisation can often be observed.

The cement between the ooids is calcite, generally present as equant crystals, with straight intercrystalline boundaries (Pl. I–5, I–6). These, according to Bathurst (1958), indicate filling of open voids. Staining with a mixture of alizarin red S and $K_3Fe(CN)_6$ results in a faint mauve to purple colour for the calcite. This indicates ferroan calcite (Evamy, 1963, 1969). There are often concentrations of ferruginous material, presumably hematite, in the enveloping layer of the ooids (Pl. I–4). In the outermost layer of the ooids this can produce a dark 'halo' (Pls. I–5, I–6), as was observed in the northern part of the autochthonous area, and in section 17 in the vicinity of La Ercina (Pls. I–5, I–7). Outside some ooids (Pls. I–3, I–6) a light 'halo' of purely calcitic blade-shaped crystals is visible. The presence of dolomite is revealed by staining tests. Much is ferroan dolomite in composition. It occurs both as rhombohedral crystals and as patches in the oolite. The rhombohedral dolomite crystals occur mainly in the inter-ooid space (Pl. I–8). They often have a hematitic coating. Through tiny cracks in this coating the dolomite itself can be seen. The relation between the cement and the rhombohedral dolomite crystals often suggests a secondary origin of the latter. The patchy dolomite is irregularly distributed over the ooids and the cement. The relation between the patchy dolomite and the dolomite rhombohedra is uncertain, although simultaneous formation seems most logical (Pl. I–8).

A few authigenic quartz crystals are also present. They grow through nuclei, altered by neomorphic degrading processes (Folk, 1965), and through enveloping layers (Pl. I–8). Sedimentary quartz grains also occur. These are often somewhat corroded, and show replacement phenomena (Pl. I–5).

The pelletoidal, ostracodal, calcispheral, gastropodal and brachiopodal mudstone, wackestone and packstone: Facies b

Under the petrographic microscope with low magnification we see generally a very fine-grained rock (calcisiltitic to very fine fossiliferous micritic), but large fragments of corals and of stromatoporoids may be present. These generally show signs of transport. Pellets are common (Pl. V–2). They often show significant differences in size (Pl. XII–2). Their distribution sometimes accentuates laminated structures. In a few cases they were observed inside as well as outside fossil fragments (Pl. XII–3). Limeclasts (intraclasts and lumps) are also present (Pl. XII–5). In section 27 breccious fragments were observed, presumably related to an erosional surface (Pl. XIV–1). The commonest components in this facies, however, are the bioclasts. Ostracods, especially thin-walled ones, calcispheres, brachiopods, styliolinas, gastropods and unidentified debris of other organisms occur in abundance (Pl. V–3). Some significant microscopic features will be dealt with separately.

Voids. – The presence of small voids in a micritic or a fossiliferous micritic ground mass, and sometimes filled with both internal sediments and cement, sometimes only with cement, is characteristic of this facies (Pls. VIII–1, VIII–2). Calcispheres were often observed in association with these voids (Pls. VIII–2, VIII–3, VIII–4). The space inside the voids is filled with cement, consisting of sparry calcite with equant crystals sometimes having straight intercrystalline boundaries, sometimes curved ones. Dimensions of the crystals grade from fine to medium. Blade-shaped crystals partly coat the walls of the voids (Pl. VIII–3), in which case a light 'halo' is present.

In a few cases the voids are the result of selective solution of fossil fragments (Pl. VIII–1) and subsequent precipitation of cement. They are sometimes present along stylolites, filled with earthy material and with minor amounts of ferruginous matter. These voids are filled with medium equant calcite crystals without blade-shaped calcite crystals being present as a rim. Another origin for 'cavity-like structures' (Pl. V–4) is presumably replacement of the original ground mass, as is indicated by veinlets and fossil fragments that are still visible as ghosts in the cement. The 'cavity-like' sparry calcite patches consist of medium equant crystals with straight intercrystalline boundaries. No evidence was found of rim cement or of bladed crystals around the 'walls'.

The most interesting, however, are those voids with a basal sediment (Pls. VIII–1 and inset, VIII–2, V–3). This basal sediment is fine bioclastic and micritic material and in a few cases hematitic, and micritic pellets are present. It becomes thinner towards the walls of the voids and indicates depositional top and bottom. It must have been introduced mechanically after the formation of the cavity and before precipitation of the calcite cement. In one case (Pl. V–3) poorly visible parts of fenestrate bryozoans occur, floating in the cement which fills the remaining space in the void. The zoecia are still filled with micrite and calcisiltite. This suggests a selective internal solution origin for this void type (Orme & Brown, 1963; Dunham, 1969). Fractures cross sometimes the cemented voids (Pl. VIII–2) and postdate these.

Fragment diminution. – This term is used to indicate the continuously active demolition of larger fragments by various processes. Fragment diminution occurs in packstone, wackestone and mudstone. The fine-grained character of the majority of the rock fragments is conspicuous. Larger fragments present, not yet diminished in size, were evidently transported into the depositional environment where this diminution is common. They give an indication of the strength of the currents that are occasionally present. The contacts between the larger fossil fragments and the ground mass are generally stylolitic (Pl. V–3). The brachiopods in Pl. IX–4 are much larger fragments than the groundmass, and illustrate the character of the material, being demolished. They are sometimes only present in a ghost-like form, floating in the matrix. Many indications for scavengers were found (Pls. V–6, IX–7). These scavengers are presumably responsible for the demolition of the fragments. Burrow tubes are preserved, filled with dusty calcite cement and showing traces of lineations resulting from the locomotion of the burrowing animal (Pl. V–6). In some cases a high concentration of brown ferruginous material and tiny rhombohedral dolomite crystals occur preferentially in the burrow tubes (Pl. IX–7). Since no dolomite and only traces of ferruginous material were observed around these tubes, it is supposed that the ferruginous material, and the Mg^{2+} ions required for dolomite formation, were introduced more or less penecontemporaneously with the formation of this tube. The primary depositional structures are sometimes only slightly disturbed, indicating slow deposition or quick induration of sediment.

Algal coating. – In the southern part of the Esla autochthonous area a sample was collected in Member C (Pl. IX–5) with algal coated grains. The size of these (approximately 2 mm) and the crenulated appearance are widely accepted recognition criteria of algal coating (cf. Wolf, 1965b, p. 15, 36, and Bissell & Chilingar, 1967, p. 164). Finely crystal calcite is precipitated around the grains. Other bioclastic fragments (Pl. IX–6) also show traces of algal activity. Algae have penetrated, bored and encrusted the fragments after deposition.

Dolomitization and dedolomitization. – In a few cases dolomitization, dedolomitization and recrystallization obscure the depositional texture (Pl. V–5). Dolomite is present in two forms: as a patchy, irregularly distributed, anhedral replacement dolomite, preferentially replacing bioclastic fragments, and as subhedral to euhedral (= rhombohedral) dolomite crystals, preferentially present in the interparticle space. This latter dolomite especially requires further discussion. Rhombohedral dolomite crystals are sometimes present in great quantities. They are often coated with brownish material, presumably hematite, and can be slightly corroded. Hematitic material enters the rhombohedral crystals along fractures. Sometimes medium to fine crystalline mosaics of neomorphically calcite or dolomite crystals, attack these interparticle subhedral to euhedral (rhombohedral) dolomite crystals (Pl. XI–1). In other cases it is convincing that originally rhombohedral dolomite crystals have selectively been removed by leaching. In those cases only the rhombohedral shape of the pores remains. Evamy (1967) explained similar features by accepting a dedolomitization phase in which, prior to leaching, dolomite is changed into calcite. The pores can easily be rendered visible by moving the microscope tube. The surrounding neomorphic calcite or dolomite shows a relief difference and has often 'halos' of iron minerals. In those cases in which the surrounding material is dolomite, it is of the patchy anhedral replacing type (Pl. XI–3).

Recrystallization. – A characteristic feature of facies b is the very fine to medium granular calcisiltite which, in many places, is changed by neomorphism into very fine or fine crystalline calcite and sometimes into fine to medium coarse euhedral to subhedral dolomite crystals (Pls. X–8, XI–4, XI–6, and XII–4). The dolomite crystals as well as the calcite crystals (Pl. XI–5) have straight intercrystalline boundaries. Another aspect of this facies is the presence of drusy linings that coat voids and bioclastic fragments with calcitic and ferroan calcitic crystals. Sometimes a rim composed of neomorphically formed tiny equant calcite crystals is found around the bioclasts that occur in a ground mass of hematitic material. The crystals of these rims are all similar but they are reduced in size in comparison with other neomorphically formed crystals (Pl. XI–6).

Silicification of the original calcisiltite and of fossil fragments (Pl. XII–1) and filling of primary voids in fossil fragments by chert and hematitic material is common in this facies. Idiomorphic authigenic quartz crystals too were often found.

The coral, bryozoan packstone-boundstone with argillaceous interstitial material: Facies c

The commonest sediments in this facies are biosparudites and biomicrudites. Microscopically recognizable fossils do not differ from those already recognized in the field (Pl. VI–5). Texture is chiefly packstone, but in a few cases it is boundstone. In the Esla autochthonous area wackestone textures were occasionally observed. Sorting and roundness of the bioclastic material is extremely variable (Pl. V–7). Pellets are present in minor quantities in the Peña Corada area, but in the Esla autochthonous area the quantity of the pellets increases in significance. Due to fractures the appearance of this sediment in some places resembles breccias. Matrix is mainly calcisiltitic to calcilutitic bioclastic material (Pl. VI–6). Especially in the Alba syncline area, the interstitial material is often red siltitic to lutitic limestone, and contains dolomite and detrital material. This material is commonly trapped by binding animals, or it is a result of sheltering effects (Pls. VI–5, VI–7). Fairly large amounts of detrital angular to subangular quartz grains and authigenic quartz crystals were found. The authigenic crystals can be partly replaced by calcite (Pl. VI–4). These crystals are always present in the vicinity of stylolites and/or fractures in the rock. They are surrounded by a film of hematite. The parts of the authigenic quartz crystals not yet replaced show a straight extinction.

Cement between the bioclasts is both purely calcitic and ferroan calcitic. Voids in larger fossil fragments or voids brought about by sheltering effects are filled with drusy calcite crystals (Pls. V–7, VI–2). In the centres of these voids equant medium (rather blocky) crystals are present (Pl. VI–8). Inside and outside of a primary void (e.g. an ostracod) the cement can display a different aspect (Pl. XIII–4). Growth zones and cleavage in the calcitic, ferroan calcitic, dolomitic and ferroan dolomite crystals, are often visible. Matrix is occasionally recrystallized into pseudosparite (Pl. V–8). Recrystallization attacks mainly sedimentary and textural inhomogenous parts of the rock and partly destroys the depositional packstone and wackestone texture (Pl. XII–4). Grain growth (syntaxial rim cementing) is common (Pl. VII–1). Geopetal structures and (pressure?) solution phenomena are typical of this facies. The larger bioclasts have stylolitic contacts with other fragments and with the matrix. In the stylolites hematitic detrital particles (Pl. VI–8) and earthy material (Pls. VI–3, IX–8 and X–1) are often present. Dolomite also shows a close relation to stylolites.

Dolomite is present as intergranular, hematite-coated, rhombohedral crystals and as subhedral to anhedral crystals, replacing bioclastic material. The mainly fine calcarenitic and calcisiltitic material, in between the larger bioclastic fragments, is often attacked by dolomitization processes, resulting in rhombohedral dolomite crystals. These crystals often have three dusty hematite zones (Pl. VII–1), alternating with rather light zones. They are interpreted as growth stages of the crystals. The centres of the crystals often consist of (ferroan) calcitic material. Evamy (1967) explains similar features by assuming dedolomitization. The original rhombohedral crystals now often have irregular surfaces, possibly due to replacement (Pl. VI–3). Rhombohedral dolomite crystals occur preferentially in very fine interstitial bioclastic material or in very fine mosaic to microcrystalline pseudosparite to microsparite (Pl. XII–6). They are seldom found in larger bioclastic fragments, in contrast to the patchy dolomite. Dark brownish hematite coated rhombohedral dolomite crystals, surrounded by clear white rims of fine sparry calcite, were observed (Pls. XIII–2, XIII–3). The clear white rims were formed after formation of the dolomite crystals and after the associated 12.1 volume percent shrinkage.

Silicification, too, is fairly common in this facies. Fossil fragments (Pls. XII–7, XIII–1) are preferentially silicified (especially the corals), but in the interstitial material patchy chert also occurs though less commonly. Chert often contains perfectly preserved rhombohedral dolomite crystals, possessing straight intercrystalline boundaries when the rhombohedral dolomite crystals have no hematitic coating. Otherwise the intercrystalline contacts are serrated (Pl. XII–8).

The coral, stromatoporal, boundstone-packstone with a biohermal aspect: Facies d
Biohermal structures are primary field features. Detailed mapping and measuring of stratigraphic sections is required for revealing them. The faunal composition on a microscopic scale does not differ from that on a macroscopic scale. Texturally we are mainly confronted with packstones and boundstones.

Dolomitization and dedolomitization often occurs. Rhombohedral dolomite crystals of the same type as in facies c occur (Pl. VII–2). Calcite replaces parts of some rhombohedral dolomite crystals, while quartz crystals, still vaguely displaying idiomorphic shapes, have most probably grown in situ and have been replaced by dolomite. The calcite replacing the dolomite is occasionally leached. The seams containing the largest degree of dedolomitization coincide with the erosion levels described (p. 183). The ground mass in which the dolomite crystals are present mainly consists of slightly recrystallized bioclastic material and pseudosparitic to microsparitic patches, the mosaic or the granular crystals of which show curved and irregular intercrystalline boundaries.

The encrinal (encrinite) bryozoan grainstone: Facies e
Grains are generally bioclastic in facies e. They are well rounded (partly because the original skeletal material was spherical) and generally fairly well sorted. The encrinal (Pl. III–1) or encrinic (Pl. III–2) appearance is typical of this facies. Sometimes much detrital well-rounded siliciclastic quartz grains are concentrated in stylolites (Pl. III–1). This material is concentrated presumably according to the mechanism supposed by Trurnit (1967, p. 189; 1968a and b). Stylolitic contacts between bioclastic fragments, possibly the result of pressure solution, are common.

A general characteristic of this facies is the grain supporting texture (Dunham, 1962). In the field and under the binocular microscope certain samples could be recognized as grainstones, while during point-counting and examination under the petrographic microscope it became clear that up to 5 vol. % of micrite and very fine-grained material is present. In those cases the field determination was maintained. In the grainstones destroyed bioclastic fragments were often observed which, just as the very fine material can be explained as the result of scavengers' activity. The grains in this facies are cemented with calcite which in a few places is slightly ferruginous. Syntaxial overgrowths are fairly common (Pl. III–2). Patches of irregularly distributed mosaic cement sometimes occur.

The encrinal, bryozoan, coral, brachiopodal packstone and wackestone, with marly intercalations: Facies f
Significant amounts of bioclastic and pelletoid grains occur in this facies. Their dimensions grade from fine to coarse. Sorting is poor. In some cases bioclasts are difficult to identify, but elsewhere crinoids, bryozoans, brachiopod fragments and ostracods were recognized. There are examples of lamellae, composed of pellets, alternating with bryozoans which trap and bind the pelletal material (Pl. III–3). Bioturbation commonly occurs (Pl. III–4). The matrix of these sediments consists of very fine-grained bioclastic debris, and is often fairly argillaceous (Pl. III–4). Detrital siliciclastic grains are present, as well as idiomorphic authigenic quartz crystals (Pls. III–4; III–6). Tiny calcite crystals occur, formed by aggrading neomorphic processes (pseudosparite, Folk, 1965). These neomorphic patches can cross depositional structures (Pl. III–5). Recrystallization may have changed the original textures significantly, and the depositional texture becomes uncertain (Pl. XIII–5). Dolomitization occurs, resulting in both subhedral and rhombohedral crystals (Pl. III–6). In a few cases dolomite crystals, pseudomorph after idiomorphic quartz (visualized by staining), are present, together with rhombohedral dolomite crystals which are coated with hematitic material (Pl. III–7). Sometimes stylolites form the contacts between the bioclasts, and fractures, presumably brought about by pressure, are present. In one case a slightly cleaved rock was produced with flattened particles, showing orientation. The oolite also crosses facies f (Pls. IV–1, IV–2). The radial fibers of the ooids display a preferential dolomitization.

The bioclastic packstone and wackestone, admixed with significant quantities of siliciclastics: Facies g
This facies is present in limestones and in siliciclastics. From the latter, samples for investigation were difficult to collect because the sediment consists mainly of very fine-grained and argillaceous, loose, material (Pls. II–5, II–6, II–8). The siliclastic part of facies g was investigated qualitatively in 15 samples. The matrix contains plagioclase and microcline, and there are also some rock fragments; chiefly complex quartz and clay pebbles. Muscovite is also present in the matrix. The quartz grains belong to the fine-sand fraction, and are subangular. This is partly due to the generally small size of the grains, partly to the replacement by calcite. Sorting is poor, presumably due to the effects of bioturbation. The results of the investigation of these 15 samples are presented in Table 3. The limestone component of facies g is mainly packstone. Recognizable bioclasts are bryozoans, crinoids, ostracods, brachiopods and gastropods, and, especially in those areas where the limestones are admixed with significant quantities of siliciclastics, zaphrentoid rugose corals, blastoids, various ramose and fan-shaped bryozoans and various colonial corals occur in addition. Tentaculites and trilobites are present in minor quantities (Pl. VII–3). Pellets and burrowing structures occur (Pl. III–8). The bryozoans

A: Primary components	B: Secondary components	C: Accessories
1) Quartz 2) Matrix: Plagioclase Microcline Muscovite Biotite (?) Rock fragments: Complex quartz Clay pebbles	1) Calcite 2) Quartz 3) Titanite/Leucoxene 4) Iron minerals: Limonite (?)	1) Titanite/Leucoxene 2) Zircon 3) Tourmaline 4) Iron minerals: Limonite Hematite Goethite (?) (red-brown) Magnetite (?) (black) 5) Uncertain green mineral [Diopsid (?), Pyroxene (?)]

Table 3. Primary, secondary and accessory components in the siliciclastic parts of facies type g, qualitatively investigated from 15 samples (by Mr. G. Gietelink).

sometimes trap the fine-grained material. Sheltering effects by larger fossil fragments are common. The ground mass contains a great amount of ferruginous material. Many bioclastic particles show corrosion features, in most cases presumably due to physico-chemical processes, in one case due to boring and encrusting effects of blue green algae (Dr. J. J. de Meyer, pers. comm., 1971; Pls. VV–4, VII–5; cf. Chilingar et al., 1967, Pls. XV and XVI).
The encrusting algal material is micrite-like and in the bioclastic material tiny boring holes are visible (Pl. VII–4). The interstitial material is blocky calcite, mainly with straight intercrystalline boundaries (Pl. VII–5). The irregular boundaries between the intergranular calcite and the coated bioclastic grains prove that the calcite is a primary void-filling precipitate. Recrystallization has slightly altered the texture. The overall picture is that of a coated-grain grainstone/packstone (Pl. VII–6). In a few cases the coating and replacement has intensively altered the originally bioclastic material (Pl. X–2). Algal micritic pellets were then formed. In the same facies we find replacement rims and dolomite rhombohedral with a yellowish brown colour and calcite embayments, proving that dedolomitization processes have taken place (Pl. VII–3). Some rhombohedral dolomite crystals are partly replaced by ferroan calcite and have a worn appearance (Pl. X–4). Ooids and pseudooids occur in some sections. Siliciclastic material, sometimes very well rounded (Pl. X–2), is mainly concentrated in stylolites and in depositional lamellae. Contacts between the clasts are often stylolitic (Pl. VII–3).

Transitional facies
Sometimes transitions were observed between the facies discussed above. A few of these merit a separate discussion.

Transition between facies b and f. – Section 7; Member B (Pl. VII–7).
The faunal composition in the sample, pictured in Pl. VII–7 is transitional. Rugose coral fragments, fenestrate bryozoan fragments, ostracods and crinoidal material are present together. Sorting is poor. Grain size is generally medium, in a few points it is coarse but, due to secondary effects, it can sometimes be fine (fragment diminution?). Texturally Pl. VII–7 shows a packstone. Zoned rhombohedral dolomite crystals are present (Pl. VII–2). From the outside to the centre we see a clear brown coating, a pure white band, a dusty band that is light blue after staining with a mixture of alizarin red S and $K_3Fe(CN)_6$, indicating a ferroan dolomite (Evamy, 1963, 1969), and in the centre we finally find a Turnbull's precipitate, indicating a ferroan dolomite with more than 50 % Fe^{2+} ions. Dedolomitization processes, too, have taken place. The corroded aspect of the rhombohedral dolomite crystals is the result of partial replacement of dolomite by calcite. The rhombohedral dolomite crystals are present in a ground mass of fine calcarenitic to calcisiltitic particles, possibly the result of fragment diminution.

Transition between facies c and d. – Section 20; Member B. In the field we recognized the same transition. Bioclastic material is crinoidal and ostracodal. Micro-laminations are visible and are accentuated by an alternation of bioclastic material, ooids and siliciclastic material. The ooids have no nucleus and the crystals in the layers have a radial orientation. The siliciclastic material is subangulair and in a few cases even idiomorphic. Some quartz crystals are authigenic. Granular cement is present in the interstices and sometimes in the centre of the voids, the walls of which are coated with fine-bladed crystals. Two generations of stylolites are present partly filled with ferruginous matter.

Transition between facies b and g. – Section 26; Member B. Bryozoans, crinoids and some fragments of trilobites were identified. Some ooids and pellets are present. Silicified fossils and bioclasts are common. Rhombohedral, corroded, dolomite crystals are poikilotopically present in calcite and in chert (Pl. VII–8).

CHAPTER V

A REVIEW OF PROCESSES

In a general way a sediment, in its development during and after deposition, undergoes successively: (1) biological and biochemical processes, (2) physicochemical processes, and, (3) physical processes (Chilingar et al., 1967, p. 246). These overlap in time and, with increasing time, these decrease in importance for reconstructing depositional conditions. According to Wolf and Conolly (1967), syngenetic, diagenetic and post-diagenetic stages can be recognized in the history of a sediment after deposition. This division has been slightly modified in the present study (Table 4). The processes, overlapping in time, can be related to one or more of the stages. Syngenetic and early diagenetic ones (prior to cementation) are controlled by surface and near-surface factors, and largely reflect the depositional environment (cf. Chilingar et al., 1967, p. 250; Krumbein & Sloss, 1963, p. 266, 267; Nagtegaal, 1969, p. 227, 228).

In the facies** recognized, the processes have been investigated and interpreted. This has been carried out in representative thin sections. Of facies b twenty-three thin sections were investigated in detail, of facies c eleven, of facies d one, of facies e nine, of facies f eleven and of facies g four. Where possible a differentiation was made for the four different structural areas. Facies b was investigated in the four areas; facies c in three different areas (not in the Esla autochthonous area), facies d only in the Peña Corada area (in Member D), facies e in the four areas recognized, facies f in the Esla autochthonous area and in the Peña Corada area, and facies g in the Esla autochthonous area and in the Alba syncline area.

In Enclosures V and VI the relative frequency of occurrence of a process is expressed by the thickness of the line. The relative time interval during which the process acted is indicated by the range of the line. No attempt has been made to express real time intervals. In designing the lines, experience gained in examining approximately 1200 samples acted as a guide.

** The oolite has characteristics of facies e, f and g because the ooid beds are present in these facies without preference. Therefore the ooids, although discussed separately on their characteristics in the preceding Chapter, will be regarded as belonging to one of the three mentioned facies.

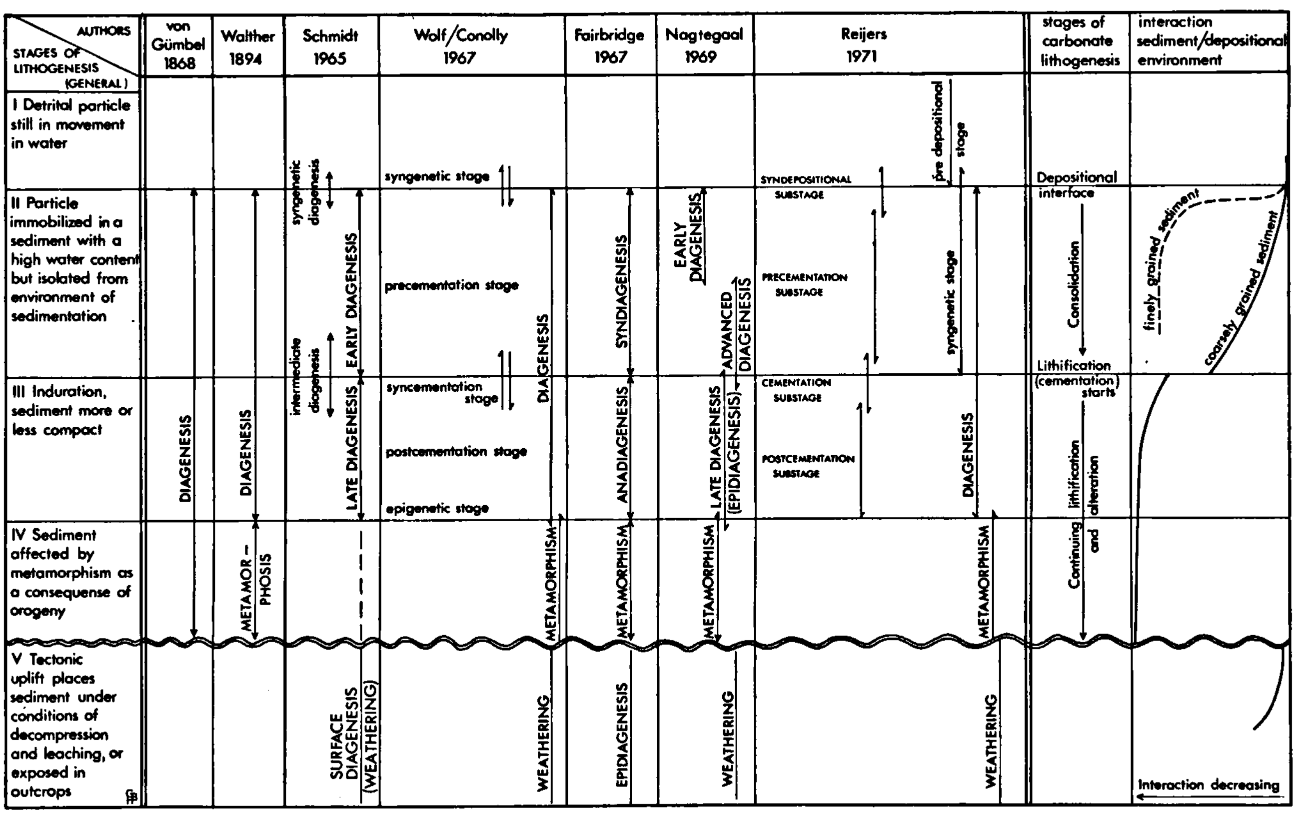


Table 4. Stages of carbonate lithogenesis recognized in present study and compared with those of other authors.

Vestiges of similar processes in various thin sections from one facies in one member in one of the four structural areas could be correlated with a fair degree of certainty, and were pinpointed on a certain relative time interval. This is carried out in Enclosures V and VI in the column showing the 1st level of uncertainty. This element of uncertainty is introduced because processes, recognized in thin sections strictly speaking, are only of local value as has been pointed out by Chilingar et al., (1967, p. 246). The similarity of processes in identic facies over the whole area investigated, however, make the correlation reasonable.

A correlation between processes acting in similar facies in more than one member in one area is established in the column showing 2nd level of uncertainty. Here an extra time element, the constancy of the processes, is introduced. No more levels of uncertainty were used in the Enclosures.

After separate discussion of some of the most significant processes an environmental interpretation will be presented, in which processes, more or less significant for a certain facies, are tabled and discussed in relation with the facies.

BIOCHEMICAL AND ORGANIC PROCESSES

A division into biochemical* and organic processes will not be made. Bioturbation, pellet formation, algal activity (mainly encrusting; some boring) and biochemical disintegration of bioclastic material have been recognized.

Bioturbation and formation of pellets. – (Pls. III–3; III–4; V–2; V–6; V–7; VI–6; VII–5; VII–8; VIII–1; VIII–2; VIII–3; VIII–4; IX–6; IX–7; IX–8; X–4; XI–6; XII–2; XII–3). Biogenic microstructures are common in the entire Portilla Limestone Formation, but occur preferentially in the b, f and g facies sediments. In a few cases burrowing disturbs the sediment fairly late in the precementation stage. Tubes, made by the burrowing organisms, can then be preserved in a slightly indurated sediment. The locomotion of the burrowing animal can sometimes be proved, since lineations and drags are preserved (Pl. V–6). In general, burrowing occurs in the syndepositional stage. Primary authigenic iron minerals and rhombohedral dolomite crystals are often selectively accumulated in these burrow tubes (Pl. IX–7).

Pellets are extremely common, both in recent environments as reported by various authors and in ancient ones (Beales, 1965, among others). They have a polygenetic origin and can among other possibilities be formed by organic (sometimes faecal) agglutination (Illing, 1954), by inorganic precipitation and by cementation. The combination of pellets and bioturbation structures is a common one in some limestones encountered in the Portilla Limestone Formation (Pls. V–6; X–4). It is therefore concluded that most of the pellets are of organic origin (faecal). Pellets also occur often in bird's eyes limestone (Pls. VIII–1; VIII–2; VIII–3 and VIII–4), where they may be related to algae and/or to decayed organic material. Gas bubbles, as a result of decaying, could form the voids that were subsequently filled with cement and that are nowadays present as bird's eye vugs. Formation of pellets is assumed to have taken place at some moment during the precementation stage or, more specifically, during the syndepositional substage.

Algal coatings. – Algal origin of micritic material (Pls. IX–5; IX–6; cf. Wolf, 1963, and 1965a, c) may be inferred from the attacked appearance and the wavy surface of some coated grains. Sometimes only the 'attacked' appearance points towards algal activity. Coatings presumably are formed in the syndepositional substage.

The processes discussed previously often result, to some extent, in disintegration of carbonate particles. Various other processes, not dealt with or recognized here, show similar results. Moreover, products of biochemical disintegration processes are often difficult to separate from products of mechanical disintegration processes and of physicochemical disintegration processes. Together or separately these processes can be important in mud formation.

PHYSICOCHEMICAL PROCESSES

This group is subdivided into a group of processes resulting in authigenic products, and into a group of processes not leading to newly formed minerals but nevertheless physicochemical. Authigenesis (Kalkowski, 1886) is the formation of any newly formed or secondary mineral. Any mineral formed in situ is referred to as authigenic and any process producing such a mineral is an authigenic process.

Physiochemical processes recognized are indicated on Table 5.

Authigenesis of chert (Pls. VII–8; XII–1; XII 7 and XIII–1) *and quartz.* – (Pls. I–7; I–8; III–6 and VI–4). In the field chert is present in nodular patches of varying dimensions. These are often concentrated in layers. Repetitions of up to six or seven times are common. Another form in which chert occurs is as preferential silicification of fossil fragments, especially of ramose and platy tabulate corals and of stromatoporoids. Both types of chert occur most frequently in facies c (Pls. XII–7 and XIII–1). Rhombohedral dolomite crystals may be present in a silicified matrix (Pl. VII–8). They have irregular outlines because dedolomitization attacked the crystals moderately to severely. Sometimes the calcitic groundmass is selectively replaced by chert, while

* 'Biochemical' is used in the sense of Chilingar et al., 1967, p. 184.

authigenic processes	non-authigenic processes
Precipitation of quartz and chert Precipitation of primary ferruginous material Precipitation of mud (aragonite, calcite) ↓ micrite ↙ ↘ microsparite → pseudosparite Formation of syntaxial cement rims Precipitation of cement → fibrous and blade-shaped crystals Precipitation of cement → equant crystals Dolomitization: Rhombohedral crystals → dedolomitization Replacements → Patchy dolomite → dedolomitization Replacements → Quartz → Calcite Replacements → Calcite → Quartz	Formation of ooids Coating of particles by ferruginous material Physicochemical disintegration of bioclastic material Microfractures and stylolites Expulsion of impurities

Table 5. Authigenic and non-authigenic processes recognized and discussed in present study.

dolomite is unaffected, and is selectively present in non silicified places (Pls. XII–1; XIII–1). Dusty zones, visible in this chert, indicate the outlines of replaced fine grains or fine crystals of the ground mass. From the sizes of the replaced fine crystals it becomes probable that some micrite → microsparite recrystallization occurred before silicification. Dolomitization then generally postdates silicification. Sometimes, however, it coincides with it, as can be seen in Pl. XII–1 where impurities expelled by the growing rhombohedral dolomite crystals are present in the surrounding chert. With regard to the place of chert formation, Sleumer (1969, p. 24) came to the conclusion that silicification in stromatoporoids took place before complete lithification of the surrounding sediment. This may indicate a near surface environment.

Fairbridge (1967, p. 58) discussed the origin of silica in marine sediments. Dust from deserts, volcanic glasses and organic remains are less likely to be responsible of silica in the Portilla Limestone Formation than liberated soil silica from a hinterland. Desert dust and volcanic glasses are not probable, because no palaeogeographic indications have been described, till so far, of deserts or of volcanism in the Devonian of northwestern Spain; organic remains are not probable, because no traces of Radiolaria, diatoms and/or sponge spiculae have been seen. The supply of silica gels from soil-forming processes into the shallow seas in which it was stabilised on certain places, has been discussed by Fairbridge (1968, p. 58), who also lists a series of examples, that such silica gels at times other than the present, are numerous. The question of the mode of stabilization and the preference for certain places, poses problems. In the Portilla Limestone Formation especially coelenterates are often silicified (Pls. XII–7, XIII–1). Sleumer (1969, p. 24), discussed the silicification of stromatoporoids and concluded that a relation could exist between silicification and structures of the coenosteum such as latilaminae, holes and the surface of the coenosteum before burial in the sediment (cf. his Figures 13–2, 13–4, 13–6 and 13–7). It is thought that during life the entire surface of a coenosteum was covered with living tissue (Shrock & Twenhofel, 1953, p. 113). After death this decaying tissue could produce pH-lowered surroundings in which chert preferentially coagulated. This possibly applies also to other forms of silicification (cf. Fairbridge, 1967, p. 59). Other (invisible) organic remains could serve as coagulation nuclei for patches of chert (Pl. VII–7) occurring today as bedded nodular chert and not displaying any visible relation with large fossil fragments. Coagulation must be succeeded by a period of dehydration in order to obtain chert as we find it today. Dehydration could occur in short periods of emergence as described by Rutten (1957, p. 436), which periods possibly also caused the dedolomitization previously mentioned.

Apart from chert, idiomorphic authigenic quartz crystals are present in the Portilla Limestone Formation (Pls. I–7, I–8, III–6 and VI–4). Evaluation of euhedralism may indicate growth in pore spaces or replacement (Siever, 1959). Where inclusions occur, a replacement origin is most likely. This is the commonest form observed (cf. Sleumer, 1969, p. 25 and his Figures 14–3, 14–4, 14–5). The quartz crystals are often, more or less, replaced by calcite or by dolomite. Walker (1960, 1962) related replacement processes to the precipitation of chert. The replaced silica could serve as the source of the silica required for silicification. This relation cannot be proved for the Portilla Limestone Formation, and if it could be proved it would be insufficient to explain the quantities of silica required for silicification.

Precipitation of authigenic ferruginous material. – (Pls. III–4; IX–7; X–8; XI–2; XI–6; XV–1; XV–2; XV–3; XV–4; XV–5; XV–6). Colloform, clotty aggregates (Pls. IV–1; IV–2) and very fine particles, both grains and crystals, of ferruginous** matter (Pls. IX–7; XI–2) are commonly present in a dispersed way in the sediments of the Portilla Limestone Formation, which results in a faint reddish-brown colour. X-ray analysis revealed that it mainly consists of amorphous material, but in the diffraction pattern some peaks may indicate the presence of goethite in small quantities (Mr. R. O. Felius, pers. comm., 1971). Electron microscope photographs (made by Mr. W. de Priester) revealed idiomorphic (authigenic) crystals of extremely small size, which are often present in relation with clotty aggregates (Pl. XV). This ferruginous material occurs as coatings in and around ooids (Pls. I–5, I–6 and I–7) and bioclastic grains (Pl. X–8), and large quantities preferentially occur in burrow tubes (Pl. IX–7). A syngenetic formation therefore is evident. It can also be visualized in those cases in which other crystals by their growth push aside the finely dispersed ferruginous material. This results in 'halos' (Pls. I–5; VI–4). According to Schellmann (1959), as quoted by Nagtegaal (1969), hematite crystallizes at low pH values and, under the influence of absorbed positive ions (Mg^{2+}, Ca^{2+}), at neutral pH values.

From the foregoing the presence of a mobile form of ferric hydroxide gel can be deduced. This most probably passes through a dehydration phase in which the clotty colloform habitus of hematite aggregates was formed. Some authigenic goethite crystals are still present (Pl. XV). The iron required for the formation of the ferric hydroxide gel is most probably derived from a weathered hinterland.

A few differences between environments of deposition of silica, and of authigenic ferruginous material must be discussed. Chert and quartz can be formed both under reducing and oxidizing conditions; ferric oxides will generally be formed under oxidizing conditions. Green layers, containing mottled and bleached patches and pyrite, alternate with reddish layers in various facies in the Portilla Limestone Formation. The former character indicates reducing conditions whereas the latter character is typical of oxidizing conditions. Dehydration of the ferric hydroxide gel, resulting in the hematite aggregates, occurred in periods of slight emergence. The quantity of ferruginous material was visually estimated for all samples investigated and some samples were chequed by point counting. The estimates are indicated in Enclosures I–IV.

** 'Ferruginous' is used in a general sense for all iron minerals. Ferruginous material with regard to its origin can be divided into two groups; detrital ferruginous material (Pls. VI–6; IX–6 and XIII–4) and authigenic minerals. The detrital material nearly completely consists of hematite. The authigenic minerals were originally (hydrous or non-hydrous) iron oxides, e.g. goethite. Now these, too, are mainly hematite, but goethite can still be present (Pl. XV).

Lime-mud, micrite, microsparite and pseudosparite. – The origin of lime-mud is the subject of detailed investigations by many investigators. In general two origins are discussed: direct chemical precipitation caused by inorganic or by organic processes on one hand, and disintegration of certain types of algae or coccoliths, or abrasion of non-algal skeletal parts on the other hand. The disintegrated organisms can be composed of aragonite or of calcite (Wolf et al., 1967b; Graf, 1960). Modern muds are chiefly or entirely composed of aragonite. Folk (1965), among others, discussed the formation of micrite, microsparite and pseudosparite. In the preceding chapter this process was often loosely referred to as recrystallization. Now these processes have to be strictly separated and the 'dating' in the paragenetic sequence is our object. Chilingar et al., 1967 summarized diagnostics of these products.

(1) Micrite consists almost uniformly of calcite. Lime-mud is mainly precipitated as aragonite. Aragonite to calcite inversion sensu Land (1967, p. 915) is therefore supposed to be an important process before real micritisation begins. Micritisation is supposed to have taken place by elimination (solution) of the smallest (inverted) calcite grains. Larger grains grew, possibly starting from widely spaced centres, and replaced the original mass of grains and pores (porphyroid neomorphism, Folk, 1965, p. 35, 36). At the end of micritisation the original mud particles are completely digested; the rock is lithified, the pores are mainly sealed and the precementation stage is past.

(2) Microsparite (Pls. XI–5; XII–1; XII–5; XII–8), is either directly produced by coalescive processes, from calcite (original calcite or an inversion product of aragonite), or it is directly produced aragonite microsparite which subsequently inverts into calcite (Folk, 1965, p. 41). The present author agrees with Folk that the former process appears more logical. Coalescive processes occur in the solid state, probably with the aid of interstitial solutions (Folk, 1965, p. 40). This places micro-sparitization in the cementation stage of the paragenetic sequence. The main characteristics of microsparite is the uniform size (Pls. XI–5; XII–8) and the simple loafish shape of the crystals (Pls. XII–1). Probably there is a relation between this process and the presence of shale (Folk, 1965, p. 39).

(3) In the chain of neomorphism as described here the final product may be pseudosparite crystals, the result of continued coalescive processes from microsparite, and it

must therefore be concluded that the formation of pseudosparite is active late in the cementation stage or in the postcementation stage. Pseudosparite may also be formed directly by inversion of organic skeletons. This may occur at any moment in the diagenetic history. Since it happens only exceptionally (Pl. XII–3) it is given no special attention in the present work (Pls. I–5; V–4; V–8; IX–4; XI–1; XII–3 and XII–4). It is difficult to identify it with certainty because of the striking similarity to normal pore-filling cement. Criteria for recognition are (1) transsection of allochems (Pl. V–8), (2) occupation of large areas unsupported by allochems (Pls. V–4; XII–3; XII–4) and (3) the presence of undigested inclusions.

Syntaxial rim cementation; syntaxial grain growth rims. – (Pls. III–1; III–2; III–6 and XIII–6). Both diagenetic fabrics are present in the Portilla Limestone Formation, the former being the commonest. The clean appearance, the fairly planar boundaries between the rim and the adjacent cement (Pl. III–2), the planar intercrystalline boundaries and the rather equidimensional mosaic, resulting from the syntaxial growth around well-sorted detrital particles (Pl. III–1; III–2), as well as the indication that the cement must have passed through a fluid state (Pl. XIII–6), point towards a direct precipitation origin (Chilingar et al., 1967). Evamy & Shearman (1965) defended the opinion that syntaxial rim cements are the result of an early phase of cementation. The presence of iron-bearing and iron-free zones (Pl. XIII–6) shows that solutions must have played a role, and supports this. In a few cases (Pls. III–5; III–6; IX–5) the rimming material and the host particles, sometimes partly rimmed by cement, show irregular contacts with the (recrystallized) matrix. In that case we are dealing with syntaxial grain growth rims. The syntaxial rim cement occurs most in facies c and e; the syntaxial grain growth rims occur more in facies f and g. This is in accordance with the decreasing susceptibility to grain growth for the mudstones, mudstone-pellets and oolite (cf. Bathurst, 1958, p. 31). Formation of syntaxial grain growth rims starts as soon as the solution-deposition process had produced adequate intergranular boundaries and a sufficiently low porosity (Bathurst, 1958, p. 30, 31). In terms of the paragenetic sequence here dealt with, syntaxial grain growth rims are found more or less simultaneously with micritisation, e.g. at the end of the precementation stage.

Precipitation of cement. – (Pls. I–3; I–6; III–3; V–3; V–7; VI–2; VIII–1; VIII–2; VIII–3; IX–4; XII–2). Precipitation of primary cement or 'orthosparite' (Wolf, 1963), 'directly precipitated calcite' (Folk, 1965) and 'granular cement' or 'drusy mosaic cement' (Bathurst, 1958) is the opposite of 'aggressive' formation of cement by which a carbonate sediment or a non-carbonate sediment is replaced (pseudosparite; Wolf, 1963). The morphological variations in the (fibrous, bladed and equant) crystals reflect different generations of cement-precipitation in the cementation substage (e.g. Pls. I–3; I–6; VIII–3). These morphological variations are a useful instrument in establishing detailed paragenetic relations (Chilingar et al., 1967). In the crystal lattices of calcite trace cations, e.g. ferrous ions can be revealed by staining (Evamy, 1963, 1969; Neal, 1969).

Rhombohedral dolomite crystals. – (Pls. I–8; V–5; VII–1; VII–2; VII–3; VII–7; XI–1; XI–2; XI–3; XII–1; XII–6; XII–8; XIII–1; XIII–2; XIII–3). Dolomite can be primary (syngenetic, or penecontemporaneous) or secondary (diagenetic) in origin. Primary dolomite has not been recorded in the sediment investigated. Among the diagenetic dolomite a morphologic division can be made into euhedral crystals (rhombohedra) and into patches of subhedral to anhedral replacement dolomite crystals. Relics, included during growth, prove that the rhombohedral dolomite crystals grew in situ (authigenic; first cycle product). Moreover, authigenic ferruginous material out of the ground mass may be pushed aside by the crystallization forces of the crystals, and may form a hematite film around the crystals. In cases in which the rhombohedral crystals have a worn aspect, this was caused by later processes, e.g. dedolomitization.

Dolomites owe their origin to hypersaline conditions; primary (syngenetic) dolomite to hypersaline depositional environments, secondary (diagenetic) dolomite to hypersaline interstitial fluids. In the sediments investigated often a close connection was observed between dolomitization and fractures and/or stylolites. It was also often noted that dolomite displays a preference for biostromal layers (facies c). Possibly the fractures and/or stylolites, after sealing of the sediments, created easily passable seams, through which the hypersaline fluid could enter. The sediments in the biostromal layers had presumbly an original porosity, partly due to pores in fossils (cf. Pl. VII), which was greater than that in the surrounding sediment. After disappearance of the porosity of the surrounding layers during and after cementation, porosity still could occur in the biostromal layers. Hypersaline fluids here were concentrated and caused preferential dolomitization. A syncementation to postcementation formation can be assumed for the formation of the rhombohedral dolomite crystals.

Replacements. – Opposed to authigenic processes resulting in products directly precipitated or formed in situ, stand authigenic processes resulting in products also formed in situ but in a more indirect way, viz. by replacement of formerly present material. Examples of the first type of authigenesis are: formation of idiomorphic quartz crystals, of chert, of primary ferruginous material, of cement, of syntaxial rim cement and of rhombohedral dolomite crystals. Examples of the second type of authigenesis are: inversion of aragonite to calcite; formation of syntaxial grain growth rims, of microsparite and pseudosparite, of patchy anhedral dolomite crystals and subhedral dolomite crystals, of 'dedolomite', and silicification and desilicification. Formation of dolomite (the patchy type), of dedolomite and of calcite replacing quartz will be discussed presently.

Patchy dolomite. – (Pls. I–8; III–7; IV–1; IV–2; V–8; XI–4; XI–5; XIII–2; XIII–3). Replacement either occurs molecule by molecule, involving 12.1 volume percent shrinkage (Pls. XIII–2; XIII–3), or volume by volume. The former type is only recognized where after shrinkage, clear and fine calcite crystals were precipitated, that are now present as a bright 'halo' (Pls. XIII–2; XIII–3). The latter type shows no volume shrinkage. Replacement dolomite crystals may have rhombohedral shapes (Pls. I–8; XIII–2; XIII–3), this being the more exceptional case. They are generally anhedral to subhedral crystals, clustered in irregularly distributed patches. Indication of replacement (Friedman & Sanders, 1967; Chilingar et al., 1967) can be found in:

(1) Skeletal remains, now composed of dolomite, which are known to have been originally secreted as aragonite or as calcite.
(2) Dolomite crystals containing relics of distinctive calcium carbonate particles (Pl. XI–5) or present in originally calcium carbonate particles, e.g. ooids (Pls. IV–1; IV–2).
(3) Irregularly distributed dolomite patches in limestones (Pls. V–8; XI–5).
(4) Dolomitization of idiomorphic quartz crystals replaced by calcite (Pls. III–7; XI–4).

The hypersaline fluid, responsible for the precipitation of rhombohedral dolomite crystals, was presumably also responsible for the formation of the patchy dolomite here discussed. From various samples (e.g. Pl. XI–5) it is clear that replacement occurred after cementation.

Dedolomitization. – (Pls. VII–2; VII–7; VII–8; X–3; XI–1; XI–2; XI–3; XII–4; XII–6; XII–8). Processes in which solutions with a high Ca^{2+}/Mg^{2+} ratio react with dolomite to form calcium carbonate dedolomite (cf. Evamy, 1967) are often referred to as dedolomitization processes (Evamy, 1967; Friedman & Sanders, 1967; Folkman, 1969; Smit & Swett, 1969). De Groot, (1967) published interesting laboratory observations, strongly suggesting that dedolomitization is a near-surface process. The significance of dedolomitization for regional interpretations was emphasized in studies by Schmidt (1965) and Folkman (1969). Considering their findings, considering the presence of intraformational planes of discontinuity, sometimes of minor importance (p. 183), in several places in the Portilla Limestone Formation, and considering the presence of cherty beds (presumably the products of short periods of emergence) in which rhombohedral dolomite crystals can partly be dedolomitized, the present author agrees with de Groot, Schmidt and Folkman, and also relates the dedolomitization processes in the Portilla Limestone Formation to sub-areal exposures. Leached dedolomite (cf. Evamy, 1967, Fig. 9) can easily be recognized in some samples of the Portilla Limestone Formation (Pls. VII–8; XI–2; XI–3), and results in pores. It can be related to weathering. With regard to the moment of dedolomitization all that safely can be said is that it is weathering, which can occur from the syncementation substage on, until very late in the postcementation substage.

Replacement of quartz by calcite. – (Pls. I–7; VI–4). The precipitation of quartz and of chert was possibly facilitated by local nuclei of lowered pH value during the initial stage of syndiagenesis, as is argued elsewhere (p. 191). Krauskopf (1959) produced a curve in which solubility is compiled as a function of pH, indicating that with a pH value increasing over 9, SiO_2 enters into solution. Equilibrium with $CaCo_3$ is reached at a pH value of 9.8 at 25° C (Siever, 1957b). These observations are useful in explaining the phenomena recognized in Pl. VI–4. Authigenic idiomorphic quartz crystals during growth push away authigenic primary ferruginous material that is now present in the matrix. This material originally was goethite and is now changed into hematite which surrounds the quartz crystals as a film. Parts of the quartz crystals are replaced by calcite. In general the pH value of connate water rises when a sediment becomes buried. Dissolution, due to the higher pH value, begins in locations where the hematite coating is not, or incompletely, present. Since the growing crystals were able to push aside (unhardened) material from the matrix, and since silica-containing fluids were present and could move in this matrix, the idiomorphic authigenic quartz crystals are assumed to have a precementation origin. The subsequent replacement of the quartz crystals by calcite most probably occurred in the late precementation stage or in the early syncementation stage.

Recrystallization. – (Pls. III–2; III–5; V–5; IX–8; XIII–5). In a strict sense recrystallization refers to changes in grain size, in morphology or in orientation of mineral species, although in a petrographic microscopical sense these mineral species remain identical (Folk, 1965). It occurs when 'nuclei of new unstrained grains appear in or near the boundaries of the old, strained grains'. These nuclei grow until the old mosaic has been entirely replaced by a new, relatively strain-free mosaic with a nearly uniform grain size. Its coarseness depends on the density of the initial nucleation. When the nuclei are widely spaced, there is an intermediate prophyroblastic stage (Bathurst, 1958). The presence of old strained grains is essential. Recrystallization is therefore principally a syncementation/postcementation process.

Some processes, evidently physicochemical, but not leading to newly formed minerals, neither to morphologically changed ones, remain to be considered. Examples are: formation of stylolites by pressure solution; physicochemical disintegration of bioclastic material; the 'coating' of crystals by pushing aside impurities out of the ground mass; impregnation of particles by ferruginous material and genesis of ooids. Two of these processes actually are discussed; others, although recognized, will not be treated with in detail (for a detailed discussion cf. Chilingar et al., 1967).

Genesis of impregnated and/or coated grains, and zoned crystals. – The morphological division of ooids (p. 168), as discussed by Rusnak (1960), is as follows:

(1) Ooids with an unoriented cryptocrystalline coating (the product of rapid precipitation).
(2) Ooids with a radial arrangement in their coating (the product of weak agitation and slow precipitation; Pls. I–3; I–4; I–5; I–6; I–7; I–8; IV–1; IV–2.
(3) Ooids with a concentric arrangement of the crystals in their coating (the product of strong agitation and very slow precipitation); Pls. I–6; I–7; IX–5; IX–6.

All three types were observed in the Portilla Limestone Formation. Ooids with combinations of two or three types of coatings (e.g. Pls. I–6; I–7; I–8) provide information concerning energy fluctuations during formation. Ooids are formed in the precementation stage, more precisely in the syndepositional substage (cf. Bathurst, 1968, among others).

Some bioclasts, especially crinoidal fragments, show infilling of channels and/or pores (Pls. I–5; I–7; X–8) by ferruginous material (often of authigenic origin) before enveloping or cementation took place (cf. Sleumer, 1969, p. 25 and his Pl. 14–7, where he notes similar features for cellulae of stromatoporoids). Sometimes coating and infilling occurred during enveloping (Pls. I–6; I–7). It can therefore be deduced that this must be an early diagenetic process. It is claimed that this process took place in the syndepositional substage. Around rhombohedral dolomite crystals one or more zones of ferruginous material were often observed. Sometimes rhombohedral pores (result of dedolomitization and leaching) are present, encircled by a fuzzy blackish halo of ferruginous material (Pl. XI–3). This supports the hypothesis that during growth of the rhombohedral dolomite crystals the very finely divided authigenic ferruginous material, present in the ground mass, was pushed aside (Pls. I–8; V–5; VII–1; VII–3; XI–3; XII–1; XII–6; XIII–2; XIII–3). The repetitious character of the zonation (Pls. V–5; VII–1; VII–3; XII–6) demonstrates arrested and subsequently renewed stages of dolomite growth (Chilingar et al., 1967, p. 291; Bissell, 1970, Pl. XXXIII). This possibly indicates pulsation in the supply of hypersaline brines responsible for dolomite genesis. The coatings and zones must have been formed together with the rhombohedral dolomite crystals, viz. in the syncementation to postcementation stage.

Formation of (micro) stylolites. – Two opinions prevail with regard to the time of formation of stylolites. One (Shaub, 1949) advocates an early diagenetic (precementation) origin. Stylolites and solution stringers more or less parallel to the stratification (Pls. VI–5 and X–4) possibly reflect moments of deposition of more argillaceous material. The other school (Stockdale, 1922; Park & Schott, 1968) claims a postcementation origin of the stylolites. Serrated contacts between clasts, cutting of clasts by stylolites, concentration of residual material in the stylolites and displacements of veinlets support this latter theory, which in most cases applies in the present study (Pls. I–6; III–1; III–2; VI–3; VI–4; IX–5; X–1; XII–7; and XIII–5).

MECHANICAL PROCESSES

This group is subdivided into a group of constructive processes and into a group of destructive processes (Enclosures V, VI). The formation of limeclasts, the mechanical filling of internal open spaces and the washing out require more detailed discussion.

Formation of limeclasts. – (Pls. VI–6; XII–5; XII–7). The ball-like limeclasts in Pl. XII–5 contain ostracods and fossil fragments, difficult to identify, but essentially of the same type as those in the surrounding rock. The colour on these limeclasts is somewhat lighter (less ferruginous material) than that of the surrounding rock. The outlines are worn and contacts with the surrounding microsparitic matrix are microstylolitic. Here we are dealing with partly consolidated material torn from the surrounding sediment as a result of occasionally higher to extremely high energy conditions (storms?), before authigenesis of iron minerals was completed. In Pl. VI–6 a more angular type is pictured. Transport over significant distance and redeposition in other environments, as supposed by Roehl (1967, p. 2008) for similar sediments, certainly did not occur in the Portilla Limestone Formation because the fauna content inside and outside the limeclast indicates a similar shallow marine origin and the angular habitus indicates deposition in place. A possible genesis is penecontemporaneously fracturing of a semi-consolidated sediment followed by immediate infilling of the fractures with sediment, which preserved the angular habitus. Fracturing may have been caused by moments of extraordinary high energy.

Mechanical filling of internal open spaces. – (Pls. V–3; VII–6; VIII–1 and VIII–2). The internal sediment observed in the vugs in the sediment forms floors which consist of micritic or very finely fragmented bioclastic matter. Hematitic pellets, too, are present (Pl. V–3). remaining space of the void is filled, sometimes with a rim of blade-shaped sparry calcite crystals followed by equant sparry calcite crystals, sometimes only with the latter. Origin of the basal sediments is uncertain, but since some induration occurred in the sediment in which the voids are present, a supergenic-exogenic origin (Wolf's terminology, 1963b; cf. Chilingar et al., 1967, p. 81) of less indurated or non-indurated sediments is presumed. A similar mechanism was proposed by Dunham (1969) for early 'vadose' silt with which the basal sediment is comparable. Voids with internal sediment were recognized in sections 26, 29 and 32 (Enclosure III). Connections with the depositional interface and with the depositional environment in which very fine muddy sediment was deposited, and which contained clotty aggregates of hematite (Pl. VIII–1, inset), continued to exist during a fairly long time. In Pl. VII–6 the origin of the void and of the basal sediment is quite clear. Some spots were sheltered by large fossil fragments, and very fine particles could settle in the calm micro-environments. These particles could not be deposited in other places, since energy was there too high. Cracks filled with calcite cement postdate sedimentation and slight induration of the basal sediment (Pl. VIII–2). Mechanical filling occurred in the precementation stage of the paragenetic sequence.

Washing-out of accumulated carbonate sediments. – (Pls. I–3; III–3; V–7; VII–1). Depending upon energy, among other factors, more or less fine-grained to very fine-grained material is deposited in the sedimentary basin. Washing-out modifies a sediment after first deposition of the particles. Therefore it is here considered as a syndepositional process. Whether washing-out occurs or not depends on the general environmental energetic conditions after first deposition of the particles, on sheltering (Pls. I–3; VII–1), or on the presence of encrusting organisms (Pls. III–3; V–7).

b facies 23 thin sect.	4c.	1, 2a, 2b, 2c, 3, **8**, 9, 10, 11, **18**, **28**
	4ex.	21, 27
	3c.	14
	2c.	19, **29**, **30**
	Misc.	**4**, **5**, **6**, **7**, **17**, **32**
c facies 11 thin sect.	3c.	1, 2a, 2b, 2c, 9, 10, 11, 15, **28**
	2c	**8**, **17**, **18**, **29**
	2ex.	3, 27
	Misc.	**5**, **7**, **25**, **30**
d facies 1 thin sect.	c.	1, 2a, 2b, 2c, 3, **7**, **8**, 9, 10, 11, 15
	ex.	**18**, 21, **25**
e facies 9 thin sect.	4c.	1, 2a, 2b, 2c, 9, 10, 11
	3c.	3, **4**, **8**, 16, **17**, 21
	3ex.	26
	Misc.	**31**
f facies 11 thin sect.	2c.	1, 2a, 2b, 2c, 3, 14, 15, **4**, **6**, 7, 9, 10, 11, 27
	2ex.	**17**
	Misc.	**8**, **30**
g facies 4 thin sect	2c.	1, 2a, 2b, 2c, 9, 10, 11, 3
	2ex.	16, **17**
	Misc.	**8**, **20**, **25**, **30**, **31**

figures in circles refer to processes recognized (Enclosures V–VI)
figures in black circles refer to characteristic processes
4 c. means: 4 areas investigated, common processes
2 ex. means: 2 areas investigated, exceptional (but characteristic) processes

Table 6. Preference of diagenetic and other processes for the six facies investigated.

ENVIRONMENTAL INTERPRETATION OF FACIES

Two levels of uncertainty have previously been discussed. An attempt will now be made to bring together the results from the 2nd level of uncertainty (Enclosures V and VI) in a diagram showing the significant processes for the various facies in the four areas distinguished. Subsequently the main diagenetic trend of each facies can be established.

In table 6 processes were indicated, acting in more than half of the samples investigated and in more than half of the areas investigated. Further more, processes significantly indicative for the environment of deposition but acting in less than half of the areas investigated or in less than half of the samples investigated, were also included, but are referred to as miscellaneous. The numbers in Table 5 coincide with those in the legend of Enclosures V and VI and indicate the processes recognized. The same numbers will return in Fig. 19. The black circles in the table indicate the processes that are regarded as characteristic. From this table and from the Enclosures interesting characteristics of the facies can be derived.

Facies b. – This facies is characterized by syngenetic processes like washing-out on some spots, locally introduction of non-carbonate material, formation of some voids, formation of intraclasts torn from the surrounding sediment in a very early stage, formation of voids, sometimes with internal sediments and formation of algal borings. Authigenesis of ferruginous material and of chert nodules indicate alternating oxidizing and reducing conditions in the depositional environment with a low to neutral pH value. Dehydration occasionally occurred in short periods of emergence during which the voids (possibly results of desiccation, possibly of algal activity) could form. Intraclasts, ooids and washing-out indicate periods with fairly high energy.
Bioturbation (associated with formation of faecal pellets) is common. Herefrom a supratidal to intertidal environment is deduced. Cementation is sometimes characterized by rims of fibrous sparite crystals. Equant sparite crystals, however, are the most important cement type. Simultaneous with cement, between the grains, rhombohedral dolomite crystals were formed out of the hypersaline interstitial fluid. Recrystallization, stylolite formation and occasional formations of fractures, partly obscured the earlier diagenetic textures.

Facies c. – Characteristic syngenetic processes are the formation of intraclasts, of bioturbation (simultaneous with formation of faecal pellets) and of voids, sometimes filled with internal sediments. Authigenesis of ferroan material is locally extremely abundant. Silicification of fossil fragments is very common. Very characteristic is dedolomitization, which could occur at any moment during history of the sediment, but which association with bioherms and biostromes is striking and which is therefore considered as a result of an early emergence of the sediment. For sediments with these characteristics an intertidal and incidentally supratidal environment applies best. Energy was especially high during emergence (reef breccia and intraclasts), but generally it was medium, as is showed by the diminished influence of washing-out in comparison with facies b. Equant sparite crystals are the most important cement type. Dolomite is present in two ways; intergranular, as rhombohedral crystals, and in fossil fragments as a patchy, subhedral to anhedral crystalline replacement dolomite. It was formed during and after cementation. Fractures are in fairly great amounts present, sometimes obscuring earlier diagenetic textures and giving the sediment a breccia-like appearance.

Facies d. – Syngenetic washing-out of the framework of the sediments is fairly common. Silicification of fossil fragments and formation of authigenic ferruginous material is most dominant in this facies in comparison with facies b, c and f. Regional widespread dedolomitization on a few levels is important. From field observations it has already been deduced that bioherms on certain moments emerged for short periods. This is confirmed by the syngenetic processes mentioned. Energy on those moments was high. Environment of deposition varies between intertidal and occasionally supratidal. Cement consists predominantly of equant sparitic crystals. Rhombohedral dolomite crystals occur in the interstices. Many of these are dedolomitized. Microstylolites are formed at a late moment in the sediments' history, possibly as a result of the overburden of the bioherms.

Facies e. – Syngenetically formed internal sediments (sometimes partly authigenic ferruginous material) is present. It is typical for subtidal areas. For loose sediments to penetrate into a (semi) indurated limestone framework, turbulent conditions and surging powerful currents are necessary. Washing-out occurred often. Concentric ooids, indicating strong agitation and very slow deposition, fit in this picture of medium to high energy sediments. Non-carbonate material, too, is often deposited. Cementation with syntaxial cement rims around the crinoidal fragments occurred fairly late in the history of the sediment because many of the serrated, microstylolitic contacts of the fossil fragments must have been formed before or during cementation. Dolomite of the patchy, replacement type was also formed fairly late.

Facies f. – Syngenetic introduction of non-carbonate sediments becomes fairly important. Ooids, mainly of the radial type, indicate slow precipitation of the coating material, and weak agitation. Washing-out of sediment in the limestone framework still occurs but is less frequent than in previously mentioned cases. The environment is clearly subtidal. The amount of authigenic ferruginous material is unimportant. Cement consists mainly out of directly precipitated equant sparite crystals. Sometimes these are preceded by rims of fibrous sparite crystals. Patchy, anhedral dolomite crystals and idiomorphic authigenic quartz crystals are of replacement origin. Together with fractures they are a fairly late phenomenon.

Facies g. – Syngenetic introduction of non-carbonate sediments culminates in this facies which is for a great part a fairly pure sandstone. Some limestone lenticles, however, are intercalated and are locally changing in facies with argillaceous limestones. In these latter washing-out occurs fairly often, together with ooids (of the concentric type), algal coated grains and indications of algal boring. This suggests medium to high energetic conditions basically in a subtidal environment. In some limestones, however, voids are present, sometimes filled with internal sediments. These, possibly, indicate occasional subaerial exposure. Authigenic ferruginous material becomes negligible in this facies. Cementation is characterized by the grain growth rims and by some syntaxial cement rims. Features that occurred late in the sediments' history are dolomitization (of the patchy type) and occasionally dedolomitization.

Summarizing, the facies according to their depositional environment and energy, can be arranged into three groups; viz. a medium energetic (locally high energetic) supratidal to intertidal environment (facies b), a medium to high energetic intertidal environment (facies c and d) and a low to medium energetic subtidal to intertidal environment (facies e, f and g).

CHAPTER VI

REGIONAL FACIES DISTRIBUTION AND BASIN-FILL HISTORY

GEOMETRY OF FACIES

Field and microscopic characteristics of lithofacies as recognized and discussed in Chapter IV, serve as a basis for correlations established in facies diagrams (Figs. 9–14). A short discussion will be devoted to the nature and position of the main facies changes. Each member will be reviewed separately.

Veneros Member

The Peña Corada area (Fig. 10). – Four major lateral changes in facies occur between sections 11 and 12, 16 and 17, 17 and 18, 19 and 20. In section 21 uppermost Huergas sediments are postulated to have been deposited simultaneously with sediments of the Veneros Member elsewhere. From the diagram it is visible that the Veneros Member can be divided into a lower and an upper part. The lower part is generally composed of facies f sediments; the upper part mainly of facies e sediments. In the western part of the Peña Corada area the influx of significant quantities of siliciclastic material ends, and in the easter part (sections 19 and 22) some lenticles and tongues, containing siliciclastics, are present, both at the transition from the lower to the upper part, and in the upper part of the Veneros Member. The ooids show no preference either for facies f or e.

FACIES DIAGRAMS B MEMBER AND D MEMBER

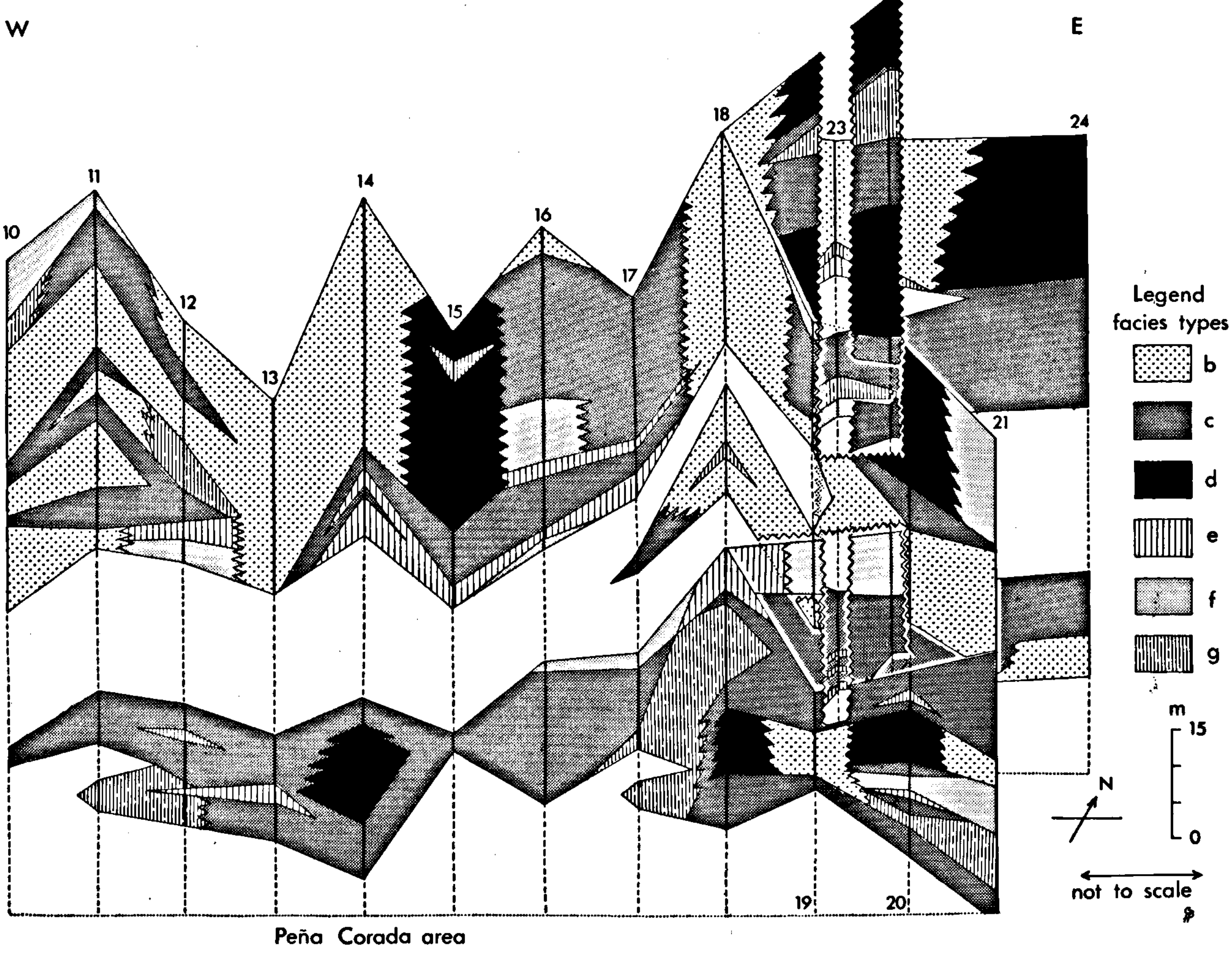


Fig. 9. Lateral correlation of facies recognized in Member B and Member D in the Peña Corada area.

The Esla autochthonous area (Fig. 12). – Here the situation is more complicated. Near sections 28 and 29 some important facies changes occur. The lowermost part of the sequence is rather uniform, and is a composition of facies e and f. This uniform sequence is the only part of the Veneros Member occurring north of section 31. South of this section the sequence is followed by a lens-like siliciclastic interval. This interval possibly has an E-W strike which cannot been seen with certainty in a roughly N-S section. In section 29 facies b overlies these siliciclastics which, in section 28, are laterally equivalent to biostromal sediments (facies c) because they show interfingering. To the top the fairly uniform facies pattern, as occurs at the base, returns. The siliciclastics and the biostromal sediments deviate in nature from the other sediments in the Veneros Member. They are present as a lentil. In sections 25 and 26 the member has been divided into an upper and a lower part, separated by a tongue of Member B. Here, too, the ooid bed, without preference, passes through facies e and f.

The Alba and Pedroso synclines (Fig. 14). – The lowermost part of the Portilla Limestone Formation in section 7 is covered by Carboniferous sediments. In comparing the sediments present east of this section with those to the west, an important facies change becomes discernible. West of section 7 the lowermost part of the Veneros Member is predominantly present as the siliciclastic facies g, east of this section mainly as facies e and f. Only in section 9 were some sediments observed, belonging to facies b. The lowermost exposed part of section 7 belongs to facies c (Member B) and does not show any lateral relation with sections 6 and 8. Another important facies change is found between sections 8 and 9, where facies f (section 8) changes into facies c (belonging to Member B).

In the Pedroso syncline we find rather uniform sediments (facies e and f). Only in section 4 does a slight deviation occur in the upper part. Some sediments here belong to facies g and b. Ooids here show a preference for facies e. In the Alba syncline no ooids have been seen.

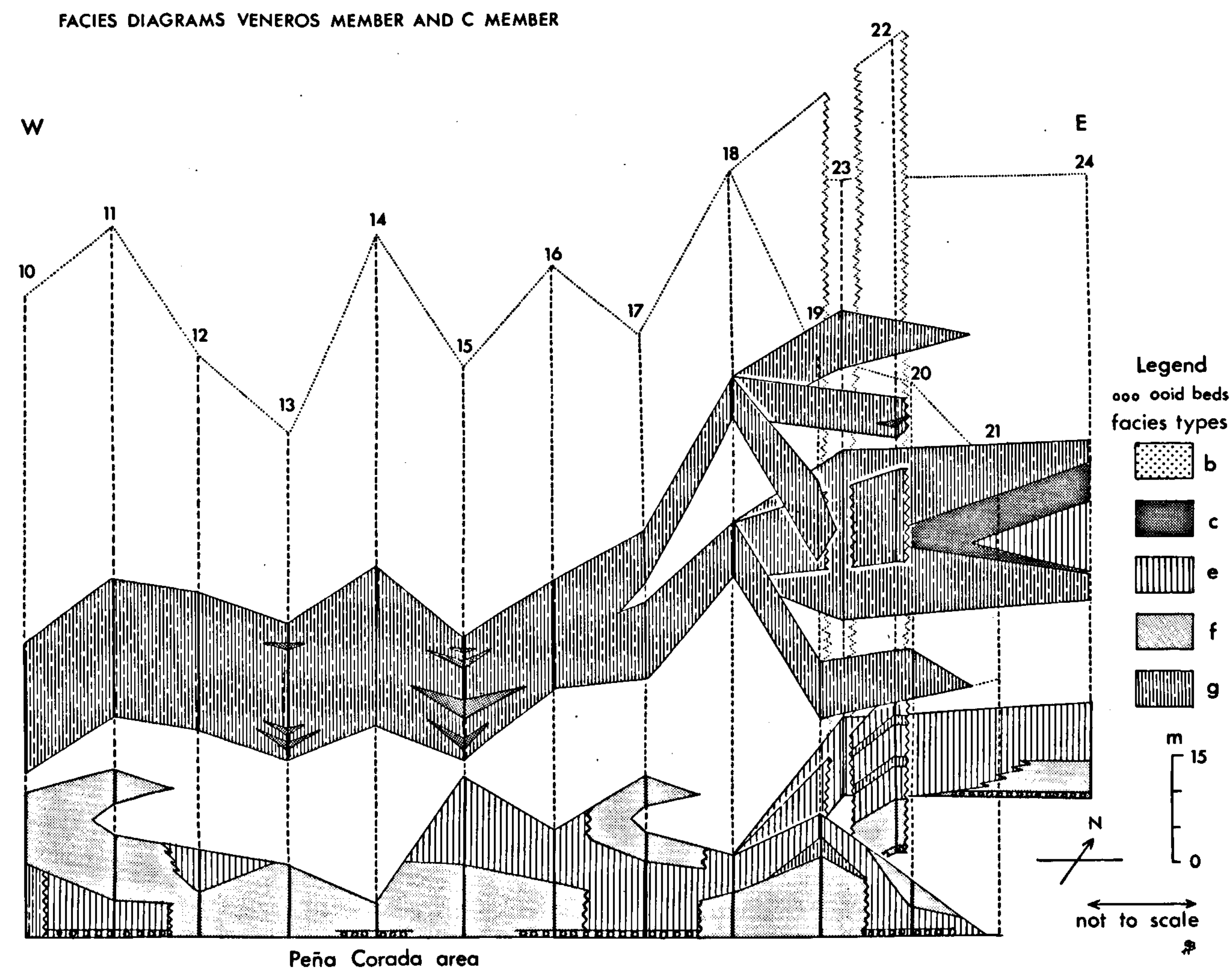


Fig. 10. Lateral correlation of facies recognized in Veneros Member and Member C in the Peña Corada area.

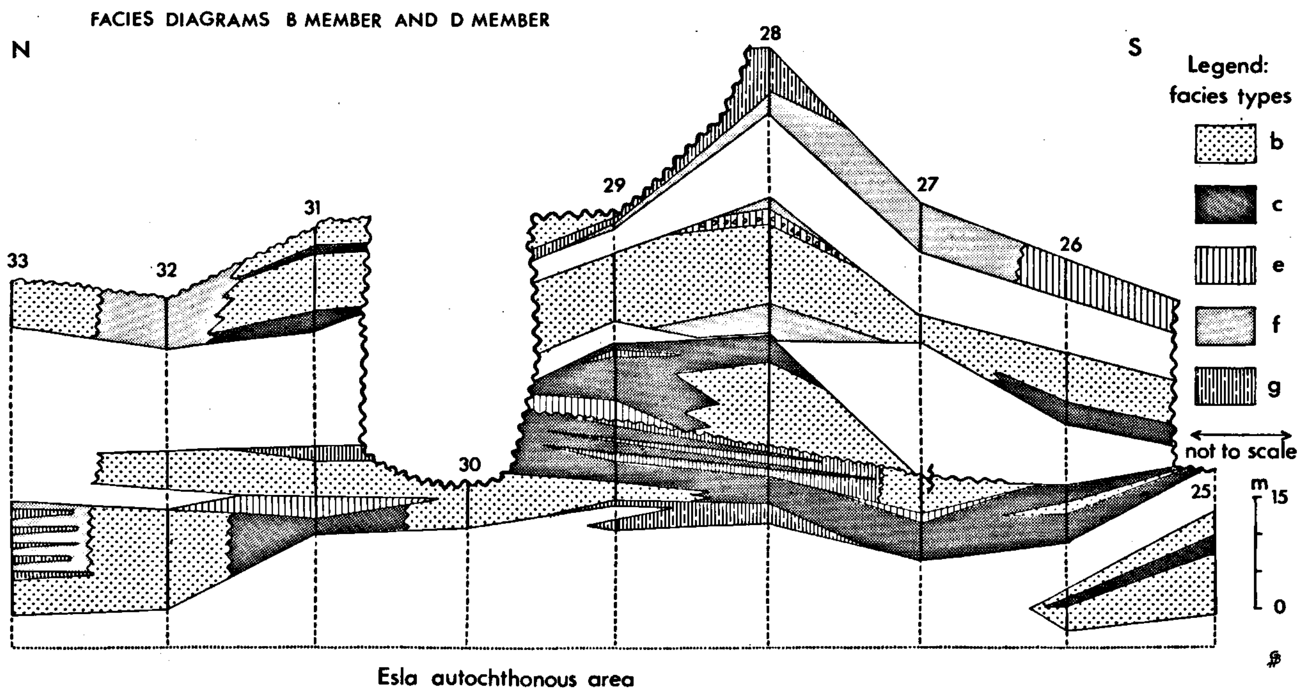


Fig. 11. Lateral correlation of facies recognized in Member B and Member D in the Esla autochthonous area.

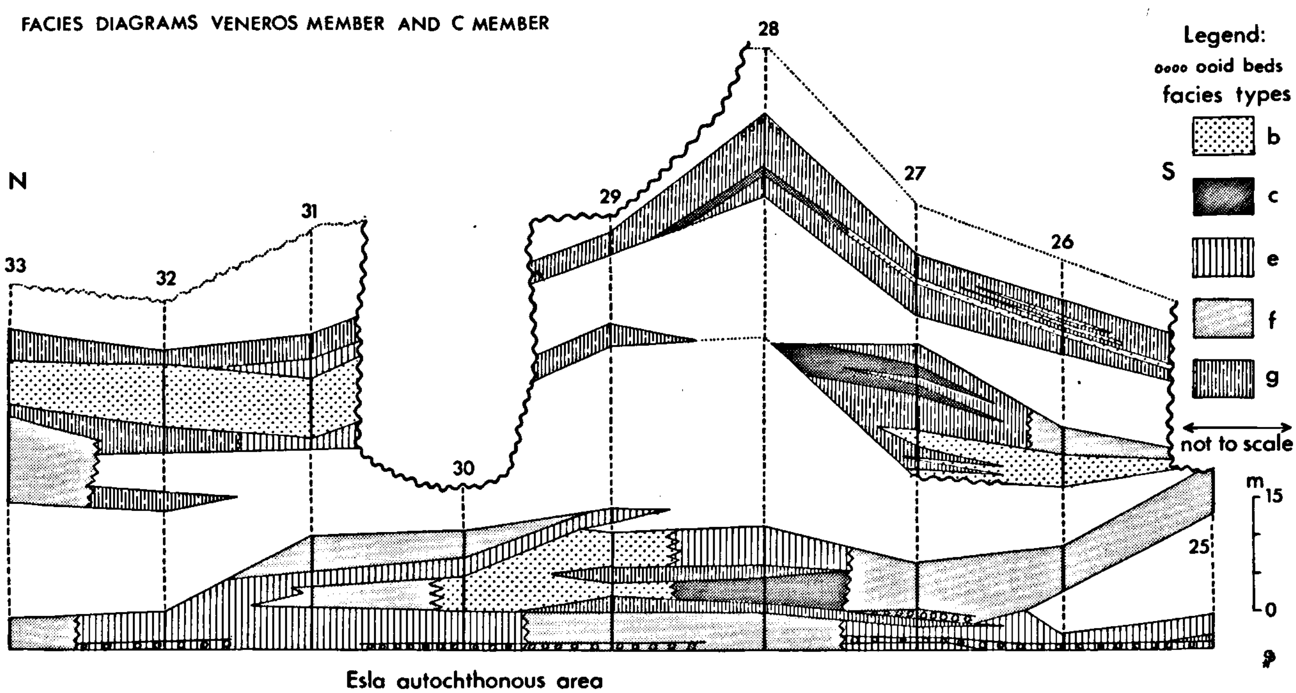


Fig. 12. Lateral correlation of facies recognized in Veneros Member and Member C in the Esla autochthonous area.

Member B

The Peña Corada area (Fig. 9). – Of the four major facies changes in the Veneros Member (cf. Fig. 10) only those between sections 16 and 17, and between 17 and 18 are still present in Member B. Moreover, a locally important facies change is found between sections 12 and 13. From section 10 towards section 16 there is a rather uniform development of biostromal sediments showing interfingering, in section 14, with a small bioherm. East of section 16 a rather complex interrelation exists between bioherms (sections 18 and 20) and the sediments of facies b, present between the bioherms. Biostromal sediments envelope and underlie both biohermal and facies b sediments, and proceed from east to west, as can be revealed after close study of the stratigraphic relations. The biostromal/biohermal area in the east is separated from the biostromal area in the west by facies g (sections 17 and 18). From section 16 towards the east the entire complex is continuously overlain by facies f and e.

The Esla autochthonous area (Fig. 11). – In this area Member B is a mosaic of facies b, c, e, f and g. Due to erosion sections 25 and 30 are incomplete. North of section 30 facies b predominates, with relatively unimportant lentils of facies c, d, e, f and g. From section 30 towards the south the influence of facies c increases, especially in the lower part of Member B. Here the division into a lower and an upper part of Member B is easy, since an intraformational unconformity of local importance cuts sections 26–29 (Fig. XIV–1; cf. p. 207).

The lower part of Member B of sections 29 and 28 is mainly composed of thick-bedded to very massive boundstone (facies c). Towards section 27 the influence of the well-bedded, thinly layered mudstones of facies b, increases. In section 26 biostromal sediments (facies c) return. Lower in this section a separate tongue is present of facies b.

The Alba and Pedroso synclines (Fig. 13). – The only important change in facies in the Alba syncline occurs east of section 7. No sediments belonging to Member B were found here. In the other sections mainly facies c is present. In section 9 a small lenticle of facies b has been observed.

In the Pedroso syncline Member B sediments were not deposited in sections 1 and 2. In sections 3 and 4 chiefly facies b has been deposited, simultaneously with sediments of the Veneros Member in sections 1 and 2. In section 4 the influence of facies c (biostromes) increases.

Member C

The Peña Corada area (Fig. 10). – Member C consists almost exclusively of facies g. In sections 13 and 15 some minor lenticles of facies c and b are present. From section 17 onwards the member is divided into two tongues, pinching out towards the east. In section 21 no sediments belonging to Member C were observed. Presumably deposition of Member B sediments continued. Sections 23 and 24 are not divided into two separate tongues. Sediments of facies e in section 24 are enveloped in sediments of facies c and these in turn are enveloped in sediments of facies g.

The Esla autochthonous area (Fig. 12). – Due to erosion, Member C of sections 25 and 30 is entirely absent. North of section 30 Member C consists of a sequence, chiefly containing facies b and g. The changes in facies between sections 31 and 29 and between sections 27 and 26 are of local importance. South of section 30 Member C consists of two separate parts. The lower part mainly contains facies g, further to the south alternating with facies c. In section 26 facies b and f are present. The transition from facies g to facies b is gradual; from facies g to facies f sudden. The upper part of Member C chiefly contains facies g. A biostromal layer grading into facies b occurs in this part. It has an extension of several kilometres and acts as a marker bed. Facies g in the upper part of Member C, south of section 30, appears to link up with the highest beds in the northern part of the autochthonous area, and so form an easily recognizable marker bed. The two separate beds of Member C are arbitrarily chosen and the division is so drawn, because of the appearance in the field.

The Alba and Pedroso synclines (Fig. 14). – In the Alba syncline area one important facies change is found between sections 6 and 7. East of section 7 the sediments belong to facies b, west of this section they are mainly facies c, changing into facies g and enveloped by facies b sediments. The sediments of facies c in section 5 gradually change into facies g and b. In the Pedroso syncline only facies g is found, with in section 4 a few small lenticles of facies e.

Member D

The Peña Corada area (Fig. 9). – Throughout Member D, rather important changes in facies exist between sections 12 and 13, 14 and 15, 15 and 16, 17 and 18; in the upper part of Member D between sections 18 and 19, 18 and 22, 23 and 24, and in the lower part of Member D between section 18 and 23. Two subareas are distinguished, containing mainly biostromal/biohermal sediments (facies c, d); two facies b subareas and one mixed biostromal/facies b subarea. The latter subarea is to be found near section 10, 11 and 12. Here we see a regular alternation of biostromal sediments (facies c) and facies b. Local changes in facies also occur here. From section 12 towards section 13 the influence of facies b gradually increases; in section 13 only the latter sediments are present. Further to the east the lower part of the sequence (facies c) grows more important and interfingers with facies b. The latter sediment type envelopes facies e. In section 15 bioherms (facies d) are present. Laterally, towards the east, we notice a change into biostromes (facies c) which in section 18 are completely replaced by facies b. In the southern part of the investigated area facies b remains in the lower part of Member D; in the northern part of the studied area they

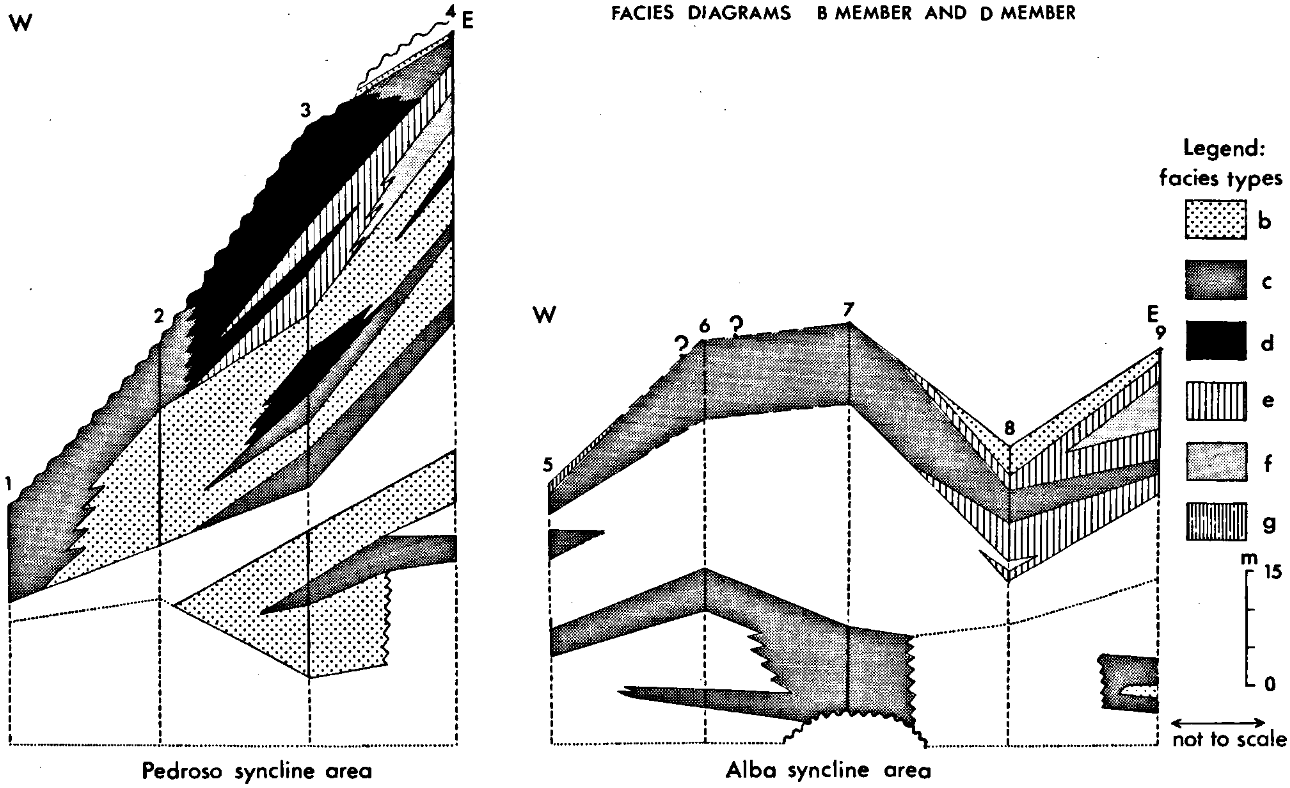


Fig. 13. Lateral correlation of facies recognized in Member B and Member D in the Alba and Pedroso synclines.

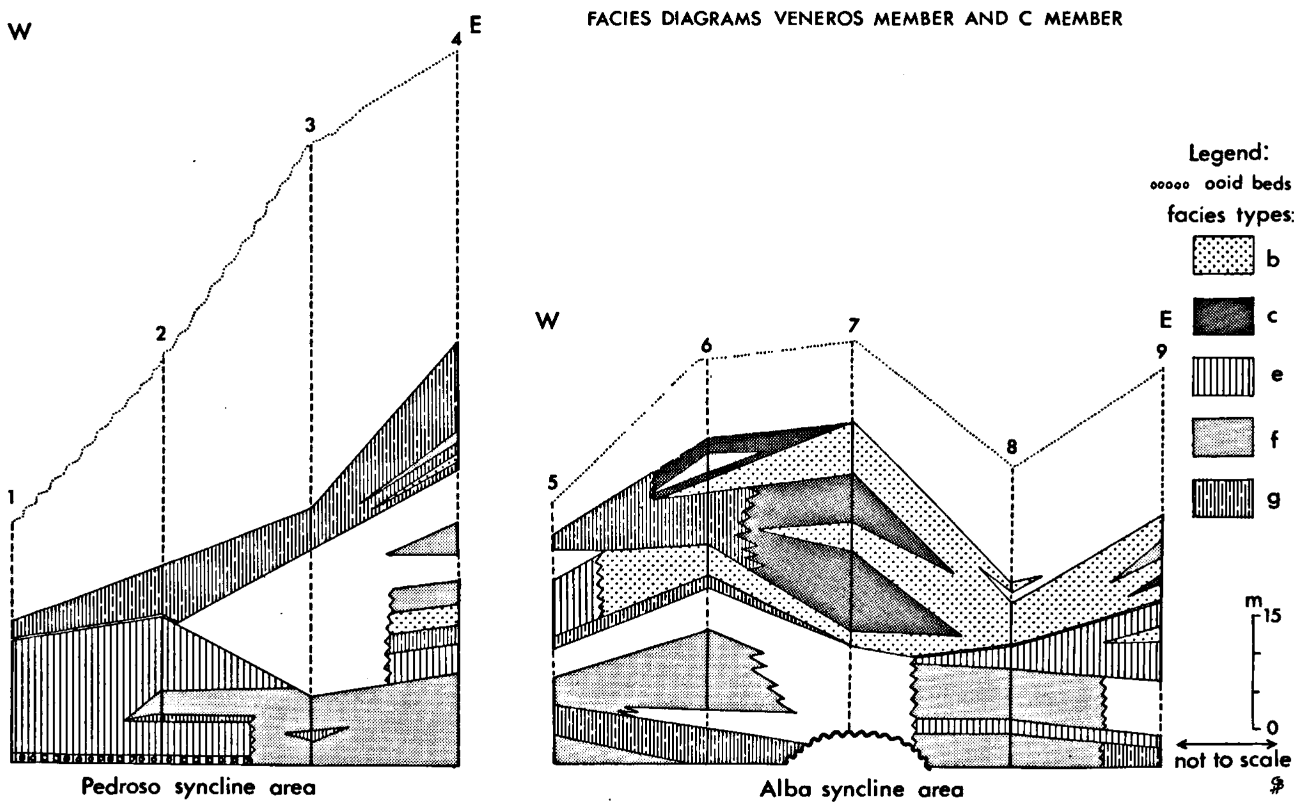


Fig. 14. Lateral correlation of facies recognized in Veneros Member and Member C in the Alba and Pedroso synclines.

remain in the upper part of the sections. In section 22 facies b completely changes into biostromes and bioherms. In the southern part these sediments overly facies b. It is remarkable that facies f overlies facies b and c in section 21. In section 23 facies b is underlaid by facies c and g. In section 24 the lower part consists entirely of biostromal deposits overlain by a bioherm.

The Esla autochthonous area (Fig. 11). – North of the completely removed Member D of section 30, two separate areas are present, with facies b and facies f; south of section 30 Member D is divided into two separate parts. The lowermost one consists mainly of facies b, with only in section 28 some layers of facies e and f. In the lower part of section 26 some sediments of facies c also occur. The upper part of Member D in section 29 is not complete. The author has the impression, however, from the parts present, that the sediments are of the same type as in the upper part of the sections north of section 30. From this section towards the south facies b changes into facies f, and from section 27 onwards these sediments are changed into facies e.

The Alba and Pedroso synclines (Fig. 13). – In the Alba syncline area a change in facies occurs between sections 7 and 8. West of section 8 facies c is present as far as can be seen (access is difficult). East of this section we see sediments of facies f (enveloped by sediments of facies e and b). In the Pedroso syncline the main change in facies is near section 2. West of this section only biostromal sediments (facies c) are present as far as could be seen (access is difficult and the Upper Devonian erosion surface cuts through the sequence). East of this section the lowermost part of Member D consists of an alternation of sediments of facies c and b, with one small bioherm, and the upper part consists of a bioherm, showing interfingering relations with facies c.

STRUCTURAL RELATIONSHIPS

The present author does not accept the Peña Corada Unit as a transported part of the Esla Nappe, as suggested by Rupke (1965). This opinion is primarily based on facies gradients in the four areas.

The Peña Corada area. – (Figs. 1, 9, 10, 15–18; Enclosures I, II).
(1) Almost all sections contain the ooid bed in the Veneros Member (Figs. 10, 15); in the west the ooids are predominantly concentric, in the east they are mainly radial.
(2) Bioherms and biostromes are abundantly present, especially in Members B and D (Figs. 9, 16, 18).
(3) Member C is composed of siliciclastic sediments. The proportion of the siliciclastics strongly increases towards the west (Figs. 10, 17), where it is present as one layer. Towards the east it is divided into two layers that are pinching out.
(4) Thicknesses of the Veneros Member and of Member C decrease towards the south, and thicknesses of Members B and D increase towards the south (Figs. 15–17).
(5) Biostromes and bioherms sometimes emerged, as is proved by dedolomitization. This is associated with layers in which fossil debris occurs, which alternates with layers in which faunal elements are in growth position (Fig. 8).

The Esla autochthonous area. – (Figs. 1, 11, 12, 15–18; Enclosures III).
(1) Almost all sections contain the ooid bed in the Veneros Member (Figs. 12, 15). In the south and north (due to the S shape of the outcrops this corresponds with the eastern part of the area) radial ooids are present, in the centre (western part of area), mainly concentric ooids are present.
(2) Biostromes are common, but emphasis in this area lies on facies b (Figs. 11, 12, 15–18).
(3) Siliciclastic sediments occur far less in comparison with the Peña Corada area. Nevertheless, the small amounts present increase in importance towards the north where it is present as one layer. In the south it is present as two layers that generally pinch out towards the south (Figs. 11, 12, 17).
(4) In the Veneros Member thickness decreases towards the east; in Member B thickness increases towards the central zone, and decreases both towards the southeast and the northwest. In Member C thickness decreases and then increases from the northwest towards the southeast. Member D is incomplete (Figs. 15–18).
(5) Many indications suggest shallow depositional conditions and moments of possible emergence in the northwest.

The Alba syncline. – (Figs. 1, 13–18; Enclosure IV).
(1) No ooids were found (Figs. 14, 15). Mr. M. Mohanti (pers. comm., 1971) does not report any either.
(2) All types of facies were observed but especially the organic barriers, composed of broken and oriented branching tabulate corals, are abundantly present (Figs. 13–18). The same holds for the entire area studied by Mr. M. Mohanti (pers. comm., 1971).
(3) Subsidence sometimes caused extreme thicknesses (Mohanti, in prep.) in comparison with other sections.
(4) Sediments containing red ferruginous material (ferric oxides and hematite) are abundantly present. Siliciclastic material is admixed.
(5) As far as could be inferred the centre of the basin seems to be situated in the vicinity of sections 6 and 7 (Figs. 15–18).
(6) Depositional environment becomes protected towards the east.

The Pedroso syncline. – (Figs. 1, 13–18; Enclosure IV).
(1) The ooid bed in the Veneros Member is present in the western part of this area, and can be traced at least as far as Mallo via the outcrops near Mirantes (Figs. 14, 15).

(2) One bioherm and a number of biostromes are present in the east, but sediments deposited on an open marine shallow subtidal platform strongly predominate (Fig. 13).
(3) Subsidence was here not so large a phenomenon as in the nose of the Alba syncline (Mohanti, in prep.). In all members however, a close spacing of the isopachs is visible between sections 3, 4 and 5. This suggests syn-sedimentary movements.
(4) Member C is composed of a fairly pure cross-bedded quartz sandstone.
This member, too, can be traced at least as far as Mallo, via the Mirantes outcrops (Figs. 14, 17).

Comparison of the sedimentary features and the stratigraphic succession of the Peña Corada area and of the Esla autochthonous area, shows a continuous gradation. The transition from the Peña Corada area with biostromes and bioherms into the Esla autochthonous area with b facies sediments, the continuous zonation of ooids with their specific characters, and the diminishing influence of siliciclastic sediments towards the north and the northwest (regarding the entire succession) are only three examples of facies gradients which illustrate that no reasons are present to accept an allochthonous origin of the Peña Corada area, 25 km to the south (cf. footnote 1). Another indication that the Peña Corada area is allochthonous can be found when we regard the mathematical approach of the relationship between the Esla autochthone and the allochthone as presented by Rupke (1965, p. 57). He assumes time equivalence of transition from the Portilla Limestone Formation into the Nocedo Formation in the supposed area of sedimentation (cf. footnote 1) and in the present autochthonous area. In Chapter III, however, evidence has been brought foreward that the base of the Portilla Limestone Formation is somewhat time transgressive in a north-south direction. This makes the results of the mathematical approach of the relation between the allochtone and the autochthone uncertain.

Finally, Rupke accepts the extension of the Upper Devonian erosion plane far south of the Sabero-Gordón line. Nowhere in the Cantabrain Mountains has any evidence been found in favour of this assumption.

Both uncertainties make the calculated distance of the nappe structure improbable and the unfavourable circumstance that the original area of sedimentation of the sediments of the nappe accepted by Rupke, is now covered by Cainozoic sediments, preclude a check on his hypothesis.

The present author is aware that all the difficulties are not solved by stating that a nappe origin of the Peña Corada Unit is improbable. Moreover, he only studied the Portilla Limestone sediments in close detail. The stratigraphic evidence presented here, however, allows him to accept the Peña Corada area as a not transported area.

PREPARATION OF LITHOFACIES MAPS

It is possible to distinguish three essentially different groups of carbonate sediments among the facies distinguished (cf. Chapter V).

(1) Sediments in which the intergranular space is filled with micrite, mud or very fine fragmented material. In Dunham's classification these sediments are packstones, wackestones and (fossiliferous) mudstones. The proportion and distribution of mud basically depends on two factors, viz. the quantity of mud available and the energy conditions of the depostional environment. In the Portilla Limestone Formation we are dealing with intrabasinally formed carbonate mud. Only energy is a variable factor. Low energy sediments generally have a grain/micrite ratio of 1 or lower. In the Portilla Limestone Formation we encounter this group of sediments in facies b.

(2) The group of sediments in which the intergranular space is not filled with mud/micrite but mainly contains precipitated sparry calcite cement. These grainstones (Dunham) may be composed of different types of grains and of interstitial cement, precepitated where mud had been washed out or could not settle. Sediments of this type generally represent environments with moderate to high energy (cf. Maxwell, 1968, p. 98). The grain/micrite ratio is between 1 and 9. Facies e and f belong to this group of sediments.

(3) The group of sediments that are evidently biostromal and/or biohermal, and that macroscopically or microscopically often show evidence of binding or encrusting by organisms. These sediments often can be classified as boundstone (Dunham). A high grain/micrite ratio as sometimes may be expected, is not always present. Facies c and d belong to this group.

A difficulty arises with facies b and g since, with regard to energy, there are no significant differences. The field characteristics (Chapter IV) and the processes recognized in these facies (Chapter V), however, allow us to distinguish between the two.

The facies, arranged according to the three different groups of carbonate sediments mentioned above, can be used as three end members in 100 percent facies triangles (Krumbein & Sloss, 1951), selected in this way in order to visualize the energy differences in the sediments in the various members. With the help of these facies triangles, lithofacies maps can be designed. The available control points were plotted in the facies triangles in order to observe the distribution of points in terms of the changing composition of the member. The geographic location of the control points situated in four subareas is expressed by using different symbols for the subareas. The regionally different composition of the members and especially the regional changes in energy are so expressed. Lithofacies maps were designed for the four members, and were combined with isopach maps (Figs. 15, 16, 17 and 18). A separate discussion will be devoted to these maps and an attempt will be made to reconstruct the basin-fill history.

REGIONAL INTERPRETATION OF THE OBSERVED FACIES RELATIONS

After establishing the structural relationships of the areas, the lateral relation of facies can now be discussed in a more regional way.

Veneros Member (Fig. 15)
Marked variations in thickness occur in the Veneros Member. The thicknesses in the Peña Corada area compare fairly well with those in the Esla autochthonous area. A general tendency is the increase in thickness towards the Porma fault. In a zone crossing the Esla autochthonous area approximately from north to south, thickness rapidly decreases towards the east. An irregularity in this picture is found in sections 19 and 22 where the thickness increases rather suddenly.

The value [(f+g)/(c+e+f+g)] 100 was calculated for each control point, and the 25 %, 50 % and 75 % lines were constructed. The value expresses the presence, in a percentage, of facies f and g. In the western part of the area east of the Porma fault, conditions seemed to be unfavourable for deposition of sediments belonging to these facies, as is reflected by the decreasing percentage. Contrasting with this, we here find the thickest parts of the Veneros Member, mainly consisting of facies e and exceptionally of facies c. The average composition for the various areas is shown in Fig. 4. The grain/micrite ratio is high. In the western part of the area east of the Porma fault the ooids are mainly of the concentric type. In a few cases there is a radial coating directly around the nucleus. Cross stratification often occurs here. More to the east, a belt is present in which radial ooids predominate. Age determinations of the lowermost member of the Portilla Limestone Formation reveal a base progressively younger towards the south, while nothing can be said with certainty about the base in an east-west direction (cf. Chapter III).

Interpretation. – The western part of the area east of the Porma fault was most probably a shallow subtidal flat, adjacent to a source area that occasionally supplied minor amounts of siliciclastic material. This flat was directly connected with a somewhat deeper environment with open marine conditions. Crinoids, bryozoans and brachiopods suggest a prolific crinoid-bryozoan biocoenosis. Sedimentation presumably started in the north (and the northwest?) and shifted progressively towards the east. There accumulation of bioclastic debris diminished. The faunal elements are often more fragile in composition. This points towards different depositional conditions. The Veneros Member decreases in thickness, and in section 21 even non deposition occurs. The sediment in the east mainly belong to facies f and g. The radial ooids here indicate moderate energy conditions (Rusnak, 1960). The transition from facies e and f in the western part of the area to facies f and g in the eastern part of the area is fairly abrupt. In general facies f and g sediments seem to have been deposited under somewhat deeper open marine conditions. Towards the southeast, conditions for the crinoid-bryozoan biocoenosis remained favourable, but energy decreased due to increasing depth. Sections 19 and 22 are exceptions. Here somewhat higher energy conditions prevailed.

The relation of an increasing thickness of the member and a decreasing importance of facies f and g also holds in the areas west of the Porma fault. The spacing of the isopachs between sections 3 and 4 possibly indicates synsedimentary instability of the depositional area. In the western part of the Pedroso syncline, and via the exposures near Mirantas and Mallo further towards the west, radial ooids are present, indicating moderate energetic conditions. Sedimentation first began in the area of section 7 (possibly already during the Eifelian). In the Givetian it shifted towards the west (and the east?), presumably because the centre of the basin moved in that direction. In the vicinity of the Bernesga River almost uniformly (75 %–100 %) facies f and g were deposited. Somewhat deeper conditions prevailed here. In the eastern part, near the Porma fault, thickness decreases and the high energy facies types predominate. The picture obtained from this area does not differ significantly from that of the area east of the Porma fault. The depositional environment becomes somewhat deeper, quieter and more open marine towards sections 5 and 6. The subtidal flat, on which accumulation of the crinoidal-bryozoan association occurred, presumably continued uninterrupted essentially north of the investigated area. Only in the northwest and the northeast, remnants of this subtidal flat deposits were found. In the north the Upper Devonian erosion has removed the Portilla Limestone Formation and has cut deeper into the Devonian sequence, as is shown by the mixed Emsian fauna found in section 5 in the Portilla Limestone Formation, and by 'San Pedro' pebbles reported by Evers (1967, p. 96, 102).

Member B (Fig. 16)
In Member B thickness decreases both towards the northwest and towards the east in the area east of the Porma fault. In between, approximately in the same place where, in the Veneros Member, the transition is situated between the area of predominantly facies e and of predominantly facies f and g, a belt is present with greater thickness, especially in the south. This belt strikes roughly north-south, and bends towards the east in section 18. The zone of maximum thickness runs from section 28 via 18 towards 20 and 21.

For each control point the value [(c+d)/(b+c+d+e+f+g)] 100 was calculated. For the Esla autochthonous area the value [(b)/(b+e+f+g)] 100 was also calculated, and for the area north of Cistierna the value [(b)/(b+c+d)] 100 is important. The first value indicates

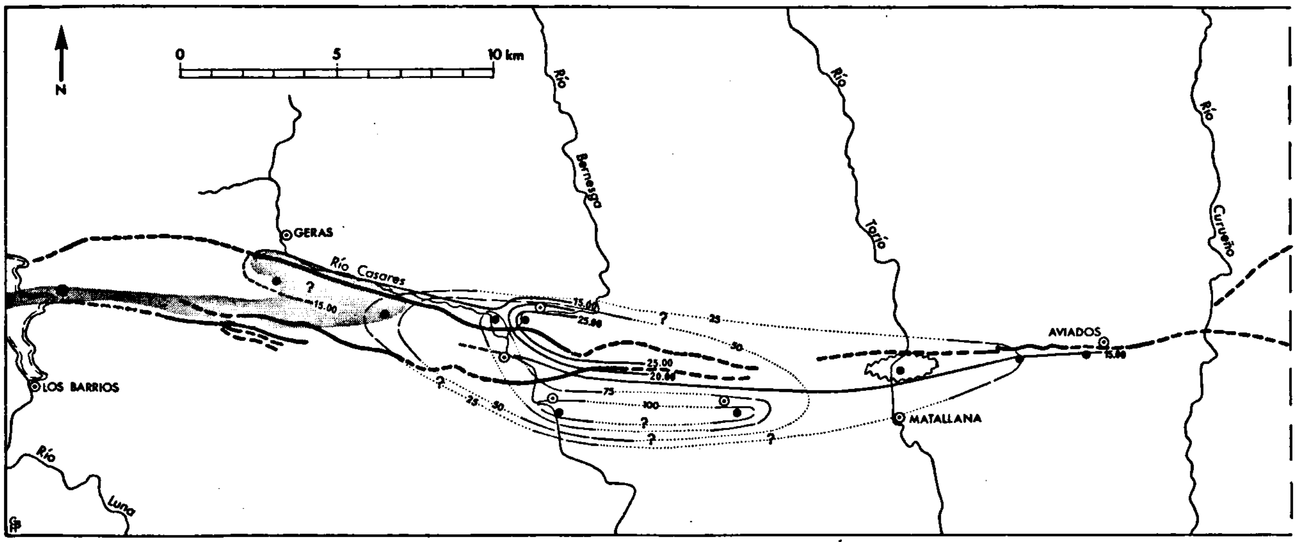


Fig. 15. Isopach-lithofacies map of Veneros Member.

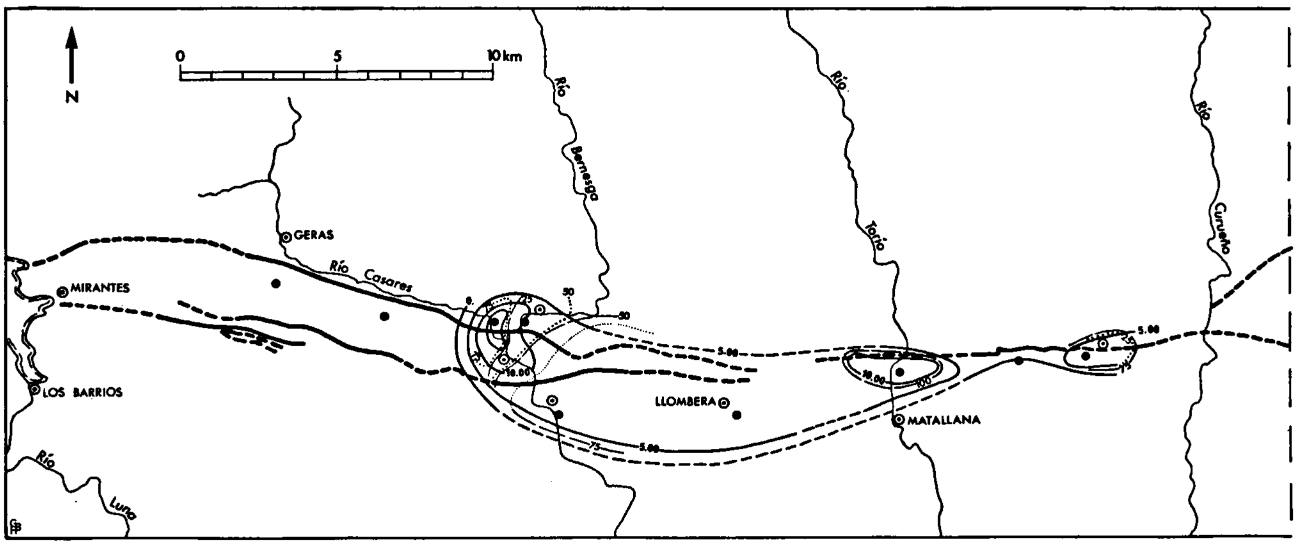
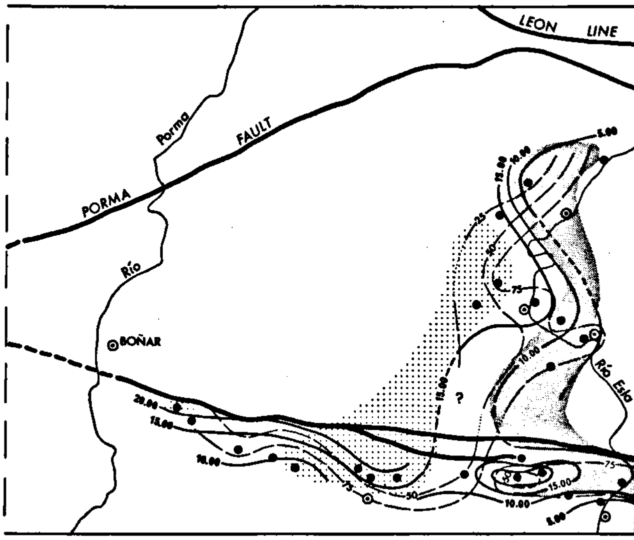


Fig. 16. Isopach-lithofacies map of Member B.

the importance of the biostromal and the biohermal sediments. The percentage lines of 100 %, 75 % and 50 % were constructed. The second value expresses the importance of facies b (deposited under fairly quiet conditions) as compared with facies e, f and g (deposited under moderate to quiet conditions). Here the percentages lines of 100 %, 50 % and 25 % were constructed. The third value expresses the importance of facies b as compared with facies c and d (deposited under moderate to high energetic conditions).

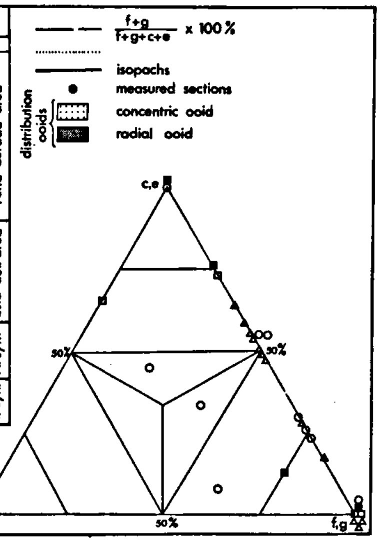
The average composition of Member B for the various areas is expressed in Fig. 5. Tabulate corals are common, especially in those parts of the areas where biostromes and bioherms occur. Stromatoporoids and platy tabulate corals are characteristic of parts of the Esla autochthonous area, especially of section 28, where they are of large dimensions. In the northern part of the Esla autochthonous area large quantities of pellets, relatively many ostracods and indications of bioturbation are present. Pyrite and bituminous matter were seen in section 28. In sections 10, 14 and 32 small slumpings were observed. Important slumping features are present on a large scale in sections 27–29 (Plate XIV), where they seem to be connected with a plane of unconformity of local importance. This plane of unconformity is subject of separate discussion. Especially towards the north the amount of siliciclastic material increases as well as the quantity of zaphrentoid corals. In analogy with the Veneros Member, a slight time difference in sedimentation is supposed between the north (earlier

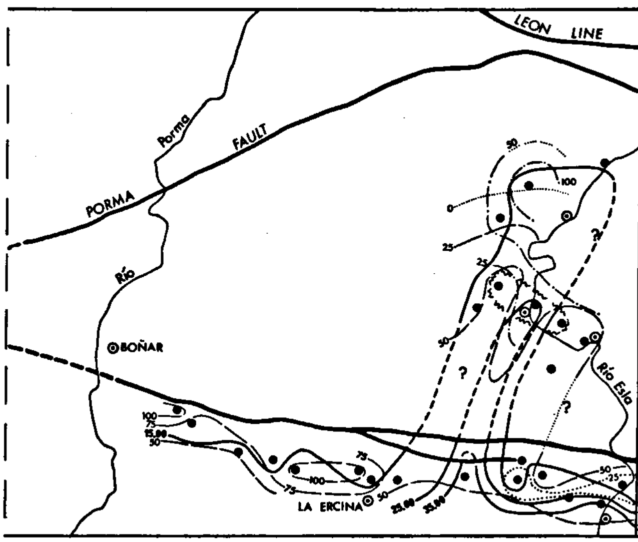


b	c	d	e	f	g	section	
				10.20	10.20	10	Peña Corada area
			5.40	13.00		11	
			6.00	6.60		12	
				10.50		13	
				4.50		14	
			12.00	10.20		15	
			7.00	7.50		16	
			10.70	8.50		17	
			5.20	6.30		18	
			2.00	12.00	2.00	19	
			3.00	3.00		20	
						21	
			14.10	7.50	2.70	22	
			10.20			23	
			8.40	4.80		24	
			5.30	2.60	2.30	25	Esla aut. area
			2.10	5.10		26	
			1.50	7.30	1.80	27	
	3.00		5.60	4.70	2.10	28	
10.10			1.80		5.20	29	
5.30			7.20	3.50		30	
			6.90	8.40		31	
			4.00			32	
				3.00		33	
4.20			10.20			9	A. syn.
			11.10	3.00		8	
						7	
				5.70	5.10	6	
				5.10	4.20	5	
1.50			3.00	18.30	2.10	4	P. syn.
				7.00	0.60	3	
			14.10	3.90	0.90	2	
			15.00			1	

Thickness of facies in Veneros Member

△ Peña Corada area
○ Esla autochthonous area
□ Alba syncline
■ Pedroso syncline

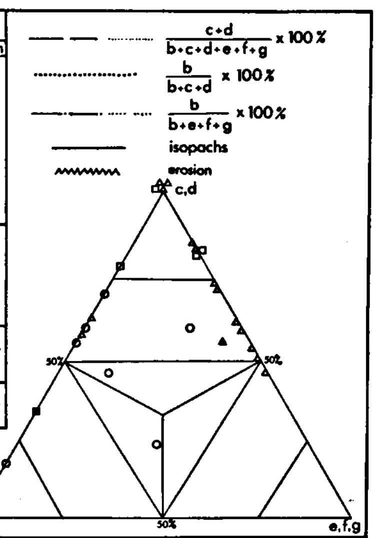




b	c	d	e	f	g	section	
	2.10					10	Peña Corada area
	6.90				4.50	11	
	8.40		1.50		6.90	12	
	11.10		2.10			13	
	10.80	13.20				14	
	1.80					15	
	16.60		1.20	1.20		16	
	8.50			1.50	6.00	17	
	14.00	9.00	6.00		9.70	18	
5.50	7.50					19	
	11.00	7.00	1.20	1.20	4.00	20	
2.40	10.80			6.00	7.00	21	
	8.40			9.30		22	
	9.80				4.00	23	
5.40	9.00					24	
4.50						33	Esla aut. area
18.00						32	
4.20	2.40		3.00			31	
5.40						30	
2.40	13.30		1.00		4.70	29	
10.80	11.40				3.00	28	
4.00	4.00	1.20				27	
5.50	7.70					26	
10.30	2.40					25	
1.50	5.40					9	A. syn.
						8	
	11.40					7	
	6.30				1.20	6	
	3.00				0.60	5	
6.60	3.30					4	P. syn.
16.30	2.00					3	
						2	
						1	

Thickness of facies in Member B

△ Peña Corada area
○ Esla autochthonous area
□ Alba syncline
■ Pedroso syncline



sedimentation) and the south, but presumably the difference was not very important, since sediments show a lateral connection and interrelation.

Plane of unconformity. – Directly below the plane of unconformity folding has been seen, developed in an irregular way. No cleavage, slickensiding or fractures, filled with quartz or calcite, were observed. Colour of the sediment in the immediate vicinity of the unconformity plane is reddish to reddish brown. Distinct karst phenomena were not observed. On either side of the erosion surface identical facies types are found. Above the erosion plane in section 29 thick-bedded biostromal sediments (facies c) were deposited, changing towards section 28 into facies b in which chert and silicified fossils are fairly common. In section 27 erosion removed the upper part of Member B. In section 26 an alteration occurs of facies c and facies b in both parts of Member B. Fossils in reversed position, and coral debris in general, often of large sizes, were often found. These observations can be interpreted in the following way.

The unconformity level found is definitely not of tectonic origin, as is shown by the perfectly similar facies types found on either side of the unconformity level, and by the absence of cleavage, fracturing and slickensiding. It must be a subaqueous erosion level. It might be the result of slight local uplift of the bottom before lithification of the sediment. The result of these more or less synsedimentary movements was the slumping and the folding observed (Pls. XIV–3, XIV–4).

From the measured strikes and dips the gravitational origin of the feature is evident. The facies changes (Pl. XIV–2) at this location supported the mobilization of the unlithified sediment during the movement of the bottom. An indication of this relatively unconsolidated state is found in the plastic deformation of the strata (Pls. XIV–3, XIV–4). Parts of this zone acted as a barrier during deposition of the sediments belonging to Member B and can best be compared with a boulder track of modern reefs.

Interpretation. – The 'transitional belt' of the Veneros Member in the area east of the Porma fault most probably formed a small slope during deposition of the sediments of Member B. Deposition of the sediments occurred progressively from north to south. Conditions on the slope and on the edge of the slope (in sections 14, 18, 20 and 28) were favourable for the growth of corals, especially of platy and branching tabulate corals and flat stromatoporoids. In analogy with results of investigations in recent reefal provinces (especially those of Maxwell, 1968) these branching and platy coral forms can be interpreted as typical of the reef front (Maxwell, 1968, p. 154–160, 229). In the northern part of the Esla autochthonous area cross-bedding, small gullies and pinching-out structures over rather small distances indicate the presence of currents, possibly in channels, while the general aspects of the sediments indicate moderate to low energetic conditions. This results in deposition of facies b, alternating with facies e, f and g, due to irregular introduction of siliciclastic material. Fairly large amounts of zaphrentoid corals and many burrowing tracks are also present. Essentially south of the Sabero-Gordón line we find another area with pronounced thickness differences. In sections 14, 18 and 20 bioherms are partly held responsible for the greater thicknesses. Slumps are nearly always present close to the bioherms, which indicates that slopes were already present during deposition. In the area northwest of Cistierna the influence of facies b becomes perceptible. Pyrite, bituminous matter, black shales and splintery black material indicate reducing environment and stagnant water during deposition. It was found in sections 14, 15, 16 and 28. In those places biostromal and biohermal sediments are quantitatively important. In shallow pools on the reef top and in interreefal sheltered areas temporarily stagnant water is imaginable. In the vicinity of section 28 the barrier previously mentioned, could temporarily cause such a stagnant water area.

The general picture, obtained from the environment in which Member B was deposited in the areas east of the Porma fault, does not differ much from that of the Veneros Member. In the western and northern parts of the area mainly subtidal, but occasionally intratidal flats occurred on which predominantly facies b sediments were deposited. The transitional zone distinguished in the Veneros Member acted as a slope. Small patchy reefs were formed in this zone, and slumping occurred. The southern limits of the area under consideration is formed by the hinge zone of the possibly active Sabero-Gordón line where conditions for reef growth were favourable and where slumping occurred. Between the reefal zones interreefal stagnant water sediments were deposited.

In the area west of the Porma fault only a fairly small number of control points are available. The situation sketched (Fig. 16) is not in conflict with the data collected in the available control points, and illustrates the author's idea that an important zone of the sedimentary basin was situated around the Sabero-Gordón line. The instability during deposition, as already recognized in the Veneros Member between sections 3 and 4, continued during deposition of Member B. Definite proof of this conception cannot be given since the northern part of the presumed basin was removed by the Upper Devonian erosion and the southern part is covered

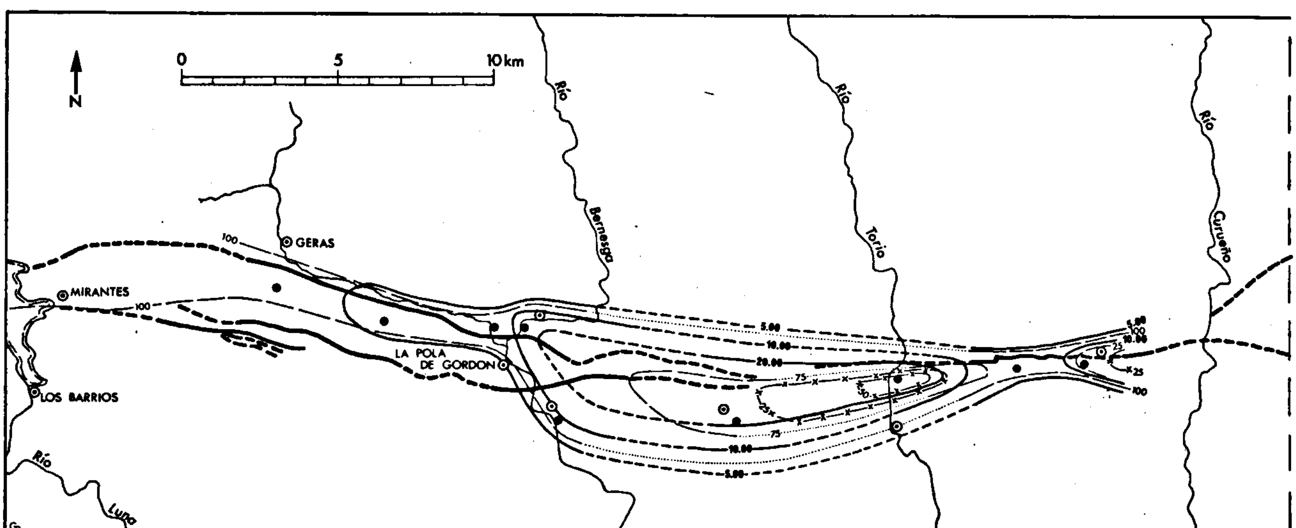


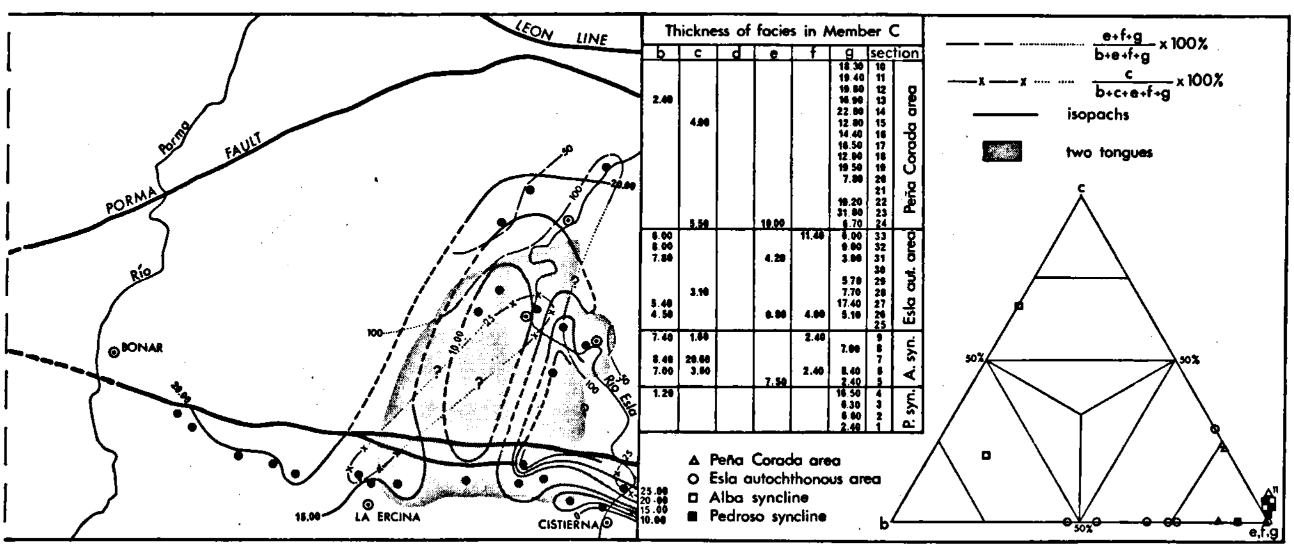
Fig. 17. Isopach-lithofacies map of Member C.

by Cainozoic sediments. Sedimentation started near Matallana. Here fragments of branching corals, typical of the reef front, accumulated in banks. Deposition of these fragments occurred more or less in place. Sedimentation gradually shifted towards the east and the west. In the west quiet-environment sediments (facies b) accumulated. Non-deposition occurred in the areas of sections 1 and 2. Sediments belonging to the upper part of the Veneros Member are laterally equivalent with the sediments of Member B in sections 3 and 4. The same is the case in section 8. Veneros Member sediments here are laterally equivalent with the sediments of Member B in sections 9 and 7.

Member C (Fig. 17)
East of the Porma fault, Member C first decreases, and then increases in thickness from west to east. Approximately in the place where the 'transitional belt' in the Veneros Member was recognized and where incipient reef growth occurred during deposition of Member B, thickness of Member C is smallest. From here eastward thickness increases rapidly, but Member C is now composed of two different tongues, the C_1 tongue and the C_2 tongue (Table 1). This holds for the Peña Corada area as well as for the Esla autochthonous area. Thickness also diminishes in the area northwest of Cistierna, and for section 21 non-deposition is assumed.

The values [(c)/(b+c+e+f+g)] 100 and [(e+f+g)/(b+e+f+g)] 100 were calculated for each control point. The former expresses the importance of biostromes. The latter expresses the importance of sediments deposited under moderate to quiet conditions on an open marine shallow flat. The quantity of this latter type of sediments is greatest in Member C, when compared with the other members. From the field observations and from the values calculated it may be inferred that biostromal sediments are present, starting approximately from the point where Member C is divided into two tongues. The percentage lines of 25 % were constructed. As may be inferred from those intervals of Member C accessible for investigation (exposure is poor in the Peña Corada area and important intervals are covered by debris), energy conditions were fairly high. Grain/micrite ratio is high. Crinoids formed the bulk of the fossil material. They were also observed in the siliciclastics. Brachiopods too, are a fairly important component (Fig. 7). The sedimentation of siliciclastic material is supposed to have taken place more or less synchronously throughout the entire area. This is not in conflict with the palaeontologic data (Chapter III).

Interpretation. – In the Esla autochthonous area facies b is characteristic. Bird's eye limestones, silicification and bioturbation, are significant. These are often indications of intertidal to supratidal sediments (cf. Chapter V). These features, together with minor slumping in the northern part of the autochthonous area and the erosion plane between Members B and C, lead to the conclusion that the depositional environment, already very shallow during deposition of the sediments of Member B, remained shallow, and that some slopes were present in the northern part of the area. The C_2 tongue of Member C is a strongly cross-bedded oolite (Pl. IX–6), composed of concentric ooids. The picture obtained from these observations is that of an open marine shallow subtidal flat in the south separated from a sheltered subtidal to intertidal and occasionally supratidal area in the north by an incipient organic barrier possessing true slopes. The source area of the siliciclastic material is unknown, but since the siliciclastic layers pinch out towards the southeast, it is most probably situated in the northwest. Presumably minor movements in the source area, disturbing the equilibrium, cause the supply of this siliciclastic material. The zones recognized in the underlying member were flooded by the siliciclastics. Reef builders re-established themselves during sedimentation of the



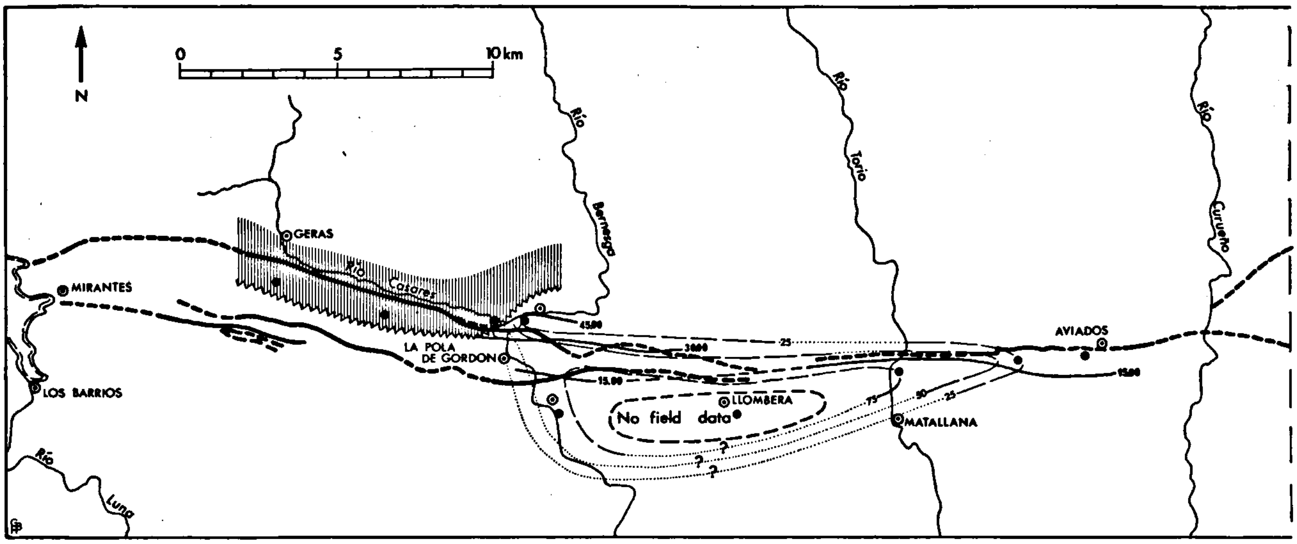


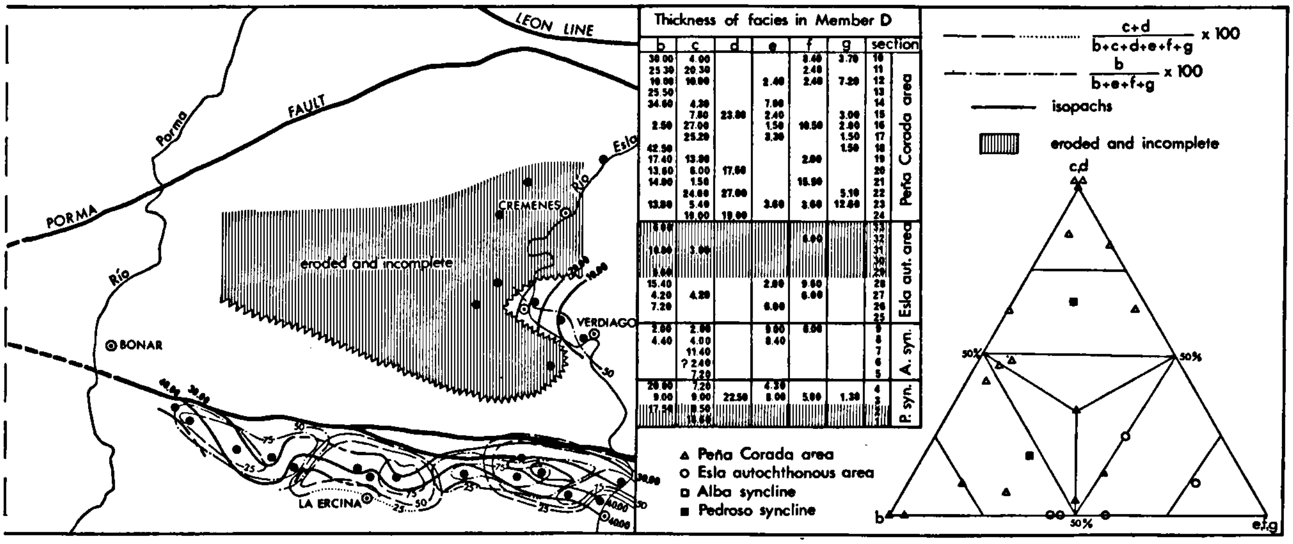
Fig. 18. Isopach-lithofacies map of Member D.

siliciclastics and formed biostromes. This happened more or less in the same zone where they were formerly present. Facies b is laterally associated with these biostromes. A second tongue (C_2) of siliciclastic material was subsequently formed, and flooded the first tongue and the biostromal reefs with associated sediments.

In the area west of the Porma fault, the situation as pictured for Member B has not essentially changed. Siliciclastic material is present throughout the whole area considered, interbedded and admixed in the limestone sequence. In the Pedroso syncline, however, it becomes a thin but very well traceable cross-bedded, rather pure sandstone member. The 15 samples qualitatively investigated (Table 3) are compositionally fairly immature (clay, plagioclase, microcline, muscovite and biotite are admixed). Together with grain size distribution and bioturbation features, this suggests a shallow depositional environment, not far from the source area. A different depositional environment probably occurred towards the centre of the basin (near section 7) where appreciable accumulations of branching tabulate coral fragments are found and where sedimentation started and kept pace with subsidence. Depositional conditions of Member B were not substantially changed in Member C. Only the quantity of red material (ferruginous and dolomitic) increases significantly. Sedimentation spread both towards the east and the west. Energy conditions in the east were quiet and facies b sediments were predominantly deposited. The bioclastic material deposited in the section near Matallana is supposed to have acted as a barrier between these facies b sediments and the siliciclastic material present especially to the west.

Member D (Fig. 18)
With regard to the thickness distribution of Member D it should be remembered that the upper part of the Portilla Limestone Formation in the northern part of the Esla autochthonous area is removed by the Upper Devonian erosion cutting progressively deeper into the Devonian sequence towards the north. The sediments removed include the entire Frasnian stage (Chapter III). For the northernmost part of the area no isopach nor lithofacies data are presented. In the southern part of the area east of the Porma fault, thickness seems largest just south of the Sabero-Gordón fault zone. Sections 17 and 23 show maximum thickness.
The $[(c+d)/(b+c+d+e+f+g)]$ 100 and $[(b)/(b+e+f+g)]$ 100 values were calculated for all control points. The former indicates the presence of biostromal and/or biohermal sediments. The 75 %, 50 % and 25 % lines are constructed. The latter value indicates the presence of the sediments, deposited under quiet to very quiet conditions.
Here, too, the 75 %, 50 % and 25 % lines were constructed. Biohermal and biostromal sediments occur mainly in sections 15, 16 and especially 17, as well as in sections 19, 22, 23 and especially 20 and 24. The facies b sediments chiefly occur in the areas in between. The presence of large amounts of chert, concentrated in layers in which large size nodules are present, must again be stressed. Silicification, especially of fossil fragments, is also common. Dolomitization and dedolomitization, too, are fairly common diagenetic processes.

The characteristics of the biostromal to biohermal sediments were dealt with extensively in Chapter IV. Beds in which allochthonous fossil fragments are present, alternate with beds containing fossils in situ in the screes. The first ones correlate with minor erosion surfaces with karst-like features (Fig. 8). Microscopically visible dedolomitization also suggests occasional emergence of the bioherm. The composition of this member is shown in Fig. 8. Corals and stromatoporoids are important faunal components. The considerable quantity of pellets present in various sections (especially



area	b	c	d	e	f	g	section
Peña Corada area	30.00	4.00			8.40	3.70	10
	25.30	20.30			2.40		11
	10.00	10.00		2.40	2.40	7.20	12
	25.50						13
	34.60	4.30		7.00			14
		7.80	23.00	2.40		3.00	15
	2.50	27.00		1.50	10.50	2.00	16
		25.20		3.30		1.50	17
	42.50					1.50	18
	17.40	13.00			2.00		19
	13.60	6.00	17.60				20
	14.00	1.50			15.90		21
		24.00	27.00			5.10	22
	13.00	5.40		3.60	3.60	12.60	23
		19.00	19.00				24
Esla aut. area	[illegible]						33
					[illegible]		32
	10.00	3.00					31
							30
	[illegible]						29
	15.40			2.00	9.60		28
	4.20	4.20			6.00		27
	7.20			6.00			26
							25
A. syn.	2.00	2.00		9.00	6.00		9
	4.40	4.00		8.40			8
		11.40					7
		? 2.40					6
		7.20					5
P. syn.	20.60	7.20		4.30			4
	9.00	9.00	22.50	8.00	5.00	1.30	3
	17.50	8.50					2
		13.60					1

coinciding with facies b) is also characteristic. The grain/micrite ratio is generally moderate. Facies b deposits and facies c and d sediments are inversely proportional.

Interpretation. – The picture obtained from these observations is fairly clear. The zone already recognized in Members C and B, in which incipient reef growth occurred, especially the part just south of the Sabero-Gordón zone, again acted as a zone in which reef growth, this time more strongly developed, is present. The bioherms in sections 15, 20, 22 and 24, recognized in the field, formed real build-ups that sometimes even emerged above sea level. In the exposures minor slopes could be measured after correction for the tectonic dip. In between these reefal areas low-energy sediments were deposited. These regions also incidentally emerged, or became extremely shallow. The dedolomitization described in these areas can thus be explained, as well as the abundance of nodular chert (Chapter V). These areas were possibly intratidal to subtidal lagoons.

In the area west of the Porma fault, the number of suitable control points is much smaller. The Upper Devonian disconformity cuts through the western sections of the Pedroso syncline. Only section 3 is almost complete. The uppermost portion of section 6 is inaccessible. From the remaining control points one obtains the impression that the sedimentary basin configuration did not essentially differ from that of Member C. In sections 3 and 4 sediments belonging to facies b occur in fairly important quantities. The only bioherm west of the Porma fault is present in section 3. The relatively close spacing of the isopachs in the valley of the Bernesga River may indicates synsedimentary bottom movements.

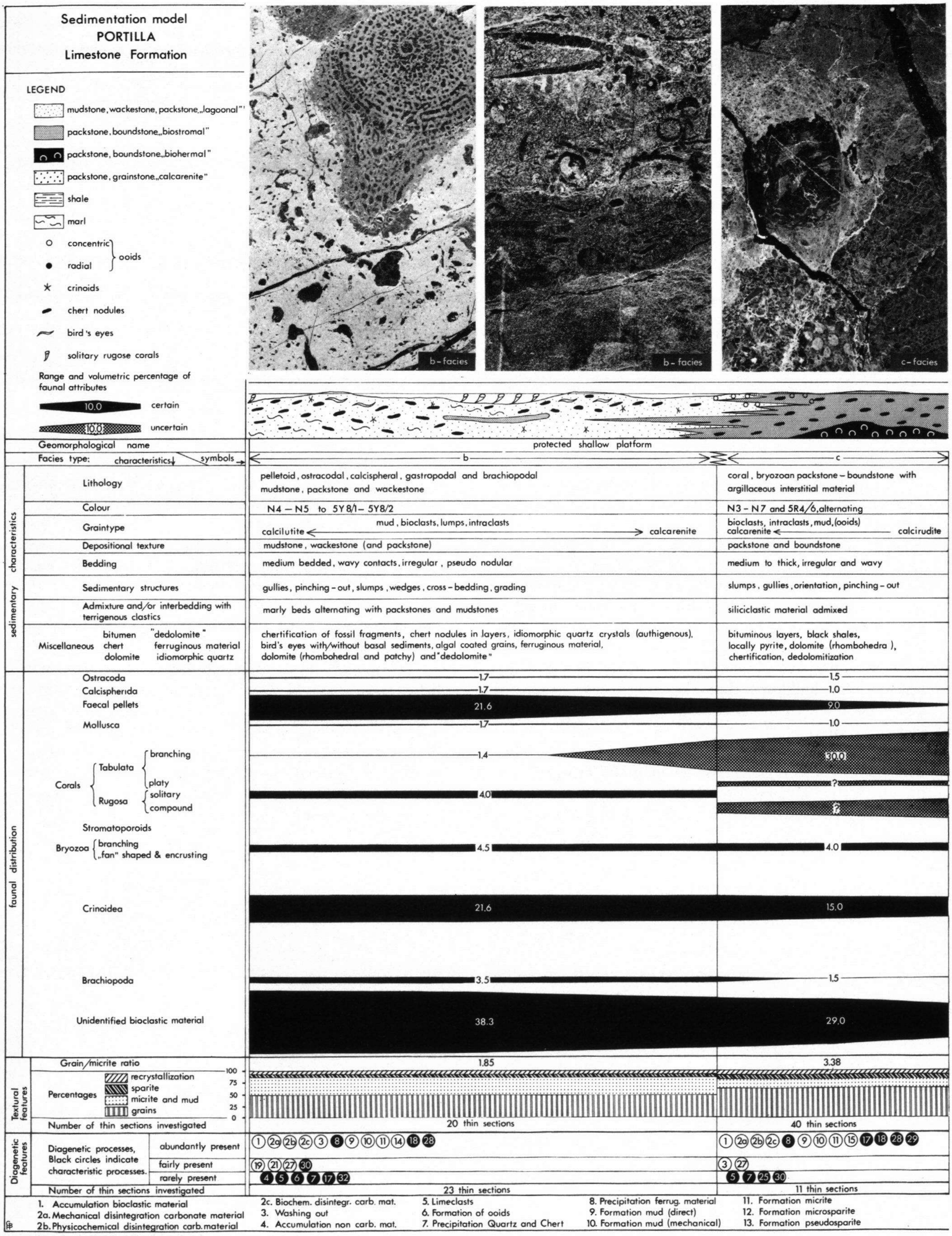
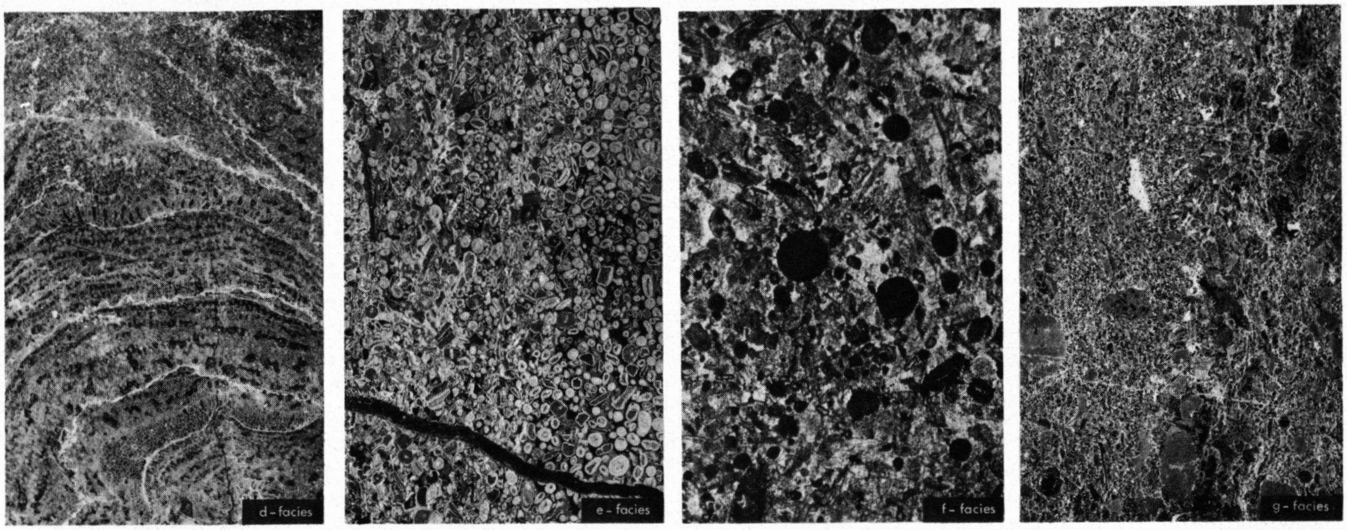
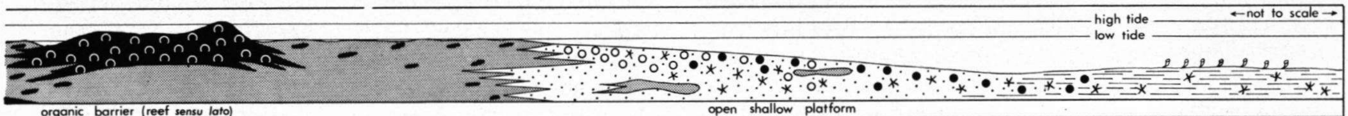


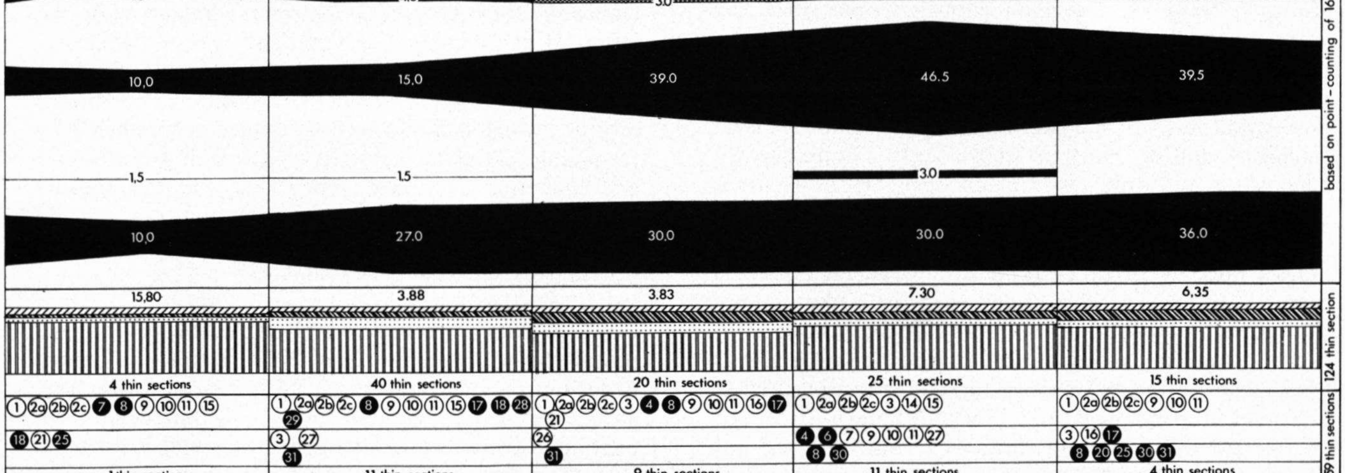
Fig. 19. Sedimentation model Portilla Limestone Formation.





d	c	e	f	g
coral, stromatoporal boundstone-packstone with a biohermal aspect	coral, bryozoan packstone-boundstone with argillaceous interstitial material	encrinal (encrinite) bryozoan grainstone	encrinal, bryozoan coral, brachiopodal packstone and wackestone with marly intercalations	bioclastic packstone and wackestone admixed with important amounts of siliciclastics
N8-N9 alternating with seams of 10YR8/6	N3-N7 and 5R 4/2 - 5R 5/2	N4-N5 and 5Y6/1 - 5Y8/1	5GY7/2 - 5YR4/1	N3-N2 and 5YR 3/2
bioclasts, intraclasts, mud calcirudite ←→ calcarenite	bioclasts, intraclasts, mud, ooids calcirudite ←→ calcarenite	bioclasts, intraclasts, (reef breccia) mud, ooids calcirudite ←→ calcarenite	bioclasts and ooids calcarenite ←→ calcarenite	bioclasts, mud, (ooids and pseudooids), calcarenite
packstone and boundstone	packstone and boundstone	packstone and grainstone	packst.-(boundst.)-(wackest.)-(grainst.)	packstone and wackestone
thick to very thick sometimes indistinguishable	medium to thick, irregular and wavy	medium, well bedded	irregularly bedded, thin-medium	moderately to well bedded, sometimes lamination
slumps, pressure phenomena, debris formation	slumps, gullies, orientation bioclasts, pinching-out	cross-bedding and gullies	cross-bedding, gullies, slumps	cross-bedding, gullies, grading, trapping
non siliciclastic material	siliciclastic material admixed	some well rounded quartz grains		silt-sized siliciclastics, admixed and interbedded. sometimes predominantly
pressure phenomena, erosion surfaces (10R 5/4), exfoliation, chertification, dolomitization-dedolomitization	dolomitization-dedolomitization idiomorphic authigenous quartz chertification		idiomorphic authigenous quartz dolomite (rhombohedral crystals)	ferruginous material, well rounded detrital quartz, ooids, (sheltering), algal boring

field observations



14. Precipitation fibr. cement crystals
15. Prec. equant cement crystals
16. Syntax. rim cem. & syntax. gr. growth rims
17. Replacement (patchy) dol.
18. Formation of DoRh crystals
19. Recrystallization *sensu lato*
20. Replacements Ca⇆Qz
21. Formation (micro) stylolites
22. Impregnated grains
24. Expulsion of impurities
25. Dedolomitization
26. Infilling (2nd phase) ferruginous mat.
28. Formation of bioturbation
29. Formation of pellets
30. Formation of voids
31. Mech. infilling open space struct.
32. Algal coating
33. Flatting carbonate particles

CHAPTER VII

SEDIMENTATION MODEL

'A conceptual carbonate model constitutes a synthesis of all pertinent data, necessary to represent the areal configuration, composition and environmental milieu of the major facies, into (sic!) a simple and orderly form.'

J. Dooge, 1966, p. 48

An attempt will be made to incorporate data so far collected and discussed, in a model showing the sedimentary environment of the Portilla Limestone Formation. Such a model has many advantages. It serves for representing in a simple and orderly form all pertinent data (Dooge, 1966, p. 48), and once the uniformity has been established and checked, it can be applied to other carbonate sequences. In Fig. 19 an idealized cross-section is sketched through the carbonate shelf under discussion. The three essentially different large groups of sediments are represented by the three distinctly different environments pictured; a quiet protected shallow subtidal and occasionally intertidal depositional environment with scattered patches of biostromes; an organic barrier, mainly subtidal, but occasionally intratidal, consisting of biostromes with some scattered bioherms, and a shallow platform, essentially subtidal, deepening somewhat in an open marine direction. From the regional distribution of the facies (cf. lithofacies maps Figs. 15–18) and from geometry (cf. Figs. 9–14) dealt with in the preceding chapter it appeared that facies belts, to a certain degree of predictibility, occupy more or less fixed positions on the shelf. Herefrom, the presented lateral succession and relation of facies is deduced. Because facies are more or less identical in the various members, but not present in equal quantities in the investigated subareas, the general aspect of these members always differs somewhat (Figs. 4–7). Uniformity, however, was good enough to recognize and correlate lithostratigraphic units. The variation in the overall aspect is a response to deposition according to an essentially uniform sedimentation model that shifts somewhat in space and time as a result of minor sea level changes and/or bottom movements. Consequently deposition occurred more in an open marine environment or more in a protected environment in a particular location, depending upon nature and quantity of the facies present. In order to illustrate general lithologic aspects of the various facies (negative) photomicrographs, characteristic of the facies, are given in Fig. 19. Point-counting, carried out on thin sections selected as the most representative ones for the facies, resulted in volumetric percentages of biota. Although it is not pretended that the representation in Fig. 19 is the last word, the author believes that the point-count results give a fair picture of the lateral variation in biota in the facies distinguished. This also holds for the diagenetic and other processes recognized in the various facies. Those indicated by a black circle are presumed to be characteristic of the facies under which they are tabled.

With regard to the problem of including the Huergas Formation, underlying the Portilla Limestone Formation, in one continous sedimentation model, we have to bear in mind the essential difference between the sedimentation patterns of the Huergas sediments and the Portilla Limestone sediments, viz. the supply of sediments (presumably terrigenous) from a hinterland. During deposition of the Huergas sediments conditions were not favourable for carbonate sedimentation. At the transition of the Huergas Formation into the Portilla Limestone Formation, supply of terrigenous siliciclastic material decreased. Sediments were mainly deposited as facies e and f, grading into the more open marine (Huergas?) shales. Where facies g predominates in the lower beds, transition is gradual from the Huergas Formation into the Portilla Limestone Formation; elsewhere it is fairly abrupt. The ooids present at the transition between the two formations give a fair impression of the changing environmental conditions. The Veneros Member has mainly been deposited under conditions prevailing on the open marine shallow subtidal platform. During deposition of sediments of Member B, incipient reef growth occurred in places favourable for reef builders. These favourable areas are often structurally controlled. Upgrowth of some small bioherms was possibly stimulated by slight movements of the sea bottom or by changes in the sea level. At the end of Member B sedimentation conditions suddenly changed. Siliciclastic sediments in Member C locally obscured the carbonates, elsewhere they became intensively admixed with the latter. Most probably the hinterland again supplied terrigenous material causing the death of many reef builders. In the overlying Member D depositional conditions returned which had previously prevailed in Member B. In favourable places (structurally controlled?) biostromes and bioherms developed. They acted as real organic barriers and gave rise to the accumulation or associated sediments, deposited in subtidal to intratidal and occasionally supratidal environments.

Nowhere does a complete lateral succession occur of the model here proposed. This is a result of general depositional conditions being fairly constant over vast areas. The model presented could be reconstructed from continuously repeating lateral relations and interfingerings. One fairly complete facies spectrum is present in the Veneros Member and in Member B in the Alba syncline, in an East-West direction. Here we pass from facies b via c to f and g. In terms of the model we pass from the protected subtidal and occasionally intertidal sediments, via the organic barrier sediments, to open marine subtidal shallow platform sediments. The example given suggests that the vicinity of the Porma fault acted as a land (?) area or as a submarine high in the direction of which quiet shallow conditions prevailed. The same conclusion can be deduced from the sediments east of the Porma fault. It is therefore concluded that the Porma fault area acted as a real high during sedimentation of the Portilla Limestone Formation. From similar relations the same conclusion can be derived for the León line in the north of the Esla autochthonous area.

This sedimentation model most probably applies to more formations in the Devonian of the Cantabrian Mountains, but remarks with regard to this question are beyond the scope of the present study.

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