LEIDSE GEOLOGISCHE MEDEDELINGEN, Deel 46, Aflevering 2, pp. 157–235, 17-2-1971

THREE UPPER CARBONIFEROUS, LIMESTONE-RICH, HIGH-DESTRUCTIVE, DELTA SYSTEMS WITH SUBMARINE FAN DEPOSITS, CANTABRIAN MOUNTAINS, SPAIN

BY

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ABSTRACT

In the eastern part of the Cantabrian Mountains, northwestern Spain, Upper Carboniferous strata crop out. In the Pisuerga area an Upper Moscovian (≈ Westphalian D) limestone-rich sequence is exposed in a number of structural units. Correlations between the various structural units are based on lithostratigraphic characteristics and paleontological dating with fusulinids and calcareous algae. The various limestone units show rapid lateral transitions into siliciclastic deposits. The siliciclastic deposits are mainly interpreted as deltaic deposits.

Three distinct delta systems are distinguished which, like the recent Rhône delta, are all of the wave-dominated, high-destructive type. The oldest delta system is associated with important turbidite deposits, which indicates that the delta prograded into relatively deep water (i.e. ⩾ 125 m). The second delta system is associated with only minor turbidite sequences but is relatively rich in coals. These two delta systems prograded from approximately SW to NE. To the NW open marine deposits with limestones were formed, to the NE fine-grained shelf deposits and limestones were deposited. After deposition of the second delta system, tectonic tilting in the southwestern part of the area caused formation of the Vergaño disconformity. In the remainder of the area this level can be recognized as a deepening or transgressive sequence. In the northeastern part of the area important slumping movements were caused by the tectonic movements.

After this tilting the river forming the delta was diverted to another area and the only source of siliciclastic sediments was a competent longshore drift system. The existence of such a longshore drift system is proved by the presence of quartz arenitic pebbles and cobbles in compositionally mature sandstones, as the deltaic deposits do not contain such coarse siliciclastics and are furthermore of lithic arenitic composition. During this period a zone with shallow marine deposits in the southwestern part of the area can be distinguished from submarine canyon and fan deposits (turbidites) in the southern and northeastern parts of the area. The third wave-dominated, high-destructive delta system prograded from SW to NE over these deposits. Open marine, shelf and shelf slope deposits are again present to the NW and NE. After deposition of the third delta system the Leonian phase gave rise to the Leonian disconformity, and a new basin configuration resulted.

In the interval studied a basin margin to the southwest of the area studied can be inferred from the facies distributions. The presence is proved of a synsedimentary fault, which separates the present Casavegas Syncline from the remainder of the area. This fault mainly influenced thickness distributions, and had but little effect upon the facies distributions.

The limestones were deposited in a wide range of environments, i.e. lagoonal to open shelf. The rapid lateral transitions into coarse siliciclastics are interpreted as indicating that the limestones were formed in slightly shallower water than the surrounding siliciclastics. Together with the generally muddy character, the absence of an organic framework and the presence of all kinds of algae, this indicates that the limestones are biogenetic bank deposits.

The data collected have led to a redefinition of the Vañes Formation and to replacement of the Sierra Corisa Formation by the Vergaño and Covarres Formations.

SUMARIO

En la parte oriental de la Cordillera Cantábrica afloran rocas pertenecientes al Carbonífero superior. En la región del Pisuerga existe una serie rica en calizas expuesta en distintas unidades estructurales, de edad Moscoviense superior (≈ Westfaliense D). Las correlaciones entre estas unidades estructurales se basan en características litoestratigráficas y en dataciones paleontológicas por medio de fusulínidas y algas calcáreas. Las distintas unidades de calizas muestran unas transiciones laterales rápidas a depósitos siliciclásticos. Estos últimos han sido interpretados como sedimentos deltaicos.

Se distinguen 3 sistemas deltaicos que, por analogía con el delta reciente del Ródano, son del tipo destructivo, dominado por el oleaje. El sistema deltaico mas viejo esta asociado a depósitos importantes de turbiditas, lo que indica que el delta ha avanzado hacia aguas relativamente profundas (⩾ 125 m). En el segundo sistema deltaico estas turbiditas solamente se encuentran esporádicamente, habiendo una abundancia relativamente grande de estratos de carbón. Estos dos sistemas de delta han avanzado aproximadamente del suroeste al noreste. En el noroeste fueron formados depósitos marinos con calizas, mientras que hacia el noreste fueron depositados sedimentos de *shelf* de grano fino y calizas. Despues de la deposición del segundo sistema deltaico se produjo un basculamiento de la parte suroccidental de la región, formándose la disconformidad de Vergaño. En el resto de la región este nivel puede ser reconocido como una secuencia transgresiva o de profundización. Hacia el noreste movimientos tectónicos dieron lugar a la formación de importantes *slumps*.

Después del basculamiento el río que formaba el delta se desvió hacia otra región, por lo que los sedimentos siliciclásticos fueron abastecidos por un sistema de corrientes con un rumbo paralelo a la costa. La existencia de tal sistema de corrientes se de-

muestra por la presencia de cantos y bloques cuarzo-areníticos en areniscas de composición madura, mientras que los depósitos deltaicos no contienen tales sedimentos siliciclásticos de grano grueso y además son de composición lítica-arenítica. Durante este periodo se pueden distinguir una zona con depósitos marinos poco profundos en la parte suroccidental de la región y depósitos de cañones submarinos y aluviones (turbiditas) que se encuentran hacia el sur y el noreste. El tercer sistema deltaico avanzó, cubriendo estos últimos depósitos, desde el suroeste hacia el noreste, dándose condiciones marinas abiertas y de ladera de *shelf* en el noroeste y el noreste. Después de la deposición del tercer sistema deltaico la fase Leonesa dió origen a la disconformidad Leonesa, de lo que resultó una nueva configuración de la cuenca sedimentaria.

En el intervalo estudiado se ha podido deducir de las distribuciones de facies la existencia de un margen de cuenca, localizado hacia el suroeste de la región de estudio. Igualmente se demuestra la presencia de una falla sinsedimentaria, que separa el actual sinclinal de Casavegas del resto de la región. Esta falla principalmente dió lugar a los espesores sedimentarios pero tuvo poco efecto sobre las distribuciones de facies.

Las calizas fueron depositadas en una escala amplia de medios ambientales, desde lagunas hasta un *shelf* abierto. Las rápidas transiciones laterales a rocas siliciclásticas de grano grueso han sido interpretadas como una indicación que las calizas fueron formadas en aguas un poco menos profundas que aquellas. Su caracter generalmente fangoso, la ausencia de una construcción orgánica y la presencia de todos los tipos de algas, indican que las calizas son bancos biogenéticos.

Los datos recogidos han conducido a una redefinición de la Formación de Vañes y a un reemplazo de la Formación de Sierra Corisa por las Formaciones de Vergaño y Covarres.

SAMENVATTING

In het oostelijk deel van het Kantabrisch Gebergte, noordwest Spanje, zijn sedimenten van bovenkarbonische ouderdom goed ontsloten. In het Pisuerga-gebied is een kalkrijke opeenvolging van Boven Moscovien ouderdom ($\approx$ Westfalien D) aanwezig in een aantal synclinalen en anticlinalen. Correlaties tussen de verschillende structurele eenheden zijn gebaseerd op lithostratigrafische kenmerken en op paleontologische dateringen met behulp van fusuliniden en kalkalgen. De verschillende kalkafzettingen vertonen vrij abrupte overgangen naar siliciklastische sedimenten. Deze siliciklastische sedimenten zijn grotendeels afgezet in deltaische milieus.

Drie verschillende deltasystemen zijn te onderscheiden welke, net zoals de recente Rhône delta, van het door golfwerking gedomineerde, sterk destructieve type zijn. Het oudste deltasysteem is geassocieërd met belangrijke turbidietafzettingen wat er op wijst dat de delta gevormd werd in relatief diep water ($\geqslant$ 125 m). Het tweede deltasysteem is met minder belangrijke turbidietafzettingen geassocieërd, maar is relatief rijk aan kolen. Deze twee deltasystemen groeiden van het zuidwesten naar het noordoosten toe aan. In het noordwesten werden kalken en ondiep mariene afzettingen gevormd en in het noordoosten eveneens kalken, maar daar tesamen met fijnkorreligere "shelf" sedimenten. Na de afzetting van het tweede deltasysteem werd in het zuidwestelijke deel van het gebied de Vergaño-disconformiteit gevormd als gevolg van tektonische opheffing. In andere delen van het bestudeerde gebied kan dit niveau herkend worden als een transgressieve of dieper wordende sequentie. In het noordoostelijke deel van het gebied hadden deze tektonische bewegingen het afglijden van dikke kalklichamen ten gevolge.

Na deze tektonische bewegingen volgde de rivier die de delta had gevormd een andere loop, en siliciklastische sedimenten werden toen enkel aangevoerd door een stranddriftsysteem. De aanwezigheid van kwartsarenietische rolstenen, in qua samenstelling rijpe zandstenen, bewijst dat een dergelijke stranddrift bestaan heeft daar in de onderliggende deltaïsche afzettingen dergelijke grove bestanddelen ontbreken en de zandstenen bovendien een lithies arenietische samenstelling hebben. Gedurende deze periode werden in het zuidwestelijke deel van het gebied ondiep mariene sedimenten afgezet, terwijl in het zuiden, midden en noordoosten submariene canyon en waaier afzettingen (turbidieten) werden gevormd.

Het derde, door golfwerking gedomineerde, sterk destructieve deltasysteem dat deze afzettingen bedekt, groeide, net zoals de twee eerste systemen, van zuidwest naar noordoost toe aan. In het noordwesten en noordoosten werden weer ondiep mariene, "shelf" en "shelf slope" sedimenten afgezet. Na de vorming van dit derde deltasysteem werd ten gevolge van tektonische bewegingen, die aan de Leonese fase worden toegeschreven, de Leonese disconformiteit gevormd. Na deze bewegingen vormde zich een nieuw sedimentair bekken.

De verspreiding van de diverse faciëstypen geeft aan dat ten tijde van de afzetting van het bestudeerde sedimentpakket, er een bekkenrand ten zuidwesten van het bestudeerde gebied lag. De aanwezigheid van een synsedimentaire breuk, die de huidige Casavegas Synclinale van de rest van het gebied scheidde, wordt aannemelijk gemaakt. Deze breuk beinvloedde voornamelijk de diktes van het bestudeerde pakket en had slechts geringe invloed op de verspreiding van de faciëstypen.

De kalken werden in een groot aantal verschillende milieus afgezet. Deze milieus varieërden van lagunair tot open marien. De snelle overgangen naar siliciklastische sedimenten worden geïnterpreteerd als een aanwijzing dat de kalken in iets ondieper water werden gevormd dan de aangrenzende siliciklastika. Tesamen met het kwantitatief zeer belangrijke aandeel van kalkmodder, het ontbreken van een organisch rifgeraamte en de aanwezigheid van vele soorten algen in vaak grote hoeveelheden, geven deze snelle overgangen aanleiding tot de interpretatie van de kalken als biogenetische bankafzettingen.

Op grond van de verzamelde gegevens bleek het wenselijk om de Vañes Formatie opnieuw te definiëren, de naam Sierra Corisa Formatie te laten vervallen en de namen Covarres Formatie en Vergaño Formatie in te voeren.

CONTENTS

CHAPTER I

INTRODUCTION

GENERAL ASPECTS OF THE CANTABRIAN MOUNTAINS

The Cantabrian Mountains in the north of Spain consist of a Paleozoic core, bordered in the east and in the south by Mesozoic and Cainozoic deposits. In the west there is a transition into the Cryptozoic and Variscan igneous and metamorphic rocks of Galicia.

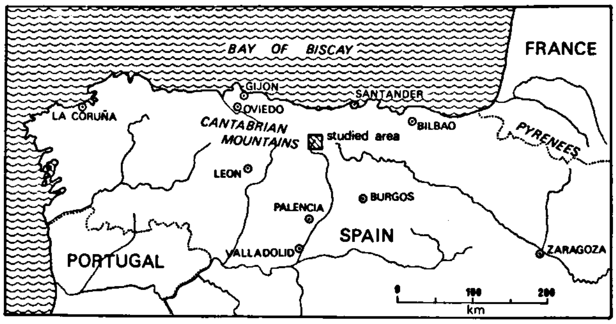

Fig. 1. Index map of northwest Spain.

Their present topographic expression is mainly due to an uplift during the Cainozoic (de Sitter & Boschma, 1966).

A synthesis of the development of the Variscan orogene, of which the present Cantabrian Mountains form only a part, was given by Matte (1968). Wagner (1970) presented a general outline of Carboniferous stratigraphy in this area.

For purposes of a general background it is sufficient to state that from the Lower Cambrian until the end of the Devonian, sedimentation in this area was typical of a stable shelf environment. Formations can be traced over tens to hundreds of kilometres and most sediments are evidently shallow water deposits. This entire sedimentary pile has a thickness in the order of 2 to 3 km.

Towards the end of the Devonian, uplift and erosion (de Sitter, 1965; van Adrichem Boogaert, 1967), resulted in an important hiatus in the central Asturian area and along part of the León line. This uplift was followed by another period of very uniform sedimentation during which limestones were deposited in most of the Cantabrian Mountains.

This relatively uniform picture changed drastically during the lower and middle Westphalian, when the important Curavacas folding phase (Palentian phase of Wagner & Wagner-Gentis, 1963) occurred. The

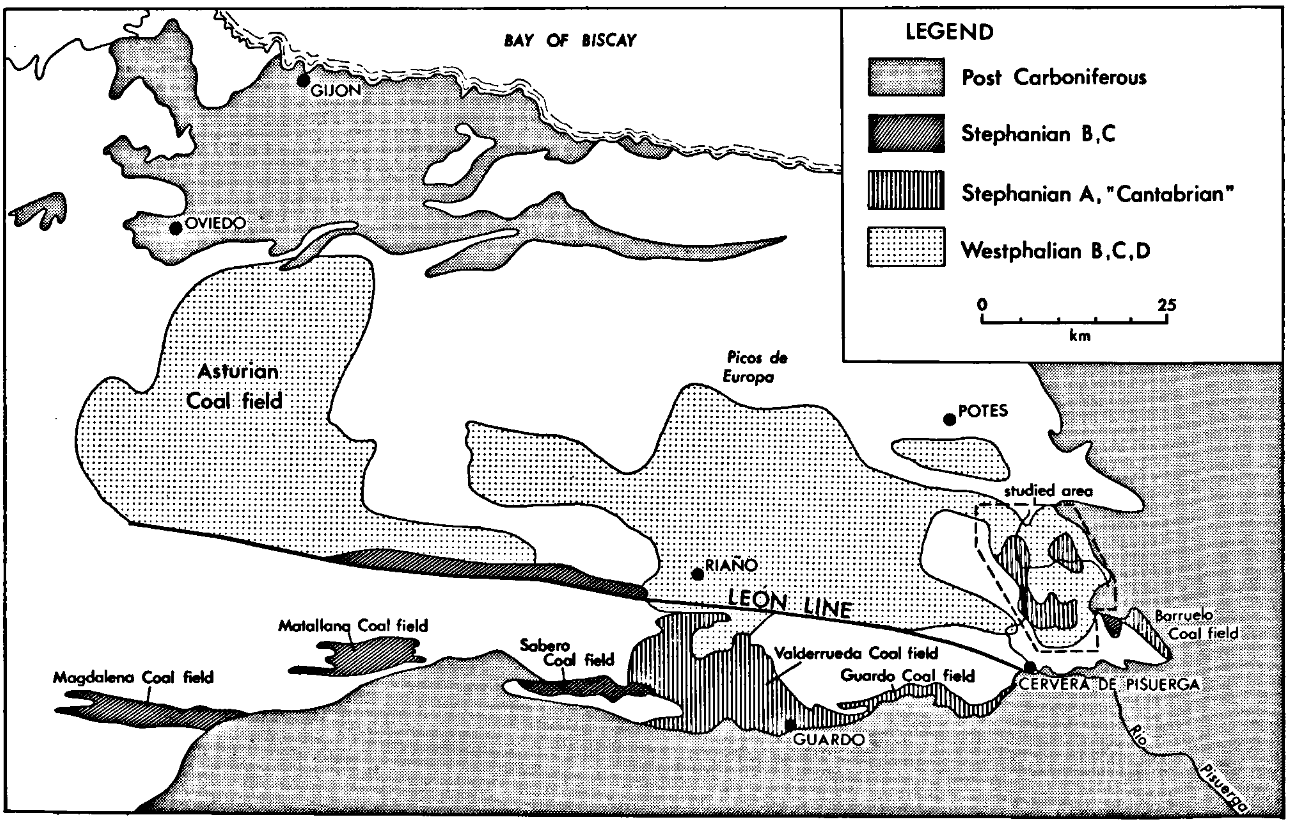

Fig. 2. Approximate distribution of preserved Upper Carboniferous strata in the Cantabrian Mountains. Lower Carboniferous and older deposits not differentiated. Modified after de Sitter (1965) and Wagner (1970).

sediments deposited after these tectonic movements are partly preserved in a number of so-called "basins" (de Sitter, 1965). These "basins" (Fig. 2) are mainly tectonic units which are merely fragments of the originally much more extensive sedimentary basins.

These Upper Carboniferous sediments contrast strongly with the earlier ones in that they are typified by great lateral and vertical variability and by their great thickness. The Westphalian alone, for instance, is over 4000 m thick in many places.

SUBJECT OF STUDY

In these Westphalian deposits marine and non-marine strata show rapid lateral and vertical transitions into one another. In the marine parts of the sequences limestones are common. Some of these limestones can be traced for several kilometres; others, however, abruptly wedge out or interfinger with siliciclastic deposits. Isolated lenses, sometimes forming a distinct band, may also occur. The cause(s) of these irregularities in the limestone deposits was (were) the original subject of this study. It was soon realized that these causes could not be understood unless the limestones were studied together with the surrounding siliciclastics on a regional scale. The Pisuerga area, from which published paleontological data were available, was selected for this study.

PREVIOUS WORK ON THE PISUERGA AREA

The Pisuerga area forms the easternmost part of the Paleozoic core of the Cantabrian Mountains (Fig. 2). General aspects of this area were discussed by de Sitter & Boschma (1966). Their map and paper constituted a useful basis for the present study. Paleontological data permitting of stratigraphic correlations between the various structural units have been published by van Ginkel (fusulinids, 1959, 1965), Wagner (continental floras, 1955, 1960, 1962, 1966) and Rácz (calcareous algae, 1966). Rugose corals were studied by de Groot (1963), but did not yield significant stratigraphic data.

With a view to completeness it should be mentioned that work on continental floras (Wagner, Sheffield), brachiopods (Winkler Prins, Leiden), conodonts (Varker, Leeds), fusulinids (van Ginkel, Leiden), goniatites (Wagner-Gentis, Sheffield) and lamellibranchs (van Amerom, Heerlen) from this area is still in progress. As most of these studies are concentrated on younger deposits they need not concern us here. In the near future their work may nevertheless give rise to further refinements of correlations in the upper part of the interval studied.

Other important papers on the Pisuerga area are by Brouwer & van Ginkel (1964), Frets (1965), Nederlof & de Sitter (1957), Nederlof (1959), Wagner & Wagner-Gentis (1963), Wagner & Winkler Prins (1970), Wagner & Varker (1970) and Reading (1970). Of these papers only the ones by Nederlof (1959) and Reading (1970) attempt an analysis of sedimentary facies. The former author made a number of general statements on facies associations and their distribution, but concentrated on a statistical analysis of turbidite sequences. The latter presented detailed descriptions and interpretations of sedimentary facies but, in view of the nature of his study, was unable to give a regional interpretation of their distribution. Such an interpretation, however, was attempted by van de Graaff (1970a). In the other papers mentioned, views on stratigraphical relationships and basin development in the Pisuerga area are also presented.

SELECTION AND DELIMITATION OF INTERVAL STUDIED

With the published paleontological data available it was possible to select the Upper Moscovian ($\approx$Westphalian D) strata as the most promising deposits for studying the relationships between siliciclastics and limestones. The reasons for this choice were: (1) the relative abundance of limestones, (2) the abundant fusulinids and calcareous algae which permit regional correlations to be quite reliable, (3) the paucity of structural complications outside the central part of the area.

The approximate distribution and general tectonic setting of the Upper Moscovian deposits is shown in Figure 3 together with the names of the respective limestone members.

The lower boundary of the interval studied was taken at a black shale/mudstone horizon below the Socavón and Camasobres Limestone Members. This shale/mudstone is well mappable and appears to occupy a relatively constant position in the sequence. For two reasons the base was not taken at the base of the Socavón Limestone Member. First of all because of the interesting facies sequences below the Socavón Limestone Member, which are very important for the interpretation of the entire area. Secondly because of the discontinuous nature of the Socavón Limestone Member, whereas the black shale/mudstone is easy to trace. Probably the same black shale/mudstone is present below the Agujas Limestone Member. Because of poor exposure, section measuring in the Redondo Syncline was usually begun at the base of the limestone itself. A few of the sections depicted in Enclosures 1 and 2 begin below the black shale/mudstone horizon (Sections 1, 4, 11, 15). This was done in order to include evidence of lithostratigraphic correlations with the Redondo area.

The upper boundary of the interval studied was drawn at the Leonian disconformity, as described by Wagner & Winkler Prins (1970) and Wagner & Varker (1970) and accepted by van de Graaff (1970a). This disconformity was first described by Nederlof (1959) as a local diastem. Although it is best exposed SE of the Cabra Mocha hill (Section 3) it can be traced around most of the Castillería Syncline. The locally very irregular disconformity surface is in most places covered by well washed and well sorted quartz arenites. A seatearth is sometimes present in this sandstone. In the central part of the area the presence

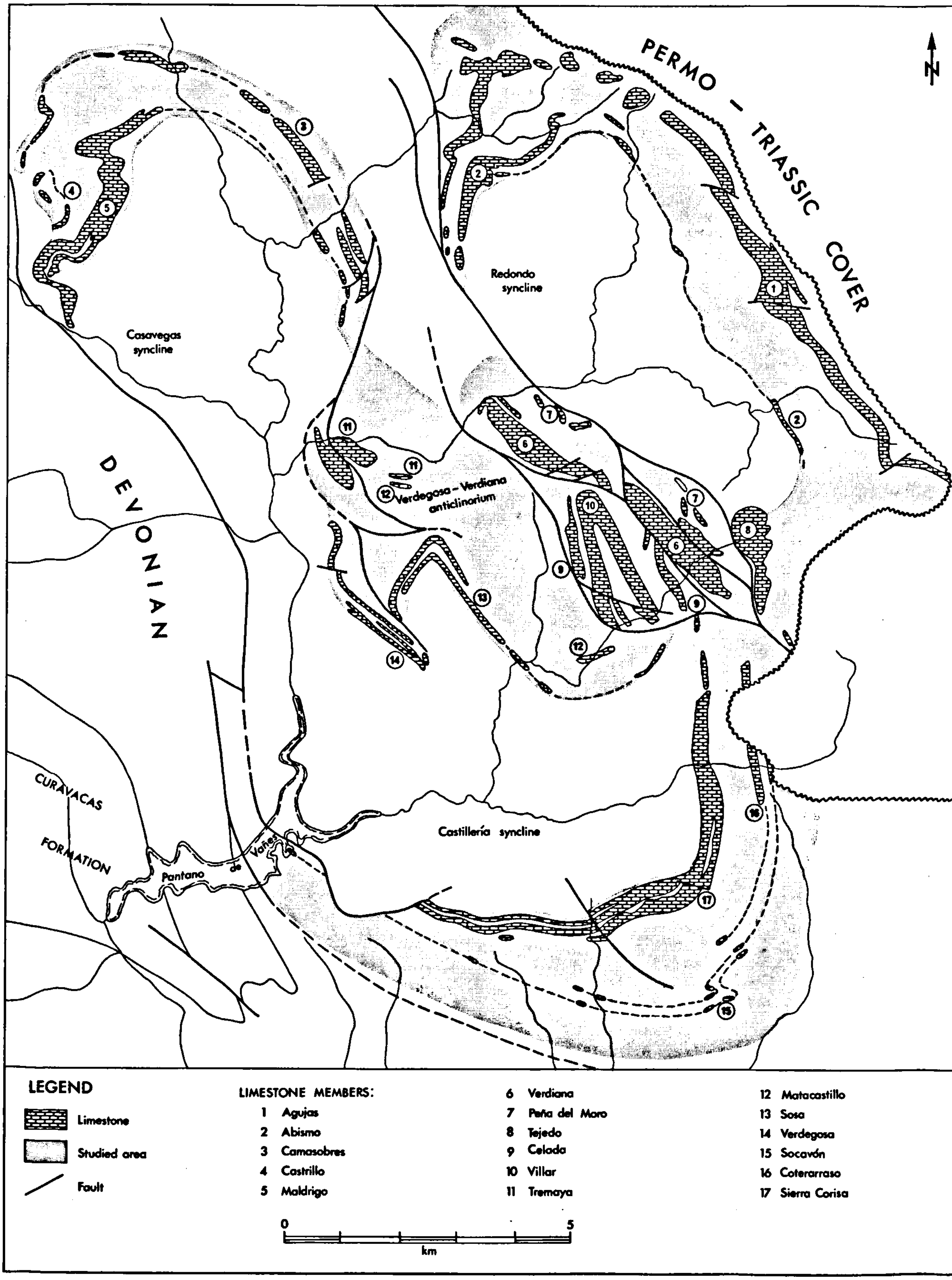

Fig. 3. Approximate distribution of Upper Moscovian strata in the Pisuerga area.

of such a mature quartz arenitic sandstone at approximately the presumed horizon in the succession is considered to indicate the disconformity. In the Casavegas Syncline no clear disconformity was observed, with the possible exception of Section 11. In this area Wagner & Varker (1970) place the disconformity at the level of some limestone breccias just above the Maldrigo Limestone Member. In the Redondo Syncline there is a very good exposure of the disconformity at the location at Section 25, but more to the north it cannot be traced. The upper boundary of the interval studied is there placed at the top of the Abismo Limestone Member or its correlatives. The correctness of this is indicated by the presence of a reworked tuff a few tens of metres above a lens of the Abismo Limestone Member in Section 21. This tuff is compositionally very similar to a sill occurring in Section 3 (447–450 m – not depicted). Although this is a sill and not an extrusive rock it suggests a correlation between Sections 3 and 21 which was already presumed by Wagner & Varker (1970) on the basis of the paleontological data.

It should be noted here that I do not follow de Sitter & Boschma (1966) in their contention that their Figure 7 (p. 214) represents a true unconformity (the Yuso-Cea unconformity at San Cristóbal hill). I agree with Wagner & Winkler Prins (1970) that it is more likely to be a fault contact. Which interpretation may be correct is, however, quite immaterial, for even if it is a true unconformity it is well above the upper boundary of the interval studied as defined here.

PETROGRAPHIC CLASSIFICATION SYSTEMS USED

Gilbert's (1954) sandstone classification is used in an adapted form as proposed by Dott (1966). As Dunham's (1962) [1] classification is used for the carbonates, confusion may occur owing to the similarity in the terminology. In Figure 4 the two classification systems are compared. As it only makes sense to compare clastic sediments, boundstones and crystalline carbonates are omitted from this comparison.

A major difference between siliciclastic and calciclastic sediments is the extrabasinal origin of the former and generally intrabasinal origin of the latter. This intrabasinal origin makes it rather meaningless to consider compositional maturity in addition to textural maturity when studying calciclastic instead of siliciclastic sediments. For this reason the mud content, i.e. the textural maturity, is the basis of Dunham's classification.

Dunham (1962) uses the name grainstone for mud-free calciclastic sediments. According to him a cal-

[1] In a few cases the name according to Folk (1959, 1962) is also used.

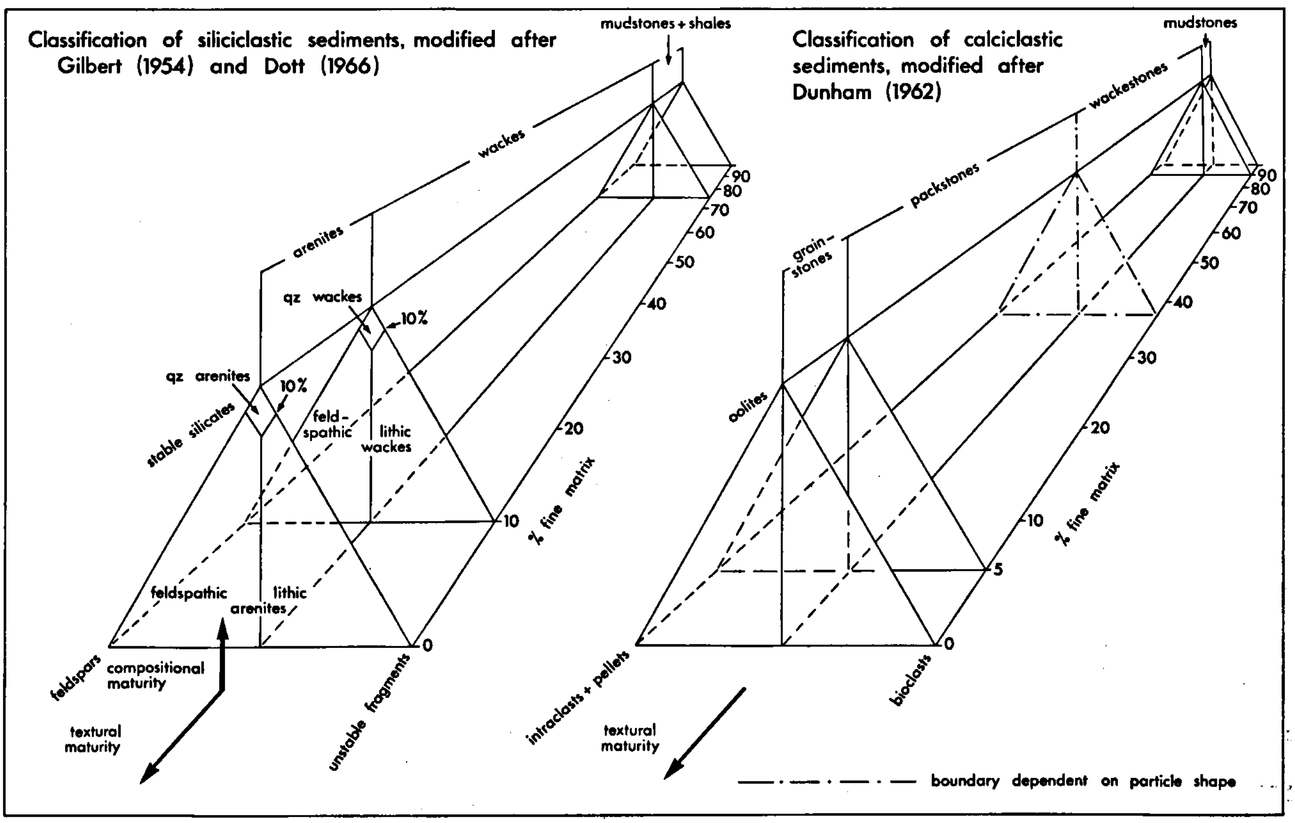

Fig. 4. Comparison of the classifications of Gilbert (1954) and Dunham (1962).

carenite should be called a packstone if containing more than 1% mud. Compared with the 10% matrix which may be present in Gilbert's arenites, or the up to 5% matrix which may occur in Pettijohn's (1957, p. 296) orthoquartzites, this 1% boundary seems too strict. A 5% boundary line is suggested. The division between grain-supported and mud-supported clastics has no equivalent in Gilbert's classification. This boundary cannot be accurately fixed, as the shape of the particles has so great an influence on the maximum porosity of grain-supported sediment. For randomly packed spheres (e.g. oolites) this boundary should be set at about 40% mud. For some fossiliferous sediments, however, it might be set at 70 to 80% mud (Dunham, 1962, Plate II). Because of this variability the boundary is not drawn in full line in the diagram. The boundary between wackestones and mudstones is arbitrarily taken at 10% grains.

It is obvious from the foregoing that Gilbert's wackes are the siliciclastic equivalents of most of the packstones and wackestones. The siliciclastic mudstones comprise both wackestones and mudstones as Dott (1966) arbitrarily takes the wackes-mudstones boundary at 25% grains.

Influence of limestone clasts on the classification of siliciclastics

In Gilbert's classification the abundance of unstable fragments determines the relative compositional maturity of the sediment. Limestone clasts are in general very unstable, and if present in a sandstone it should, for instance, be classified as a lithic arenite. This is, however, misleading, as the limestone clasts are commonly of intrabasinal origin. In the area studied this intrabasinal origin can commonly be proved. For this reason the compositional maturity of sandstones is determined without counting the limestone clasts present.

Further remarks

The term mudstone will be used in this paper for those siliciclastic deposits which contain a fair amount of macroscopically visible grains in a fine matrix. Shale will be used for uniformly fine-grained deposits with few or no macroscopically visible grains. Siltstone will be used for deposits of the appropriate grain sizes. It is fully realized that in many cases these distinctions are very hard to make on account of the complete gradations between these categories, but nevertheless it seems useful to make an attempt.

Another difficulty when trying to classify such fine-grained sediments is the possibility of extensive recrystallization at low temperatures, which may be accompanied by the development of cleavage. As axial plane cleavage is developed in the northwestern part of the area, this is a distinct possibility. In some sandstones there are also indications that the matrix has to some extent been recrystallized.

METHODS AND TECHNIQUES USED

As the lateral variability precluded a complete reliance on measured sections for an understanding of the stratigraphic problems in the interval studied, some mapping had to be done. This was partly done on enlargements at a scale of 1:10,000 of sheets 81, 82, 106 and 107 of the Mapa de España 1:50,000. As these maps are of rather poor quality, most of the mapping was carried out on aerial photographs with an approximate scale of 1:27,000. Control points and geological boundaries in most parts of the area studied were indicated on these photographs in the field. The photogeologic map was then projected on the topographical map by means of a sketch master. The resulting map at a scale of 1:25,000 was reduced and simplified to the form shown in Enclosure 3.

Sections were logged at scales varying from 1:10 to 1:1000. The scale used depended on the amount of variation in the interval measured and on whether fundamentally new facies types were present or not. In order to save time and to standardize data collecting, check lists were used when measuring limestones and turbidite sequences.

Sampling was directed at obtaining representative samples of the coarse siliciclastics and of the limestones. Shales and mudstones were neglected. At an early stage in the survey, sampling was relatively intensive for some sections. This was done in order to verify field descriptions of the lithologies. Later, fewer samples were considered sufficient to check field descriptions.

In total about a thousand thin sections were studied, half of which were from limestones. In addition about a hundred acetate peels and polished surfaces were made of limestone samples. Staining of the limestone samples with Alizarine Red-S was helpful in distinguishing between calcite and dolomite. The relative proportions of the various components of the limestones were estimated with the aid of percentage charts. The composition of the siliciclastics, however, was determined by point counting. Three hundred points per thin section were counted for over 90 representative samples.

The paleontological dating of the samples collected was carried out on fusulinids and algae. Van Ginkel (Leiden) studied the fusulinid samples and only after he had assigned a relative age was the stratigraphic location of the sample revealed to him. Wishful interpretation was thus avoided. The calcareous algae were studied by myself. As I did not take the trouble of giving the slides a random code number the relative ages assigned to them may have been influenced by the knowledge of the stratigraphic horizon at which they were collected. Other groups of fossil organisms did not permit of any further refinement of these datings.

Storage of collections and additional data

The samples collected and studied were deposited in the collections of the Afdeling Stratigrafie en Paleontologie of the Geologisch Instituut, Garenmarkt 1[b], Leiden. Most of the macrofossils collected were, however, put at the disposal of the respective specialists mentioned in the section on previous work in the area.

Additional data such as field maps, Sections 2 and 24, thin section descriptions, faunal and floral lists of the various fossiliferous horizons were deposited in the files of the Afdeling Stratigrafie en Paleontologie and are available to interested workers.

ACKNOWLEDGEMENTS

It is a pleasure for me to thank those people and institutions not connected with the Rijksuniversiteit Leiden, who in one way or another contributed to the development of the ideas presented in this thesis. First of all I should like to thank Dr. H. G. Reading, who invited me to spend a couple of months at the University of Oxford, and who contributed more to my geological education than any other single person. I also wish to thank the staff and students of the Department of Geology at Oxford for their stimulating discussions and hospitality. My stay in Oxford would have been impossible without the financial assistance of the Netherlands Organization for the Advancement of Pure Research (Z.W.O.) for which I am most grateful. I thank Dr. R. H. Wagner and Mrs. C. H. T. Wagner-Gentis (Sheffield) for their cooperation in the field, and for putting field maps and other unpublished data at my disposal. The discussions with Dr. R. H. Wagner, which resulted from our sometimes conflicting views, were a constant challenge and stimulus to me. I also wish to thank Mr. H. Dupont (formerly Louvain University) who kindly showed me the Waulsortian "reefs" of the Dinant basin. I am grateful to Dr. C. F. Winkler Prins (Leiden) for studying the brachiopods collected during the present survey. The comments by Dr. S. van der Heide (Haarlem) and Mr. H. W. J. van Amerom (Heerlen) on the pelecypods collected have been very useful to me. Mr. A. V. Zimmermann contributed to the readability of this paper by correcting the English. Mr. B. van Hoorn kindly translated the abstract into Spanish. And last but not least I thank Mrs. J. E. A. Bergman-van de Kuinder who typed the main part of the manuscript.

CHAPTER II

FACIES DESCRIPTIONS

INTRODUCTION

The various facies groups and types are distinguished and described on the basis of characteristics thought to be of genetic importance. The grouping together of several facies types in this chapter implies a genetic relationship. However, this genesis as interpreted from the combined properties of a facies will be dealt with separately from the descriptions in Chapter III. The facies groups and types differentiated are, as far as possible, natural units which should be recognizable in the field by subsequent workers. Nevertheless, many arbitrary boundaries had to be drawn where contacts between facies types were gradational or where they had many properties in common.

General remarks on the colours of the various sediments

In the area studied a distinction can be made between lithic arenites and lithic wackes on the one hand and quartz arenites and quartz wackes on the other. This is not a clearcut distinction, and sandstones of intermediate composition are common. Taking into account the transitional nature of the boundary line, although it is sharply defined in the classification system used, some remarks can be made concerning colours of fresh and weathered siliciclastic deposits. The lithic sandstones tend to be dark grey in fresh condition but weathering produces greyish green or greyish brown colours (N3 to 5Y5/2 or 5YR3/2). When fresh, the quartzose sandstones are of a light grey to white colour. This colour may be somewhat altered by weathering but it may also be changed into rusty colours (N7 to 5Y7/2 or 10YR7/4). The shales and mudstones are usually either dark grey to black, or light grey to greenish grey. (N2 to N6).

Limestones are light grey to dark grey. If dolomitic they may have a rusty brown colour when weathered (N3 to N7, 5YR6/1).

Sideritic concretions are dark grey when fresh and, when weathered reddish brown to dusky red (N3 to 5R3/4, 10R3/4 or 5YR5/6).

All colour numbers according to Rock-Color Chart of Geological Society of America.

FACIES GROUP I – GRADED ARENITES

(Plates Ia, b, c, d, e, IIa, b)

Alternations of arenitic to rudaceous beds with shale/mudstone beds. The coarse-grained beds have sharp bases, often with inorganic, sometimes with organic sole markings. Grading is commonly obvious in the field. The beds are laterally extensive and most frequently are of a blanket form, although channelling may occur. A sequence of internal structures which runs as follows: massive (graded) interval – laminated interval – rippled interval, very rarely with convolutions – laminated interval, may be discerned. This sequence is generally incomplete but its order does not seem to vary. Sometimes part of the shale/mudstone overlying this sequence can be shown to have been deposited during the same single sedimentary event. The inorganic sole markings recognized are flute casts, groove casts and rare prod marks, all of which may be deformed by loading. Load casts which are not deformations of one of the other kinds of sole markings are rare. The grading is in general due to a combined decrease in the maximum and mean grain size and an increase in the amount of matrix in an upward direction. Apart from rare channelling, no macro cross-stratification occurs in this facies. Body fossils are very rare and, if present, normally broken. If the fossils had a calcareous shell they are commonly leached out in the outcrops. Plant fragments may be very common. Especially in the lower laminated interval, bedding planes can be completely covered with comminuted plant debris. The fragments are commonly orientated but no attempt has been made to determine the relation of this orientation to other current direction indicators. Trace fossils are not very common. They may be present as burrows and/or trails (?) at the soles of the sandstone beds, and as burrows which penetrate the bed from above. Body and trace fossils are extremely rare in the interbedded shales/mudstones.

Several sub-types are distinguished.

Type Ia.–The sandstones are of a lithic arenitic to lithic wacke composition. Apart from clay pellets, which may be several centimetres in diameter, the maximum grain size is coarse to very coarse sand. A few notable exceptions, however, do occur in the northeastern part of the area. Bed thickness varies from a few decimetres to several metres. Channelling is relatively frequent and amalgamation is very frequent. The sequence of internal structures is incomplete. Especially the thick and very thick-bedded sandstones are often massive and ungraded with a sharp top. In the medium-bedded sandstones, grading can sometimes be discerned and parallel lamination may be present. Normally very little shale/mudstone is interbedded between these sandstones which constitute some prominent mappable units. Except in very fresh exposures, spheroidal weathering is characteristic of these sandstones. Apart from comminuted plants no body fossils or calcareous debris were found in these sandstones.

Type Ib.–The sandstones and siltstones are of the same composition as those of type Ia, although wackes are more common in these finer-grained clastics. Maximum grain size is coarse sand. Bed thickness varies from laminated to medium-bedded. Individual beds are generally of the blanket type. Sequences of internal structures which run as follows: massive graded interval – parallel laminated interval – rippled interval, or parallel laminated interval – rippled interval are often recognizable. Gradational tops are the rule in this facies type. The sandstone/shale-mudstone ratio, as estimated in the field, varies in general between 1:5 and 5:1. In most exposures, however, weathering renders it difficult to make an accurate estimate.

Type Ic.–The sandstones are of a quartz arenitic to quartz wacke composition, or may be transitional into the lithic varieties. In general, the maximum grain size is coarse sand. In this size fraction calcareous debris is common. Bed thickness ranges from a few decimetres to over two metres. Channelling, though rare, does occur. Most beds, however, are of a tabular to blanket shape. Generally little or no shale/mudstone is interbedded with the sandstones of this facies. Commonly the sandstone beds are amalgamated. If not amalgamated they tend to have sharp tops. Grading is often indistinct or absent and the sequence of internal structures is incomplete. Upon the massive interval a thin (10 cm) laminated, and an equally thin rippled interval may sometimes be discerned. Included in this facies are conglomeratic beds which occur interbedded in this facies or in type Id. These conglomerates contain well-rounded quartzitic pebbles and cobbles with rare boulders up to 30 cm in diameter. The pebbles and cobbles may form a framework or may "float" in a quartz arenitic matrix. Bedding is indistinct and only indicated by the interbedded sandstones. These conglomerates are only distinguished from other sandy conglomerates on the grounds of the associated facies.

Type Id.–The sandstones and siltstones are again mainly of quartz arenitic to quartz wacke composition, although wackes are more common than in the generally coarser-grained deposits of type Ic. Maximum grain size is normally in the coarse sand grade – the coarse particles are commonly calcareous. The bed thickness generally varies from thin-bedded to medium-bedded, although some sandstones included in this facies are over one and a half metres thick. The layers are laterally very extensive and little thickness variation occurs in a single bed. Especially in the thicker beds a complete sequence of internal structures can be discerned: massive, graded basal interval (often calcareous) – lower parallel laminated interval – rippled, sometimes convoluted interval – upper parallel laminated interval – transition into overlying shale/mudstone. Beds possessing an incomplete sequence, as for instance: massive, graded interval – laminated interval – rippled interval – sharp top, lower laminated interval – rippled interval – upper laminated interval – gradational top, or rippled interval – laminated interval – gradational top, are much more common. The sandstone – shale/mudstone ratio, as estimated in the field, generally varies between 10:1 and 1:10. *Chondrites* and *Spirodesmos* (?) are the only determinable trace fossils in this facies. The former occurs in the sandstone layers, the latter along the soles of the beds. Other types of burrows may occur both in the upper part and along the soles of sandstone beds of this facies.

Type Ie.–The sandstones and siltstones of this type generally have a quartz wacke to quartz arenitic composition although transitions into the lithic varieties are probably common. Maximum grain size is in the fine to medium sand grade. The small grain sizes make it difficult to ascertain the precise composition. Bed thickness varies from thinly laminated to thin-bedded. Sometimes it is difficult to see whether a layer has a distinct, sharp base or not. Grading is commonly indistinct and the sequence of internal structures is very incomplete: rippled interval – laminated interval – gradational top, or laminated interval – gradational top. Sandstone – shale/mudstone ratio ranges from 5:1 to 1:30.

Types Ia and Ib, and types Ic, Id, and Ie are gradational to each other. Type Ie is also gradational to type IIIa. These five subdivisions of facies group I are quantitatively very important. The following types are less important.

Type If. – Medium to coarse-grained quartz arenitic sandstones, which are medium to very thick-bedded. The beds have sharp tops which may be masked because of amalgamation of the beds. In the upper five to ten centimetres of a bed clay clasts and/or well-rounded quartzitic pebbles occur, which are absent in the lower and middle part of the layer. Grading was not observed. Sole markings may be present and the only internal structures observed are the peculiar dish structures as defined by Stauffer (1967) (Plate IIa). This facies type occurs interbedded in types Ic and Ie.

Type Ig. – Commonly graded calcirudites to calcarenites with a fine-grained, siliciclastic matrix, which are often overlain by mudstone with "floating" calcareous debris of the fine sand grade. The lower part of the beds is often of a very localized occurrence in strongly erosive channels. The upper parts are much more extensive. The complete sequence, which was only found once may be several metres thick. The following sequence could be recognized from bottom to top: poorly rounded cobbles consisting of a calciruditic microbreccia, which are "floating" in shale/mudstone – an interval consisting of angular limestone fragments up to five centimetres in diameter, which mainly "float" in a fine-grained siliciclastic matrix; in this interval contorted shale/mudstone lenses several decimetres in diameter occur – a graded interval in which the limestone clasts form a grain-supported framework (wackestone to packstone) – an interval consisting of well-graded packstones sometimes with parallel lamination – sharp top – an interval of greyish green mudstone, often several metres thick, with "floating" silt and sand-sized limestone fragments – gradational contact with the overlying blackish shale/mudstone. Common variations are for instance: graded interval with the overlying greyish mudstone, or an ungraded interval with or without associated mudstone. In the former case the sometimes extremely local occurrence of the graded interval in pockets and channels several decimetres deep, and often only a few metres or less wide, is very striking. Especially as the mudstones can be traced for several hundreds of metres.

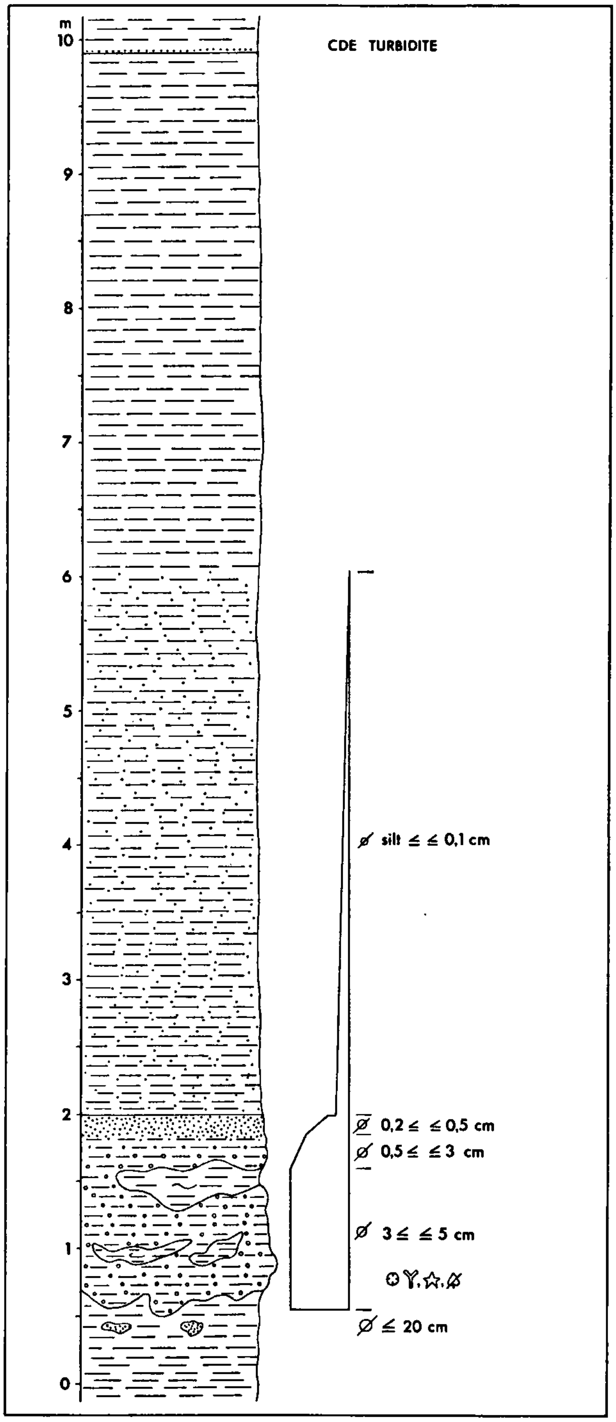

Fig. 5. Most complete development of facies type Ig. Exposure approximately 150 metres above the top of Section 20.

Type Ih. – Graded calcirudites to calcarenites with little fine-grained siliciclastic material as a matrix. The base of the beds is commonly erosive and may have a channelling appearance. Limestone clasts 10–15 cm in diameter may occur in the basal part. Stylolitic contacts are abundant in the coarser-grained parts.

Bed thickness ranges from medium-bedded to very thick-bedded. A bed may consist of one thick graded interval, the upper part of which is normally a coarse calcarenite. This massive, graded interval, however, may be followed by: lower parallel laminated interval – rippled interval – upper parallel laminated interval – gradational top. In a few cases the coarse ruditic part is ill-developed or absent.

A few more rare variations of graded calcarenites will not be described.

FACIES GROUP II – CONTORTED AND FRAGMENTED SEDIMENTS (Plates If, IIc, d, e, f)

Shales, mudstones, sandstones and limestones which are deformed, contorted, or broken and fragmented, often in a chaotic manner. Folding and breaking are not accompanied by the presence of cleavage, slickensides or fractures filled with vein quartz or calcite. Folding, if developed, is irregular, and pull-apart, hook-shaped and roll-up structures are often present. Small unconformities are sometimes associated with these deposits.

Type IIa.–Sandstone – shale alternations of type Ic or Id, deformed in a chaotic manner. The thickness of the sequence involved in deformation may range from a few decimetres to several tens of metres.

types may also be present, but they are quantitatively unimportant. In some cases it may be shown that bedding in the limestones forms an acute angle to the original depositional surface, or geopetal structures prove that the blocks are overturned. Figure 7 is a map of a few of these limestones "floating" on and in type IIa and covered by relatively undisturbed sediments of facies types Id and Ie. No soft sediment deformation of the limestones has ever been observed. The lowermost Sierra Corisa limestone tongue in Section 4 is also included in this facies type, although it is still continuous with the limestone in situ. The reason for this is the fact that it is imbedded in facies IIb (see Plate IIc, d, and Fig. 6). At the locality of Section 4 this lowermost limestone tongue has an extremely sharp and irregular lower surface. In the lowermost few metres of this limestone, cracks occur which are filled with quartzose sandstone. These sand-filled cracks are absent in the remainder of this limestone.

Type IId.–Limestone breccias with a shale/mudstone matrix. The limestone fragments, which are very angular to angular, may "float" in the matrix, but may also form a grain-supported framework. Maximum diameter of the limestone fragments is normally in the pebble to cobble range. Other components are fragments of sideritic concretions and clay pebbles. Quartz

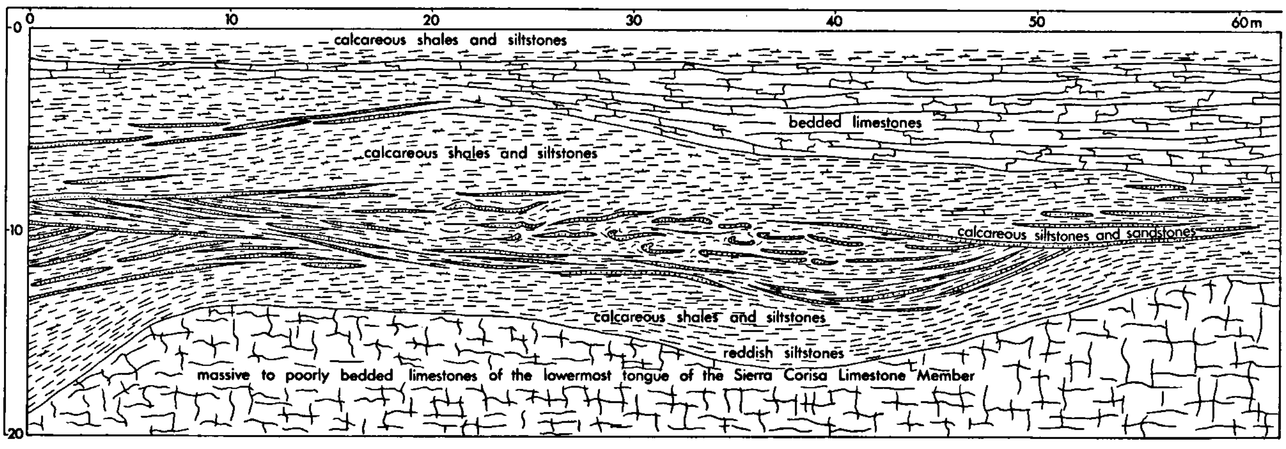

Fig. 6. Small-scale unconformities on top of the lowermost tongue of the Sierra Corisa Limestone Member (Section 4; approx. 1590–1620 m).

Type IIb.–Sandstones, siltstones, shales, calcareous shales of facies type IIId, which are chaotically deformed, and which may exhibit small-scale unconformities (Plate IId and Fig. 6) This type was only observed in Section 4, where it is closely associated with the lowermost Sierra Corisa limestone tongue.

Type IIc.–Isolated limestone blocks varying in diameter from 0.5 to 250 m which "float" in or on shale/mudstone or in facies types IIa and IIb. These limestones consist mainly of facies types Xb, c (massive or poorly-bedded mudstones to wackestones with occasional boundstones). The other, much rarer limestone

arenitic pebbles rarely occur. The maximum lateral extension of these breccias is a few hundred metres. Thickness may be upto several tens of metres. In a few cases their basis was observed to be erosive, and similar breccias in slightly older deposits exposed near San Martin de Perapertú (east of the map area) are channel fills. Plate IIf shows a very coarse example of this facies from slightly younger deposits. Contorted sandblocks occurring associated with these limestone breccias are included in this facies.

Type IIe.–Channel fill consisting of a sandstone breccia with a shale/mudstone or quartz to lithic

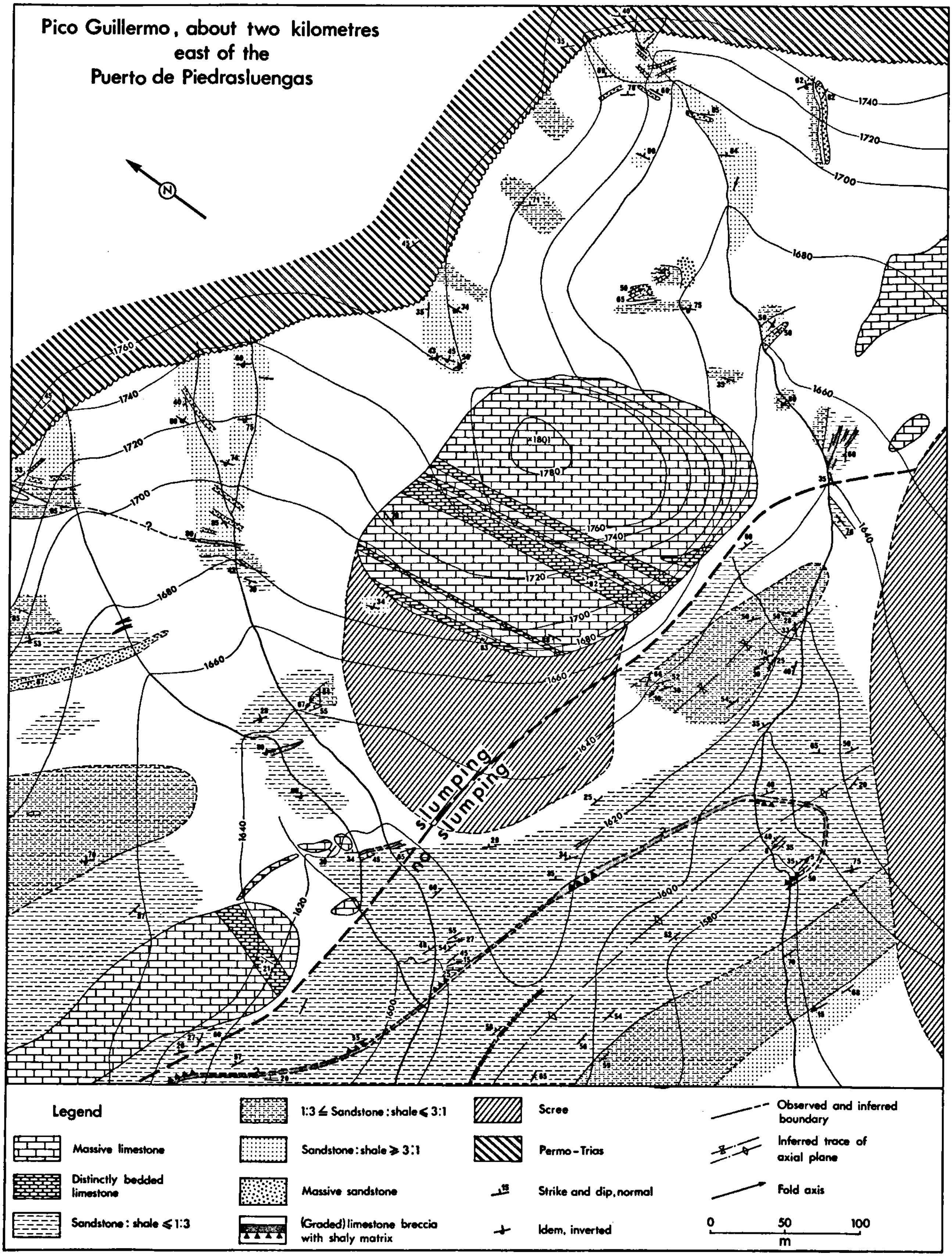

Fig. 7. Isolated limestone knolls east of Piedrasluengas.

wacke matrix. The very angular lithic arenitic and shale clasts locally form a grain-supported framework. The fragments are several centimetres in diameter. No bedding, grading or imbrication was observed in this channel fill. The channel was cut at least 8 to 10 m down into facies Id. Lateral extent was not observed.

Type IIf.–Lenses of cobble to boulder-sized conglomerates with a shale/mudstone matrix. The cobbles may "float" or form a grain-supported framework. The cobbles, which are of a quartz arenitic composition, are very well rounded. Maximum thickness of the lenses ranges from one metre to about 5–10 m. Lateral extent is in the order of a few tens of metres.

The latter two types are very rare, while the others occur in several sections. Types IIc and IId may be gradational to each other. Type IId is also gradational to type Ig.

FACIES GROUP III – FINE-GRAINED SILICICLASTICS (Plates IIg, h)

Shales, mudstones, argillaceous siltstones and/or fine-grained sandstones. These fine-grained sediments are often difficult to distinguish in the field. It is proved by the associated facies types that superficially very similar argillaceous sediments may have been deposited in widely differing environments. Because of this, most of the types of this group are distinguished primarily on the basis of the associated facies types. In general, bedding is difficult to discern in this facies. It is often only recognizable because of the interbedded sideritic concretions or silt and sand laminae. The quantitative importance of this group is obvious after a glance at the sections, especially taking into account the fact that most of the unexposed parts are largely argillaceous. Furthermore, about $\frac{1}{4}$ to $\frac{1}{2}$ of the total thickness of facies group I is occupied by these fine-grained deposits.

Type IIIa.–Dark grey shale/mudstone with sometimes abundant sideritic concretions which indicate the bedding plane. These concretions are flat spheroids with a maximum diameter of about 30 cm. They occasionally form a nearly continuous bed but normally they are isolated from each other. Sand or silt laminae with sharp or gradational bases may be present, but if the sandstone-shale/mudstone ratio exceeds 1:30 the deposit is, for instance described as type Ie. Thin graded calcarenitic beds may also occur. Bioturbation was rarely ascertained. This facies type is closely associated with sediments of group I. It is gradational to types IIId, IVa and c. Lenses and beds of marine limestones may sometimes be intercalated.

Type IIIb.– Well-bedded or laminated to homogeneous, dark grey shales or mudstones which overlie coarsening upwards sequences of type Vb or e, or type Va. Bioturbation can sometimes be recognized in the lower part because of the sandy fill of the burrows. Sometimes abundant drifted plant fragments occur. In one instance marine lamellibranchs occur in great numbers. (Section 7, 1011 m). Pyrite is common in this facies type. This facies type is at the utmost only a few metres thick and commonly grades into another coarsening upwards sequence.

Type IIIc.–Indistinctly bedded to homogeneous, argillaceous sandstones and siltstones, and often somewhat calcareous mudstones. Spheroidal weathering may mask any bedding structure present. Marine fossils such as bryozoans, brachiopods, crinoid ossicles often occur. In polished surfaces bioturbation may be visible.

Type IIId.–Calcareous shales, mudstones, siltstones and argillaceous, fine-grained sandstones, which are generally intensively bioturbated. The only determinable forms are *Spreitenbauten* of the *Zoophycos* group. Although complete homogenization may have occured, bedding is usually still discernable. At the main occurrence of this facies type (Section 4 and eastwards) the lowermost few metres of mudstone, which immediately overlie the lower Sierra Corisa limestone tongue, are locally dusky red (5R3/4). The transition into the overlying greenish grey deposits is very irregular. This facies type is exceptionally fossiliferous at the location of Section 4. The most remarkable fossils are the sphinctozoan sponges (van de Graaff, 1969), but brachiopods, crinoids, solitary rugose corals, bryozoans, fusulinids, trilobites, lamellibranchs, gastropods, nautiloids and calcareous algae also occur.

Type IIIe.–Dark to light grey calcareous mudstones with locally abundant marine fossils such as brachiopods, solitary rugose corals, gastropods, bryozoans and, rarely, sphinctozoan sponges. Bedding is indistinct and varies from thin to medium.

FACIES GROUP IV – COMMONLY RIPPLED AND CROSS-BEDDED SILICICLASTICS

Generally well-bedded or laminated mudstones, siltstones and sandstones which do not form part of a fining upwards or coarsening upwards sequence (Allen, 1965; Oomkens, 1967). Bedding contacts are usually gradational but erosive contacts may occur. In the fine-grained sandstones ripple cross-lamination and ripple drift are common features. Bedding or lamination planes can usually not be traced for more than a few metres. This facies group, too, is differentiated because of the associated facies types. Types IVa, b form a sub-group associated with quartz arenitic coarsening upwards or fining upwards sequences, and with limestones. Types IVc, d form a second sub-group associated with lithic arenitic coarsening upwards and fining upwards sequences.

Type IVa.–Laminated to thinly-bedded siltstones and fine-grained sandstones of quartz arenitic to wacke composition. Bedding planes are often covered with micas and/or finely comminuted plant debris. Ripple cross-lamination and ripple drift are common. Sandy or silty streak (de Raaf et al., 1967) rarely occur. Both marine body fossils and trace fossils occur.

Type IVb.–Thin to thick-bedded quartz arenitic sandstones. Bed shape ranges from tabular to blanket. In some cases a sharp base to a bed may be discerned, but gradational contacts dominate. A bed may be laminated in itself and occasionally low-angle macro cross-bedding occurs. Marine fossils sometimes occur in this facies type.

Type IVc.–Laminated to thinly-bedded mudstones, siltstones and fine-grained sandstones of lithic arenitic to lithic wacke composition. Micas and finely comminuted plant remains may cover the bedding planes. Rippling is common in this type. Sandstones of type Ib or channel fills of type VI occasionally occur in this facies.

Type IVd.–Rippled, laminated to medium-bedded lithic arenitic sandstones of fine to medium grain size. Contacts are usually gradational but sharp erosive contacts do occur and low-angle macro cross-stratification can sometimes be discerned. Although many borderline cases exist, channel fill with or without fining upwards sequences is described as type VIb.

Type IVe.–Thick-bedded to massive, very coarse-grained quartz wackes. Especially the larger grains are very well rounded. No sedimentary structures apart from indistinctive bedding planes were observed. This facies type occurs as intercalations in limestones in Sections 20 and 21.

FACIES GROUP V – SILICICLASTICS WITH PLANT ROOTLETS

This facies is characterized by the presence of plant rootlets which are preserved as carbonaceous sheaths. These rootlets pierce the sedimentary laminations and bedding. *Stigmaria* may be recognizable in good outcrops. Apart from rootlets, burrows may also pierce the sedimentary layering and it may be hard to distinguish the two, especially if the sediment is homogenized. Bedding contacts are usually gradational in these deposits. Sometimes coal laminae or beds overlie such a rootlet containing horizon. Determinable plant fossils, if present, occur above the coal or directly on top of the rootlet bearing layer. Again, two sub-groups may be distinguished. The first comprises types Va and b and is associated with lithic arenitic coarsening upwards and fining upwards sequences of types VIIa and VIb. The second consists of type Vc and is associated with quartz arenitic coarsening upwards sequences of type VIIc and with type IXa.

Type Va.–Laminated and rippled shales, mudstones, siltstones and fine sandstones, the latter of lithic arenitic or wacke composition. These deposits, which are normally not more than a few decimetres thick, are often characterized by brownish weathering (limonite?) concretions probably of sideritic composition (Hemingway, 1968). Often a distinct sequence can be distinguished in these deposits. If they form the upper part of a fining upwards sequence of type VIb or if they form a thin fining upwards sequence of their own they are included in this facies type. They are also included in this type if no clear sequence can be discerned.

Type Vb.–This type differs only from type Va in that the rootlet bearing deposit exhibits a coarsening upwards sequence on a decimetre to metre scale. It may contain slightly coarser and better sorted sandstones than type Va, which has fine sand as its coarsest component.

N.B. Fining upwards and coarsening upwards sequences are here used in a slightly different sense than in facies groups VI and VII. In types Va and b they are on a decimetre to metre scale while in the other examples they are on a larger scale. Furthermore, in the case of the fining upwards sequence no channel fill is meant (cf. type VIIIa), but a gradationally based sequence with an upward increase in argillaceous material which is either interbedded or homogeneously distributed in coarser sediments. Usually the only depositional structures recognizable are rippling and lamination. In the case of a coarsening upwards sequence no macro cross-bedding or channel fills occur in the upper part. Only ripples and laminations are to be expected.

Type Vc.–The upper part of the well-washed, well-sorted quartz arenitic sandstones of types VIIc and IXa, if penetrated by rootlets and/or *Stigmaria*. No small-scale fining upwards or coarsening upwards sequences are associated with the occurrence of rootlets as in types Va and b.

FACIES GROUP VI – SILICICLASTIC FINING UPWARDS SEQUENCES (Plate IIIa)

Sandstones and conglomeratic sandstones which do not form part of a definite coarsening upwards sequence of facies group VII, and which exhibit macro cross-bedding and erosive bases. A channel fill shape and/or fining upwards sequence are recognizable. Again two sub-groups may be distinguished. The first consists of type VIa and is associated with quartz arenitic deposits of types IVa and b, Vc, and VII c and d. The second consists of Type VIb and is associated with lithic arenitic deposits of types IVc and d, Va and b, and VIIa and b.

A sequence is called a fining upwards sequence if the following features can be distinguished in the field. 1) A sharp, often erosive base. 2) An upwards decrease in both maximum and median grain size. 3) A concomitant decrease in the size of cross-bedding, if present, from large-scale lateral accretion sets to medium-scale (dm-m) cross-stratification, generally of scoop/trough type, and thence to ripples. A smaller type may occur superimposed on the preceding larger type. Parting lineation may occur associated with the two larger types of cross-bedding. Ordinary laminations consisting of, for instance, alternations of silt and fine sand occur associated with the small-scale ripples. 4) A gradational top, which characteristic may be masked by later erosion. Such fining upwards sequences differ from graded beds of, for instance, type Ic or d, in that they have large-scale cross-bedding of a fore-set type; in their clear association with, for instance, facies group V; their smaller lateral extensions, etc. (see also Allen, 1965).

Type VIa.–Quartz arenitic sandstones with occasional conglomerates. The pebbles are dominantly of quartz arenitic composition too, but lydite pebbles and clay galls may also occur. Some of the coarser sandstones of this facies have a lithic arenitic to quartz arenitic composition. The content of unstable fragments, however, is around 10–15%, while in the true lithic arenites of comparable size grades the unstable fragments may form 20–35% of the rock (e.g. sediments of types Ia or VIb). Channel fills and more tabular deposits exhibiting a fining upwards sequence both occur. The channel fills do not necessarily show a clear fining upwards sequence. One type of channel cuts down into facies type IVb for over a metre and is completely filled by low-angle ($\leqslant 15°$) cross-strata. Some poorly exposed conglomerates are included in this facies type. This is done because of their relatively clean and well-sorted sandy matrix, which is of quartz arenitic to lithic arenitic composition. Contacts with surrounding strata and sedimentary structures could not be used because of the poor exposure.

Type VIb.– Medium to coarse-grained sandstones of lithic arenitic composition. Clay galls and drifted plant fragments form the only coarser components. Scoop to trough cross-stratification on a dm to m scale are the commonest types of cross-bedding. Low angle to high angle tabular cross-stratification on a 2–5 m scale also occurs. If so, it is present below the smaller scoop and trough cross-stratification.

FACIES GROUP VII – COARSENING UPWARDS SEQUENCES (Plates IIIb, c, d, IVa, b, c)

Shales, mudstones, siltstones and sandstones, which show a definite coarsening upwards sequence. Again two subgroups may be distinguished. One characterized by lithic arenitic sandstones, the other by quartz arenitic sandstones.

A sequence is called a coarsening upwards sequence if it shows the following characteristics from base to top: 1) A progressive increase in both maximum and median grain size, beginning with argillaceous deposits and ending with sandstones. Minor reversals of this trend do occur, for instance in the form of clay or silt laminae in sandstones. In that case, however, the sand/shale ratio will still increase upwards. 2) Concomitant with these changes are changes in the kind of sedimentary structures present. In the lower part of the sequence the deposits are generally laminated to medium-bedded. Bedding contacts are gradational. Seemingly unbedded, homogeneous intervals may, however, also occur. Higher up, rippling and ripple drift may occur. Still higher, well-bedded to well-laminated, fine to medium-grained sandstones may be present. Bedding and laminations are very regular and can often be traced for several metres. Several types of macro cross-bedding begin to occur at this level. Small channels (dm-m scale) and tabular low-angle cross-bedding are most common. The sequence may end with a gradational top, a sharp top or it may be unconformably covered by a channel fill or fining upwards sequence of type VIa or b. The thickness of these sequences varies from a few metres to over a hundred metres. Their lateral extension ranges from a few hundred metres to several kilometres.

Type VIIa.–Lithic arenitic sandstones, siltstones and shales/mudstones showing a coarsening upwards sequence on a scale varying from a few metres to over a hundred metres. If the size exceeds 25 m they may be called major c.u.s., if not, minor c.u.s. Not considering clay galls and comminuted plant debris, the coarsest deposits are in the coarse to very coarse sand grades. Because of weathering effects, small structures like ripples may be difficult to distinguish in the lower to middle parts of these sequences. They may also be truly absent. Above the laminated and rippled interval a much better and more evenly laminated interval with coarser sandstones may follow. This is especially the case in some of the major examples. (Section 4, 510 m, Section 13, 355 m). In this well-bedded part, low-angle, tabular cross-stratification on a dm-m scale may begin to occur, and small channels may also be present. Both phenomena become more frequent in upward direction. This sequence is typically covered, with an erosive contact, by a channel fill or a fining upwards sequence of a thickness varying from a few decimetres to a few metres. This channel fill or fining upwards sequence is still included in this facies type for reasons explained in Chapter III. In one case (Section 7, 1040 m) no distinct erosive base could be discerned below the scoop and trough cross-bedded part. There seems to be a gradual increase in the number of scoops and troughs. In the upper part of this example, low-angle, tabular cross-bedding on a 10–20 m scale occurs with superimposed scoop and through cross-bedding on a metre scale (Plate IVa, b). This cross-bedding is cut off sharply by finer-grained deposits which are of facies type IVc. Low-angle, tabular cross-bedding on a similar scale, but without superimposed smaller cross-stratification, occurs in Section 13 (360 m) (Plate IVc). In a few cases it may be observed that only the lower part of a coarsening upwards sequence is developed, i.e. the channel fill or fining upwards sequence is absent and the sequence may then either gradually or sharply be covered by fine-grained deposits. Laterally, however, a channel fill forming the top may sometimes be observed (Section 4, 730 m).

Type VIIb.–Coarsening upwards sequences which differ from type VIIa in that the sandstones are transitional from lithic arenites into quartz arenites. Furthermore they are closely associated with facies types VIIc and d and other deposits with marine faunas. The upper part of one example (Section 10, 590 m) is actually formed by deposits of type VIIc. In the middle part of this particular sequence a number of peculiar channel fills occur. They consist of sandstone bodies about 5 to 50 m in lateral extent and up to one metre in thickness. They differ from other channel fills in these sequences not only because of this greater lateral extent, but also owing to the fact that the thin sandstone beds composing the channel fill are intensively rippled. A low-angle ($\leqslant 5°$) tabular cross-bedding on a metre scale can sometimes be discerned. (Plate IIId). Some of these regularly and thinly-bedded sandstones show sole markings. Another difference from type VIIa is that in the fining upwards sequence which forms the top, conglomerates may occur.

Type VIIc.– Mudstones, siltstones and sandstones showing a coarsening upwards sequence on a scale varying from 5 to 20 m. The sandstones are of a quartz arenitic composition and the maximum grain size rarely exceeds the fine sand grade. This type of coarsening upwards sequence is furthermore characterized by a sharp top which may be penetrated by plant rootlets and/or burrows. Complete homogenization of the upper few centimetres may have occurred. Tree trunks, up to five metres long and half a metre wide, which floated in, were observed in a few examples of this facies type. Characteristically the lower part consists of rippled and laminated mudstones to sandstones. The middle part is well-laminated and most frequently shows distinct low-angle tabular or trough/scoop cross-stratification on a dm-m scale. Lamination planes are sometimes covered with fine plant fragments and/or micas. In the upper part, bedding and/or lamination is less easily recognized. This is often accompanied by the presence of rootlets and/or burrows. In a few examples the upper surface shows smooth ripples on a metre scale. Fossil debris is occasionally recognizable in these deposits.

N.B. The upper parts of deposits of this type are described as type Vc if penetrated by rootlets.

Type VIId.– Mudstones, siltstones and quartz arenitic fine sandstones showing a coarsening upwards sequence on a 10 to 50 m scale. The deposits are generally laminated to thin-bedded, and individual beds are tabular to blanket shaped. Bed thickness and distinctness of individual beds show a slight increase in upward direction. Rippling on a centimetre scale is common, but larger types of cross-bedding are rare or absent. In a few cases the ripples were observed to have symmetrical cross-sections. Marine fossils are found occasionally. Burrowing may occur especially in the lower parts. These sequences may have a sharp top but in a number of cases (Sections 12, 13) a gradational top is observed. In the latter case they are covered by sediments of type IIIa or c, i.e. the same facies type in which these sequences began.

FACIES GROUP VIII – FINING UPWARDS SEQUENCES WITH MARINE FOSSILS AND LITTLE OR NO CROSS-BEDDING (Plate IVd)

Type VIIIa.–Sometimes calcareous, quartz arenitic sandstones to siltstones of a tabular to blanket shape. These sandstones have sharp bases and vary in thickness from less than one metre to over thirty metres. Marine fossils are common, especially in the upper part, as is burrowing. Upwards this facies commonly grades into either black shale/mudstone of type IIIa or into limestones. A combination of an upward decrease in maximum and median grain size, a decrease in distinctness of bedding and the restriction of major cross-bedding or rippling to the lower part warrant the name fining upwards sequence. The common presence of marine fossils and the much greater possible thickness indicate, however, that this type is fundamentally different from those of group VI.

FACIES GROUP IX – QUARTZ ARENITES ON ERODED LIMESTONES (Plates IIIe, f, IVe, f)

Type IXa.–Sometimes calcareous, quartz arenitic sandstones overlying an often strongly erosive contact with limestones. The sandstones fill pockets and pipes which penetrate the limestone. In one example already described by Nederlof (1959, p. 662), these pipes, which are up to 50 cm in diameter, penetrate the underlying limestone for at least seven metres in a direction perpendicular to the bedding. Normally, however, the pockets are only a few decimetres or less in depth. Associated with these pipes and pockets may be numerous cracks and vugs in the limestone which are also filled with quartz arenitic sand. These cracks and vugs are usually a few centimetres in diameter. Limestone conglomerates occur locally in this facies type. Sometimes a fining upwards sequence of type VIIIa is present, but a sharp top penetrated by plant rootlets and/or burrows may also occur. These sandstones, which are normally not more than a few metres thick, are generally very well washed and sorted. Not counting some dissolved carbonate(?) material, some of the sandstones consist for over 95% of quartz. The only drawback is that overgrowth has occurred of the originally well rounded grains. Nevertheless these sandstones are the most mature in the entire area studied. Lateral tracing of different examples showed that some of them grade laterally into quartz sand bearing lime grainstone after a few hundred metres. Others, however, can be traced for kilometres.

FACIES GROUP X - LIMESTONES (Plates V, VI, VII, VIII, IX)

Although the relationships of the various limestone members with the surrounding siliciclastic deposits form a main theme of this paper, only one limestone body was studied in detail. This is the Agujas Limestone Member which crops out in the Redondo Syncline. The following facies descriptions, with the exception of Xn, are all based on observations on the Agujas Limestone Member. During a cursory examination of the other limestones, however, it appeared that in most aspects these are very similar to the Agujas Limestone Member. Their facies patterns, however, are much less complicated. The causes of these differences will be dealt with in Chapter V.

From the following facies descriptions it will be obvious that several types distinguished are very similar to each other. In some cases the only difference is the direction in which a certain trend changes. As the genetic significance of such changes is still obscure, it seems important to note these minor differences although they are as yet difficult to interpret.

Type Xa.–Irregularly bedded limestones with a brecciated to nodular appearance in the outcrop. In thin sections the clasts may be seen to have pressure solution contacts. Along the pressure solution seams dolomitization is frequent. Texturally these limestones are at present coarse ($\leqslant$3 cm) wackestones to packstones. This type is locally very fossiliferous. Brachiopods, crinoids, echinoids, solitary rugose corals, sphinctozoan and other sponges, goniatites, nautiloids, trilobites, bryozoans, ostracods, gastropods, radiolarians, calcareous algae and oncolites were recognized. In Section 21 (530 m) a few phosphatic nodules occur in this type. This facies type is most common in the Redondo Syncline, where it occurs at the base of the Agujas and Abismo Limestone Members. It also occurs in isolated lenses in Sections 1 and 4. Laterally and vertically it shows transitions into shale/mudstone of types IIIa or IIIe.

Type Xb.–Indistinctly, irregularly bedded mudstones, wackestones and rare packstones. Bedding is most frequently indicated by pressure solution seams along which dolomitization occurred. Fossils are not abundant but crinoid and echinoid fragments, fusulinids and other foraminifera, dasyclads and other calcareous algae, chaetetid, "lithostrotionid", auloporid and solitary rugose corals occur. Crinoids and fusulinids are often somewhat concentrated along the pressure solution seams. Chert nodules occur locally. This type is quantitatively very important in nearly all the limestone units of this area. It is transitional into most other facies types of this group, i.e. Xa, Xc, Xe, Xf, Xi, Xj, Xl, Xo, and Xp.

Type Xc.–Indistinctly to massive bedded mudstones, wackestones with some boundstones and rare packstones. Fossils are rare. Crinoid ossicles, foraminifera, algae (mainly *Komia*) and fenestrate bryozoans occur. It is this facies type that, together with type Xb, constitutes most of the limestone bodies in this area. This type, too, shows transitional contacts with many other facies types such as Xa, Xb, Xd, Xe, Xf, Xh, Xh, Xj, Xk.

Type Xd.–Light or dark coloured sparite patches of irregular but smooth shapes. The dark colour of some of these patches is caused by the presence of organic material. Single patches vary in diameter from a few mm to about 10 cm. In one example radiating tube-like structures are visible in the sparite patches. The sparite patches either consist of fibrous calcite (dark variety) or of a fibrous calcite rim with more equidimensional crystals in the centre. In a few examples no fibrous calcite crystals can be recognized. In some specimens undulating fronds of platy codiacean algae can be recognized. In thin section the fibrous calcite rim as well as the equidimensional sparite are seen to emanate from these fronds (Plate VIb). In the depressions of these fronds mudstone/wackestone is sometimes present, forming a clear geopetal structure. In most specimens collected, however, codiacean or other algae are not present in a recognizable form. In those examples the sediment in between the sparite patches consists of mudstone, wackestone, packstone and rare grainstones and boundstones. In rare examples the geopetal fill is seen to overlie a thin fibrous calcite rim. In most specimens studied brittle deformation and fragmentation occurred of the sediment surrounding the sparite patches. If this is the case, the fragments are seen to rest upon each other and the sparry cement in the fractures belongs to the same generation as the cement in the larger patches. This facies type is quantitatively unimportant but nevertheless fairly common. It is transitional into type Xc and occurs most frequently surrounded by this facies type.

Types Xe and Xf.–Facies types Xb and Xc often show very gradational transitions into one another. In a number of cases a clear trend-like change was observed in the number of dolomitized pressure solution seams which are the main characteristic of type Xb. If such a change from Xb to Xc occurred in an upward direction the sequence was called type Xe. If it occurred in a downward direction it was called type Xf. If the transition is relatively sharp, only Xb and Xc are differentiated.

Type Xg.– Massive-bedded to well-bedded oolite packstones and grainstones with a sharp, sometimes clearly erosive base. Clasts of underlying limestone deposits and abraded fossils such as coral fragments may be present. Sometimes "mottled" pelsparitic grainstones and packstones occur. The "mottled" appearance of these pelsparites is due to minor differences in the proportion of matrix and cement. These differences are accentuated by pressure solution phenomena. These "mottles" range in shape from equidimensional spots to platelike layers which may have a thicknesswidth ratio of at least 1:20. These "mottled" pelsparites occur in the upper part of this facies type. The coarse-grained part of this facies type contains a relatively rich flora and fauna. Fusulinids and other foraminifera, crinoids, echinoids, bellerophonid and other gastropods, dasycladacean and codiacean algae are all very common. A peculiar phenomenon in this facies type in the Agujas Limestone Member is the occasional occurrence of quartz arenitic pebbles. The largest pebble observed was about 20 centimetres in diameter. All pebbles are spherical and superbly rounded. The occurrence of these pebbles is all the more surprising as few or no fine-grained siliciclastics were observed as an admixture in the limestone. Only in Sections 20 and 21 does a coarse-grained quartz wacke occur as a thick intercalation in the limestone but here little or no mixing of the two lithologies took place. The upper boundary of this facies type may be sharp or gradational.

Type Xh.–Gradationally based, massive to well-bedded packstones and grainstones while "mottled" pelsparites may also occur in the upper parts. In fact, apart from the sharp base the description of this type is identical to that of type Xg. The lower part, however is transitional from mudstones and wackestones of types Xb, Xc or Xe. The mudstones and wackestones change, in an upward direction, into packstones and grainstones which gradually become oolitic. In the lower parts oncolites do occur. Both type Xg and Xh are commonly overlain, with a sharp contact, by type Xi.

Type Xi.–Irregularly but distinctly bedded mudstones, wackestones and rare packstones and boundstones, which are often rich in fine-grained insoluble material. Pressure solution seams with attendant concentration of insolubles and dolomitization are common. Big auloporid colonies (up to 1.5 m in diameter and 40 cm in height), chaetetid and "lithostrotionid" colonies, solitary rugose corals, oncolites, dasycladacean and codiacean algae are characteristic of this facies type. Although sometimes transported, the corals, especially the colonial ones, are often in position of growth as are the algae. Although faunal elements like crinoids and fusulinids may also be present, the above-mentioned association is characteristic especially if all the elements are present. This type is transitional in upward direction into types Xb or Xe, and commonly overlies Xg or Xh with a sharp contact.

Type Xj.–Wackestones, packstones and rare grainstones, which show distinct and regular bedding on a medium scale. Chertification often occurred along the bedding planes. Crinoid ossicles, echinoid spines and plates, algal fragments and foraminifera are the most important fossils present. Bioturbation was observed in a number of exposures. Indistinct indications of channeling are present in Section 17. Laterally this type seems to be transitional into type Xl.

Type Xk.–This type is a combination of types Xb and Xj. The basal part of a sequence called Xk consists of facies type Xb which grades upwards into regularly and well-bedded packstones and grainstones of type Xj. Apart from the upward increase in median and maximum grain size, a decrease and eventual disappearance of dasycladacean and other algae can be observed. Furthermore, the number of dolomitized pressure solution seams and the amount of insoluble material decreases in upward direction. In the packstone-grainstone part of the sequence chert is rare.

Type Xl.–Regularly and thinly-bedded mudstones and wackestones which contain few macroscopic fossils. In thin section sponge spicules can sometimes be recognized but other fossils are rare. Chertification along bedding planes is common. Due to discontinuous exposures, the lateral equivalents of this type are difficult to ascertain. Nevertheless, it seems likely that it grades into types Xj, Xb, e, f on the one hand and Xm on the other.

Type Xm.–Dark coloured, laminated to thinly bedded, fetid mudstones and wackestones which are completely non-fossiliferous. Laminae and bedding planes are extremely regular and can in some cases be traced for several metres. No indications of bioturbation were observed. Low angle cross-bedding on a cm to dm scale sometimes occurs. In these cross-bedded parts few erosional features are visible. The dark colour is caused by a relatively high content of organic material. Chertification sometimes occurred along bedding planes. In thin section this type consists of microspar to pseudospar (Folk, 1965). Laterally this facies type grades into types Xl or Xb. In Sections 20 and 21 quartz wackes of type IVe are associated with this facies.

Type Xn.–Beds of mudstones and wackestones with abundant codiacean algae and furthermore crinoids, brachiopods and lamellibranchs. Some of these fossils are pyritized. These beds occur intercalated in dark grey shale/mudstones of type IIIa or IIIb (Section 7, 1013 m). Lateral tracing of this bed indicated its close association with a seatearth/coal horizon.

Type Xo.–Limestone breccias to poorly rounded conglomerates which occur in facies types Xb or Xp. The clasts may be up to 30 cm in diameter and have the same composition as the immediately underlying deposits. The lateral extension of these breccias is in the order of a few tens of metres, the thickness is up to about 1.5 m. Texturally they are wackestones to packstones. Some grading of the maximum grain size may be present.

Type Xp.–This facies type comprises those sequences characterized by an upward increase in the number of dolomite partings, which do not start in type Xc like Xe, but in Xl. In its upper part it is in places similar to Xa or Xi.

CHAPTER III

INDIVIDUAL FACIES INTERPRETATIONS

INTRODUCTION

In order to keep the more factual observations as presented in Chapter II as separate as possible from the more subjective interpretations, these interpretations will be given here in a separate chapter. In this chapter the interpretations of the individual facies types and groups will be presented without attempting any areal interpretations. Areal and regional interpretations which can only be made after further subjective assessment of the basic data will be presented in the following two chapters. This grouping into three chapters is really a grouping on the basis of the number and importance of the uncertainties, inaccuracies and the subjectiveness of the various levels of interpretation. In this chapter some facies groups and types the interpretation of which is considered to be fairly straightforward will be only be touched upon. Other groups or types which are more difficult to interpret or which are of paramount importance for establishing the environment of deposition, will be dealt with more exhaustively.

FACIES GROUP I – TURBIDITES

The features enumerated such as grading, the presence of part of, or of the entire ABCDE sequence of Bouma (1962), the presence of solemarks, great lateral extension, etc., are characteristic of what may be called sequences with flysch-type sand beds. Kuenen (1967) reviewed the possible interpretations of these deposits and concluded that deposition by turbidity currents entering a realm of clay sedimentation is the most likely interpretation for these, often graded, sand beds. Admittedly (see Kuenen, 1967) some features are difficult to understand in the light of this hypothesis but any other mechanism of deposition postulated presents far greater inconsistencies. With the exception of type If, all types in this group are considered to have been deposited, at least partly, by turbidity currents, i.e. they are turbidites.

Type Ia.–These, often massive, sandstones are difficult to interpret if they occur in isolated exposures. The main reason for including sandstones of this type in group I is their close association with type Ib and the lack of features like macro cross-bedding. In some exposures sole markings, indistinct grading and gradational tops are recognizable. Amalgamation of several beds is a common feature. In this way very thick beds may originate. In some exceptional exposures amalgamation is indicated by multiple grading. The close association of this type with more characteristic flysch-type sand beds and the presence of part of their typical features indicate a similar but not completely identical depositional mechanism. Similar sandstones have been called fluxoturbidites (Dzulynski et al., 1959) or proximal turbidites (Walker, 1967). These sandstones are considered to have been deposited by immature turbidity currents in which grain segregation had not yet occurred (Walker, 1965). As the character of a turbidity current and the resulting deposit is a function of a number of factors, viz. magnitude of the current, character of sediment load, time of travel, distance to source, rate of deceleration, Reading (1970) is followed in his contention that the terms mature and immature turbidites are to be prefered to distal and proximal turbidites. The relative proximity to the source point can only be deduced from a regional analysis of a turbidite sequence, not from the character of a deposit at any single place. Reading (1970) also mentions the possibility that a number of these sandstones were emplaced by a different mechanism like grain flow (cf. type If).

Type Ib.–This facies conforms much better to descriptions of well developed flysch-type sand beds (e.g. Bouma, 1962) and it is therefore interpreted as having formed by the action of mature turbidity currents which entered an environment of quiet clay deposition.

Type Ic.–Sandstones of this type differ from those of type Ia in their composition and in attendant features such as spheroidal weathering. All other characteristics of the finer grained varieties of this type are quite similar to those of type Ia. Because of this similarity these sandstones, too, are interpreted as immature turbidites. The conglomeratic sandstones present in Section 5 (570 m) are included in this facies solely on the ground of their close association with the sandstones typical of type Ic or Id and the lack of features permitting of their recognition as type If. Their genesis, however, is much more likely to be the one inferred for type If, as boulders up to 30 cm in diameter are unlikely to have been transported in suspension (Dott, 1963, p. 123). The association of coarse conglomerates with flysch-type sand beds is not unusual. A recent example in the La Jolla fan valley was described by Shepard et al. (1969).

Type Id.–Like types Ia and Ic, type Ib and this type differ mainly in the composition and the attendant weathering behaviour of the sandstone beds. This type, too is interpreted as having been deposited by mature turbidity currents in an area of clay deposition.

Type Ie.–The characteristics of this type indicate that the sandstones were deposited by very mature turbidity currents. However, if grading becomes indistinct and the interbedded material becomes silty it is very difficult to differentiate between turbidites and normal current deposits. This type has no lithic arenitic or wacke equivalent like Ic or Id. The reason for this is that because of weathering effects the thin bedded sandstones of lithic composition often have a much more massive appearance. Because of this difficulty of recognition these thinner sandstone beds were included in type Ib (for instance Section 7: 0–150 m).

Type If.–The absence of grading and the occurrence of pebbles in the upper part of sandstone beds of this type preclude interpretation as turbidity current deposits. Sandstones with dish structures and big clasts in the top part of the bed are described by Stauffer (1967) in a sequence which ranges from deep water turbidites to shallow-marine and nonmarine deposits. These sandstones with associated conglomerates occur in between the turbidites and the shallow-marine deposits. Stauffer interprets them as grain flow deposits. Sanders (1965, p. 208) states that in inertia flow layers (= grain flow) larger particles tend to move to the top of the flow. Dill (1964) describes moving sheets of sand flowing down steep submarine slopes. These flows start as traction carpets under the influence of shoaling waves and once moving down slope, continue as flowing-grain layers which are kept in motion by the tangential component of gravity. Sanders (op. cit.) argues that such grain flows could keep moving on very gentle slopes. Shepard, Dill & von Rad (1969) consider sand flow (= grain flow) to be an important process in both La Jolla submarine canyon and fan valley. In the fan valley they describe graded sand beds and sand beds showing the ABC part of the Bouma sequence. From the foregoing I conclude that an interpretation as a grain flow deposit does indeed seem appropriate for this facies type, which occurs interbedded in types Ic and Id.

Type Ig.–This type, too, is difficult to explain as the deposit of a simple turbidity current. The ungraded to poorly graded, matrix-rich lower part with features of soft sediment deformation, the sudden jumps in grain sizes, the strongly erosive and channelling appearance of the lower part, all suggest a different mechanism. The very muddy lower part with lime clasts "floating" in a matrix together with contorted shale fragments and lenses, indicate deposition from a viscous suspension (Sanders, 1965) or a mudflow. In this interpretation the graded parts are thought to have been deposited from a turbid suspension associated with the mudflow. The slightly calcareous shale above the coarse-grained part of the turbidites indicates that part of the shale/mudstone was also deposited from the turbidity current. The great thickness of the graded deposit and its relatively great lateral extension suggest that this suspension cloud was a true turbidity current and not a mere cloud hovering on top of the mudflow. The mudflow deposits are sometimes clear channel fills and this proves strong erosion prior to deposition.

Type Ih.–Sanders (1965) argues that the coarse part of this type of deposit was deposited by grain flow layers and that only the upper part of these graded beds was deposited from turbidity currents. Considering the grain sizes of up to 10–15 cm this would seem a likely interpretation for this facies type. Van Hoorn (1970) arrived at a similar conclusion for such graded limestone breccias.

Thick sequences of the types discussed above are inferred to have been deposited in the recent submarine canyon, fan valley and fan systems off the coast of California (e.g. Gorsline & Emery, 1959; Hand & Emery, 1964; Shepard, Dill & Von Rad, 1969 and many others). The facies types discussed are thought to have been deposited in similar environments although not necessarily at the same depth. This applies especially to the laterally and vertically extensive occurrences of these facies types. For the smaller and thinner sequences a different setting is more likely (see following chapters). The differences in composition of the various types are due to a different history of the sediments before resedimentation. The lithic sandstones were resedimented before reworking in a littoral zone could eliminate the unstable components. The compositionally more mature sandstones came either from a different source area or lost their unstable components, i.e. fine-grained rock fragments, during their stay in a littoral zone. The facies types and the fossils present as clasts in the limestone-rich varieties of this group indicate that they are of very local derivation.

FACIES GROUP II – SLUMP AND MUDFLOW DEPOSITS

The intensive deformations observed in sediments of this group are not tectonic, as shown by the absence of cleavage, slickensides, etc. This means that deformation must have occurred shortly after deposition when the sediment was not yet lithified. This relatively unconsolidated state is proved by the plastic deformation of the strata. The limestones, however, do not show indications of simple plastic deformation but of brittle fragmentation instead. All varieties can occur, from barely deformed to completely broken-up and homogenized. The group as a whole is considered to have been deposited by slumping (i.e. sliding) and mudflow (plastic mass flow to viscous fluid flow) mechanisms (Dott, 1963).

Type IIa.–The general interpretation of this group cannot be amplified for this facies type.

Type IIb.–The small-scale unconformities depicted in Figure 6, Plate IId constitute the best proof of the synsedimentary origin of the associated soft sediment deformation. These unconformities are definitely not of tectonic origin because of the absolutely identical facies both above and below the unconformity surface. Their small areal distribution in the present-day outcrops as shown almost completely in Figure 6 also suggests a sedimentary origin instead of a tectonic one. The occurrence of type IIb above and below the lowermost tongue of the Sierra Corisa Limestone Member indicates that the limestone itself underwent some displacement. Proofs of this interpretation are shown in Plate IIc, where a slump ball is visible below the sharp, irregularly scoured, base of the limestone, and in Figure 6 where the unconformities on top of the limestone are depicted. The sharp, irregular base of the limestone cannot be interpreted as an erosive channelling base of a clastic limestone, as the limestone consists of types Xb, c and d. The sand-filled cracks which are restricted to the lower part of the limestone tongue are interpreted as having originated during the displacement of the limestone. During or after these movements the cracks were filled by injected quick sand, i.e. a process very similar to the formation of sand dikes. The displacement of the limestone which is still connected to limestone lying in situ, occurred in several phases, as is indicated by the presence of several unconformities. The total displacement may have been very small, possibly in the order of a few tens of metres.

Type IIc.–Limestone blocks, lenses and beds described as type IIc are surrounded by or associated with types IIa and b, and can often be shown to have experienced movement relative to these strata. A spectacular example is shown in Figure 7. This limestone block, called the Pico Guillermo, in contrast to the Sierra Corisa example, has broken off its adjoining deposits and slid down into a completely different environment.

Type IId.–This type is transitional between IIc and Ig. Normally deposits of this type are intensively contorted, and original bedding is rarely recognizable. Flow lines are occasionally visible in these deposits (Plate IIf). The homogenization is interpreted as being due to deposition from a mudflow. The general absence of sand beds or sandstone fragments in these deposits, which are often clearly erosive, is thought to be due to two different factors. Firstly, the general absence of sand beds in the eroded strata and secondly the unconsolidated nature of the sand beds eroded.

Type IIe.–This type and type IIf, although containing clasts of a very different composition, are texturally identical to type IId, i.e. the coarse to very coarse clasts "float" in a muddy matrix or form a grain-supported framework with such a matrix. In this type the clasts are mainly sandstone fragments which are identical in composition to turbidites of type Ia occurring nearby. The one exposure of this type is interpreted as the deposit of a mudflow which originated locally and which fills a channel cut into type Id turbidites.

Type IIf.–In contrast to type IIe, the quartz arenitic cobbles and boulders of this type cannot have originated locally. The superb rounding testifies to a long history of transport of the clasts before they were finally deposited in a mudflow.

The different types of this group all indicate unstable conditions of the original depositional slopes. Whether this instability was due to sedimentary overloading or to tectonic movements can be deduced partly from the lateral and vertical extensions of the deposit. Only regional correlations can lead to a more definite interpretation of these deposits as due to tectonic or to local causes.

FACIES GROUP III – FINE-GRAINED SILICICLASTICS WHICH FORMED IN QUIET WATERS

These fine-grained sediments were all deposited in quiet water. The water depth at which deposition took place, however, varied greatly. Fossils indicate that all deposits of this group were formed in marine to marginally marine environments.

Type IIIa.–The main clues permitting of interpretation of these deposits are the close associations with limestones of types Xa, b and c, on the one hand and the gradational relationships with types Ic, d and e on the other. In fact the shale/mudstone interbedded with the turbiditic sandstones is identical to deposits of this type. These relationships indicate that the depth of deposition is intermediate between that of the shallow-water limestones and the relatively deep-water turbidite sequences. The sideritic nodules so common in some deposits of this type do not permit of further refinement of this interpretation as they may have been formed in a great variety of environments.

Type IIIb.–This type overlies seatearth horizons with or without associated coals, which form the upper part of coarsening upwards sequences. The seatearth levels were formed in very shallow water or above the water table (see types V). These shales/mudstones contain a clearly marine fauna. Such a sequence can only be interpreted as a transgressive sequence. In the following pages it will be argued that coarsening upwards sequences of type VIIa, were formed by prograding delta distributaries. Taking into account this interpretation, sediments of this type are considered to have formed in lagoons on the delta top (Oomkens, 1967).

Type IIIc.–The association of this type with types Ic, IVa and VIId indicates intermediate to shallow depths of deposition. The intensive bioturbation observed in polished samples indicates that the indistinctly bedded to homogeneous appearance is probably due to this bioturbation. Such mottled or secondarily homogeneous deposits are indicative of relatively low rates of deposition (Moore & Scruton, 1957, fig. 13). The often fragmentary bryozoans and crinoids probably also owe their appearance to scavenging and bioturbating activity, as is inferred from the occasional concentration of fossil debris in distinct burrows.

Type IIId.–The extremely rich fauna and flora locally present in deposits of this type testify to conditions very favourable to life and to the preservation of these organisms. The generally preserved but somewhat indistinct bedding (see Plate IId depicting types IIb and IIId) suggests less intensive reworking of the sediment by organisms than in the case of type IIIc.

The best depth indicators present are the dasycladacean algae and the seatearth level about 20 to 40 m above this facies type (Section 4:1590–1630 m). Both features indicate water depths of a few tens of metres at the utmost. The presence of *Zoophycos* burrows indicates relatively quiet conditions at shallow to intermediate depths (Seilacher, 1967).

Type IIIe.–This type yields a fauna similar to the one found in IIId but on the other hand it is transitional to type IIIa. The main difference with IIId is probably the absence of sand-sized siliciclastic material, indicating very quiet conditions. The latter inference is supported by the observation that solitary rugose corals are sometimes preserved in an upright position, i.e. probably in a position of growth.

FACIES GROUP IV – SILICICLASTICS DEPOSITED IN AGITATED WATERS

This group comprises a large part of the sediments included in Reading's (1970) agitated basin association. The body fossils found indicate that most, if not all, strata of this group were deposited in fully marine to marginal marine environments. The distinction of two sub-groups is based on the compositional maturity of the sandstones. It should be noted that the rare marginal marine pelecypod faunas occur mainly in the mudstones and siltstones of types IVc and d.

Type IVa.–This type is associated with mature sandstones and is considered to have been deposited in an environment characterized by slightly agitated water. Extensive reworking of the sediments by waves and currents is the cause of the absence of unstable components (fine-grained rock fragments) in the sandstones.

Type IVb.–This type is closely associated with type IVa and is interpreted as having formed in the same environment but under more agitated conditions.

Types IVc and d.–The environment inferred for these two types is similar to the ones invoked for IVa and b, i.e. shallow and slightly agitated to agitated waters. The main difference is the immature character of the sandstones. This is probably due to a difference in marine influences as the more marginally marine faunas occur in these types or are associated with them.

Type IVe.–This type is very difficult to interpret as no sedimentary structures have been observed in these deposits. In outcrop deposits of this type have a massive, homogeneous appearance. In thin section the salient features are the superb rounding of some of the sand grains, up to 2.5 mm in size, and the poor sorting. Its marine origin is proved by the limestones in which it is intercalated. Its homogeneous appearance may be due either to very rapid deposition or to bioturbation (Moore & Scruton, 1957).

FACIES GROUP V – DEPOSITS WITH AUTOCHTHONOUS PLANT GROWTH

The presence of rootlets in a sediment of this type is due to autochthonous plant growth. The importance of this group in interpreting other facies types is very great as, for instance, noted by Reading (1970) who states (p. 22):

"Although limited in amount, this association is of disproportionate importance because it is the one certain indicator of depth, proving, if not actual emergence, very shallow water".

The presence of a seatearth below a coal is taken to indicate an autochthonous origin of the coal. If no seatearth was observed below a coally bed, this was considered to be of allochthounous origin.

Type Va.–The fining upwards sequence often recognized in deposits of this type is interpreted as a gradual decrease in competency of currents to carry and deposit sand-sized material. Reading (op. cit.) states the following concerning this feature:

"In some cases the presence of rootlets in rippled sandstone and the decreasing grain size upward through the seatearth suggests that plant growth began under water and current competence decreased as plants became established".

The presence of sideritic concretions does not permit of further refinement of the interpretation of the depositional environment.

Type Vb.–This type of coarsening upwards sequence, with a seatearth on top, yielded marginally marine pelecypods in one outcrop (Section 1,807 m). This means that such a sequence is a shoaling or regressive sequence. Ferm & Cavaroc (1969) show the possible development of such a sequence in their Fig. 9. The lower part of this facies type is thus interpreted to be a marine bay fill deposit, whereas the upper part was deposited on the bay side of a levee or in a coarse overbank environment. In either case subaerial exposure was long enough to permit of autochthonous plant growth.

Type Vc.–In contrast to types Va and b, the sandstones of this type are very well sorted and washed.The associated marine fossils suggest that the maturity of the sandstones is due to intensive reworking by marine waves and currents. This means that this type is to be considered as a deposit formed in emergent beaches or barriers. Another feature supporting this interpretation is the occasional presence of very big, abraded tree stems (max. size observed 5×0.5 m) in this facies type, which is a common phenomenon on recent beaches.

Recent environments claimed to be facies analogues of ancient coal measures of paralic character are, for instance, coastal mangrove swamps, the Everglades of Florida (Spackman et al., 1969, Scholl, 1969) and the delta systems as, for instance, formed by the Mississippi River (Frazier & Osanik, 1969).

FACIES GROUP VI - DEPOSITS FORMED IN (MIGRATING) CHANNELS

Fining upwards sequences showing upward decreases in maximum and median grain size, combined with changes in the size and kind of the associated sedimentary structures, have been interpreted by, for instance, Allen (1965a, b) and Visher (1965a, b) as representing the deposits of migrating rivers. Similar deposits can also be formed by migrating channels in marine environments. The sequences observed can be interpreted in terms of flow regime and of position relative to the channel. The parting lineation has been formed in the lower part of the upper flow regime, the large cross-stratification in the upper part of the lower flow regime and the laminated and rippled sands and silts in the lower part of the lower flow regime and by vertical accretion. The macro cross-bedded sands and the sands with parting lineation are both in-channel deposits. The micro cross-bedded sands and silts are partly in-channel and vertical accretion deposits. The overlying laminated sands, silts and mudstones are also vertical accretion deposits. A lag deposit is sometimes present at the base of these sequences. The character of these lag deposits, together with the faunas of the associated facies types and the presence or absence of autochthonous plant growth in the vertical accretion deposits, form the most important clues for deciding upon a fluvial or a marine environment respectively.

Type VIa.–Deposits of this type are closely associated with facies types containing marine fossils (Section 3, 152 m; Section 5, 304 and 407 m; Section 12, 390 m). This association indicates an origin in a marine environment for this facies. This would also agree with the relatively mature composition of the sandstones. The lower parts of these sequences, sometimes conglomeratic, indicate that competent currents must have occurred. In a marine environment such strong currents are likely to be at least partly due to tides. The presence of this facies type is one of the few indications of the influence of tides in the entire interval studied as features like herring-bone cross-stratification or flaser und linsen structures have not been identified.

Type VIb.–Fining upwards sequences of this type can be identified as being of fluvial origin by the absence of marine organisms in the lag deposits, by the presence of rootlets in the upper part of the sequences and by the abundant transported plant fragments (Visher, 1965b). Notable in this facies type is the immature composition of the sands which is a common feature of fluviatile sands. The presence of clay galls several centimetres in diameter and of tree logs up to a metre in length, proves that the rivers were able to transport coarse material. Nevertheless, the coarsest quartz grains and lithic fragments are not over 2 mm in maximum diameter. This is taken to indicate that no coarse material was supplied by the rivers to the depositional basin. If fining upwards sequences on top of coarsening upwards sequences were not included in the latter, this facies type would be much more common. Now it is mainly restricted to Section 1, 750–860 m and Section 4, 1100–1145 m.

FACIES GROUP VII - DELTAIC AND MARINE REGRESSIVE SEQUENCES

The upward increase in grain size with the attendant change in the kinds of sedimentary structures present, can be interpreted in terms of an increase in water agitation. Such a sequence is either a regressive marine sequence or a deltaic sequence (Visher, 1965b). Both are due to a predominance of sedimentation over subsidence which causes shoaling and eventually emergence because of the prograding of the coast, i.e. these are regressive sequences. A simple marine regressive sequence usually contains well washed littoral deposits and lagoonal sediments, whilst the deltaic sequence is characterized by the presence of fluvial deposits.

Type VIIa.–This type of coarsening upwards sequence (c.u.s.), which locally contains marine fossils at the base, is usually topped by channel fills or fining upwards sequences identical to the deposits of type VIb. In other words, these are fluviatile deposits which form the top part of the c.u.s. The sandstones below the distinctly fluvial part of the sequence are also lithic arenites which means that these sandstones were not reworked by waves or currents. The deltaic sequence has been discussed by, among others, Scruton (1960), Visher (1965b), Fisher et al. (1969). With their data available it is possible to differentiate between the various parts of these sequences according to the respective depositional environments. This differentiation is only feasible for the major c.u.s., for the minor sequences slightly different settings have to be invoked. The pro-delta environment is characterized by shales/mudstones and silts which may show lamination, rippling, some graded bedding and some burrowing. When measuring a section, the selection of the horizon at which a c.u.s. is considered to begin is arbitrary because of the gradational nature of this contact. In general graded beds were not considered to belong to type VIIa. The delta-front deposits consist of rippled and laminated sands, low-angle tabular cross-bedded sands, sands deposited in minor channels, and trough cross-bedded sands. These sediments were deposited by migrating distributary mouth bars, subaqueous levees, distal bars and marginal sheet sands. The only delta top deposits included in this type are the fining upwards sequences in the upper part of the sequences. These are included in this type as their formation on top of a c.u.s. is the logical consequence of a single process, i.e. the extension of a delta distributary because of sedimentation at its mouth. Although the bases of these fining upwards sequences are erosive, extensive erosion of the underlying mouth bar deposits is rare. The water depth in which these major c.u.s. were formed is considered to have been of the same order as the length of the sequence. The reason for this assumption is the dominance of sand and silt sized material in these sequences, which materials do not compact as much as finer grained sediments. The major coarsening upwards sequences are considered to have formed because of the prograding of major delta distributaries into a relatively deep hydrographic basin. The minor sequences included in this type only occur above major sequences. This indicates that they were formed in one of the delta-top environments. The prograding of minor distributaries into delta-top lagoons, or crevassing of major distributaries are invoked to explain their presence.

Type VIIb.–The mature composition of the sandstones and the association with sediments with marine faunas are the main differences to type VIIa. The reduced importance of unstable fragments in the sands suggests reworking of the sands by waves and currents. Seatearth levels may be present in closely associated sediments. If this is the case, however, especially the sandstones are very different from those inferred to be of fluviatile origin. The main occurrence of this type (Section 10, 500–597 m) can only be interpreted as a regressive marine sequence. Only in the next chapter will a more detailed interpretation be possible. Another occurrence of this type (Section 12, 255–280 m) differs from the first in that channel fills are present at the top. This is more like the normal deltaic sequence. The marine fauna at 295 m and the mature character of the sandstones indicate, however, that this is not a normal fluviatile delta. The features mentioned are best explained if this example is interpreted as a tidal delta.

Type VIIc.–This type of coarsening upwards sequence comprises compositionally mature sandstones which are commonly well sorted. Further characteristics are the often evenly laminated aspect and the close association with marine faunas. The presence of a seatearth (Vc) is indicative of emergence. This combination of features is characteristic of recent beach deposits. In recent delta systems a number of beach forms can be distinguished (Todd, 1968; Hails & Hoyt, 1968; Hoyt, 1969). The forms of sandy beaches relevant for interpreting this facies type are:

(1) barrier beaches which, according to Hoyt, are formed by slow submergence of pre-existing beach ridges. These may form very extensive deposits which are up to several tens of metres thick;

(2) spits are formed by longshore currents and may give rise to a barrier. Hardly distinguishable from 1) in stratigraphic sections;

(3) cheniers are formed by periodic reworking of a prograding muddy coast. They do not form very extensive sand deposits. The thickness of the Lousiana examples does not exceed 8 m (Hoyt, 1969);

(4) "normal" beach ridges in deltaic areas are here considered to be those sandy beaches which form due to a delicate balance of waves, currents and sediment supply at a specific place and which are not produced by reworking of a pre-existing muddy shore. These deposits may also be quite extensive and relatively thick. The recognition of the various forms is extremely difficult in stratigraphic sections as the lateral equivalents of these deposits can seldom be observed. Even if exposures are favourable, the exact time equivalence of strata has yet to be proved. Considering only the thickness it seems likely that the examples present in Section 10, 598–643 m, are chenier deposits. For all other occurrences of this facies type, each of the alternatives mentioned seems equally likely to me.

Type VIId.–The absence of seatearth levels indicates that the upper part of this facies type was never emergent for long periods. The alternation with the relatively fossiliferous deposits of IIIc suggests open marine conditions. Because of these features this type is interpreted as having formed as offshore bars.

FACIES GROUP VIII – TRANSGRESSIVE SEQUENCES

Type VIIIa.–The upward decrease in grain size of this type is interpreted as a gradual decrease in water agitation. In the two best exposures of this facies type (Section 4, 1148–1190 m; Section 10, 643–646 m) it overlies sediments with seatearth horizons and in turn is overlain by sediments containing marine fossils. In other words, these sediments form a deepening sequence. As they begin on former land surfaces (the seatearth horizons) they may also be called transgressive sequences. Visher (1965b) discussed the aspects of different forms of transgressive sequences. Fisher (1961) described the sequences at present developing on the Atlantic coast of New Jersey. With their data the type under discussion can be interpreted. Fisher considers that in basins with tides of limited magnitude and not too great wave energy a transgression would leave the following record:

"A stratigraphic section through such a transgressive sequence would show a transition from lagoonal shoreline deposits to deeper lagoonal deposits, then up into lagoonal shoreline deposits, disconformably overlain by the sediments of the beach and of the open sea".

In the examples observed, lagoonal deposits (for instance IIIb) are thin or absent. This indicates extensive erosion of part of the sequence. This means that the sharp bases of some of these sandstones are disconformity surfaces which can be of regional importance (Oomkens, 1967).

A transgressive sequence is not necessarily developed as a siliciclastic sequence as dealt with above. Limestones may also occur in the lower part (Section 5, 200–280 m; Section 12, 580 m ??). Aspects of these various developments will be dealt with in the following chapter.

FACIES GROUP IX – KARST INFILLINGS

Type IXa.–The presence of originally vertical pipes of up to seven metres depth in limestones with associated limestone conglomerates, and development of secondary porosity can only be explained by inferring subaerial erosion and solution by meteoric waters. As such, the presence of this facies is indicative of prolonged emergence of the limestones on which it is present. The karst holes are filled with sands containing a marine fauna not present in the limestones. This indicates that erosion probably occurred in near-littoral environments (cf. Newell & Rigby, 1957, Plate XX, Fig. 1). The sequences observed in the sandstones are commonly of type VIIIa or VIIc, the latter with or without indications of autochthonous plant growth (Vc). This is interpreted as follows: after a more of less prolonged emergence the karstified limestone surface was transgressed by the sea. In this shallow sea the sands were extensively reworked before deposition as is proved by the extremely mature composition of the sandstones. These newly deposited sands first filled in the depressions of the surface and part of the secondary porosity. Locally emergent beaches developed which could support plant growth but after some time these, too, were submerged and deposits of type VIIIa, developed.

In areas where limestone was not emerging the contact with the overlying sands of this facies is either sharp, but not clearly erosive, or gradual.

FACIES GROUP X – MARINE LIMESTONES

The fossils found in all limestone deposits in this area prove that these deposits are all marine limestones. The algae, which are present in great quantities in most facies types and which in general occur throughout the entire section, prove that the limestones were deposited in shallow water. Especially the presence of dasycladacean algae indicates very shallow water. With the data and interpretations presented by Rácz (1966a, b) available it may be concluded that the algal assemblages observed in the various limestone members indicate waterdepths not exceeding 50 to 100 m. If dasycladacean algae and oncolites are present, they indicate waterdepths not greater than 10–25 m. From sedimentological evidence presented later it even becomes obvious that waterdepths in the order of a few metres were common. From the mere presence of the generally fairly pure limestone deposits one more general conclusion can be drawn: in this area the supply of siliciclastics at many points and moments was negligible.

Type Xa.–The complete lateral and vertical gradation with siliciclastic mudstones with approximately the same fossil content, the highly characteristic fossil content itself, the lack of current-induced sedimentary structures, the fine-grained matrix in between the "clasts" cannot be explained satisfactorily by assuming a normal, i.e. clastic mode of deposition of this facies type. All these characteristics obliged me to consider other interpretations. Garrison & Fisher (1969), in their description of the Jurassic Adnet Beds, figure a number of specimens of these nodular limestones. These photographs (Figs. 5, 6, 7(1) are very similar to our Plate Va. Garrison & Fisher argue, that the nodular appearance of these Adnet Beds is due to a combination of subaqueous solution of the fine-grained carbonate muds, alternating with subaqueous lithification and possible shrinkage-cracking of these sediments. Their contention is that these processes produced a rubble of semi-consolidated to consolidated nodules which were embedded in an insoluble residue. Paleontological evidence proves in this particular case that the rate of deposition was very low. Apart from the pure paleontological evidence, the preservation of ammonites, the presence of manganese and hematite crusts also indicate low sedimentation rates. In the case of type Xa, relatively low sedimentation rates are not unlikely as is indicated by the presence of a few phosphatic nodules and by the preservation of a goniatite found in Section 21 (532 m). This undeterminable goniatite had a fairly well preserved lower side and a corroded upper side. Most other fossils, however, are well preserved, which is due to an oncolitic coating on most specimens collected. The oncolitic coatings and the divers fauna, in addition to the calcareous algae found in this facies type, indicate an important difference to the environment of deposition invoked for the Adnet Beds. While the latter are thought to have been deposited at depths of up to 4,500 m, the former were certainly formed in much shallower water. The presence of algae makes waterdepths exceeding 50 to 100 m most unlikely. The presence of diagenetic dolomite in the insoluble-rich parts of this type, in combination with interpenetrating contacts between "clasts", makes it likely that post-depositional pressure solution phenomena were important. Nevertheless, I consider it likely that the mechanisms invoked by Garrison & Fisher played an important part in producing the rock fabric observed. The occurrence of this type at the base of the Agujas and Abismo Limestone Members or as isolated lenses in the southern flank of the Castillería Syncline indicates, that it was likely to form as the first carbonate deposit in areas with a diminishing supply of siliciclastic material.

Type Xb.–This facies type is considered to have been deposited under relatively quiet conditions because of the high carbonate mud content. The poorly bedded and often apparently homogeneous appearance are possibly due to extensive bioturbation on both a macro and a micro scale. In the same manner, scavenging activity of some of the faunal elements causing homogenization of the sediment is invoked to explain the widely scattered occurrences of crinoid ossicles, broken shells, etc.. The only problem is that, except in type Xj, distinct burrows or tracks have rarely been observed in any of the limestone facies types. However, from the flora and fauna present it is obvious that conditions were suitable for the occurrence of a rich benthonic fauna both upon and in the sediments. Moore & Scruton (1957) showed that in recent siliciclastic deposits complete homogenization may occur as a result of the burrowing activity of organisms which do not leave any other trace of their presence (see also interpretation of type Xm). The dolomitization observed in this facies type is connected with pressure solution seams and as such is considerd to be diagenetic. That it is not of synsedimentary (=primary) origin is also indicated by the fact that the dolomitized parts are fine to medium crystalline, whilst primary dolomites tend to be much more fine-grained. Furthermore primary dolomites are much more extensive than the dolomite streaks of this facies which are at best a few decimetres long. Nevertheless, the restriction of this dolomitization to laterally extensive sequences suggests an association with some primary factor influencing the composition of the sediment.

Type Xc.–This facies type, too, is considered to have been deposited under relatively quiet conditions because of the predominance of carbonate mud. The more massive character is thought to indicate more continuous processes of deposition and homogenization. However, the local presence of non-bedded, algal boundstones suggests that the massive homogeneous appearance may at least partly be an original feature. The rare occurrences of birdseye structures with geopetal fills prove that organic or inorganic consolidation of the sediment occurred locally immediately after deposition. Shinn (1968) argues that birdseye structures are good evidence of a supratidal or intratidal depositional environment. Whether this holds good for every kind of birdseye remains to be proved but their presence further corroborates the argument in favour of a shallow water origin of the limestones under discussion.

De Meijer (1969) proved the presence of non-calcareous algae in similar limestones from the slightly older Lois-Ciguera Formation. He also found these non-calcareous algae in a sample from this facies type. He considers these non-calcareous algae, which were also observed in other facies types, to have been important agents in the entrapment and consolidation of these carbonate sediments.

Type Xd.–The sample showing the most important clues as to the mode of formation of this type was not collected from the Agujas Limestone Member but from the Sierra Corisa Limestone Member. This sample was collected about 150 m west of the line along which Section 4 was measured, at the point where the limestone interfingers with fine-grained siliciclastics. This sample is shown on Plate VIa, b. In these photographs it can be seen that the "primary" components of the rock are the undulating fronds or "leaves" of codiacean algae (phylloid algae in the terminology of Pray & Wray, 1963). According to Pray & Wray, these phylloid algae probably grew as upright plants (p. 216). The reason for this assumption is the bilateral symmetry of the internal structures. This means that the sub-horizontal position of the "leaves" indicates that they have fallen over. Because of the small quantity of sediment present in the sample depicted, the fronds necessarily formed a grain-supported framework. The geopetal structures are proof that the "leaves" formed an open framework through which overlying sediments could filter down. The remainder of the pore space was then filled with fibrous and equant sparry calcite, i.e. typical void-filling sequences. It is obvious that the proportion of crystalline calcite cement and original carbonate sediment may vary greatly. Furthermore, it can be seen that even in the specimen shown on Plate VIb the fine structures and textures which make the algae recognizable as such, are ill-preserved. In the other examples depicted, algae are not recognizable and the amount of sediment in relation to the amount of crystalline cement is much greater. In some examples (Plate VIIa) only vague masses of fibrous calcite are present, pervading a mudstone/wackestone. On Plates VIf, g partly and almost completely brecciated examples of this facies type are shown. This brecciation is interpreted as a synsedimentary to early diagenetic collapse of sediments of this type under the influence of overlying sediments. That brecciation is indeed a very early phenomenon is proved by the fact that only the sediment and not the crystalline calcite is brecciated. The very sharp edges of the breccia fragments and the lack of indications of plastic deformation prove that this sediment was consolidated and possibly lithified very soon after deposition. The example pictured on Plate Vh is of a slightly different type because of the radiating tube-like structures in the sparite filled cavities. Furthermore, the cavities have much greater dimensions than in any other example.

However different the various occurrences of these sparite patches may look at first sight, they nevertheless have in common that they all occur closely associated with or surrounded by facies type Xc. Furthermore, any single occurrence shows complete gradations to facies type Xc. Although no such series can be observed in the field, from the samples collected a continuous series can be arranged showing gradations from one form to another. This strongly suggests a generally similar mode of formation with minor differences in environment for most occurrences, which caused the different aspects of this type. Organisms, in some cases proved to have been algae, played an important role in the formation of these sparite filled cavities and the attendant entrapment of sediment. Although it has been shown that the algae are probably no longer in position of life, the indications of early consolidation of the entrapped sediment and the occurrence of specimens as depicted on Plate Vh lead me to consider the rocks of this facies type to be boundstones. Admittedly, it is possible to prove this interpretation for only a minority of the examples studied. Therefore it is wiser to mention the interpretation of very similar deposits presented by Pray & Wray (1963). They consider their examples to be biostromal deposits formed by essentially in situ clastic accumulation of algal fragments in a relatively turbulent environment. In the case of the Agujas Limestone Member and the other limestones studied, however, I consider it most unlikely that these deposits were formed in a relatively turbulent environment because of their association with mudstones and wackestones. Packstones and oölitic grainstones truely indicating agitated water are not closely associated with this facies type.

Some of the sparite patches discussed agree with the definition of the term stromatactis given by Chilingar & Bissell (1967, p. 167), and this term has been used loosely during fieldwork and in Fig. 12.

Types Xe and Xf.–As the lithological characteristics of these two types are nearly identical to those of types Xb and Xc, the interpretation is identical too. The transitional contacts which characterize these two types are thought to be due to the lateral shifting of the realms of types Xb and Xc. There are as yet no clues as to whether Xe or Xf represent, for instance, a shoaling sequence or a difference in wave energy.

Type Xg.–The packstones and grainstones of this facies originated in a strongly agitated water environment (Plumley et al., 1962). This is indicated by the presence of oölites, the occasional presence of overturned and rounded fragments of chaetetid corals, of well rounded pebbles of grainstone, and of quartz arenitic pebbles. The presence of a few very angular fragments of the underlying mudstone also suggests strong currents which could erode consolidated sediment. In the best exposed example these deposits seem to grade in upward direction into Xb or Xi. Those two types represent a much less agitated environment. In Xg no distinct structures or definite sequence were observed. This type is considered to have been deposited either by a migrating channel or by a migrating shoal or bar. Other aspects such as the interpretation of the "mottled" parts will be dealt with under type Xh.

Type Xh.–The upward change from mudstones and wackestones into packstones and grainstones represents a change from relatively quiet water deposits into deposits formed in relatively agitated water. In analogy to siliciclastic deposits, such a sequence is here called a coarsening upwards sequence. In siliciclastic deposits a coarsening upwards sequence may safely be interpreted as a shoaling sequence and often it can also be proved to be a regressive sequence. In carbonates this is much more difficult to prove. The evidence for interpreting this facies type as a shoaling sequence will be presented in the next chapter, where the areal facies relationships are discussed. These areal facies relationships form the best evidence for this and other interpretations. Nevertheless, the oölitic grainstones themselves can be interpreted as deposits formed in relatively agitated and shallow water.

The quartz arenitic pebbles which occur in type Xg and in this type constitute additional evidence of strongly agitated water, but a much more important conclusion can be drawn from their presence. A very competent longshore or shallow marine drift system must have existed at the time of deposition of the Agujas Limestone Member in order for these pebbles to be deposited in this place.

Included in this facies type and in type Xg are the peculiar "mottled" pelsparitic grainstones which are depicted on Plates VIIh, VIIIa, b, c, d. As shown in these photographs, the "mottles" are sub-horizontal blebs, rods and plates of various dimensions. The difference in composition between the rounded "mottles", which are slightly more sparitic, and the intervening sediment, which contains more matrix, is primary. It is only accentuated by the presence of pressure solution seams and stylolites. For the interpretation of the depositional environment of these pelsparitic grainstones the genesis of the pellets is of importance. This origin, however, is quite difficult to establish. The subspherical shape and the good sorting of the pellets do not provide any definite clues. Folk (1962), although using pellets as a purely descriptive term, nevertheless considers a majority of those pellets to be very probably of fecal origin. He bases this view on the generally constant size, shape and the extra high content of organic matter. Beales (1965) states that pellets may be formed by organic agglutination, inorganic precipitation and cementation, recrystallization, or by a combination of these processes. Although such a variety of processes may play a role in pellet formation, the remarkable similarity and uniformity of ancient pelletiferous limestones suggest similar depositional and diagenetic histories. If the pellets are of fecal origin very complete reworking of the sediment must have occurred. Beales (p. 52) states the following concerning this point:

"A lack of coarse skeletal debris, uniform pellet size, commonly pale coloration, and a meagre insoluble residue characterize many pelleted limestones. The fine comminution of skeletal debris could be due to feverish activity of many scavengers working and reworking the sediments for every last vestige of nutrient material. Alternatively, and perhaps more likely, the shallow to extremely shallow water, which environmental interpretations seem to suggest, may have been largely stripped of nutrients in the marginal areas, and only a restricted fauna and flora could survive in the sometimes vast interior areas. Under such circumstances mud-ingesting organisms would process and pellet large amounts of predominantly precipitated sediment for the little food contained therein.

Whatever the pelleting organisms were, most of them left no trace of burrows in the majority of pelleted limestones."

This last remark points towards the main difficulty in interpreting our examples. Although the pellets may well be fecal pellets, the structures observed are definitely not burrows. Dr. Cl. L. V. Monty of Liège University suggested that non-calcareous algae played a role in their formation because of the numerous algal threads and tubes visible in thin sections under high magnifications (100 × and over) (pers. comm.). Because of the presence of these non-calcareous algae, whose importance in the production, entrapment and consolidation of sediment has been shown by Monty (1965a, 1965b, 1967), the platy structures shown on

Plate VIIIa, b are considered as algal mats. In the light of this interpretation the possibility that the pellets, too, are of algal and not of fecal origin, has to be borne in mind. A third possibility is indicated by the presence in some samples of small foraminifera of pellet size. The preservation of these foraminifera varies from good to very poor. This indicates that part of the pellets originated by micritization (=grain diminution, Wolf, 1965) of small fossils.

Disruption due to dessiccation of these mats seems a likely process to explain the occurrence of the discontinuous plates and the rounded blebs. This dessiccation was not necessarily a subaerial process. Subaqueous syneresis may well have produced the rounded shapes and the concretionary aspect of some of these "mottles".

Compared to the facies types already described the occurrence in types Xg and Xh of distinctly and regularly bedded sequences together with more massive parts is an important difference. Whether the massive, homogeneous appearance of parts of these facies is due to biogenic homogenization, extremely rapid deposition or to a very continuous, uninterrupted sedimentation process could not be established. The presence of well-bedded parts suggests, however, that at least part of these sequences were deposited too rapidly for organic homogenization to occur.

The preceding interpretations can now be combined and summarized. The coarse, often oolitic packstones and grainstones were deposited in probably sometimes emergent shoals and in migrating channels. A relatively rich and varied flora could flourish in these environments. Relatively high sedimentation rates are characteristic of part of these deposits. On top of these shoals the water was less agitated as is deduced from the decreasing amount of oolites. Most agitated conditions are assumed to have occurred at the rims of the shoals where the waves broke. Conditions of life were furthermore less favourable as is inferred from the common absence of a preserved fauna. In this environment huge amounts of pellets were formed by some mechanism (a fecal, algal or micritization origin are considered the most likely). The pelletiferous sediment was trapped and consolidated in algal mats. The algal mats have been preserved as such or were broken by dessiccation in either a subaerial or a subaqueous environment. The depth of water in which these deposits were formed is considered to have been in the subtidal to intratidal range.

Type Xi.–The rich coral fauna and the equally rich algal flora, together with the presence of other faunal elements, indicate conditions favourable to life. The presence of oncolites points towards a shallow, subtidal environment with sluggish currents (Logan, Rezak, & Ginsburg, 1964). The latter interpretation is also supported by the presence of mudstone in between the corals. The relatively high content of siliciclastic insolubles in this facies type may be interpreted in various ways. It may mean a somewhat greater supply of these materials. Its presence may, however, be equally well explained by assuming a relatively low sedimentation rate of carbonate material, causing less dilution of the siliciclastics. The preservation of the chaetetid, auloporid and "lithostrotionid" corals and of many of the dasycladacean and possibly also of the phylloid codiacean algae in position of growth raises the question whether these organisms needed a hard substrate. No indications of early lithification were observed. The only hard objects available were probably the remains of other organisms. Although no special attention was paid to this feature in the field, I have the impression that the corals and algae could settle and flourish on a, possibly consolidated, muddy substrate (cf. Rich, 1969, p. 353). The number of corals and algae present in growth position at the majority of occurrences of this type induce me to consider it to be a coralgal biostromal deposit formed in quiet water.

This facies type with its rich floras and faunas commonly occurs right on top of the agitated water deposits of types Xg and Xh. These two types were interpreted as subtidal to intratidal deposits with especially in the upper part of the sequence conditions unfavourable to most forms of life. This return to favourable conditions with a fairly great variety of life forms is interpreted as a deepening or a transgressive sequence. Because of the deepening, the oölitic and pelsparitic grainstone shoals "drowned" and their upper surface was then covered by the settling corals and algae. Rich (1969) also considers *Chaetetes* to have grown in slightly deeper and quieter water than where the oolites and coated grains were formed and deposited (Fig. 10, p. 357). Areal distribution of facies supporting this interpretation will be dealt with in the next chapter.

Type Xj.–The clastic wackestones, packstones and grainstones of this type were deposited in shallow, fairly agitated water. Bedding and vague channelling are preserved, indicating that burrowing organisms though present, were unable to rework and homogenize all of the sediment. The genetic significance of the chertification along bedding planes is not clear. From observations made on thin sections it seems likely that porosity differences played an important role in determining the location of the concretions.

Type Xk.–This type, too, shows a coarsening upwards sequence like type Xh. It differs from Xh in that it begins on deposits of type Xb instead of type Xc and that it does not contain oölites in its upper part. Like type Xh it is considered, for reasons discussed in the next chapter, to represent a shoaling sequence.

Type Xl.–This facies type was formed in quiet waters in an environment not very favourable to life. The main argument in favour of this interpretation is the observation that this type seems to be intermediate between Xb and Xm.

Type Xm.–These fetid limestones yield a relatively high quantity of insoluble organic material. This high content of organics is the main clue for interpreting these unfossiliferous and undisturbed deposits. Purdy (1964), in his discussion of the influence of sedimentary substrates on benthonic faunas states that if the content of organic matter in a sediment is too high (approx. 3–4%), decomposition produces too many toxic compounds. Especially in poorly permeable, i.e. fine-grained deposits this leads to conditions hostile to benthonic organisms. From the complete absence of a recognizable flora or fauna and the undisturbed aspect of even the thinnest sedimentary laminae, I conclude that no benthonic life could be supported by this facies type.

The cross-lamination with few erosional features and chiefly tangential contacts between the laminae, i.e. kappa-cross-stratification (Allen, 1963, Fig. 4a) is indicative of relatively high deposition rates from suspended load. Their presence indicates that the water above the depositional interface was not stagnant, at least not all the time, but that sluggish currents occurred from time to time. This favours the interpretation that toxic conditions in and upon the sediment and not a lack of circulation of the water above the sediment, were the main factor responsible for the absence of benthonic life in these fine-grained clastic deposits.

The relative depth of deposition of these quiet water deposits can, once more, only be established after considering the areal facies relationships. Nevertheless, the general similarity of this type to some of the examples described by Wilson (1969) suggests an environment of deposition at least slightly deeper than that inferred for the other limestone types.

As extensive recrystallization to microspar or even pseudospar (Folk 1965) has occurred, the original grain sizes of these sediments are difficult to establish, but are thought to have been in the clay to silt range.

Type Xn.—This facies was mainly differentiated because of its close association with shale/mudstone of type IIIb, which in turn was found overlying a coarsening upwards sequence of type VIIa. Furthermore, this facies type differs from the very similar types Xb and Xi in that it contains pelecypods which were not found in any other limestone type. Type IIIb is interpreted as a lagoonal mudstone formed after the marsh, in which the first Por si Acaso coal formed, was drowned. Because Xn overlies IIIb, and as it is of shallow water origin, it is interpreted as having been deposited in the same lagoonal environment on the top of the delta. Such lagoonal limestones are mentioned by Oomkens (1967).

Type Xo.—These breccias to conglomerates occurring in types Xb or Xp are products of local erosion and redeposition. This is deduced from the poorly rounded forms of the clasts and from their composition which is very similar to that of the underlying strata. Their occurrence in otherwise relatively quiet water deposits suggests that they may have originated due to local currents caused by storm surge (Hayes, 1967).

Type Xp.—As this type differs only from type Xe, in that it is transitional from Xl instead of from Xc, its environment of deposition is considered to be nearly the same. The environmental significance of this sequence in terms of water depth or wave agitation is not yet understood. From the combined interpretations of Xb and Xl it is clear that this sequence represents conditions gradually more favourable to life.

CHAPTER IV

AREAL INTERPRETATION OF LATERAL AND VERTICAL FACIES RELATIONS

Within the various structural units (Fig. 8) lithostratigraphic correlations are easy to make. Correlations between the various structural units are based mainly on paleontological data. These regional aspects will be dealt with in the following chapter. Only some of the more obvious lithostratigraphic correlations will be mentioned in this chapter and the facies relationships will be discussed in their stratigraphic sequence. The approximate distribution of the major facies groups is shown in Enclosure 3.

SOUTH FLANK OF THE CASTILLERÍA SYNCLINE
(Figures 8, 9a, b, c and Enclosures 1, 3, 5)

The lowermost part of Sections 1 and 4 consists of quartz arenitic turbidites. In Section 1 only mature turbidites (types Id and e) are present, whilst in Section 4 immature turbidites (type Ic) also occur. Tracing this interval laterally from Sections 1 to 4 proved that these immature turbidites make their first appearance south of Section 2 and increase in importance towards Section 4. More to the east they disappear again, and north of San Cebrián de Mudá only mature turbidites have been observed. The sole markings indicate current directions from SW to NE (Fig. 14a). This direction is oblique to the plane of a cross section through Sections 1 and 4. This indicates that the basal part of Section 4 cannot be considered to be more proximal than the equivalent deposits in Section 1. It is for this reason that I interpret the present-day outcrop of this interval as a somewhat oblique transverse section through submarine fan valley deposits. Menard (1955) and Shepard et al. (1969), among others, describe recent fan valleys in

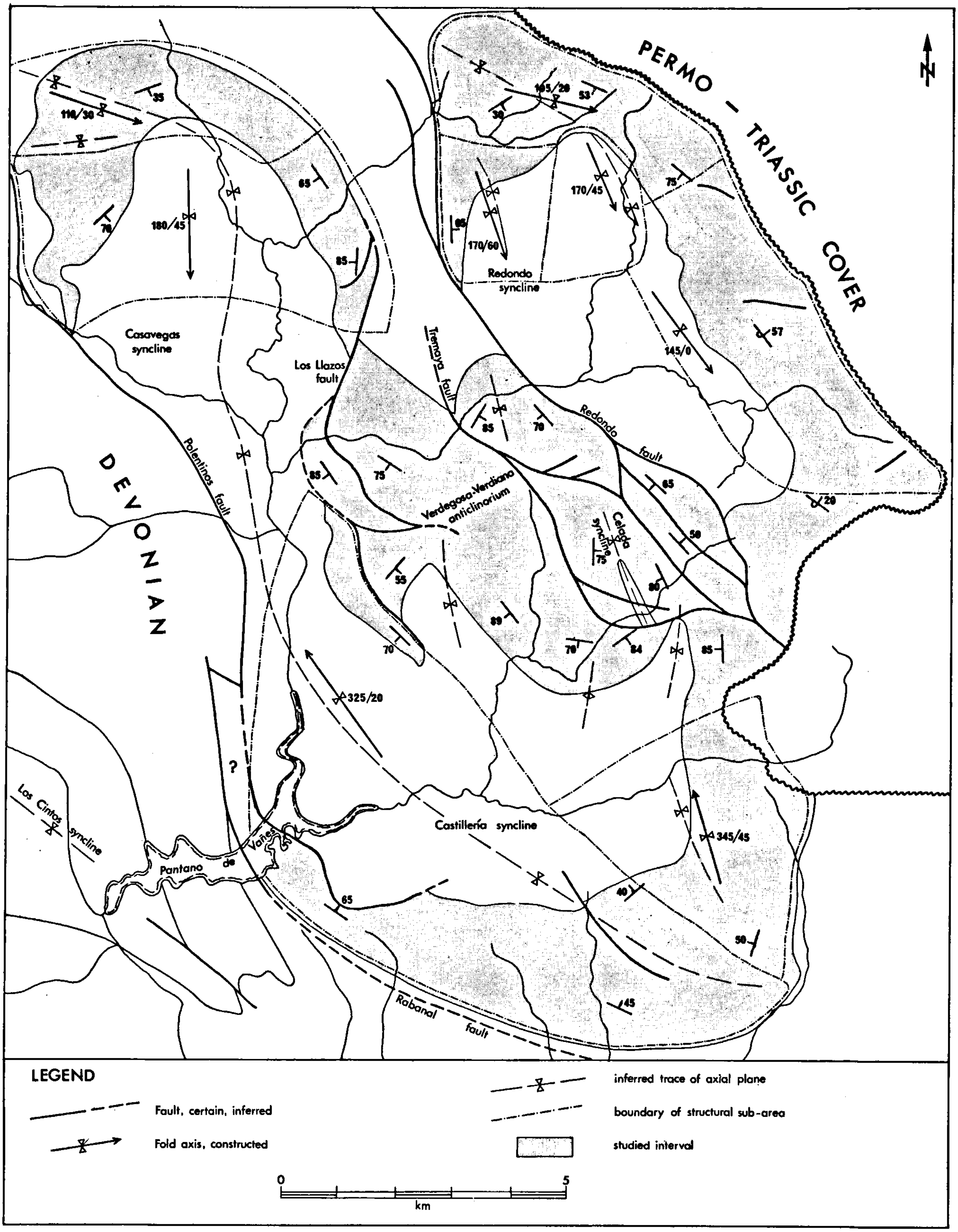

Fig. 8. Structural map of the studied area.

which coarser sands were deposited in the axial parts than on the levees. Apart from the differences in grain sizes, differences in sedimentary structures were also noted. These differences, if translated into terms of relative maturity, indicate the presence of more immature currents in the axial parts of the valleys. In the axial parts, sand flow, creep and slumping may also be important. In this interpretation Section 4 is located somewhere near the axis of the valley whilst Section 1 and the deposits north of San Cebrián were deposited nearer to or even on the levees.

The overlying shale/mudstone sequence indicates that the supply of coarse siliciclastics was negligible for a considerable time.

The succeeding lithic arenitic turbidites are compositionally identical to the deltaic and fluviatile(?) sandstones above them. In Section 4 graded sand beds are present up to the base of the deltaic coarsening upwards sequence. (c.u.s.). The turbidity currents which deposited these turbidites are therefore inferred to have originated right on the delta front where the sediment was not reworked by marine waves and currents. The absence of slumping phenomena in the lower parts of the coarsening upwards sequences brings to mind the possibility that turbidity currents originated at the river mouth during flood stages, as for instance suggested by Heezen et al. (1964), Houbolt & Jonker (1968), Nesterof et al. (1969) and van Straaten (1959). Reading (1970), discussing his river generated sandstone association, also concluded that turbidity currents may originate from rivers in flood. The sole marks measured on these turbidites indicate derivation from the west (cf. Fig. 14a). This constitutes a fairly marked difference to the direction of supply of the quartz arenitic turbidites. This difference is an indication of the important changes in the general setting of the area during the time represented by the shale/mudstone interval.

The deltaic and fluviatile sandstones immediately succeeding the lithic arenitic turbidites are mainly distributary mouth bar and channel deposits. At the shore of the Vañes reservoir (Spanish: Pantano de Requejada; also Pantano de Vañes) where the only distinct fluviatile deposits were recognized, cross-stratification indicates a westerly derivation (Fig. 14a). The well-developed seatearth levels occurring in Section 1 (approx. 800 m) were not recognized at the corresponding horizon in Section 4 (approx. 530 m), although Reading (1970) reports autochthonous plant growth in this level (his Fig. 3 – Section 9, approx. 450 m; N.B. as indicated by the explanation of Fig. 3 and Fig. 7, Fig. 3 – Section 9 and Fig. 3 – Section 10 have been reversed, i.e. in Fig. 3 Section 9 should be Section 10 and vice versa). A few hundred metres to the west I observed a good seatearth at this level. These deltaic deposits are overlain by marine mudstones with a few limestone lenses (lower level of Socavón Limestone Member) which become more frequent and thicker towards the east. The directions of supply and the facies of both siliciclastics and carbonates indicate more marine conditions towards the east.

Above the limestone horizon another interval follows, with deltaic deposits which are considered to have formed in bays and sounds on the delta. Crevasse splay deposits are probably important in this interval. The lenses of the second level of the Socavón Limestone Member overlie this interval. This limestone horizon cannot be traced as far west as the lower one. Mapping evidence indicates that the two limestone lenses below the San Cebrián coals in Section 5 probably correlate with this second limestone level. In Section 4 (approx. 800 m) and near San Cebrián, well washed, compositionally mature sandstones are present above the limestone. This is most probably the same sandstone as is present between the limestone lenses of Section 5 (100–132 m). These quartz arenitic sandstones are interpreted as deposits of the destructional phase of the underlying delta system. This interpretation means that only the lower limestone lens in Section 5 (80–95 m) correlates with the second level of the Socavón Limestone Member. The second limestone lens of Section 5 (132–140 m) is therefore considered as the third level of the Socavón Limestone Member (see also the discussion of stratigraphical nomenclature).

In Section 4 the transgressive sequence overlying the Socavón Limestone Member is followed by lithic arenitic turbidites. Laterally this turbidite sequence cannot be traced very far. The small areal extension of these turbidites may be explained by interpreting these deposits as having formed in a delta front gully (van Straaten, 1959; Shepard, 1956). The turbidite sequence is overlain by deltaic deposits which in their upper part contain minable coals. These are the San Cebrián coals which are mined near the village of the same name. These coals were formed on the delta plain and are intercalated with fluviatile sandstones and fine-grained overbank and probable bay fill deposits.

These San Cebrián coals are overlain by marine sediments deposited after a transgression (Section 4: 1145 m; Section 5: 222 m). Good exposures of the actual contact between the coal measures and the overlying transgressive sediments are rare, but the map configuration west of Vergaño definitely indicates important erosion below the transgression surface (Fig. 9a). Although the disappearance of the San Cebrián coals west of Vergaño is partly due to facies changes, the coal bearing horizon as such is probably eroded and overstepped by the overlying marine strata. In Section 4 (1145–1150 m) these transgressive marine deposits have a clearly erosive contact with the underlying delta plain deposits. This exposure, however, does not prove the erosive nature of the base of the transgressive deposits, as at this locality the quartz arenitic sandstones only fill in the upper part of a 12 m deep channel. The lower part of this channel is filled with lithic arenitic sandstones and the contact between the two varieties of sandstone is sharp. This channel fill suggests that the channel was cut by a delta distributary which eroded delta plain deposits. The channel was cut shortly before the transgression occurred and the upper part of the channel was filled with marine

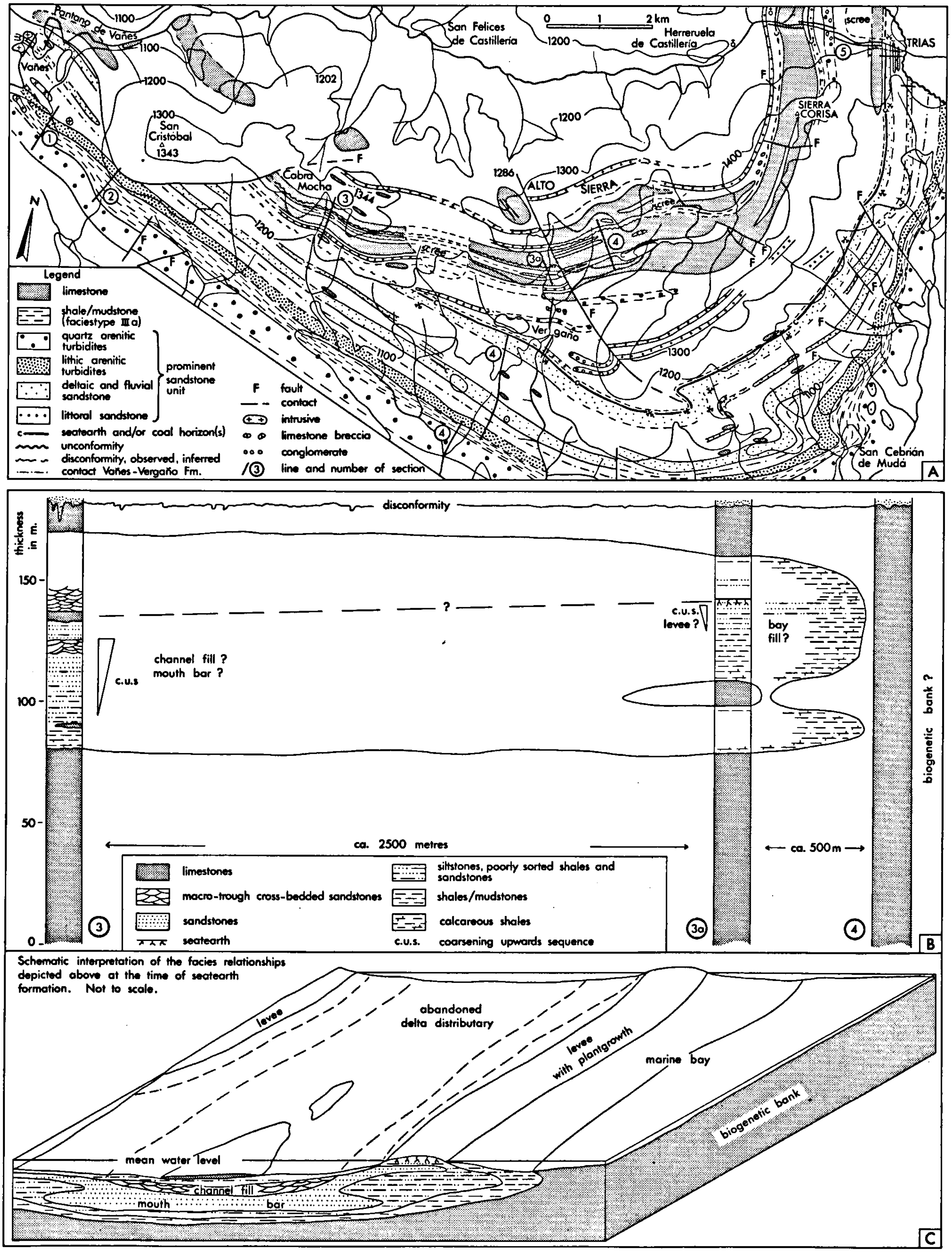

Fig. 9. A. Map of the southern flank of the Castillería Syncline. Some intervals are drawn on an exaggerated scale. B. Interfingering of the Sierra Corisa Limestone Member with siliciclastic deposits. C. Interpretation of 9B.

sandstones. Further to the west the actual contact between the deltaic sediments and the transgressive marine deposits seems to be a rather featureless surface with little erosional relief. The overstepping and onlapping relationships of the post-transgression deposits are shown on the map (Fig. 9a.).

The thickness distribution of the sediments between this transgression surface and the Leonian disconformity (Section 3: 440 m; Section 4: 700 m; Section 5: 910 m; lateral distance between Sections 3 and 5 about 5.5 km) indicates that this transgression with attendant erosion of the underlying strata is not due to a simple change in sea level or a mere cut-off of sediment supply. Tectonic tilting is considered to be the cause of the features observed. This is the more likely as for the sequence between the second level of the Socavón Limestone Member and the top of the San Cebrián coals the thinning is in the opposite direction (Section 4: 350 m; Section 5: 115 m; lateral distance between Sections 4 and 5 about 4.5 km). For the underlying part of the sequence insufficient data have been collected to ascertain whether such thinning of the strata occurs. The map configuration, however, suggests thinning from San Cebrián towards the north.

The features enumerated warrant the terms low-angle unconformity and disconformity to diastem for the transgressed surface. In the following parts this surface will be referred to as the Vergaño disconformity.

The sandstones above the Vergaño disconformity are of a quartz arenitic composition, and as the fossils indicate a marine environment this relative maturity of the sandstones is considered to be due to reworking by waves and currents. Cross-stratification and the local development of limestone (Section 5) indicate that the water in which sedimentation occurred remained shallow. A few conglomeratic sandstones are present in these shallow marine deposits (Fig. 9a, E of Vergaño). The presence of these quartz arenitic pebbles and cobbles proves the existence of a competent longshore drift system as no pebbles of this composition occur in the underlying deltaic deposits (see Chapter II, types VIb and VIIa). It is assumed that together with the pebbles this longshore drift system also supplied finer-grained siliciclastics to the basin.

Sedimentation did not keep up with subsidence and, somewhat east of the location of Section 3, a relatively deep hydrographic basin came into being. Slump and mudflow deposits together with turbidites were formed at the location of Section 4. These deposits correlate with mature and immature turbidites in Section 5. Cursory mapping north of Section 5 suggests that this interval gains in relative importance towards the north. A single sole mark reading indicates transport directions towards the north at the location of Section 4. The orientation of the axes of slump balls at the same locality suggests movements from W to E. This turbidite bearing interval is tentatively interpreted as a submarine canyon fill with associated fan valley deposits. In this interpretation the slump and mudflow deposits of Section 4 constitute the canyon fill, whereas the immature turbidites and grain flow deposits of Section 5 constitute the fan valley deposits.

After deposition of these turbidites the relative rate of sedimentation increased and a shoaling sequence was formed. Beaches formed locally in the upper part of this sequence (Section 5).

A thick limestone was then deposited at the location of Section 5. To the southwest this limestone splits up into several tongues which interfinger with siliciclastic deposits. At the location of Section 4 the lowermost tongue experienced synsedimentary displacement (Fig. 6, Plates IIc, d). Further to the west this tongue is developed as isolated limestone lenses. Between Section 4 and the top of the Sierra Corisa hill this lower tongue has a very irregular upper surface (Figs. 6, 10). The depressions in the limestone are filled with shales/mudstones (types IIIc, d and e), which at the

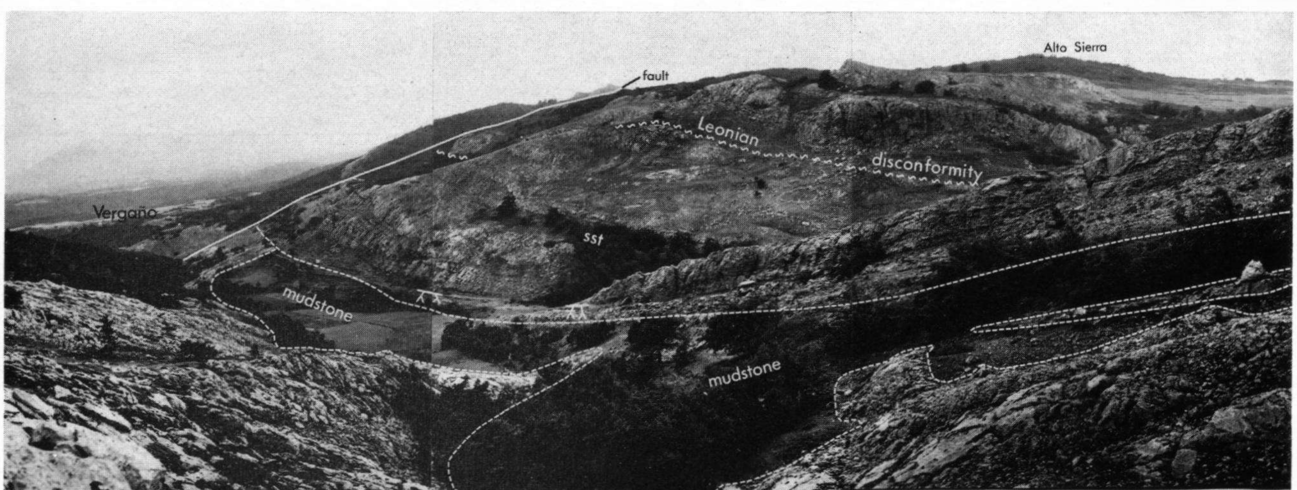

Fig. 10. Irregular upper surface of the lower tongue of the Sierra Corisa Limestone Member. Fields and wooded parts situated on siliciclastic deposits. sst-quartz arenitic sandstones; symbols as used on the enclosures 1 and 2.

actual contact often have a dusky red colour (5R 3/4).

The relation between the following tongues and the siliciclastics with which they interfinger is depicted in Figure 9. The interpretation of the coarse siliciclastics in Section 3 as a deltaic coarsening upwards sequence consisting of mouth bar and channel deposits is straightforward (Chapter III). The seatearth level in the poorly exposed Section 3a, forms the top of a c.u.s. of type Vb. In Chapter III it is argued that this type was formed as a subaerial levee or very near to such a levee. At approximately 500 m to the east of this levee deposit fully marine limestones were formed (the fault in between Section 3a, and Section 4 is considered not to have affected this distance). The fine-grained siliciclastic deposits between Section 3a, and Section 4 are interpreted as marine bay fill deposits. The relative abruptness of the limestone-siliciclastics contact and the small lateral distance between pure limestone and sand-sized siliciclastic deposits is taken to indicate that the limestones originally formed slight elevations relative to the bay fill deposits. If the limestones had been deposited at the same level as, or lower than the siliciclastic bay fill deposits, they would have been much more diluted by these siliciclastics and the zone of mixing of the two lithologies would have been much more extensive. Freeland (1969) describes recent coral reefs on the shelf near Veracruz (Mexico), which are completely surrounded by siliciclastics. From the reefs to the surrounding deeper areas with siliciclastic deposits, the percentage of $CaCO_3$ may drop from 100% to 5% within a lateral distance of 200 m. Maxwell & Swinchatt (1970) also mention the presence of coarse quartz sands within one mile of individual coral reefs of the Great Barrier Reef. In both examples the difference in topographic elevation of the two facies realms is the probable cause of the abrupt facies differentiation.

In the example under discussion only scattered patches of boundstone of type Xd are present in the otherwise mainly micritic limestone, and no trace of a reef framework, as occurs in the recent examples mentioned, is present. In Chapter III it is argued that it is unlikely that type Xd was formed in agitated waters. This contention is mainly based on the observation that oölitic packstones and grainstones are not closely associated with type Xd. For this reason I conclude that these boundstones were not originally wave resistant, nor did the constructing organisms have the ecological potential to produce a wave resistant structure. The early diagenetic compactional collapse of many examples of type Xd supports the contention that this type was not a wave resistant boundstone. Consequently the limestones under discussion cannot be considered to be true reefs. Nevertheless organisms preserved in situ, or nearly in situ, played an important part in their formation. Furthermore, the geometry of these micritic limestones as shown in Figure 9 precludes interpretation as simple clastic deposits. The general absence of calcareous debris in those deposits which formed under influence of longshore drift makes interpretation of these limestones as calciclastic deposits even less likely. For this reason I interpret these lime mudstones, wackestones, packstones with rare grainstones and boundstones as biogenetic banks, i.e. essentially in situ accumulations of organic debris from benthonic plants and animals.

Baars (1963), in his discussion of the distinctive properties of reefs, banks, bioherms and biostromes, mainly follows Lowenstam (1950) and Nelson et al. (1962) in his terminology. Baars states the following about biogenetic banks (p. 112):

"The other biogenetic carbonate deposit, which is quite the opposite of a reef deposit, is the aggregational build-up, or biogenetic bank. These accumulations of skeletal debris commonly are found in the more protected waters of the back shelf, in marked contrast to reefs, and appear to be ordinary carbonate sediments with varying amounts of lime mud and skeletal material. For this reason they usually are considered to be "poorly sorted" clastic deposits, but this is wrong, for no transport of sediment and no dynamic sorting of particles has been involved in their formation. The deposit results wholly from prolific in situ production and accumulation of skeletal material in all size-grades from mud to pebbles. The sediments normally consist of unsorted calcareous mud and skeletal particles, which have accumulated together in a low energy environment completely unaffected by vigorous wave action. All or most of the sediment is produced by organisms at the site of deposition. Much of the skeletal sediment consists of fragmental particles rather than whole shells because the scavengers and mud feeders are usually very efficient at their job of making small particles out of large ones. Textures are, for the most part, mud-supported with a fairly high amount of skeletal and pelletal particles "floating" in the fine-grained matrix. But near the margins of the accumulation, grain-supported textures become more common. This change may be due to fringing belts of specialized organisms or to the winnowing action of waves and tidal currents or to both."

Marine grasses are important agents in the entrapment and consolidation of sediment (Ginsburg & Lowenstam, 1958) and are as such very important in the construction of recent biogenetic banks in Florida Bay. Other organisms, like algae, may also be important. A few recent examples are described by Baars (1963, p. 120):

"Three prominent mounds of carbonate sediment have been constructed in situ along the inshore, well-protected margin of the shallow reef tract (Florida Keys). They are, from northeast to southwest, Rodriquez Bank, Tavernier Bank, and Snake Creek Bank (Fig. 11). All are composed of mud-size carbonate sediments with varying amounts of skeletal debris derived from organisms that live on the banks. Rodriquez and Tavernier Banks are not related to any tidal passes through the keys, so they cannot be considered deltaic in origin. Snake Creek Bank is traversed by two tidal channels which at least have altered its configuration, but the internal construction of the bank is not suggestive of current deposition. There is no apparent external source for the sediments found on the banks, and indeed, some of the organisms that have contributed skeletal debris to the banks are not found living in any other localities. It seems likely, then, that these banks were built in place by the prolific life functions of organisms; there is, however, no rigid organic frame to promote growth of the mound into the surf zone, as in the case of reefs. In fact, there is no surf although the waters are shallow, and there are no potential framebuilders – nothing but muddy sediments."

(p. 125) "Here, then, are three examples of biogenetic banks in the modern, shallow carbonate seas. They were constructed of muddy skeletal debris in the very quiet, well-protected environment of the back-shelf, and, most important, they have no rigid framebuilding components or binding agents as do the coral reefs of the fringing shelf. Yet strangely enough, these banks recently withstood the onslaught of a major hurricane better than the coral reefs (Ball, Shinn, and Stockman, 1963). Although modern biohermal banks are being constructed in situ by biologic processes, they differ markedly from coral reefs in their internal make-up, kind of constructional organisms, and geometric configuration, and most important, they differ radically in environment of deposition."

These quotations should suffice to convey the biogenetic bank concept.

The recent examples described by Baars (1963) differ from the limestones under discussion in at least one, possibly important, aspect. In the Carboniferous, marine grasses which play an important part on the recent biogenetic banks, had not yet come into existence. Algae are thought to have been the most important agents in the formation of these Carboniferous biogenetic banks (cf. de Meijer, 1969). They not only played an important part in the trapping and consolidation of the sediment, but most probably also produced most of the fine carbonate mud. Lowenstam (1955) and Stockman et al. (1967) have shown how recent algae such as *Penicillus* disintegrate upon death and produce great quantities of lime mud. Similar processes are considered to have been important in the Carboniferous.

Tavernier Bank and the other banks described by Baars (1963) stand about 2 to 3 m above the surrounding sea floor and rise to about low-tide level. Differences in elevation of the same order of magnitude are considered sufficient to explain the facies distributions depicted in Figure 9.

The shales/mudstones which fill in the depressions in the upper surface of the lower limestone tongue (Fig. 10) are very similar to those interpreted as bay fill deposits in Figures 9b, c. Because of this similarity and because of their approximate equivalence to deltaic deposits in Section 7 (950–1010 m) these depression fills are tentatively interpreted as bay fills too.

North of Section 5, the interval studied is poorly exposed and this part of the area only received a cursory examination. It seems likely that there, too, the limestone splits up into several tongues and lenses which, however, interfinger with shales/mudstones of type IIIa and turbidites. The discontinuous nature of the limestones may also partly be caused by tectonics and/or slumping.

VERDEGOSA-VERDIANA ANTICLINORIUM WEST OF THE TREMAYA FAULT (Figure 8 and Enclosures 1, 3, 5)

The sequence in this area can easily be correlated with the one exposed in the south flank of the Castillería Syncline, with the aid of only lithostratigraphic criteria. No single section covers the entire interval studied and because of the tectonic complications few data on thickness distributions in this structural unit were collected.

Section 10 begins at the approximate base of very thick, lithic arenitic turbidites which overlie a thick shale/mudstone sequence of type IIIa, with some turbidites of types Id and Ie. This shale/mudstone sequence is considered to be the equivalent of the shale/mudstone which is intercalated between the quartz arenitic and lithic arenitic turbidites of the south flank of the Castillería Syncline. This correlation implies that the quartz arenitic turbidites of the Castillería Syncline are the lateral equivalents of the Curavacas Conglomerate Beds exposed to the NE of Los Llazos and in the Casavegas Syncline. The immature, lithic arenitic turbidites form a prominent sandstone body which is an excellently mappable unit.

In Sections 7 and 10 the turbidites are followed by a c.u.s. of type VIIa or VIIb. Type VIIa and type VIIb are each other's lateral equivalent in this area. In Chapter III it is argued that the more mature composition of the sandstones of facies type VIIb is due to reworking by waves and currents. The absence of a channel fill in the upper part of this example of type VIIb provides further evidence that this c.u.s. is not a normal deltaic sequence. As lithic arenitic sands were supplied to the basin at the location of Section 7 (type VIIa), the c.u.s. of type VIIb in Section 10 is interpreted as the deposit of a prograding coast (Visher, 1965b). The coast could prograde because of a plentiful supply of sand and finer-grained siliciclastics which were eroded from nearby delta distributary deposits. In the process of erosion and transportation by marine waves and currents the unstable components were partly or entirely eliminated. Such contemporaneous development of both constructive and destructive facies of a delta is characteristic of high-destructive delta systems of which the Rhône delta is a recent example (Fisher, 1969).

Because of their position above the lithic arenitic turbidites, these coarsening upwards sequences are considered to be the correlatives of the lowermost coarsening upwards sequences in Sections 1 and 4. In Sections 7 and 10, chiefly mature sandstones which were deposited in a shallow marine to littoral environment overlie these c.u.s. deposits. Beaches (see Chapter III, type Vc), often supporting autochthonous plant growth formed locally. A number of coal seams

(Perniana coals) were formed in these strand plain deposits. The migration of the various types of beaches combined with continuing subsidence and sedimentation produced a number of minor transgressive and regressive cycles. In Section 10 a major transgression surface can be recognized at 643 m, where facies type VIIIa, overlies a sandstone with seatearth, and is overlain by limestone and shales/mudstones of type IIIa. Lithic arenitic turbidites, and a poorly exposed c.u.s. of type VIIa, follow. The top of the c.u.s. is not exposed but a major transgression must have occurred after its formation, as a thin marine limestone followed by clean sandstones with a seatearth and a thick sequence of shales and mudstones overlie this deltaic deposit. The limestone is interpreted as a lagoonal deposit, the overlying sandstones as the deposits of a migrating barrier beach the migration of which did not produce a strongly erosive base to the deposit. The actual transgression surface below the limestone is not exposed, only the secondary transgression surface on the seaward side of the barrier is well exposed and indicated in the section.

In Section 7 no lithic arenitic turbidites or deltaic sediments were observed at this level. Only marine sandstones and siltstones of a relatively mature composition were found. In other words, the constructive facies of the delta in Section 10 had its destructive equivalent at the location of Section 7. This further corroborates the interpretation of this delta system as a high-destructive one.

The transgression surface at 808 m of Section 10 can be traced laterally into transgressive deposits of type VIIIa in Section 7 at 650 m. This transgression surface is followed above by a thick shale/mudstone sequence which contains turbidites and calcareous mudflow deposits. This sequence establishes the correlation of the transgression surface discussed above with the Vergaño disconformity. This also establishes the correlation between the deltaic deposits of the constructive facies below the Vergaño disconformity in Sections 1, 4, 5 and the deltaic and shallow marine deposits of the destructive facies in Sections 7 and 10. This correlation renders the conclusion that we are dealing with the deposits of a high-destructive delta system inescapable.

From the presence of turbidites in the upper part of Section 10 inferences can be made concerning the depth of deposition of these sandstones. The turbidites are overlain by a coarsening upwards sequence of deltaic origin. These deltaic deposits are predominantly sandy and are overlain by lagoonal and barrier beach deposits. The seatearth in the barrier beach deposit provides a datum with regard to water depth. Taking into account compaction of the various sediments, the depth of deposition of the turbidites can be calculated as the formation of a deltaic c.u.s. is probably too fast a process to permit of important subsidence or uplift of the basin floor. Above the uppermost turbidites in Section 10 (710–720 m), no more than about 10 m of shale/mudstone are present. As consolidated clays have an average porosity of 50% (Pettijohn, 1957), these 10 m may originally have been some 20 to 30 m of sediment. Sandstones normally show little compaction and pore volume is usually reduced from 35–40% to 10–20% by cementation (Pettijohn, 1957). In thin sections of this interval compaction effects are plentiful. Suppose this compaction to have resulted in a loss of volume of 15%. This would mean an original thickness of about 90 m for the 78 m of sandstone preserved. The overlying 10 m of limestone and sandstone were deposited after some subsidence had already occurred, but this will be neglected as the seatearth is in the upper part of this sequence. This means that the turbidites were deposited at a depth not exceeding 125 m and possibly as little as 100 m.

West of Section 10 a hill named Peña Tremaya (Fig. 8, Enclosure 3), which consists of thick limestones, constitutes a tectonic and stratigraphic problem. Nederlof (1959) and de Sitter & Boschma (1966) considered this limestone hill as an anticlinal core. Although dip and strike readings give some support for this interpretation, it was easily disproved as top and bottom criteria in sandstones exposed NE and SW of the limestone indicated that both sides younged towards the SW. The aberrant dip and strike readings can therefore only be explained by interpreting an important fault in the limestone. A strongly recrystallized zone with slickensides was actually found in the middle of the limestone, and this zone connects the two wedges of siliciclastic material which point to the middle of the limestone hill. This doubling of the limestone is confirmed by the presence of fusulinids of the *Fusulinella* B_1 zone (van Ginkel, pers. comm.) in the upper parts of the two limestone bodies, and by the finding of a Westphalian D flora (Wagner, pers. comm.) in the sandstones SW of Peña Tremaya. Such flora indicates that this sandstone is the same as the sandstones in Section 10 (600–645 m) and not the continuation of the sandstones connected with the Aurora coal seams of Cantabrian age. This sandstone sequence is overlain by Cantabrian strata, as stated by Wagner and Varker (1970), who interpret the contact as a sedimentary one. However, I interpret another fault to the west of the Westphalian D sandstones, which is probably a direct continuation of the Los Llazos fault. This structural interpretation accounts for part of the Peña Tremaya structure. In this interpretation, however, the northeastern part of Peña Tremaya is in its original position with respect to the succession of Section 10. Mapping in this poorly exposed area indicated that this northeastern limestone block is the lateral equivalent of the interval approximately between 450 and 650 m of Section 10. No interfingering relationships between limestone and siliciclastics were observed. In view of the limestone facies types and the small distance between pure limestone and siliciclastics these limestones are interpreted as biogenetic bank deposits which at present form a bioherm.

Above the transgression surface, which is the equivalent of the Vergaño disconformity, the presence of a thick shale/mudstone sequence containing quartz arenitic turbidites and calcareous mudflow deposits has been mentioned. The turbidites disappear to the NW of Section 7.

If is accepted that in analogy with most recent turbidites, the turbidites in Section 7 (760–880 m) were deposited at the foot of the shelf slope, the dark shales/mudstones of type IIIa, in which they are intercalated, may be interpreted as shelf slope deposits.

This shale/mudstone interval with turbidites, correlates with the turbidite bearing sequences of Sections 4 and 5. The slump and mudflow deposits together with the turbidites of Sections 4 and 5 were interpreted as submarine canyon and fan valley deposits. The turbidites of Section 7 are in this interpretation considered as deposits formed on the fan or in the fan valley. Towards the northwest, i.e. towards Section 10, the hydrographic basin was probably shallower than at the location of Section 7 and this is considered to be the reason for the absence of turbidites in this part.

This interval is followed above by coarsening upwards sequences of type VIIa, which are coal bearing (Por si Acaso coals). The coal bearing parts are interpreted as delta plain deposits. Limestones of the same facies as those in the upper parts of Sections 3, 4, and 5 are intercalated with these deltaic deposits and the associated shallow marine deposits (Sections 6 and 7). They are also interpreted as biogenetic banks and they are the lateral equivalents and continuation of the biogenetic bank deposits in the south flank of the Castillería Syncline.

The destructive facies equivalents of these deposits belonging to the constructive facies of the delta system are distinctly developed north of the location of Section 6, and the beach deposits at 770 m in Section 5 probably also belong to this delta.

The most remarkable feature of this delta system is the close association and the interfingering relationship with fully marine limestones which formed as biogenetic banks. Although the older delta systems of this part of the area were also associated with biogenetic banks, no such interfingering was observed.

VERDEGOSA-VERDIANA ANTICLINORIUM BETWEEN THE TREMAYA AND REDONDO FAULTS (Figure 8 and Enclosures 1, 3, 5)

Both Nederlof (1959) and de Sitter & Boschma (1966, Fig. 10) presented too simplified an interpretation of the structures of this part of the area. Top and bottom criteria (cross-bedding; seatearth; corals and algae in position of growth) prove that the limestones forming the La Frechilla and Verdiana hills do not form the eastern flank of a simple syncline. Instead, the limestones forming these two hills belong to the opposite sides of a syncline which have shifted in such a way as to constitute each others continuation (Enclosure 3). This new structural interpretation has of course an important effect on any attempt at an palinspastic reconstruction of the area (Chapter V).

Only in the Celada Syncline could a fairly complete section of the interval studied be measured. Lithostratigraphic correlations prove that in the remainder of the sub-area only the lower part of the interval studied is present. Although corresponding to the lower 300 m of Section 8, the sequence exposed may well be thicker. North of San Juan de Redondo even lower parts of the interval studied than those exposed in Section 8 are cropping out. Near the Redondo fault a number of more or less isolated limestones crop out in the woods (Peña del Moro limestones of Nederlof, 1959, p. 627). Their surroundings are very poorly exposed, but their occurrence as isolated lenses and blocks suggests that they may be slide deposits. The map configuration indicates that these limestones may be the lateral equivalent of the shale/mudstone interval below the lithic arenitic turbidites.

In Section 8 lithic arenitic turbidites are succeeded by cross-bedded sandstones which are also of lithic arenitic composition. These are interpreted as deltaic deposits. This part of the sequence correlates with the lowermost coarsening upwards sequences in Sections 1, 4 and 7. The overlying limestone consists of the same facies types as the limestones which form the upper part of Sections 3, 4 and 5 and is therefore also interpreted as a biogenetic bank deposit. Its position above lithic arenitic turbidites and a deltaic c.u.s. indicates that it is the lateral equivalent of the Socavón Limestone Member. In its upper part it contains a few cross-bedded, calcareous sandstones which permit of correlation with the limestone of Section 9. This sequence is overlain by slump and mudflow deposits. Limestone blocks of over 20 metres in diameter occur in this interval. The presence of these deposits indicates instability of sedimentary slopes and, also, since they are overlain by shales/mudstones with turbidites, important deepening of the basin. This deepening sequence is considered to correlate with the transgressive sequence which overlies the Vergaño disconformity in the Castillería Syncline. In other words the instability of the sedimentary slopes at this location, which resulted in slump movements, is related to tectonic tilting in an area to the southwest. Above an interval with distinct turbidites, which correlates with the turbidites in Sections 4 (1340–1460 m), 5 (530–650 m) and 7 (750–875 m), follows a poorly exposed sequence consisting mainly of shallow marine deposits. These are interpreted as the lateral equivalent of the shallow marine and beach deposits in Section 5 (650–780 m). The overlying limestones form the most impressive feature of the Celada Syncline. Because of the constituent facies types these limestones are interpreted as biogenetic bank deposits correlating with the upper parts of Sections 3, 4, 5 and 6.

CASAVEGAS SYNCLINE (Figure 8 and Enclosures 1, 3, 5)

Correlations based only on lithostratigraphic criteria are difficult to make between the Casavegas Syncline and the Castillería Syncline or the Verdegosa-Verdiana Anticlinorium. Only for the lower part of the sequence can they be used with some confidence. The shale/mudstone interval underlying lithic arenitic turbidites can be traced from Tremaya (Section 10) to the road-cut north of Camasobres (Section 13). In this road-cut a few very mature turbidites are visible in the shale/mudstone. Further to the west this shale/mudstone interval, though poorly exposed, can be traced as far as the location of Section 11. At this point it has become very thin. The general environment of deposition in which this shale/mudstone interval was formed can be inferred from the sequence as exposed between Sections 10 and 13. Here the shales/mudstones overlie the Curavacas Conglomerate Beds. These conglomerates interfinger with lenticular limestones lying in situ. As the limestones contain algae, this indicates a shallow water origin for the conglomerates. The shales/mudstones are overlain by lithic arenitic turbidites. This sequence indicates that the shale/mudstone interval is a shelf slope deposit as it is enclosed by shallow water and relatively deep water deposits.

In Section 13 no distinct lithic arenitic turbidites were observed. The shale/mudstone interval is immediately overlain by a c.u.s. which contains a few slumped limestone blocks at the base. The composition of the sandstones of this c.u.s. indicates that the lower part has to be included in type VIIa, whereas the upper part belongs to type VIIb. This upper part is very much distorted by tectonics. Laterally this c.u.s. cannot be traced far.

Both east and west of the location of Section 13 the succeeding limestone unit seems to rest immediately upon the shale/mudstone interval. (Section 13 is not very representative of the situation as the thick limestone units wedge out on either sides of the valley in which the section was measured). The limestones are lenticular and the maximum thickness of the lenses decreases to the west (Enclosure 3). Their facies indicates that they are biogenetic bank deposits.

These limestone lenses are overlain by shales/mudstones which contain turbidites. In Section 13 these are lithic arenitic turbidites (270–300 m), whereas in Section 11 quartz arenitic turbidites occur. Their presence indicates that considerable deepening occurred after deposition of the limestones. In Section 12 turbidites occur at two levels between shallow marine deposits. This interbedding indicates that thin turbidite sequences may have originated in comparatively shallow water.

The overlying deltaic c.u.s. in Section 13 has a small lateral extension. Half way in between Sections 12 and 13 no trace was found of lithic arenitic sandstones of deltaic origin, and further west only compositionally more mature sandstones of marine origin are exposed. This is one more proof in favour of the interpretation of these deltas as high-destructive delta systems. The occasional presence of conglomeratic sandstones in the western part of the Casavegas Syncline is in analogy with the conglomeratic sandstones of the Castillería Syncline, interpreted as proof of the existence of competent longshore drift systems. In these areas affected by longshore drift, tidal currents were locally important. This contention is based upon the presence of deposits of type VIIb (Section 12, 250–275 m) which are interpreted as tidal delta deposits in Chapter III. In Section 11 a discontinuous limestone is present at approximately this level. This limestone, too, is interpreted as a biogenetic bank deposit.

This is followed by a sequence of bioturbated mudstones and siltstones with locally better sorted sandstones which are interpreted in Chapter III as bar deposits. This sequence, which is cyclic in Sections 12 and 13, was formed in a shallow, open marine environment. Periodically low rates of deposition enabled burrowing organisms to rework and homogenize the sediment. This does not necessarily mean that the sand bars were deposited much faster, as the presence of burrowing organisms is largely dependent on the characteristics of the substrate (Purdy, 1964). As these bars formed in at least slightly agitated water, which is inferred from their comparatively good sorting, the commonly clay and silt sized food particles were also winnowed away. The result of this winnowing is that the sediments deposited are unattractive to mud ingesting organisms, i.e. the sediment will be relatively undisturbed by such organisms.

This sequence is overlain by locally very thick limestones (Enclosure 3) which are thin to absent between Sections 12 and 13. A number of isolated limestone blocks which are overlain by a few calcareous turbidites indicate that this wedging out is probably due to deeper water in this zone. These limestones, too, are considered to be biogenetic bank deposits because of the constituent facies types and also on account of the rapid lateral transitions into siliciclastic deposits.

A minor increase in thickness of the interval studied seems to occur from east to west. This trend is parallelled by a transition from a sequence containing fluviatile (?) deltaic deposits into a sequence containing only marine deposits.

REDONDO SYNCLINE (Figures 8, 11, 12 and Enclosures 1, 2, 3, 4, 5)

The lower part of the interval studied is very poorly exposed, largely covered as it is with scree of the unconformable Permo-Triassic or with the Permo-Triassic itself. The oldest part of the sequence is relatively well exposed on the sides of the main road from Cervera de Pisuerga to Potes about 1 km south of the village of Piedrasluengas (due west of Section 16). Mature and immature quartz arenitic turbidites, which are quite intensively folded, are exposed there. This turbidite sequence is overlain by a poorly exposed shale/mudstone interval, which in turn is overlain by a limestone unit of variable thickness. This sequence of relatively deep water sediments (turbidites) overlain by shale/mudstone and shallow water limestones (algae) leads once more to the interpretation of the shale/mudstone interval as a shelf slope deposit.

If we trace this part of the sequence from the SE (Section 27) to the NW and around the synclinal nose up to Section 14 it is possible, despite the poor exposure to observe important facies changes. At the base of Section 27 a few metres of shale/mudstone are exposed and the contact with the limestones seems undisturbed. From here up to the location of Section 23 the limestone itself is continuous as is indicated by the presence of sinkholes in the scree-covered parts. At the location of Section 23 the limestone overlies a distorted sequence of type IIa. The contact itself is undisturbed. For Section 22 the same description holds good. About 1.5 km NNW of Section 22 better exposures are avaiable of the sequence below the limestone. About 100 to 150 m below the base of the limestone, slump and mudflow deposits are present (types IIa, IIe and IIf) together with mature and immature turbidites. Upwards more mature turbidites are more frequent and although the siliciclastics are somewhat distorted the limestone conformably overlies the shales with sandstones. East of Sections 20 and 21 the sequence discussed is again poorly exposed but the few outcrops permit of recognition of immature and mature turbidites and slump deposits. The siliciclastics-limestone contact is not exposed in this part of the area, but as sinkholes are generally absent in the scree-covered parts it is likely that the limestone outcrops in this area (Sections 20 and 21) are in fact isolated blocks. However, only in the Pico Guillermo area is definite proof available that the limestone outcrops actually represent isolated limestone blocks (Fig. 7). These limestone blocks "float" on and in slump and mudflow deposits. Their distribution can be likened to that of currants in a pudding. From the facies relationships depicted in Figure 7, especially the extremely abrupt contact between the limestone blocks and the siliciclastics and the unconformity between the main limestone block and the overlying sequence with mature turbidites, I conclude that the limestone blocks are allochthonous too (cf. Chapter III). This means that from Section 27 to Sections 18 and 19 we trace a limestone unit which lies in situ in the south to a zone where the same limestone is broken up and displaced by sliding movements. An intermediate zone where minor displacement without breaking of the limestone occurred can also be distinguished.

Further west it is difficult to decide as to whether the limestone is in place or displaced on account of the more complicated tectonics. Section 17 was measured in a limestone block which appears to be in place, while at the location of Section 16 the limestone certainly is in place. Further to the south, at the location of Section 15, the limestone experienced some displacement as is testified by its eroded base. A few hundred metres to the south where Section 14 was measured the limestone is again broken into several blocks.

The occurrence of sliding or slumping is indicative of unstable slopes. The association of such deposits with pebbly mudstones and immature turbidites and/or grain flow deposits suggests deposition in a submarine canyon or in the proximal parts of a fan valley. The presence of such deposits below unbroken limestones indicates that conditions favouring such a depositional environment already existed before the main slumping occurred. The limestones, however, were formed in relatively shallow waters on the shelf. They were already lithified when displaced by sliding movements, as no soft sediment deformation was observed in these allochthonous blocks. Only breaking occurred, resulting in smaller or larger fragments. As the displaced and broken limestone unit was at least 200 metres thick before the process began (Section 18), considerable instability of the slope has to be assumed in order to set such a limestone in motion. This instability may either have been caused by tectonic movements, sedimentary overloading, backward erosional valley cutting or by a combination of these factors. Considering the magnitude of the resedimentation phenomena, tectonic movements seem to be a likely cause. That such movements indeed played a part will be argued in the following chapter as this can only be deduced from regional relationships.

On the basis of the preceding inferences the features observed are interpreted as follows. In an area with predominantly limestone deposition in a shelf environment, an active or abandoned submarine canyon (or upper reaches of a fan valley) formed a deeper zone. Due to tectonic movements the slopes of the canyon became unstable and the siliciclastics which were not yet lithified began to slide and flow. Because of this slumping and flowing of the slope material the overlying, already lithified limestone unit was losing its supports. When sufficient material had flowed and slid into the deeper parts of the depression (canyon or fan valley), the lithified limestone cover broke up and was carried along on the moving substrate. Once started, such a process will probably be kept going for some time by the sheer weight of the overlying limestone which presses away the unconsolidated siliciclastics. Whether the process could get started merely as a result of such overloading is to be doubted because

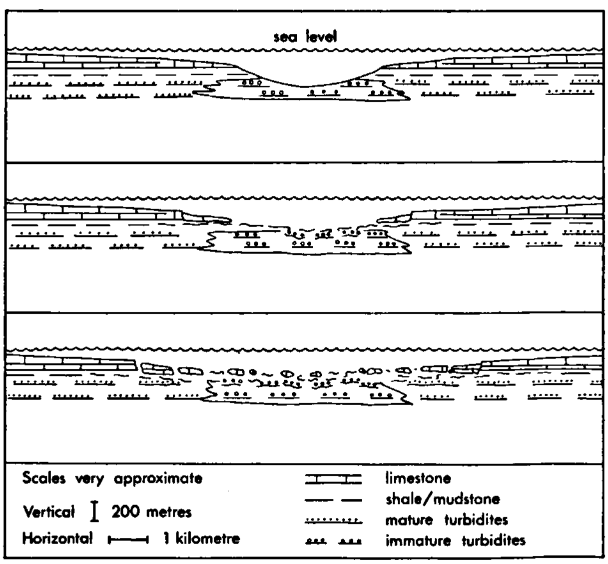

Fig. 11. Inferred mechanism of resedimentation of the Agujas Limestone Member.

of the magnitude of the phenomena described. The process invoked may be likened to the collapse of a dike due to sliding after it has been eroded at its base by currents, which is a dreaded phenomenon in Holland and Zeeland.

The time required for completion of the process described is difficult to estimate. From the description by Yamasaki (1926) of the effects of the 1923 earthquake in Japan, it can be gathered that such resedimentation can take place in a couple of hours. On page 99 he states:

"One of the most remarkable local disturbances is that which occurred in the Uraga furrow, an eastern branch of the main trench. The northern flank of the submarine range which extends to the west of Cape Suno-Saki, embracing the mouth of the Uraga Channel, experienced a great slip 10 km long, similar to a mountain slip on the land surface, filling the floor of the furrow with debris 230 m in thickness, or in other words, decreasing the depth of the furrow by just that much."

Whether the deposits discussed were indeed sedimented so rapidly is difficult to establish and it is a distinct possibility that the process of resedimentation took considerable time. Although the tectonic movements invoked caused local infilling of depressions and thus shallowing of the basin at that location, the overall effect was probably the opposite as the limestones are overlain by a turbidite sequence.

As the siliciclastic sediments do not contain any macrofossils except plant debris, the dating of the sequences is only based on the floras and faunas contained in the limestones. Despite this lack of definite data it is considered unlikely that great differences in age between the resedimented strata and the surrounding sequence exist, as they are lithologically very similar. This means that these deposits probably constitute a borderline case between slumps and olistostromes. According to Görler & Reutter (1968), slumps remain in their original stratigraphic horizon, while olistostromes contain fragments (olistoliths) considerably older than the underlying and overlying sequences.

The distribution of the various facies types in the limestone unit which was involved in the slumping will now be dealt with. The gross facies relationships are depicted in Enclosure 4. Although Section 18 was measured in the allochthonous block forming the Pico Guillermo, it is included in this diagram as this block is believed not to have been displaced over great distances (a few kilometres at the most). Moreover, its constituent facies types can be perfectly matched with, for instance, Section 22. Nevertheless, it is wiser to base the interpretations solely on the sections measured in limestones which are still in situ of nearly in situ. In the correlation diagram the top of the limestone has arbitrarily been taken as a datum level for want of a better marker. Identical facies types were correlated between the various sections. A few units were actually traced in the field but most correlation lines were drawn afterwards. The presence of quartz arenitic pebbles in Sections 18, 21, 22 and 24 constitutes something like a marker horizon. In between Sections 21 and 22 it provides a correlation which indicates that facies type Xm is the lateral equivalent of types Xa, b, c and d. In Chapter III type Xm is interpreted as having formed in quiet, probably relatively deep water, where only sluggish water movements occurred. Deposits of types Xb, c and d, occurring in the south flank of the Castillería Syncline are interpreted as biogenetic bank deposits, which biogenetic banks were slightly elevated above their surroundings. The facies relationships, as observed in the lower parts of Sections 21 and 22, fit very well into such a model. In this interpretation facies type Xm was formed in a deeper area in between slightly more elevated biogenetic banks (facies types Xb, c and d). Such a depression may be likened to the lagoons of present day coral reefs, and therefore the term "lagoon" is used in Figure 12. For facies type Xl (Section 25) the interpretation is identical with the minor difference that the sedimentary interface was not completely barren of life. The differences in elevation between the banks and the "lagoons" were very slight as is indicated by the lower part of Enclosure 4 where the limestone is depicted true to scale.

The distribution of the quartz wackes of types IVe is also easier to understand in this interpretation. As they were deposited in pre-existing depressions their lateral extension is determined by the size of these depressions How these quartz wackes were emplaced still remains a mystery.

As the limestone under discussion is both underlain and overlain by thick turbidite sequences, it evidently forms the shallowest part of the entire sequence. Its setting immediately above slope deposits without any associated coarsening upwards sequences or coarse-grained shallow marine deposits, suggests that it formed on an open shelf. Being the shallowest part of a sequence, both the culmination of a shoaling sequence and the beginning of a deepening sequence are likely to be present and recognizable in the limestone. It is on account of this inference that the Xa→Xb, c and d→Xh sequence (well developed in Sections 18, 23, 27 and less distinctly in Sections 22 and 24) is interpreted not only as representing an overall increase in wave energy but also as a shoaling sequence. The inferred development of the limestone unit under discussion is as follows. On a shallow shelf supply of siliciclastics at a certain moment becomes sufficiently low to allow of deposition of limestone (type Xa) at some points. On these first limestone deposits prolific organic (mainly algal) activity built biogenetic banks which rose slowly to the wave base. Once the wave base was reached packstones and grainstones were deposited instead of mudstones and wackestones and in even more agitated waters oolites formed. On the shallowest and quieter parts of these shoals which "capped" the real biogenetic banks, the "mottled" pelsparitic grainstones were formed. Sedimentation on these shoals was probably so slow that they "drowned" because of continuous subsidence and after deposition of a trans-

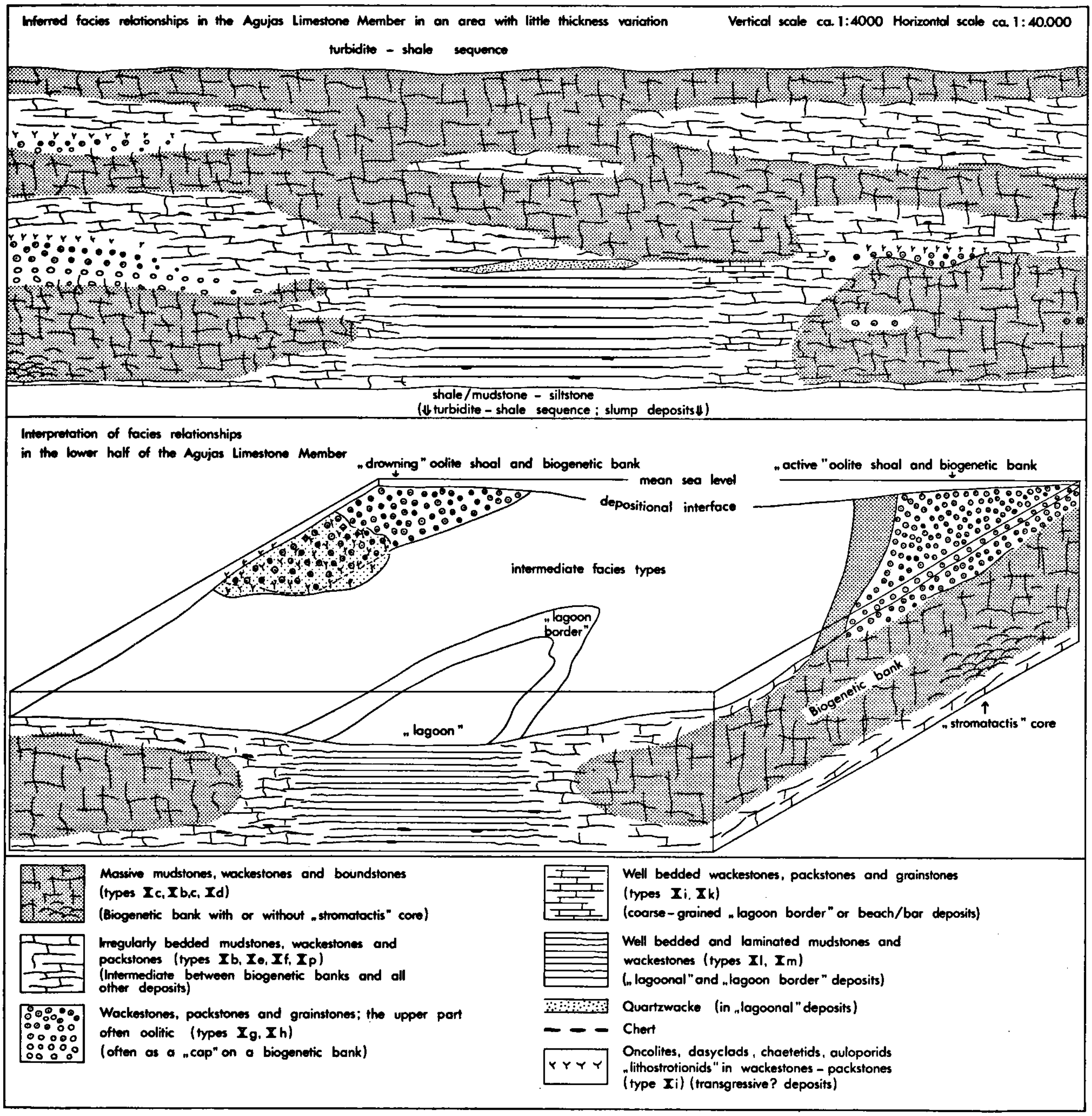

Fig. 12. Facies relationships in the Agujas Limestone Member. The upper part of the figure is a hypothetical cross-section. The terms turbidite-shale, slump deposits, etc. denote underlying and overlying deposits. The lower part of the figure is an interpretation of facies relationships during deposition of the middle part of the Agujas Limestone Member.

gressive deposit (type Xi) another biogenetic bank sequence could develop.

Incidentally, this aggradational sequence which does not contain any obvious debris of the underlying limestones is a further proof that these deposits were not reefs. Recent reefs are commonly associated with great quantities of debris produced by waves, currents and all kinds of scavenging and boring organisms, and redistributed by waves and currents. No trace of such debris zones was observed in the limestones of the area studied.

Few of the second generation biogenetic banks developed an oolite capping (Section 24) and higher in the sequence the facies distribution is much more monotonous than in the lower parts. No "lagoonal" deposits are present in between the biogenetic banks and oolite shoals are rare and poorly developed.

That the invoked "drowning" of the shoals was most probably due to local causes and not to regional changes in sea-level is deduced from the somewhat haphazard distribution of the various occurrences of facies types Xg and Xh. A regional change in sea-level, however, would be very difficult to distinguish from the local effects and its occurrence remains a distinct possibility. Of course there has been an overall change in sea-level as the limestone is overlain by turbidites but this major change is thought to be responsible for the monotonous aspect of the upper part, not for the cyclic sequences in the middle part of the limestone unit.

In the greater part of the Redondo Syncline the limestone unit is overlain by turbidites. Only in the NW is a shale/mudstone sequence intercalated between the two. This shale/mudstone is again interpreted as a shelf slope deposit. The fact that the turbidites directly overlie the limestones is taken to indicate that considerable deepening occurred in a short time. The deepening was probably related to the slumping movements discussed previously. The thick turbidite sequence is considered to be a submarine fan deposit because of its lateral and vertical extension. In the southern half of the Redondo Syncline immature turbidites occur in the upper part of the turbidite sequence (Section 25). These are interpreted as proximal fan valley deposits which formed at the foot of the shelf slope. The overlying shales/mudstones are interpreted as shelf-slope deposits intercalated as they are between turbidites and shallow marine limestones. In these shelf-slope deposits lenticular to channel-like calcareous mudflow deposits occur (type IId).

The limestones which form the upper part of the interval studied are strongly lenticular (Enclosure 3) and their facies warrants the interpretation as biogenetic banks. These banks also formed on the open shelf. In Section 21 these limestones show indications of slumping and resedimentation by grain flow and/or immature turbidity currents.

From the map (Enclosure 3) and the sections it is obvious that in the Redondo Syncline the interval studied thickens very considerably from NW to SE (Section 14: approx. 400 m; Section 25: approx. 1050 m).

CHAPTER V

REGIONAL INTERPRETATION OF THE OBSERVED FACIES DISTRIBUTIONS, BASIN FILL HISTORY AND COMPARISONS WITH OTHER DELTA SYSTEMS

The preceding chapter dealt with the interpretation of the sedimentary facies within the bounds of the main tectonic units. We may now attempt a regional interpretation of the observed facies distributions. Their present-day distribution is shown in Enclosure 3. As stated in the introduction such a regional interpretation would not be possible without the detailed correlations presented by van Ginkel (1965) on the basis of fusulinid studies. His data combined with those of Rácz (1966) and myself, permitted the following correlation chart of the various limestones to be drawn (Table I). The locations of these limestone members are shown in Figure 3.

From this chart it can be deduced that the main limestone levels in the various structural units represent the same intervals. With this knowledge a correlation diagram can be made of the sections measured. Though the correlations as such are not influenced by tectonic displacements, such displacements have to be taken into account when attempting to reconstruct the paleogeography. From de Sitter & Boschma's map and sections (1966) and from my own map (Enclosure 3) it is evident that considerable shortening of the strata has occurred. For an accurate determination of the amount of shortening a much more detailed map and a tectonic analysis of the area would be required, but this is beyond the scope of the present study. Nevertheless, a reasonable estimate based on the available data can be made. Although the displacement along the faults in the central part of the area is difficult to estimate, a shortening in the order of 30 to 45% of the original length seems likely. In this estimate the internal deformation of the sediments is neglected, as slaty cleavage is very rare to absent in the area studied. The orientation of the trend of greatest shortening is SW–NE (cf. de Sitter & Boschma, 1966, Fig. 24). It is obvious that if these tectonic displacements are not taken into account the resulting paleogeographic picture will be in error.

Apart from the shortening, another factor further complicates the construction of a fence diagram. Although most sections were measured perpendicularly to strike and dip of the strata, this does not mean that an originally vertical sequence was measured. Sections are customarily depicted as if they represent a truly vertical sequence. In most cases when a relatively uniform interval is studied this does not lead to great errors. But we are dealing with a relatively thick sequence (thick in relation to the size of the folds), which furthermore shows rapid lateral variations. This means that it is no longer permissible to depict the sections vertically (Fig. 13). For this reason all the lines of sections were rotated around the fold axes to determine their original orientation. The fold axes around which rotation took place were first deter-

		fusulinid zones	algal zones		South flank Castillería Syncline	Verdegosa-Verdiana Anticlinorium	Casavegas Syncline	Redondo Syncline
Upper Moscovian	?Cantabrian	B3	V					
Upper Moscovian	Westphalian D	Fusulinella B2	IVb	Vergaño and Covarres Formations	Sierra Corisa L.M.	Verdegosa L.M., Sosa L.M.(?), Villar L.M.	Maldrigo L.M.	Abismo L.M. Tejedo L.M.
Upper Moscovian	Westphalian D	Fusulinella B2	IV	Vergaño and Covarres Formations	Coterarraso L.M.			
Upper Moscovian	Westphalian D	Fusulinella B1	IVa	Vergaño and Covarres Formations	Socavón L.M.	Celada L.M., Verdiana L.M., Matacastillo L.M., Tremaya L.M.	Castrillo L.M.(?) Camasobres L.M.	Agujas L.M.
				-?-				
L. Moscovian	Westphalian C ?-?-	A	III	Vañes Formation				

Table I. Correlation chart of the various limestone units. Mainly after data presented by van Ginkel (1965) and Rácz (1966). Boundaries of Westphalian C, Westphalian D and Cantabrian after Wagner & Varker (1970).

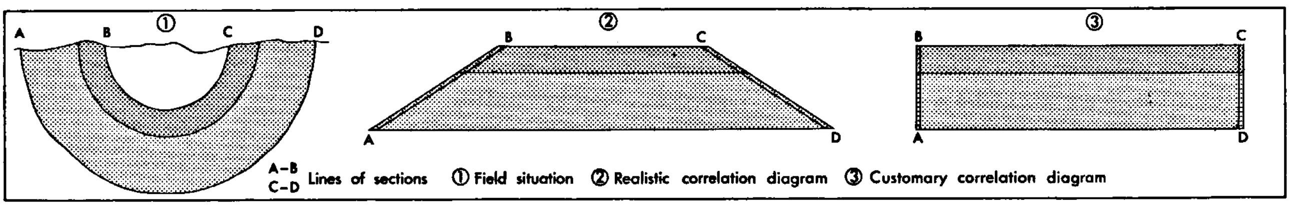

Fig. 13. Influence of folding on the orientation of a stratigraphic section.

mined by construction on a stereographic net. Because of the limited data available, the accuracy of these determinations is probably 10° to 15° for both strike and dip. The orientations of these fold axes and the areas of which they are representative are shown in Figure 8. If after rotation the sections dipped 70° or more, they were depicted vertically. Some of the shorter sections were also drawn vertically because in the projection used it would have been impossible to depict them clearly in their actual orientation. Although probably inaccurate in many points the resulting correlation diagram as shown in Enclosure 5 is thought to be realistic in its proportions.

QUALITATIVE FACIES MAPS

With the palinspastic correlation diagram as a base, a number of qualitative facies maps were drawn (Figs. 14a, b, c, d, e, f) as the accuracy of the paleontological correlations does not permit of the construction of paleographic maps. Nevertheless, Figures 14c, d and e are probably fair approximations of such maps. The distribution of the various facies groups as depicted on these maps, is primarily based on outcrop distributions, and secondly on the inferred continuation of these outcrops. When inferring the continuation of the outcrops the character of the facies group, the directions of supply and the general paleogeographic picture were taken into account. Inclusion of such data introduces a subjective element in the maps. A second, probably more important subjective element is the allocation of the siliciclastic sediments to one subzone or another. As the only dated horizons are the limestone units, the allocation of a certain interval of siliciclastics to a subzone is largely fortuitous. A third subjective element in these maps is the subdivision of the

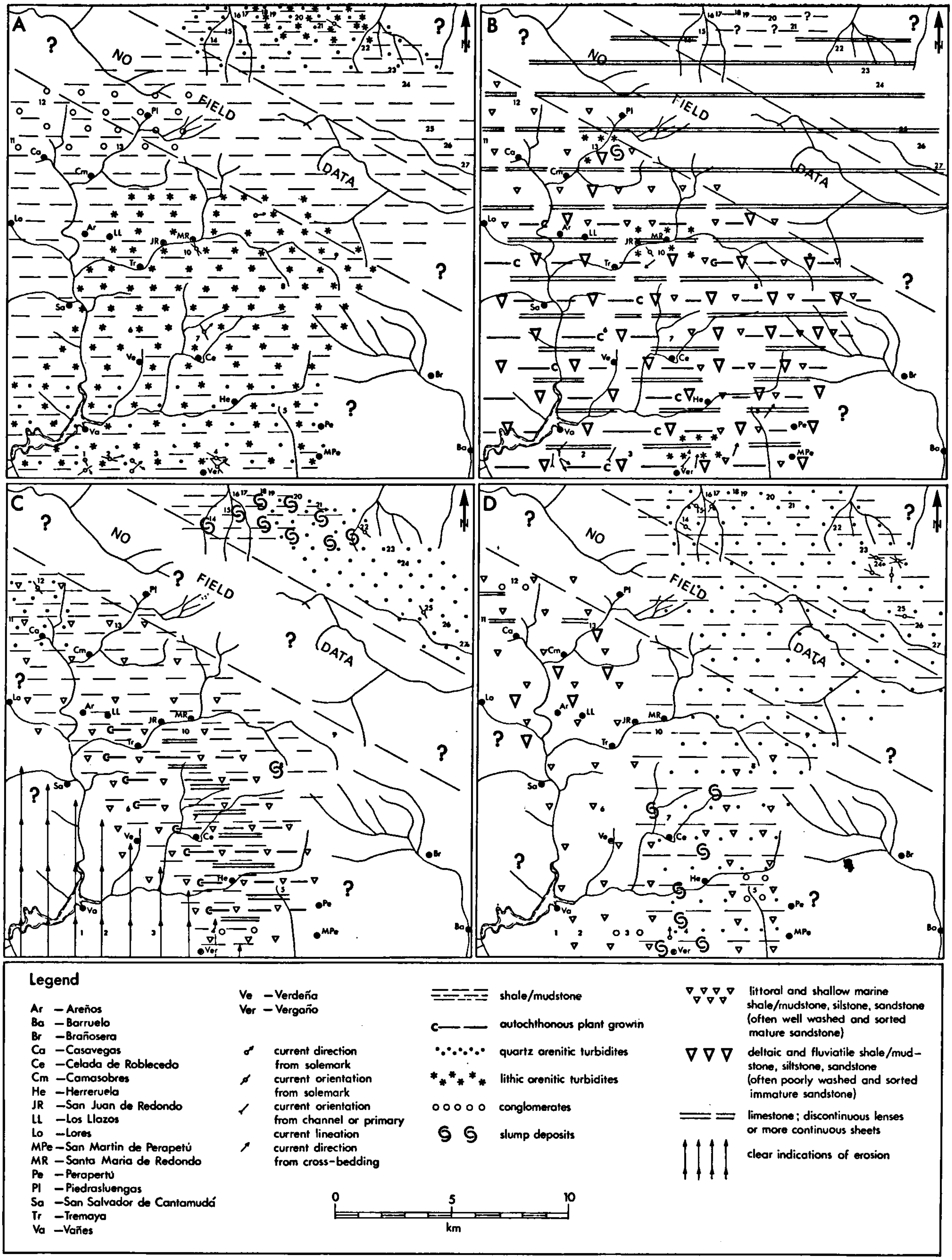

Fig. 14. Qualitative facies maps of the interval studied. A. Top *Profusulinella* B, *Fusulinella* A, base *Fusulinella* B_1 (sub)zones. B. *Fusulinella* B_1 subzone. C. Top *Fusulinella* B_1, base *Fusulinella* B_2 subzones. D. Base *Fusulinella* B_2, middle *Fusulinella* B_2, subzone. For remainder of Fig. 14 see page 203.

subzones. As van Ginkel (pers. comm.) is not able to make such subdivisions, these are based on lithostratigraphic correlations and inferences as to which facies groups are conceivably each other's time equivalent. Notwithstanding these subjective elements, Figures 14a to f give a fair impression of the overall facies distributions. The various map configurations will now be explained and interpreted.

Figure 14a. – This map possibly represents the upper part of the *Profusulinella* B zone, all of the *Fusulinella* A zone and probably the lowermost part of the *Fusulinella* B_1 subzone. No paleontological control is available for the southern part of the map. During the deposition in the northwest of shallow water conglomerates, which are considered to be tongues of the main body of the Curavacas conglomerates to the west of the area studied, quartz arenitic turbidites were deposited in the northeastern and in the southern part of the map. After this episode supply of coarse clastics to the area was halted. A shale/mudstone sequence was deposited all over the area. In some of the structural units this shale/mudstone sequence can be shown to be a probable shelf slope deposit (Chapter IV). In the southern half of the map this shale/mudstone interval is covered by lithic arenitic turbidites which are also present in the extreme northeast. Whether these two occurrences form one single lithosome could not be established. These lithic arenitic turbidites are interpreted as having been deposited on a submarine fan which prograded over the underlying deposits. As the overlying deposits are interpreted as having formed on a delta, the immature composition of the turbiditic sandstones warrants the conclusion that delta front sands were resedimented without being reworked by shallow marine processes.

Figure 14b. – This map represents nearly the entire *Fusulinella* B_1 subzone. In the southwestern part of the area a delta system developed during lower B_1 time. The fluviatile part of the delta is coal bearing at the location of Section 1. Further to the east there appears to be more marine influence in this delta system and at the locations of Sections 7 and 10 these deltaic sediments were contemporaneously reworked by marine waves and currents, and a strand plain supporting autochthonous plant growth was formed.

This delta system may therefore be classified as a high-destructive one. To the NW (Sections 11, 12, 13) and to the NE and E (Sections 8, 9 and 14 to 27) thick biogenetic bank deposits formed. Depending on the distance to the distributary mouth these biogenetic banks are associated with delta front sands (Sections 8, 9, 13), shallow marine to littoral deposits (Section 12) or fine-grained shelf to shelf slope deposits (Sections 11, 14 to 27). The development of the Tremaya Limestone Member in Section 10 and of the Socavón Limestone Member (Sections 4, 5) is proof of an important transgression over this first delta system. The continuation of deltaic sedimentation on an important scale in nearby areas during this period is indicated by the development of minor coarsening upwards sequences which were formed by minor distributaries in shallow bays and lagoons. After deposition of the second level of the Socavón Limestone Member, the main transgression took place. After this transgression a second delta system prograded over the submerged older one. The coal-bearing fluviatile deltaic deposits (San Cebrián coals) and the coal-bearing strand plain deposits (upper Perniana coals) extend over a large part of the southern half of the map. This delta system probably had a facies distribution and extension identical to the underlying one. An important difference, however, is that the first delta system is genetically associated with a thick sequence of lithic arenitic turbidites, whereas the second delta system only gave rise to the formation of relatively thin and not very extensive sequences of lithic arenitic turbidites. The depth of water in which the respective deltas formed may possibly have been the determining factor for this difference.

Figure 14c. – On this map deposits are depicted which formed during uppermost *Fusulinella* B_1 times and lowermost *Fusulinella* B_2 times. After deposition of the second high-destructive delta system an important transgression occurred. In Chapter IV (Fig. 9a) it was shown that this transgression was associated with, or may have been directly due to, quite important tectonic tilting of the southwestern part of the area. This tilting, which caused some erosion in the southwesternmost part of the area (Vergaño disconformity), apparently caused the delta forming river(s) to flow into a different area, no deltaic deposits immediately overlying the transgressed surface being known. This delta switching, which resulted in a diminished sediment supply, caused a progressive deepening in the southeastern part of the area as sedimentation could not keep up with subsidence. During the initial stage of this transgression shallow marine deposits were formed over a large part of the area. Locally (in between Sections 5 and 10) discontinuous limestones were deposited and some beach ridges supported autochthonous plant growth. This important transgression is well dated in Sections 5 and 10. In Section 5 the surface transgressed overlies the Socavón Limestone Member, the lowest level of which was dated near San Cebrián as belonging to the *Fusulinella* B_1 subzone. The surface transgressed is in turn overlain by the Coterarraso Limestone Member which was dated as *Fusulinella* B_2 subzone. In Section 10 the base of the main transgression is situated just below the Matacastillo Limestone Member, which was dated as *Fusulinella* B_1 subzone. These datings prove that in this part of the area the main transgression, which is associated with the Vergaño disconformity, occurred at the transition of *Fusulinella* B_1 to B_2 times and most probably in upper *Fusulinella* B_1 times. The Celada, Agujas and Camasobres Limestone Members, which are of *Fusulinella* B_1 age, are all three overlain by a deepening sequence (turbidites and locally slump deposits). This deepening sequence is obvious in Sections 8, 11, 14 to 27 and less distinct in Sections 12 and 13.

Taking into account the various datings it is logical to interpret this deepening, which occurred all over the area in probably upper *Fusulinella* B_1 times ,as being due to the tectonic tilting which caused the Vergaño disconformity. This correlation of the deepening of the basin at the location of Sections 14 to 27 with the tectonic tilting which caused the Vergaño disconformity to form, also provides the best evidence that the slump movements in the Pico Guillermo area were triggered by tectonic movements. An identical conclusion was already presented in Chapter IV for the slump level above the Celada Limestone Member in Section 8.

In Chapter IV it is argued that the presence of quartz arenitic pebbles in the deposits overlying the transgressed surface proves the existence of a longshore drift system. As at this time no fluviatile deltaic deposits were formed in the area studied, nearly all the siliciclastics must have been supplied by such a longshore drift system. Considering the small angle between the sediments below and above the Vergaño disconformity, it is unlikely that this part of the area supplied much sediment. The quartz arenitic pebbles provide a clue to the provenance of these post-transgression deposits. To the west of the area studied the very thick Curavacas conglomerates of Westphalian A to Upper Westphalian B age are present. The pebbles found in the present area and in these Curavacas conglomerates are both of quartz arenitic composition. It is considered very likely that the tectonic movements which resulted in the formation of the Vergaño disconformity also affected the area where these Curavacas conglomerates originally occurred. Because of these tectonic movements the conglomerates could be reworked by the sea and the erosion products were transported by longshore drift and partly deposited in the present area. The presence of quartz arenitic pebbles in the Agujas Limestone Member of *Fusulinella* B_1 age is explained in the same manner. The only difference is that no evidence is available of tectonic tilting of the area at that time. The interpretation presented for the occurrence of these quartz arenitic pebbles implies that the sediment transported by longshore drift came from western to southern directions.

Figure 14d. – On this map facies distributions during lower and middle *Fusulinella* B_2 times are shown. During this time interval the eroded area near the location of Section 3 was covered by marine sediments and could no longer be a source of sediment. In the eastern half of the map area mainly turbidites were deposited. A proximal facies with immature turbidites, grain flow and slump deposits, which in Chapter IV were interpreted as submarine canyon and proximal fan valley deposits, can be discerned at the location of Sections 4 and 5. To the NE an extensive submarine fan was formed (Sections 8, 14 to 27). Paleocurrent directions in this northern part of the area indicate derivation from the SE. Supply from the SE was also found by Nederlof (1959, p. 690) for younger deposits in the Redondo Syncline. He also noted that, considering the facies distributions, supply directions from SW to NE were to be expected. A prograding of this submarine fan and the adjoining shelf slope is indicated by the sequence found, for instance in Section 25, where mature turbidites are overlain by immature turbidites which are interpreted as fan valley deposits. These are in turn overlain by shelf slope deposits. To the west of, and partly coinciding with this realm of turbidite sedimentation in relatively deep water, shallow marine deposits were formed. The occasional

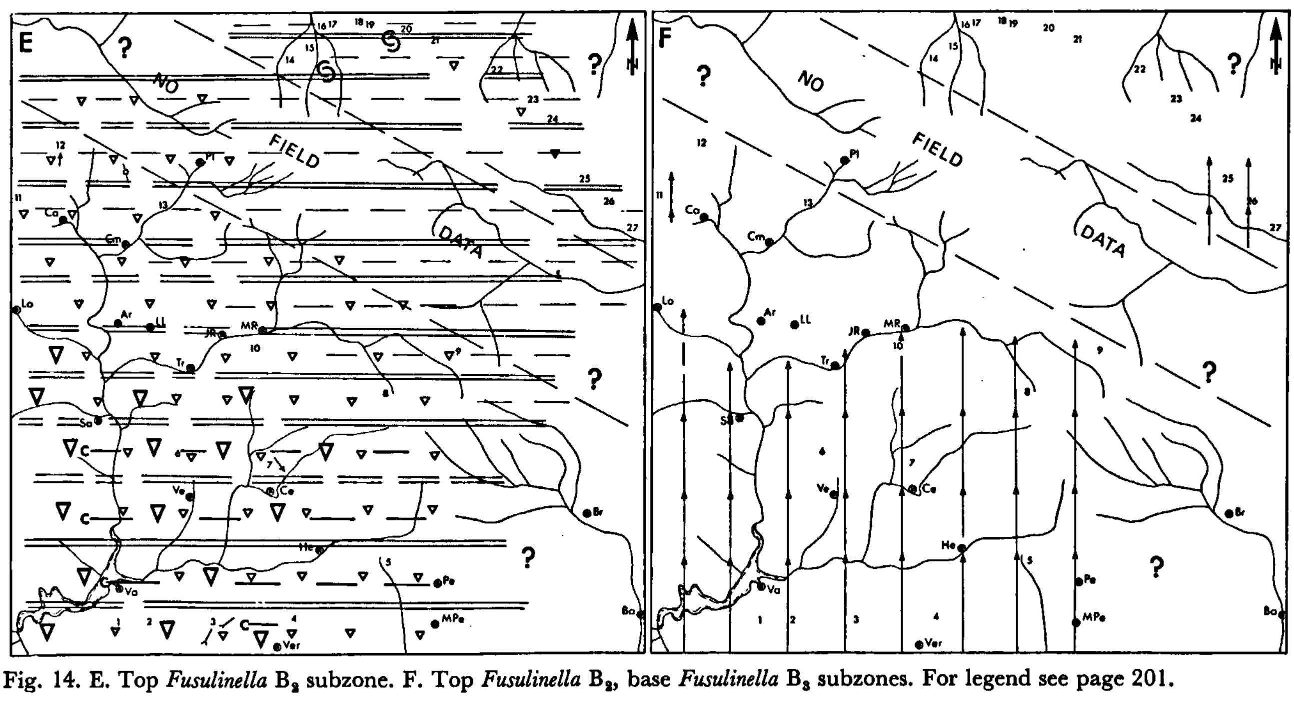

Fig. 14. E. Top *Fusulinella* B_2 subzone. F. Top *Fusulinella* B_2, base *Fusulinella* B_3 subzones. For legend see page 201.

presence of conglomerates proves that longshore drift was still important, but at the same time the first fluviatile deltaic deposits after the main transgression were formed at the location of Section 13. The small lateral extension of this deltaic deposit and its interfingering relationship with shallow marine deposits indicates that this delta, like the preceding ones, was of the high-destructive type.

Figure 14e. – This map represents the upper part of the *Fusulinella* B_2 subzone. In the southwestern corner of the map coal-bearing fluviatile deltaic deposits were formed which interfinger with biogenetic bank deposits within comperatively small distances. More or less continuous biogenetic bank deposits formed over most of the area during this period. To the N, NE and E of the fluviatile part of the delta (i.e. constructive facies), strand plain (= beach plain) and adjoining shallow marine deposits were formed (destructive facies). The strand plain deposits are locally coal bearing. In thc northeastern part of the area fine-grained siliciclastics were deposited in shelf and shelf slope environments in close association with biogenetic banks.

Figure 14f. – This map represents the situation at the transition of the *Fusulinella* B_2 into the B_3 subzone, i.e. the time of formation of the Leonian disconformity. Only the areas where more or less unambiguous signs of erosion were observed are indicated.

PALEOGEOGRAPHIC SYNTHESIS

As stated previously, every single map of Figure 14 may contain serious errors because of inaccuracies in correlations. Nevertheless, the overall paleogeographic evolution of the entire studied interval is very clear from the foregoing and may be summarized as follows. During *Fusulinella* B_1 and B_2 times (= large part of the Upper Moscovian $\approx$ Westphalian D) the area studied formed part of a more or less rapidly subsiding sedimentary basin. That it was not a small, independent basin, as according to de Sitter & Boschma's (1966) contention, is proved by the distribution of the facies and of the thicknesses. A basin margin to the southwest of the area studied is indicated by the formation of three delta systems the fluviatile parts of which are mainly restricted to this southwestern quadrant, by the presence of the Vergaño disconformity and by the fact that the Leonian disconformity is most distinct in this area. Of the three delta systems the oldest is associated with extensive turbidite sequences, the middle system with a few minor turbidite sequences and the youngest system with none at all. Each single delta system has a rim of shallow marine deposits to the NNW, N, NE and E. The presence of strand plain deposits in this rim, which were formed contemporaneously with the advance of the delta, and the rarity of tidally affected sediments indicate that the three delta systems are to be considered as wave-dominated, high-destructive delta systems. Further away from the delta fine-grained shelf and shelf slope sediments accumulated. Locally thick limestone units, which formed as biogenetic bank deposits, interfinger with siliciclastic deposits. After the formation of the Vergaño disconformity the same general distribution of land and sea is recognizable: emergence in the southwest, a rim of shallow marine deposits to the N, NE and E of this land, and further to the NE relatively deep water in which a submarine fan was formed by turbidity currents. The proximal facies of this submarine fan can be recognized in the southern part of the map. During this period sediment supply to the area took place by means of longshore drift. This longshore drift system transported the sediment from W, SW or S to the area studied.

PALEOCURRENT DIRECTIONS

The measurements of paleocurrent directions are too few in number to permit any other conclusion than that they do not contradict the paleogeographic picture presented. For several reasons no exhaustive survey of sedimentary structures in order to determine paleocurrent directions was made. First of all, folding in this area is often strongly disharmonic which makes it difficult accurately to determine the orientation of the fold axes. Ten Haaf (1959) has shown the effect such uncertainties will have upon the accuracy of the paleocurrent determinations. Secondly, it is relatively difficult to determine the current direction from trough cross-stratification which is the dominant type of cross-bedding in the deltaic deposits. Thirdly, the variability of the measurements made indicated the need for a statistical survey of large numbers of measurements. The three factors mentioned indicate that it would be very time consuming to obtain reliable results. Only sedimentary structures were therefore measured which were accidentally encountered during section measuring and which gave unambiguous results.

PROVENANCE OF THE SEDIMENT

The facies distributions strongly suggest that the source area from which the basin was supplied with sediment was situated to the S, SW or W. As a nearly fully marine sequence of *Fusulinella* B age is known about 40 km to the west (Lois-Ciguera Formation) (van Ginkel, 1965), it is most likely that the source area was located to the S or SW. That means that the source area is now covered by the Mesozoic and Cainozoic of the Castilian Meseta. The dominant heavy minerals of the Carboniferous sediments are tourmaline, rutile and zircon (Nossin, 1959). This indicates that only sedimentary rocks were exposed and eroded in the source area. From the size of the preserved deltas a very rough estimate can be made of the size of the delta forming river and of the size of its drainage area. The preserved deltaic deposits are but fragments of the original deltas. Especially the landward sides with fluviatile deposits are poorly preserved. However, in Figure 14b it is shown that about 100 km^2 of the subaerial part of the second delta system are still preserved. This means that the original subaerial delta was much larger. An estimate of 200 square kilometres seems a very conservative one. The recent Ebro delta, which has a subaerial area of approximately 250 km^2 (Shirley & Ragsdale, 1966), is of comparable size. This comparison indicates that a fair-sized river was responsible for the formation of the deltas in the area studied. Shirley & Ragsdale (1966) presented a number of data on the extension of the subaerial parts of recent deltas and the associated drainage basins. These data are given of 21 recent deltas and the drainage area/subaerial area ratios were calculated. The minimum ratio was 5 (Po delta), the maximum 210 (Ebro delta). These values indicate that the drainage area of these Carboniferous delta forming rivers was at least 1000 km^2 and probably much more.

DIFFERENTIAL SUBSIDENCE AND BASIN MARGINS

In the section on the paleogeography of the area studied it was stated that the sediments preserved represent but a fragment of the original sedimentary basin. De Sitter & Boschma (1966) contend that the studied sediments were formed in a fairly small, independent basin of about the same size as the present day outcrops. This basin, which they termed the Pisuerga basin, was bounded in the south and west by the Rabanal ridge and the Polentinos fault respectively (p. 233, Fig. 21). The facies and thickness distributions observed prove that the two features mentioned did not have the slightest influence on sedimentation and therefore probably did not exist. Only to the southwest can the presence of a basin margin be proved as is done above. During the upper part of the interval studied another ridge, the Celada ridge (= Los Llazos ridge according to Frets, 1965), which separated the present Redondo Syncline from the remainder of the area, is considered to have developed (de Sitter & Boschma, 1966). Nederlof (1959, p. 621) stated the evidence in favour of the assumption that a zone of less subsidence (Celada ridge of de Sitter & Boschma, 1966) existed. This evidence is the rapid thinning of strata from east to west in the northern part of the Redondo Syncline, and the thinning and wedging out of the limestones near San Juan de Redondo and Celada de Roblecedo. The latter argument is fallacious, as the wedging of individual limestone units merely indicates that they are laterally replaced by siliciclastics. Only complete sequences may be compared. In their Figure 22, de Sitter & Boschma (1966) draw a number of lines which suggest that, apart from the features mentioned, basin margins were present immediately north and east of the present day outcrops. De Sitter & Boschma (1966) do not present any factual evidence in the form of data on thickness or facies distributions to support their ideas. Another major weakness in their

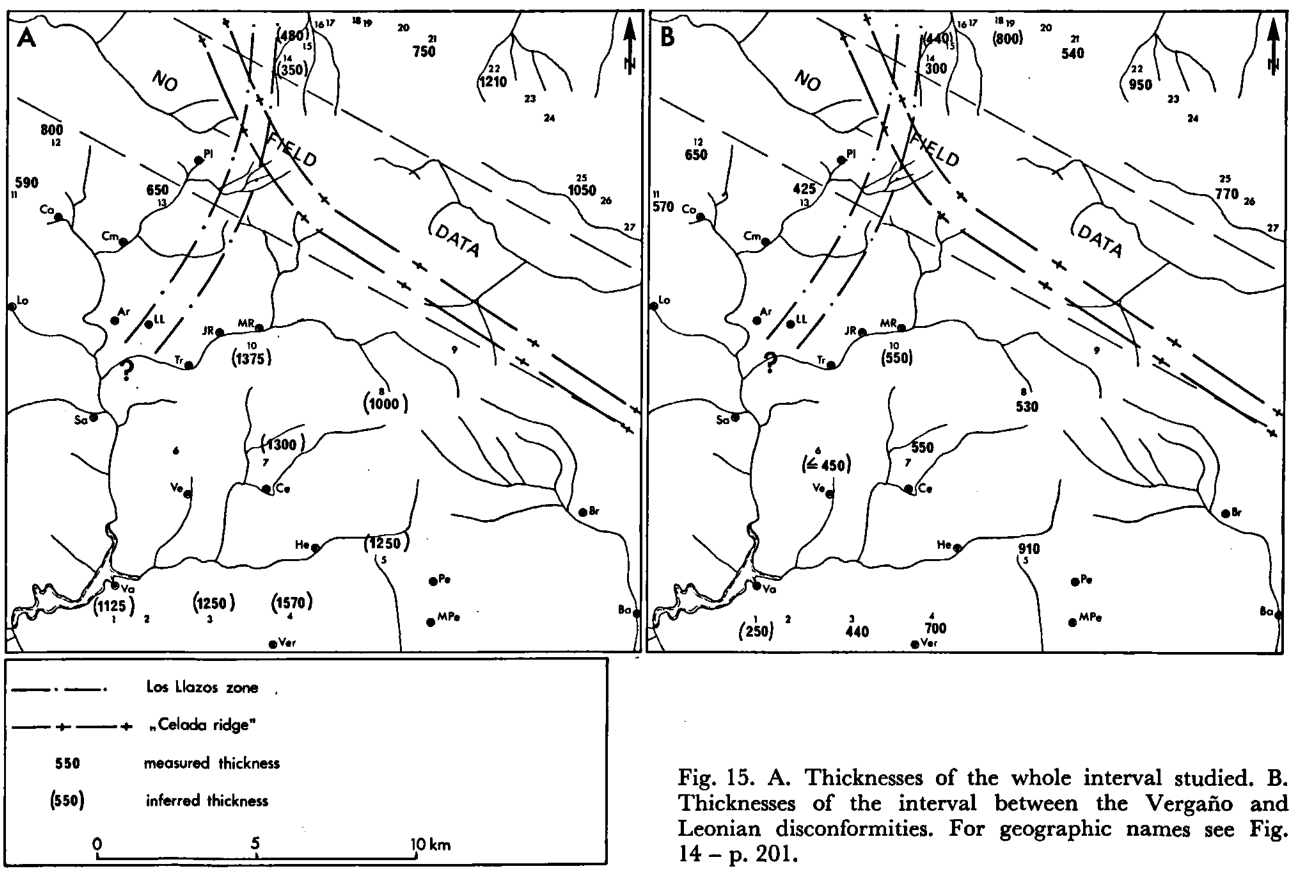

Fig. 15. A. Thicknesses of the whole interval studied. B. Thicknesses of the interval between the Vergaño and Leonian disconformities. For geographic names see Fig. 14 – p. 201.

interpretations is that they did not attempt a palinspastic reconstruction, although Nederlof (1959, p. 685) already mentioned the necessity of doing so.

With regard to the so-called Celada ridge it may be stated that the thicknesses measured do not indicate the existence of such a ridge which separated the Redondo Syncline from the remainder of the area. Especially in the upper part of the interval studied, when the ridge is considered to have developed, the thickness distributions are not very suggestive of its existence. As the facies distributions are not indicative of a ridge which separated the area into two subbasins either, evidence of the existence of a Celada ridge is lacking. I consider it much more likely that the rapid thinning of the interval studied in the Redondo Syncline is to be connected with the zone, between Sections 10 and 13, where important thickness changes occur (Figs. 15a and b). The most remarkable aspect of this zone is that for the lower part of the interval studied, sediment thicknesses are much greater to the east than to the west of the zone. For the upper part of the interval studied, however, the differences do not seem significant, but for the succeeding, post-Leonian sequence sediment thicknesses are much greater to the NW of the zone than to the SE and E. According to Wagner & Varker (1970), over 2000 m of sediment were deposited in the Casavegas area before sedimentation to the east of the zone began. This belt separating two areas which subsided at different rates was fairly narrow in between Section 10 and 13. If tectonic displacements on both sides of the zone are taken into account, it appears that the actual transition zone was probably about a kilometre wide. The sedimentary facies present on both sides of this belt suggest that only during lower *Fusulinella* B_1 times was the water appreciably deeper immediately east of the belt. During the remainder of the interval studied there was probably no surface expression of this zone. Considering the narrowness of the zone and the magnitude of the thickness differences, this zone is considered as a fault zone which was periodically active during sedimentation. During later tectonic deformation this zone was reactivated and the present day contacts are all interpreted as fault contacts (cf. Chapter IV-Peña Tremaya). As the fault zone apparently had little influence on facies distribution it never formed a topographic ridge separating the sedimentary basin into two separate areas and never was a source of sediment. This synsedimentary fault zone is referred to as the Los Llazos zone in Figures 15a and b, while the present day fault is called the Los Llazos fault in Figure 8.

CYCLIC SEDIMENTATION

In the preceding chapters it has been shown that deltas were very important in the area studied during *Fusulinella* B times. The great variety of facies, especially the cyclic occurrence of various facies types is easy to understand when thinking in terms of a delta model. The cyclicity of sequences occurs on widely different scales and in very different environments. On the one hand there are cyclic turbidite sequences which are cyclic on a centimetre to metre scale. On the other hand there are the deposits of the three main delta systems which may also be considered as cyclic deposits but on a scale in the order of hundreds of metres. These major cycles may be called megacyclothems. In both the constructive and destructive facies of the delta systems cycles on an intermediate scale may be discerned. Deltaic sedimentation with occasional diversion of the distributaries because of their overextension, combined with more of less continuous, but possibly irregular, subsidence can fully account for the cyclic sequences observed. This even holds true for the megacyclothem formed by the first and the second delta system. However, after deposition of the second delta system tectonic tilting occurred, as was shown in Chapter IV. This proves that at least on this scale cyclic sedimentation in this area was partly controlled by tectonic movements. In areas like the present one, where rapid lateral facies changes occur, it is impossible to distinguish regional or eustatic sea-level changes from local effects. The occurrence of such changes cannot therefore be excluded but is by no means required for explaining the observed cyclicity.

LIMESTONES AND THEIR RELATION TO THE SILICICLASTIC DEPOSITS

A distinctive feature of these delta systems is their very close association with extensive limestone deposits. In Chapter IV it was shown that in many cases this association was only possible on account of the slight elevation of the carbonate facies realms above their siliciclastic surroundings. As their muddy character and the absence of an organic framework precludes interpretation as reefs, they are considered to be biogenetic bank deposits. These limestones were formed in a wide variety of environments: in bays and sounds on a drowning delta system (Socavón Limestone Member); in protected lagoons behind a barrier beach (Matacastillo Limestone Member); in the open sea after passage of the seaward face of a migrating barrier (Tremaya Limestone Member); immediately laterally of active distributaries (Sierra Corisa Limestone Member); laterally of the delta in association with sand-sized sediments which were not reworked by marine agents (Celada, Verdiana Limestone Members); laterally of the delta in association with reworked sand-sized sediments which were deposited in open marine environments (Castrillo, Coterarraso Limestone Members); on the open shelf in association with fine-grained siliciclastics (Camasobres, Abismo, Agujas Limestone Members).

An important difference between the limestones formed on the open shelf, where supply of siliciclastics was low, and those formed nearer to the deltas is the character of the sediment which fills the depressions in between the biogenetic banks. In the areas sufficiently near to the deltas to receive considerable amounts of siliciclastics, the areas between the banks are filled with siliciclastic deposits. It is in this zone that the limestone units may have distinct biohermal shapes (Enclosure 3). On the open shelf, however, these depressions in between the biogenetic banks may be filled with carbonate sediments (Agujas Limestone Member) Another difference between the Agujas Limestone Member and the other limestone units is the much more frequent occurrence of oolitic grainstones in the former. This may also be related to its formation in an open shelf environment where waves had not yet broken on siliciclastic shores.

COMPARISONS WITH RECENT AND ANCIENT EXAMPLES

Before attempting such comparisons it is useful to summarize the main characteristics of these Upper Carboniferous delta systems. The contemporaneous development of the constructive and destructive facies of the deltas, and the general rarity of tidally influenced sediments proves that they are wave-dominated, high-destructive delta systems. The oldest and the middle system are genetically associated with turbidite sequences, which indicates that the deltas were formed in relatively deep water. The delta systems are furthermore closely associated with extensive limestone deposits which formed as biogenetic banks in environments ranging from lagoonal to open shelf. Between deposition of the second and third delta system, tectonic tilting occurred which is considered to have caused delta diversion. Consequently, only longshore drift systems supplied the basin with coarse siliciclastics. These sands are compositionally much more mature than the deltaic sandstones. Because of the tectonic tilting, continuing subsidence and low sediment supply, a deepening sequence developed and turbidites were deposited on a submarine fan adjacent to, and overlying shallow marine deposits. The various sedimentary processes and environments which have at one time or other occurred or existed in the area studied, can be understood in the terms of two fairly simple models. These are the delta model and the longshore drift/submarine canyon/submarine fan model. In both settings biogenetic banks could form. In Figure 16a and b these two models are depicted.

A recent example of a wave-dominated, high-destructive delta system mentioned by Fisher (1969) is the Rhône delta. It is a very good example indeed, associated as it is with a large submarine fan (Menard, Smith & Pratt, 1965). If size is taken into account, the much less studied Ebro delta would appear to be an even better example. However, no important limestone deposits are associated with these deltas. In this respect the Yallahs area near Kingston, Jamaica, which was described by Burke (1967), may be a better recent analogue. Gravel deltas supply the Yallahs basin with sediment which is inferred to be resedimented by slides and turbidity currents as indicated by several cable breaks. From the shape of the deltas it seems likely that they are of the wave-dominated, high-destructive type. Within a few kilometres of the gravel deltas and the adjoining spits, coral reefs flourish on the island shelf. These carbonate deposits also contribute sediment to the submarine fans in the Yallahs basin. As Jamaica forms part of the Caribbean loop it is frequently affected by earthquakes. This may be another similarity between the Yallahs area and the area studied, as the Vergaño and Leonian disconformities prove that tectonic tilting occurred periodically.

The limestone deposits which are interpreted as predominantly biogenetic bank deposits, are somewhat similar to the Waulsortian "reefs" of Belgium and Ireland. These Waulsortian "reefs" of Lower Carboniferous age were recently described by Dupont (1969) and Lees (1961, 1964). These "reefs" lack an organic framework and are interpreted by Lees as mudbanks which commonly have a biohermal shape. They differ from the Spanish bank complexes in several aspects. First of all, the stromatactis cavities in the Waulsortian "reefs" are of a very different type. Secondly, bryozoans are an important faunal element whereas algae are more important in the Spanish examples. Thirdly it can be shown in a number of cases that steep depositional slopes occurred on the flanks of the Waulsortian banks, which feature has nowhere been noted in the area studied. Other occurrences of "reefs" of the Waulsortian type were described by, for instance, Pray (1958) and Cotter (1965, 1966) in the Mississippian of the United States. In the Pennsylvanian of the southwestern United States, carbonate mound complexes occur which resemble the Spanish examples closer than do the Waulsortian ones. Elias (1963), Pray & Wray (1963), Heckel & Cocke (1969) who, among others, described these deposits, stress the important part played by phylloid algae in the formation of these biostromal or biohermal carbonate deposits. Apparently similar carbonate deposits are closely associated with the Cisco delta systems of Upper Virgilian and Wolfcampian age in North Texas (Brown, 1969). These Cisco deltas, however, differ from the Spanish examples in that they are of elongated to lobate high-constructive type. These deltaic deposits, which grade laterally into interdeltaic and strand plain deposits, interfinger with limestones formed on an open shelf. Near the shelf edge, carbonate banks developed. Shelf slope deposits are predominantly mudstones with subordinate limestones, siltstones and sandstones. At the foot of the slope thin fan-like sand bodies were deposited. These sandstones were possibly emplaced by turbidity currents. Although numerous ancient delta deposits have been recognized and described as such, these Cisco deltas and associated environments are the best ancient analogue to the Spanish delta systems that I am acquainted with from literature.

As recent analogues of the sediments deposited in between the second and third delta system, I consider, for instance, the submarine canyon and fan systems off

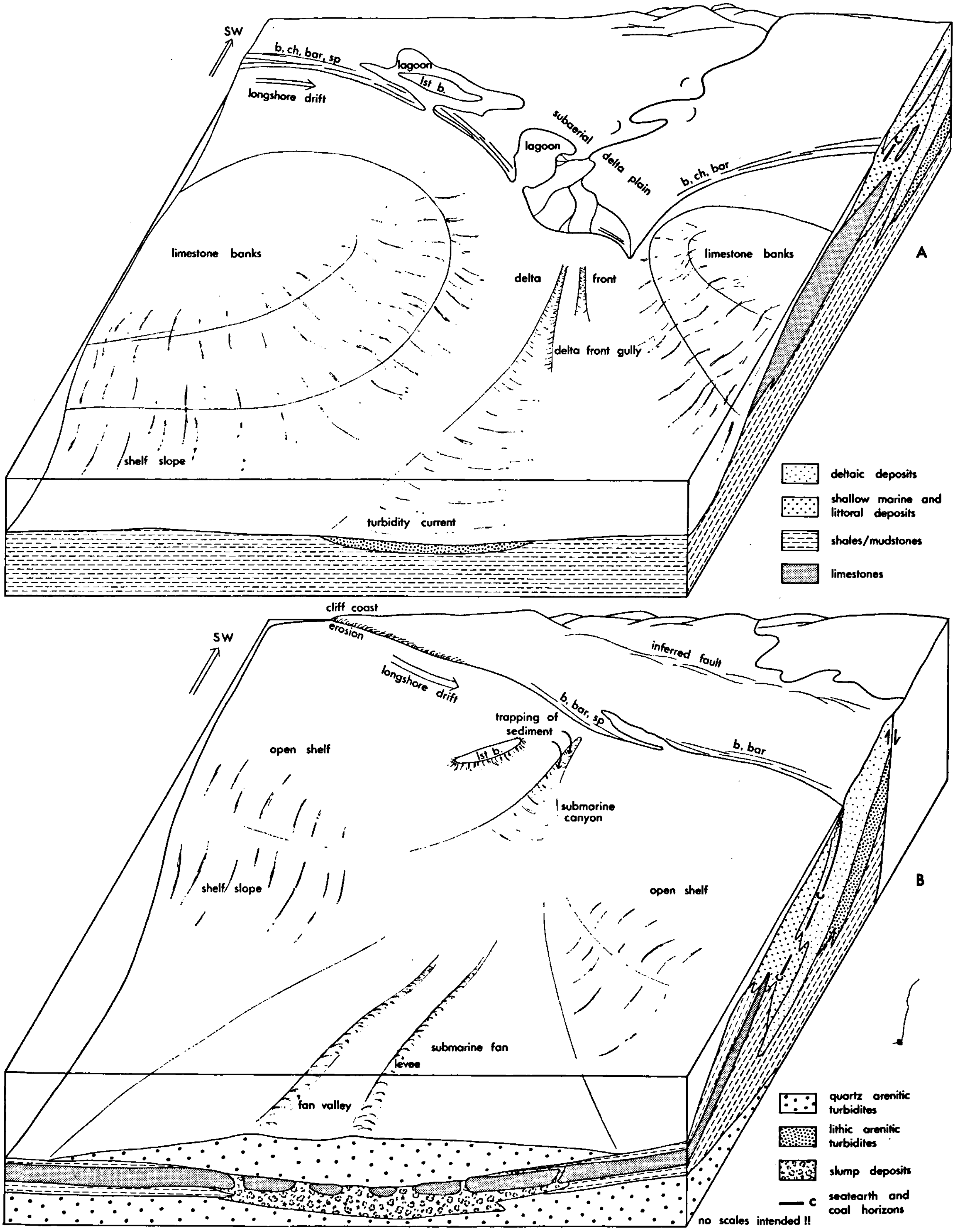

Fig. 16. A. Depositional environments during delta formation. B. Depositional environments after formation of the Vergaño disconformity. Although they are approximations of paleogeographic reconstructions these drawings mainly illustrate sedimentary processes and environments. b – beaches; ch – cheniers; bar – barrier beaches; sp – spits; lst b – limestone banks.

the coast of California. As, for instance, the Delgada and Monterey canyons are not associated with major rivers, the sediment transported down the canyons is probably supplied by means of longshore drift. Such interception of coarse sediment moving along the shore by a canyon heading near the shore line is a well-known process (Hand & Emery, 1964).

CHAPTER VI

LITHOSTRATIGRAPHIC NOMENCLATURE

Although a set of lithostratigraphic names does not help in understanding a stratigraphic problem, such names can be very useful in describing and discussing such a problem. In order to render such a set of names useful they have to be precisely defined and used in a consistent manner. Because of rapid facies changes and an inadequate understanding of the sedimentary facies of the interval studied, the formations and members used by previous authors are partly ill-chosen. Furthermore, the characteristics on which mappability of the various units is based and the boundaries used are poorly described. In some cases the name of a formation and one of its constituent members were both taken from the same geographical feature (Corisa or Sierra Corisa Formation and Sierra Corisa Limestone Member; Maldrigo Formation and Maldrigo Limestone; Abismo Formation and Abismo Limestone), which is hardly recommendable.

Table II shows the sequence as exposed near Vergaño with the formation names as used in some of the more recent papers on the Pisuerga area. In the remainder of the area studied use of these names is even more confusing. As I consider the set of formation names in current use in the entire Pisuerga area as unsatisfactory, a number of changes are proposed. The changes proposed are restricted to the interval studied, but some changes in formation names currently being proposed by Wagner & Winkler Prins (1970) and

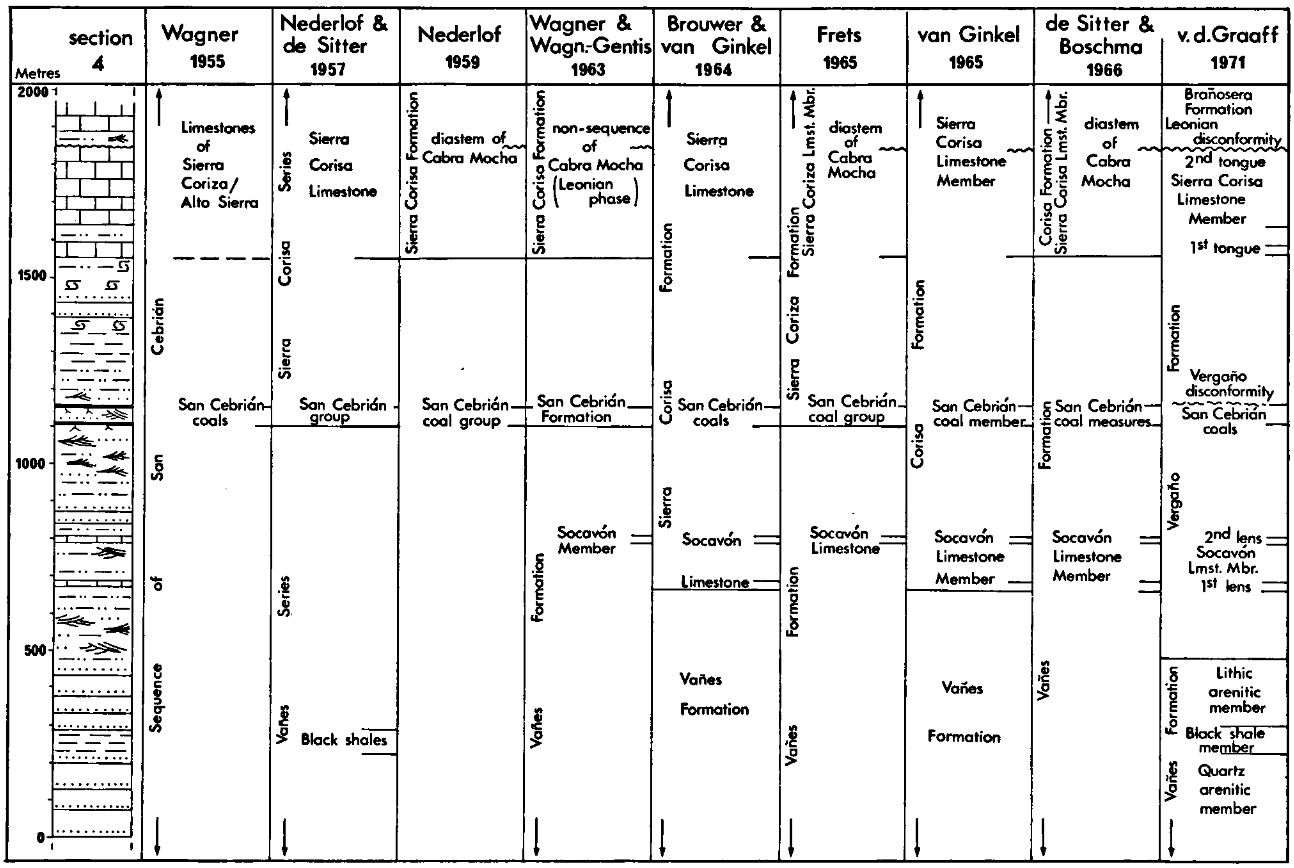

Table II. Lithostratigraphic nomenclature of the interval studied as exposed in the southern flank of the Castillería Syncline. All terms have been translated into English.

Wagner & Varker (1970) for the overlying sequence have been adopted. Most of the formally introduced members are limestone units which are easily recognizable in the field. Therefore there is little need to revise most of these limestone members. The formal and informal units recognized have been indicated in the stratigraphic sections and also partly in the correlation diagram.

VAÑES FORMATION

The name "Series de Vañes" was originally proposed by Nederlof & de Sitter (1957). The type section is situated to the northwest of Rabanal de los Caballeros (Brouwer & van Ginkel, 1964), i.e. in between Sections 2 and 3. As the type section is not very well exposed, Section 4 is selected as a reference section. The basal part of the Vañes Formation was not studied. The upper boundary, however, is here redefined and arbitrarily taken at the base of the first macro cross-bedding (decimetre to metre scale) in the sequence exposed in the type section. Even in poorly exposed areas this is an easily recognizable level and the boundary with the Vergaño Formation is therefore easily mappable. Defined as such, the Vañes Formation forms only a part of the Vañes Formation according to previous authors (Table II).

The redefined Vañes Formation comprises three informal members. A lower quartz arenitic member is distinguished consisting of a sequence of more or less distinctly graded sandstones alternating with shale/mudstone. In the preceding chapters these sandstones are interpreted as mature and immature turbidites. This lower member was mapped in the south flank of the Castillería Syncline and in the northern part of the Redondo Syncline near the Piedrasluengas pass. This lower member grades into the middle member, which is called the black shale member. The boundary is drawn where facies type Ie grades into type IIIa. The black shale member consists of dark grey shale/mudstone with rare sideritic concretions, a few burrows and occasionally silty or sandy layers of type Ie. It can be shown locally (Section 10) that this member was probably formed on a shelf slope. This middle member of the Vañes Formation is the best marker unit of the whole area studied occurring as it does in the Castillería Syncline, the Verdegosa-Verdiana Anticlinorium, the Casavegas Syncline and in the Redondo Syncline. Peculiarly enough this black shale member was distinguished in one of the earlier papers(Nederlof & de Sitter, 1957) but was not subsequently used as a map unit. In a large part of the area the middle member is overlain by the upper or lithic arenitic member. In the Castillería Syncline the contact is usually sharp, as the black shale/mudstone is abruptly overlain by thick, often massive lithic arenitic sandstones. Some of these sandstone beds are graded, and in between the sandstone beds shale/mudstone is intercalated. The upper part of this member does not usually contain any shale/mudstone layers. The presence of small-scale cross-bedding (centimetre scale) may herald the succeeding macro cross-bedding of the overlying Vergaño Formation. In the preceding chapters these lithic arenitic sandstones are interpreted as immature to mature turbidites. The upper part of this member is usually the lower part of a major coarsening upwards unit. This upper member was mapped in the Castillería Syncline, the Verdegosa-Verdiana Anticlinorium and in Section 13 in the Casavegas Syncline.

The lower member of the Vañes Formation is shown (Chapter IV, Verdegosa-Verdiana Anticlinorium) to be the lateral equivalent of the Curavacas Conglomerate Beds of the Casavegas Syncline. A similar correlation was already proposed by Brouwer & van Ginkel (1964). This also corroborates the proposed correlation between the Lechada Formation and the Vañes Formation (Boschma & van Staalduinen, 1968). On the basis of extrapolation of van Ginkel's (1965) data the age of the Vañes Formation is considered to be Lower to Middle Moscovian ($\approx$Westphalian C to Lower Westphalian D), as the underlying Albas Limestone Member yielded a fauna of the *Profusulinella* B zone, and the overlying Socavón Limestone Member (Vergaño Formation) a *Fusulinella* B_1 fauna.

VERGAÑO FORMATION

This formation replaces the Corisa or Sierra Corisa Formation of previous authors in the Castillería Syncline, the Verdegosa-Verdiana Anticlinorium and the Casavegas Syncline. A new name is introduced for two reasons: first of all, both the upper and lower boundary are changed considerably, secondly, both a member and the formation itself were named after the same geographic feature. As the dominant lithology of the Sierra Corisa hill is limestone, it is considered least confusing if the name Sierra Corisa is retained for the limestone member.

The formation is named after the village of Vergaño. Section 4 was measured along the road from Vallespinoso de Cervera to Vergaño and in the valley NNE of Vergaño. This Section 4 is designated as the type section, and because of lateral variability Sections 7 and 13 are designated as reference sections (Section 4, 491–1845 m; Section 7, 238–1198 m; Section 13, 130–655 m).

The lower boundary of the Vergaño Formation is taken at the base of the first macro cross-bedding in Section 4, which horizon is easy to map. Where no distinct macro cross-bedding was observed, as in the western part of the Casavegas Syncline, the base of the formation is arbitrarily taken at the base of the Camasobres Limestone Member which in this area directly overlies the black shale member of the Vañes Formation. Comparison of Sections 12 and 13 indicates that no great error is made in doing so. The upper boundary of the Vergaño Formation is taken at the base of the mature sandstones of facies type IXa, which indicate the level of the Leonian disconformity (cf. Chapter I). In the Casavegas Syncline, where the position of the disconformity is difficult to establish, it is arbitrarily taken at the top of the Maldrigo Limestone Member (Sections 12 and 13). Defined as such the Vergaño Formation comprises a wide variety of lithologies which at many places show a distinct cyclicity. The cyclicity may assume several forms, but repetitions of coarsening upwards sequences are the most striking.

Limestone units are common in this formation. These limestone units commonly possess sharp boundaries and clear topographic expressions and because of these characteristics are easy to map as distinct members of the Vergaño Formation. As the sequences in the various structural units are not obviously homotaxial and because of the discontinuous nature of the limestone units, these limestone members have been differently named in every structural unit. Van Ginkel (1965), though not formally proposing them as such, evidently considers the various limestone units mentioned by him as formal members. Although it is likely that some of these formal members were formerly continuous to each other (Table I, Fig. 14b, c, d, e, and Enclosure 5) it is considered advantageous to retain these names as in this manner it is easier to indicate precisely which limestone outcrop is meant. Van Ginkel (1965) recognizes the following limestone members belonging to the Vergaño Formation introduced above: Socavón Lst. Mbr., Coterarraso Lst. Mbr. (spelt Cotarazo in van Ginkel, 1965), Sierra Corisa Lst. Mbr., Camasobres Lst. Mbr., Castrillo Lst. Mbr. and Maldrigo Lst. Mbr..

The following limestone members of the Vergaño Formation are proposed here: Celada Lst. Mbr., Villar Lst. Mbr., Verdiana Lst. Mbr., Verdegosa Lst. Mbr., Sosa Lst. Mbr., Tremaya Lst. Mbr., and Matacastillo Lst. Mbr.. All members are indicated in Figure 3 and in the stratigraphic sections. Several of these names were used before in an informal manner by Nederlof (1959), viz. Verdiana (Verdiana-Frechilla) limestone, Verdegosa limestone, Sosa limestone. Although Nederlof (1959) does not state precisely which limestone unit is meant by Celada limestone, it probably comprises both the Celada Lst. Mbr. and the Villar Lst. Mbr. as used in this paper. The Villar Lst. Mbr. is named after a brooklet that crosses the Celada Syncline near its middle. The Tremaya Lst. Mbr. occurs in Section 10 and is named after the nearby village of Tremaya. This limestone unit is considered as a lens or possibly a tongue of the limestones composing the actual Peña Tremaya. The Matacastillo Lst. Mbr., which is slightly higher in the succession, is named after the flat area south of Peña Tremaya.

I consider the San Cebrián, Perniana and Por si Acaso coals as informal members of the Vergaño Formation. Of these three coal-bearing sequences the San Cebrián and Perniana coals are at least partly each others equivalent.

Field evidence and datings (van Ginkel, 1965) indicate that the Vergaño Formation is the lateral equivalent of the Covarres Formation (see section on Covarres Formation) and of (parts of) the Lechada Formation (Boschma & van Staalduinen, 1968) and the Lois-Ciguera Formation.

As the rich fusulinid faunas present in the various limestone members belong to the *Fusulinella* B_1 and B_2 subzones, the Vergaño Formation is of Upper Moscovian age. Wagner (1970, pers. comm.) collected a Westphalian D flora from cross-bedded sandstones south of the San Cristóbal hill. This appears to be the sandstone level below the first lenses of the Socavón Limestone Member (Section 1, 770–860 m; Section 4, 500–570 m), i.e. very near the base of the Vergaño Formation. In the Casavegas Syncline the Casavegas coals, several hundreds of metres above the top of the Vergaño Formation, were dated as uppermost Westphalian D (Wagner, 1970; Wagner & Varker, 1970). This proves that the Vergaño Formation represents only a part of the Westphalian D.

COVARRES FORMATION

The Corisa or Sierra Corisa Formation of previous authors was considered to extend into the Redondo Syncline. Although the limestones present in the lower part of the sequence exposed in the Redondo Syncline are very similar to the limestones of the Vergaño Formation, the associated siliciclastics are quite different. Although the sandstone/shale sequences which are interpreted as turbidites with interbedded shales, are cyclical, no rhythms of coarsening upwards or fining upwards sequences were observed. The absence of features like c.u.s., f.u.s., macro cross-bedding and seatearth or coal horizons, all characteristic of the Vergaño Formation, indicates that a different formation name is useful. The interval studied as exposed in the Redondo Syncline is therefore considered to comprise the upper part of the Vañes Formation and the Covarres Formation formally proposed here. This new formation is named after the Covarres hill in the southern part of the syncline (Enclosure 3), and Section 25 is designated as the type section.

The Covarres Formation conformably overlies the shale member of the Vañes Formation in the larger part of the syncline, and in turn is overlain disconformably by the Brañosera Formation. Three formal and two informal members of the Covarres Formation are recognized. At the base of the formation the Agujas Limestone Member is present. In the area were the limestone is discontinuous, the underlying contorted siliciclastics are arbitrarily included in the Covarres Formation. The Agujas Limestone Member is overlain by a sandstone/shale alternation which is named the sandstone member of the Covarres Formation. The sandstones are interpreted as turbidites in the preceding chapters. In the northwestern part of the Redondo Syncline a fairly thin, more shaly sequence occurs at the base of the sandstone member. The sandstone member is overlain by a shale/mudstone sequence which is called the shale member of the Covarres Formation. Lenticular limestone breccias occur locally in this shale member. The shale member is interpreted in preceding chapters as a shelf slope deposit. The

Abismo Lst. Mbr. and the Tejedo Lst. Mbr. overly, and partly interfinger with, the shale member.

The names Agujas and Abismo Lst. Mbrs. were used before in a formal sense by van Ginkel (1965). The sandstone and the shale member were already described by Nederlof (1959), who used them as informal units. The Tejedo Lst. Mbr., which is named after the Peña Tejedo, is distinguished here as a formal unit. The reason for this is its isolated position, although it is probably the equivalent of the Abismo Lst. Mbr.

The Leonian disconformity is only clearly recognizable at the location of Section 25. Elsewhere the top of the Covarres Formation is taken at the top of the Abismo Lst. Mbr.. Where the Abismo Lst. Mbr. is discontinuous, the boundary is taken at the top of an often intensively bioturbated sandstone (Type IIIc to IVa, Plate IIg), which occurs a few tens of metres below the disconformity (Section 25).

The Agujas and Abismo Limestone Members yielded a fusulinid fauna belonging to the *Fusulinella* B_1 and B_2 subzones respectively (van Ginkel, 1965), which indicates an Upper Moscovian age for the Covarres Formation. The Tejedo Lst. Mbr. yielded a fauna that appears to be intermediate between those found in the Abismo and Coterarraso Lst. Mbrs.. These datings prove that the Covarres Formation is the lateral equivalent of the Vergaño formation.

De Sitter & Boschma (1966) extended the use of the names Ruesga Group, Yuso Group and Cea Group to the Pisuerga area. The interval studied, i.e. the redefined Vañes Formation and the newly proposed Vergaño and Covarres Formations, belongs to the the Yuso Group as employed by de Sitter & Boschma (1966). These groups were originally defined by Koopmans (1962) as being separated by folding phases. The Yuso Group is considered to comprise all sediments deposited after the Sudetic folding phase and before the Asturian phase. This division into three groups may be a useful concept in the area where Koopmans worked, but in the area studied it has led to an illogical subdivision of the sequence (Wagner, 1970, p. 451). Although the use of groups based on subdivision of the sequence by tectonic phases may be practicable in the entire Pisuerga area, this should be considered in a wider context which is beyond the scope of this study.

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