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A PROGRADING DELTAIC COMPLEX IN THE UPPER CARBONIFEROUS OF THE CANTABRIAN MOUNTAINS (SPAIN): THE PRIORO-TEJERINA BASIN

BY

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ABSTRACT

In the southern flank of the Cantabrian Mountains, northwestern Spain, a sequence of Upper Carboniferous sediments is exposed in a synclinal structure, probably coinciding with the original basin, near the villages of Prioro and Tejerina. By means of palaeontological dating with several fossil groups (e.g. fusulinids, brachiopods, calcareous algae and land plants) the lower sequence of these sediments could be dated as Westphalian B/C to lower or middle Westphalian D (Yuso Group). After a relatively short time interval follows a sequence with an uppermost Westphalian D to lower Cantabrian age (Cea Group). These two groups are separated by an angular unconformity.

These sediments together represent a regressive sequence, starting with a turbidite facies and gradually passing into a shallow marine facies at the top of the Yuso Group. The Cea Group is possibly fully continental, except for a few metres of shallow marine sediments in the middle part.

Facies interpretations were made by investigation of the fossil content and the sedimentary structures. Rapid lateral facies changes could be traced from 15 stratigraphic sections through the best exposed parts. Six of these sections were sampled in detail to enable a petrographic investigation to be made. This resulted in the possibility of drawing conclusions on the lateral and vertical facies changes by means of grain-size distribution, micro-fossil content and, especially, the modal distribution of matrix-sized (< 25 microns) material.

From these data, together with field observations, the palaeogeography could be reconstructed: a deltaic complex, emerging from the northern border and supplying much material from the N, prograded into an E–W trending basin. In the deeper parts this material was transported to the E along the basinal axis. The prograding of the delta caused a gradual shallowing of the basin, which, finally, resulted in fluvial sedimentation with coal layers.

SAMENVATTING

In de zuidflank van het Cantabrisch Gebergte, noordwestelijk Spanje, is een opeenvolging van bovencarbonische sedimenten ontsloten in de omgeving van de dorpjes Prioro en Tejerina. Deze afzettingen zijn synclinaal geplooid, waarschijnlijk volgens het patroon van het oorspronkelijke bekken. Datering met behulp van verschillende fossielgroepen (o.a. fusulinen, brachiopoden, kalkalgen en landplanten) wijst op een Westfalien B/C tot onderste of middelste Westfalien D ouderdom voor het onderste deel (Yuso-Groep). Na een vrij kort tijdsinterval volgt een opeenvolging, die gedateerd is als bovenste Westfalien D tot onderste Cantabrien (Cea-Groep). Deze twee groepen worden gescheiden door een hoekdiscordantie.

Deze afzettingen vormen samen een regressieve sequentie, die begint met een turbidietfaciës en geleidelijk overgaat in een ondiep mariene faciës in het bovenste gedeelte van de Yuso-Groep. De Cea-Groep is mogelijk geheel continentaal, met uitzondering van een enkele tientallen meters dik pakket van ondiep mariene afzettingen in het middelste gedeelte.

Interpretaties van de faciës zijn gemaakt door middel van een onderzoek van de sedimentaire structuren en de fossielinhoud. Snelle laterale faciëswisselingen kunnen aangetoond worden door middel van 15 stratigrafische secties door de best ontsloten delen van het gebied. Van zes van deze secties werden uitgebreid handstukken verzameld die petrografisch zijn onderzocht. Hierdoor kunnen conclusies worden getrokken over de laterale en verticale faciëswisselingen naar aanleiding van de korrelgrootteverdeling, de aanwezige microfossielen en vooral door de modale verdeling van het materiaal dat kleiner is dan 25 micron.

Deze gegevens geven, samen met de veldwaarnemingen, gelegenheid om de palaeogeografie te reconstrueren: een delta-complex vergrootte zich vanuit het noorden naar het zuiden in een oost-west lopend bekken. Het uit het noorden aangevoerde materiaal werd in de diepere delen van het bekken verder naar het oosten getransporteerd volgens de bekkenas. Het groter worden van de delta veroorzaakte een ondieper worden van het bekken met, uiteindelijk, een volledige verlanding. In deze laatste fase werden fluviatiele afzettingen met koollaagjes gevormd.

CONTENTS

CHAPTER I

INTRODUCTION

I.1. GEOGRAPHICAL AND GEOLOGICAL SETTING

The area which is the subject of this study lies in the northeastern part of the province of León, on the southern flank of the Cantabrian Mountains (Spain). The Río Cea, one of the larger N–S flowing rivers which emerge from the E–W trending mountain chain, has its source in this area (Fig. 1).

The main subject of this investigation (see I.3) was to study marine and continental Carboniferous sediments and their transitions in a sedimentary basin of small size. In the course of the work the investigation also yielded additional data on the stratigraphy of the upper Westphalian and lower Stephanian.

The sediments studied all belong to the Upper Carboniferous. Four formations can be distinguished, the Prioro and Pando Formations (the latter being divided into three members) that belong to the marine Yuso Group (≈ Westphalian), and the Ocejo and Tejerina Formations (the latter with two members) belonging to the mainly continental Cea Group (≈ Stephanian). These two groups are separated by an unconformity.

Since the area studied lies S of the León line, a fundamental fault zone (see Chapter IX), it must be considered to belong to the 'Leonides', a region most parts of which were uplifted as blocks during Westphalian times. The area studied is one of the few in the Cantabrian Mountains where Westphalian deposits in a marine facies can be found south of this line. This must be caused by a very local subsidence to below sea level.

The León line was active, as is shown by previous investigations, from before until after the deposition of the sediments studied. For this reason, tectonic differences as well as facies differences arose at both sides of this line. We therefore took this line as the northern boundary of the area investigated. The road from Pedrosa del Rey to Besande forms the eastern limit; the road from Prioro to Tejerina the western and southwestern limit. The contact between the Prioro Formation and the Devonian formations or the Ocejo Formation in the SE part was taken as the SE boundary (Fig. 2).

The unconformity between the Yuso and Cea Groups

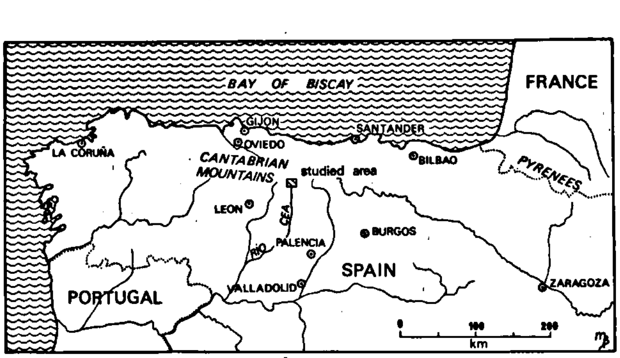

Fig. 1. Geographical setting of the area studied.

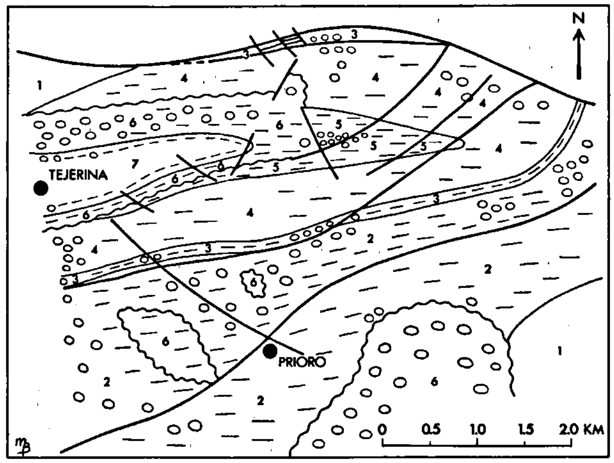

Fig. 2. Schematical geological map of the Prioro-Tejerina area. 7: Tejerina Fm.; 6: Ocejo Fm.; 3–5: Pando Fm. (5: Upper Sst. Mbr.; 4: Mesao Lst. Mbr.; 3: Lower Sst. Mbr.); 2: Prioro Fm.; 1: older fm.; —: relatively fine-grained facies; ooo: relatively coarse-grained facies.

is clearly shown on the geological map. In the field, however, it can only be ascertained in a few places that there is an angular unconformity between these groups in this region.

The sediments have been folded into a large syncline the axis of which plunges to the W.

I.2. PREVIOUS AUTHORS

Geological maps dealing with this area have been published by:

Mallada, 1892, whose map ended some 150 m S of the southern part of our Figs. 2 and 69. His – very schematical – map indicates a Lower Carboniferous age for this part of the Prioro Formation.

Mallada, 1927, noticed the occurrence of both marine Carboniferous with crinoids, brachiopods and corals N of Tejerina, and continental deposits with coal seams.

Helmig, 1965, and Rupke, 1965, together published a map of the Cea-Esla-Porma region. In their theses on the structure of this region they consider the Prioro Formation as Namurian, the Pando Formation as Westphalian, and the Ocejo and Tejerina Formations as Westphalian D or Stephanian A. Helmig presented many structural and palaeobotanical aspects of the Cea Group, which he named the Cea Formation.

Important stratigraphical data were published by:

Brouwer & van Ginkel, 1964, who described the type sections of both the Prioro and the Pando Formations, dating them as *Profusulinella* A/B subzone (tentatively, on account of its lithological resemblance to the dated Piedras Lenguas Formation) and middle *Profusulinella* B to *Fusulinella* B subzone respectively (see Fig. 3). They thought these formations to be separated by an unconformity (the same that is shown on the map by Helmig and Rupke). Our Cea Group (their Cea Formation) was divided into three members.

Van Ginkel, 1965, described fusulinid foraminifers from several Carboniferous formations, among which the Mesao Limestone Member of the Pando Formation. His datings were already given in Brouwer & van Ginkel.

Winkler Prins, 1968, investigated a number of brachiopods from Carboniferous formations, among which several parts of the Pando Formation. His dating is in rather good accordance with the dating by van Ginkel.

Van Loon, 1971, produced a rather exhaustive list of fossils from both the Prioro and the Pando Formations, indicating Westphalian B/C to Westphalian C/D ages for these formations. Fifty fossils from these formations were reproduced on eight plates, added to this paper. Several other pelecypods have been described and/or figured by van Amerom in van Amerom et al., 1970. Other, not yet identified pelecypods possibly will be described by the same author in the future.

Wagner in several papers (Wagner, 1964, 1966b, 1969; Wagner et al., 1969) demonstrated that a stratigraphic problem arose when dating the Ocejo and Tejerina Formations by means of plants. This problem will be dealt with in VII.7. Especially our section 12, N of Tejerina, has been the subject of many recent papers.

Some sedimentological aspects of the Yuso and Cea Groups in this area have been dealt with by van Loon (1970) and de Jong (1971) respectively.

I.3. SCOPE OF THIS STUDY

Our investigation began as a purely sedimentological study. It was our intention to study the sedimentary structures, the relation of these to the lithology and the possible sedimentary sequences to establish a final reconstruction of the depositional environment at several chronostratigraphic intervals.

The sequence could conveniently be divided into two parts: first, the sediments of the Yuso Group, which are fully marine and in which one megacycle may be seen. On the other hand, the sediments in the Cea Group which, except for a small marine band (the Barranquito Member), are fully continental (or only slightly affected by the sea), showing many repetitions of similar cycles.

During the work in the Yuso Group many difficulties were encountered, caused by the laterally rapidly changing lithology and the many tectonic disturbances, the effect of which could not easily be estimated for want of any marker horizon. It also soon became evident that an unconformity between the Prioro and Pando Formations, as still recently supposed (e.g. Boschma & van Staalduinen, 1968), was very improbable. To obtain more certainty, much more attention than was foreseen was paid to the fossil content. Therefore, as our study proceeded, the stratigraphic aspect became no less important than the sedimentological aspect. This resulted in a publication on the stratigraphy of the Prioro and Pando Formations for the benefit of an excursion to this region by the Subcommission on Carboniferous Stratigraphy of the I.U.G.S. (van Loon, 1971).

The Cea Group yielded less difficulties. The sequences could easily be recognized, while the sediments were much less disturbed tectonically. Besides, palaeobotanists were already very active here before.

For all these reasons, emphasis was laid on the Yuso Group and particularly on the Prioro Formation, because neither the tectonics nor the sedimentology of this formation were reasonably known, while before our work hardly any fossils were found in it, preventing a dating based on other data than lithological resemblances.

The main goal, however, still remained the interpretation of the sedimentary features and the conclusions that could be drawn herefrom on the palaeogeography during the time span in which the four formations described were deposited.

I.4. METHODS

Field work in this area, totalling nine months, was carried out in the summers of 1967 to 1970. First, rather rough structural investigations were made in order to obtain an impression of the tectonic deformation. As the study proceeded it became necessary, for correlation purposes, to have very exact data on the structure, especially in the Prioro Formation. For this reason a more detailed investigation was made, resulting in the structural map (Fig. 69).

Stratigraphic sections were made wherever possible. Their locations are shown in the maps of enclosures 1 and 2, which also show the most important topographic names. However, there are not many exposures, owing to the vegetation on the shales or mudstones which form the predominant lithology. Only along the footpaths, roads and in valleys was there a possibility of measuring the rock sequences. All in all, twelve sections were made through (parts of) the Prioro and Pando Formations, most of them in the southern flank of the syncline (enclosure 1). In the Ocejo and Tejerina Formations four sections were studied (enclosure 2), three of which are badly exposed, while the fourth was described earlier by Wagner et al., 1969, who proposed this one as the stratotype section for the lower Cantabrian (Stephanian s.l.).

These sections were studied in much greater detail than could be drawn in the sections of enclosures 1 and 2 (scale 1:1,000), but from each formation or member a small part was drawn in greater detail (enclosure 3, scale 1:10 and 1:20) in order to show all separate layers, and the structures and fossils occurring in them. These observations are also included in the 1:1,000 sections, but here it cannot be seen in which individual layers the observations were made.

From six sections (4, 7, 10, 11, 12 and 13) rock samples were taken at stratigraphic distances of 5 m. Where a thin, deviating layer was encountered, which was not included in this way, a sample was also taken. From the other sections samples were only taken of the most interesting layers.

The region between the sections only yielded a number of more or less isolated exposures. Where possible, fossils were collected and sedimentary structures studied, and where necessary rock samples were collected. All together, more than 750 localities yielded rock samples and/or fossils. These localities can be reconstructed by comparing the sample numbers of the sections with the maps in the enclosures. The localities which do not occur in the stratigraphic sections are shown in the map of enclosure 2. Rock samples, thin sections of them and fossils have the same numbers as the localities in which they were found. These numbers are indicated in the text between brackets (e.g. 700).

Since marker horizons are absent and the lithology changes very rapidly in a lateral direction, correlation of the sections was not possible in any detail by considering the types of rock only. Sedimentary structures, the relative abundance of some of these, fossiliferous horizons, or levels with a specific faunal assmeblage, had to be taken into consideration. But even with the aid of this it was in many cases impossible to give exact correlations. Thin sections, which had been made of nearly all rock samples were of no value for correlation purposes except in a few cases, where a kind of mineral zoning existed. The thin sections were studied in detail, however, to detect microfossils and uncommon minerals that might give indications on the source areas of these sediments. Minerals were – unless stated otherwise – identified only microscopically by using identification tables and mineral descriptions by Tröger (1959) and Milner (1962).

By counting the minerals and calculating the percentages in the thin sections the rock samples could be classified petrographically. For this purpose a nomenclature was used as described in I.5. The percentages were arrived at by means of point-counting. In each slide 300 non-correlated points were counted, which gives fairly reliable results (Kalsbeek, 1969). The reliability, when using the results of a series of slides to establish the petrographic composition of a whole formation, mainly depends on two things: first, does this thin section give a representative picture of the entire interval from which the rock sample was taken, and second, is this way of counting sufficiently exact? It will be shown in II.3 that this is the case. For this reason, we shall give percentages of the rock types that we distinguished (I.5). These percentages are based on the samples from the fully sampled stratigraphic sections only.

These results by point-counting analysis are more reliable than those obtained by granular analysis in the laboratory. This latter method could not yield reliable values, since the mudstones are too hard to be pounded into the original individual grains. This will be shown in II.3.

The clay minerals were identified by means of X-rays. This method was used in a few other cases, too.

Numerous fossils were collected from all over our area. They were identified by several palaeontologists:

land plants	: Dr. R. H. Wagner (Univ. of Sheffield)
calcareous algae	: Dr. J. J. de Meyer (Univ. of Leiden)
brachiopods	: Dr. C. F. Winkler Prins (Geol. Museum, Leiden)
pelecypods (marine)	: Mr. H. W. J. van Amerom (Geol. Bureau, Heerlen)
pelecypods (non-marine)	: Dr. M. A. Calver (Inst. of Geol. Sciences, Leeds)
cephalopods	: Dr. J. Kullmann (Univ. of Tübingen)
foraminifers	: Dr. A. C. van Ginkel (Univ. of Leiden)
conodonts	: Mr. W. J. Varker (Univ. of Leeds)
trilobites	: Dr. J. Gandl (Univ. of Würzburg)
corals	: Dr. G. E. de Groot (Geol. Museum, Leiden)
crinoids	: Dr. A. Breimer (Vrije Univ., Amsterdam)
sponges	: Dr. W. J. E. van de Graaff (Univ. of Leiden)

By combining the results of the stratigraphic interpretations based on their identifications, it was tried to obtain an idea of the stratigraphy of this region based on as many fossil groups as possible. Part of the results have already been published (van Loon, 1971), but new data, which became available afterwards, made it necessary to make some slight alterations.

I.5. TERMINOLOGY

To avoid misunderstandings, we shall here give an explanation of some terms used, particularly those not normally used in literature.

Structural: the larger part of the sediments studied lie between the León line and the Monte Viejo fault (Fig. 69). In this part the sediments have roughly been folded into a syncline. Unless otherwise stated, we shall use the terms 'N-flank' and 'S-flank' for the flanks of this synclinal structure between these two fault systems.

Stratigraphy: stratigraphic ranges can be given in several ways. One possibility is a biostratigraphic zone. Other possibilities are ranges according to either the West European or the Russian standard. The most probable correlations are given in Fig. 3. For the sake of reliability, the range of some fossils will sometimes be given in this

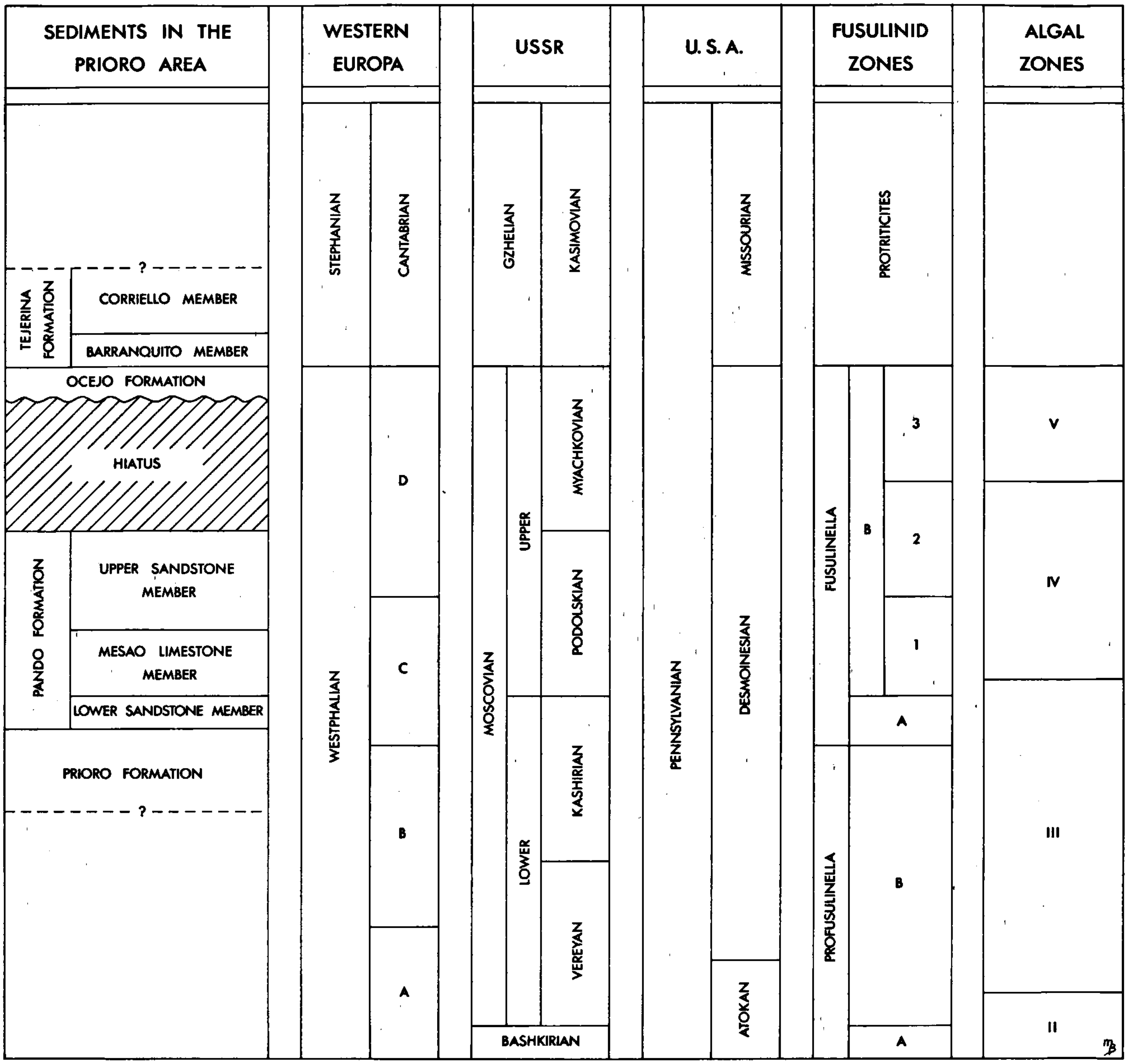

Fig. 3. Stratigraphic correlation chart, partly after van Ginkel (1965), Rácz (1965) and Wagner & Winkler Prins (1970).

way: brachiopods indicate Kashirian – Westphalian C/D. This means that the oldest species is known from levels that can be correlated with the Kashirian onwards, while the youngest species is known from deposits correlating with sediments that are dated, according to West European classification, as ranging to the boundary between Westphalian C and D. Note: Westphalian C/D means: the boundary between Westphalian C and D, while Westphalian C–D means: Westphalian C to (and including) Westphalian D. We shall use the term 'older deposits' to indicate all sediments dealt with in preceding chapters of this thesis.

Palaeontology: fossils were only collected for the purpose of dating the members and formations. The fossils collected that could be identified are given in lists for each member. The localities where they were found are added between brackets. We suppose that a more detailed sampling by a palaeontologist would result in an important extension of the lists of fossils.

Petrography: Several useful papers on petrographic classification were recently published (e.g. Travis, 1970). Since the siliciclastic sediments under study are, however, petrographically very immature, a simple classification was used according to van de Graaff (1971), who based his classification on Dott (1964) and Gilbert (1958). Compared with van de Graaff, the present author made a modification, however, concerning the percentage of matrix (Fig. 4). Microscopic study of the

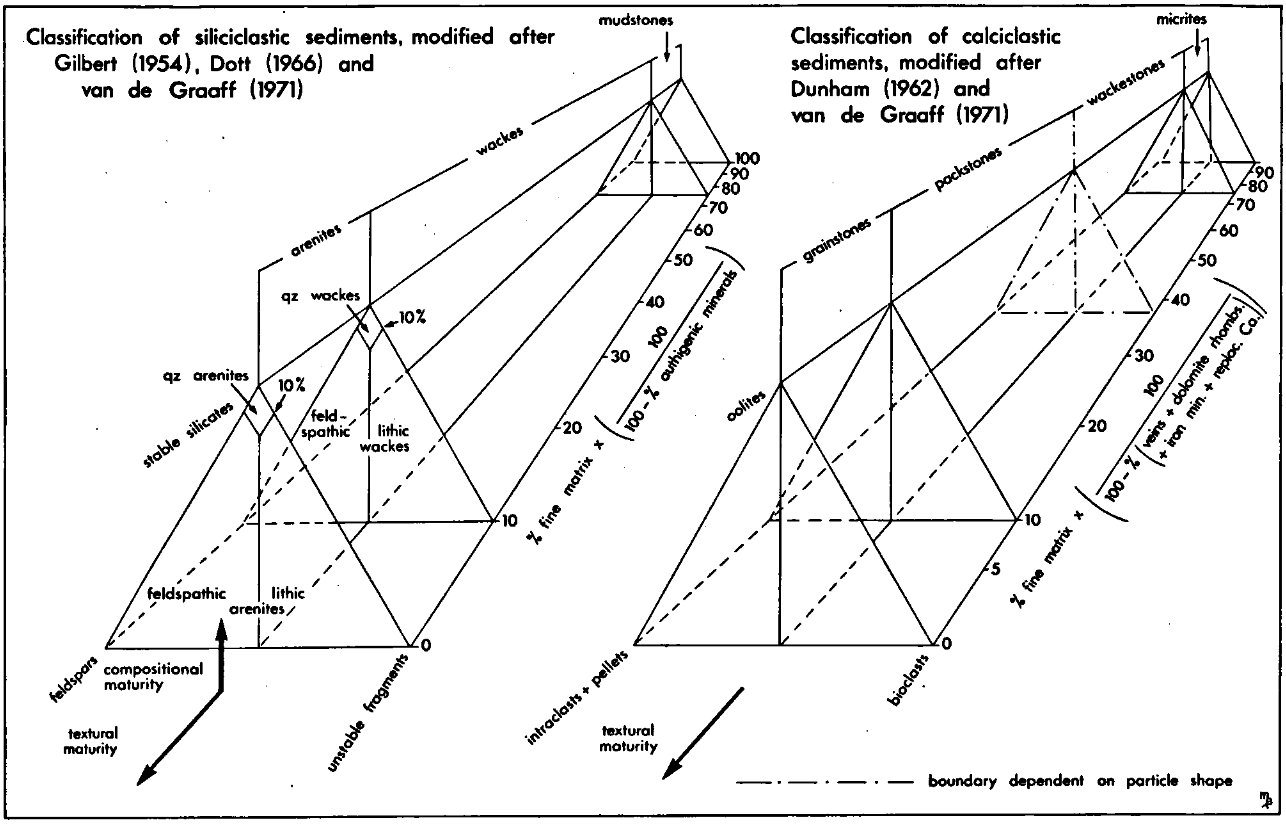

Fig. 4. Petrographic classifications used.

sediments usually does not reveal whether the matrix (or part of it) is clastic or authigenic. Moreover, in many cases it is unknown whether larger authigenic minerals are derived from larger grains or from matrix-sized material (cf. Whetten & Hawkins, 1970). It therefore seems best to suppose that these authigenic minerals were formed from matrix and grains in the same ratio as they occur now. Since we are interested in the original percentage of matrix we establish this percentage in the following manner:
% of matrix = counted % of matrix x{100/(100 – counted % of authigenic minerals)}.

According to the classification used the sediments could be divided, after the calculation of the percentage of matrix, into arenites, wackes and mudstones.

The carbonates were classified according to Dunham (1962), applying a similar calculation for the matrix. Sometimes a name was also given according to Folk (1959). This carbonate classification was adapted to the siliciclastic classification (Fig. 4).

Since it was impossible to obtain an idea of the percentage of matrix by means of a granular analysis (I.4), this percentage was also determined by means of point-counting. For practical reasons (the distance between the lines in the ocular) the boundary was laid at 23 microns virtually. As the grains, except for the micas, usually have a reasonable sphericity, although they are very angular, we may estimate the real boundary at about 25 microns, according to Friedman (1958). Other authors (e.g. Münzer & Schneiderhöhn, 1953; Chayes, 1950; Krumbein, 1950) give more or less similar calculations on empirical results for calculating the average size of spheres by measuring them in thin sections. Our method yields quite a reasonable result, for the boundary between grains and matrix is usually considered to be 20 or 30 microns. To avoid misunderstandings, we use the term 'matrix s.l.' in conglomerates to designate all material between the pebbles.

The following types of rock fragments were distinguished:

sandstone	: predominantly quartz grains with point contacts; some matrix
quartzite	: idem, suture contacts, usually less matrix
chert	: greyish, very small-sized parts with suture contacts
opal	: brownish, apparently non-crystalline
limestone	: micrite, sparry calcite, pelletiferous wackestone inter alia
phyllite	: clay-sized material with orientated authigenic micas
shale	: idem, authigenic micas not orientated
mudstone	: idem, but also silt-sized grains. No authigenic micas

clay gall : idem, apparently plastically deformed, rounded
clayflake : idem, platy
coalflake : piece of predominantly organic (here: plant) material; only present in the Cea Group

The quartzite fragments are usually easily recognizable, since the sediments studied are rarely quartzitic. Sandstone fragments, however, may have the same appearance as the surrounding material, especially when the sediment is coarse-grained. When the sediment is very fine-grained, clay flakes and galls, as well as mudstone fragments, shale and phyllite are hard to recognize (cf. Wang, 1968). The limestone fragments, as well as those with clay-sized minerals (except for the phyllites) most probably are intraclasts in the Yuso Group; in the Cea Group they may be derived from eroded older rocks.

The term 'iron minerals' is used to indicate iron oxides and/or hydroxides, viz. goethite, hematite, magnetite, ilmenite, wustite, lepidocrocite and amorphous matter.

Point-counting: Since even the grains in the Prioro and Pando Formations (and in the mudstones of the younger formations) are rather small, it was not possible microscopically to distinguish quartz and potash-feldspar. Since it is known that the potash-feldspar is scanty or absent in Upper Carboniferous deposits in this mountain chain, all these grains were counted as quartz. A percentage of the potash-feldspar could therefore never be given. To check, however, whether the percentage is indeed negligible, as assumed, some coarser grained samples were coloured, and some X-ray analyses were made. Nowhere were feldspars distinctly shown in the first way, but X-ray analyses revealed their presence in the fraction smaller than 2 microns of most samples. Plagioclase was also found frequently in this way. All our arenites and wackes are, however, lithic according to the classification used.

The percentage of the fossils was given according to their entire surface in the slides as defined by their outlines. When filled with clastic or authigenic material, these parts were also counted as fossil. When a fossil is recrystallized this new substance was not considered as authigenic material, but still as a fossil.

When two minerals, lying upon each other, were hit upon during the counting, only the rarer mineral was taken.

I.6. ACKNOWLEDGEMENTS

I wish to express my gratitude to all those who contributed to my thesis with either geological information or technical help. They are so many that it is impossible to mention them all here. However, I want to make a few exceptions. In the first place, I wish to thank Prof. J. D. de Jong for his enthusiastic assistance in the field and for all the discussions we had. Secondly I want to thank Prof. A. J. Pannekoek for his valuable criticism of my manuscript and for his many very helpful suggestions. I am also much indebted to Miss E. van der Wilk who spent much of her time helping me when petrographic problems arose. All specialists in palaeontology mentioned in I.4, who identified the numerous fossils, merit grateful acknowledgement. I am most grateful for the financial assistance afforded during the larger part of my investigation by the Netherlands Organisation for the Advancement of Pure Research (Z.W.O.).

CHAPTER II

PRIORO FORMATION

II.1. INTRODUCTION

The Prioro Formation, S of the León line, is only exposed in this area in the S-flank and in the eastern bend of the syncline. The name of this formation was derived from the village in the centre of this formation.

The maximum thickness is unknown, because the contact with the underlying formations is nowhere exposed, and a fault system (the Monte Viejo fault, probably accompanied by some parallel faults more to the north) divides this formation into two parts, of which it is not exactly known how large the onlap is (Chapter IX), although this may be only a few tens of metres. The occurrence of many small faults and of cleavage makes it still more difficult to estimate the original thickness. The part below the Monte Viejo fault (to be referred to further on as 'the older part') is nearly only exposed in isolated places. For structural reasons the thickness of this part is estimated at between 200 and 500 m.

The upper part, which is much better exposed, can be defined as the part of the Prioro Formation indicated in section 7 (reference section of this formation) of enclosure 1, where it has a thickness of 470 m.

The older part could hardly be sampled, but in the upper part three sections were sampled in detail:

section 4: 61 samples
section 7: 84 samples
section 11: 75 samples

These 220 samples were used for the numerical petrographic data. Additional data were obtained from 30 other slides. The following localities belong to this for-

Fig. 5. Lithology typical of the older part of the Prioro Fm. Mainly shales, with some sandstone turbidites. Overturned position. Loc. 590.

mation: 1–75, 110–193, 267–327, 589–592, 603–604, 608, 610, 651–652, 658–668, 696–699, 711, 725–731 and 764–766.

II.2. LITHOLOGY

In all exposures the older part appears to consist of shales (this is a field name; according to our petrographic nomenclature mudstones and wackes), in which several coarser sandstone layers (wackes) occur (Fig. 5).

The upper part consists almost entirely of shales (Fig. 6). In this part, however, coarser elements also occur:

Fig. 6. Lithology typical of the upper part of the Prioro Fm. Succession consists of graded shale layers, 20 cm thick on an average. Loc. 610.

some sandstones as well as conglomerates and pebbly mudstones. These are not distributed evenly over the entire area, but are often concentrated in certain places. The shales, however, remain predominant everywhere (II.3).

Because of the vegetation this part is hardly exposed in the westernmost part of the area. Only some very small, fully isolated exposures occur, showing the characteristics of the Prioro Formation. But the fact that there is so much vegetation here (mainly pasture and thicket) gives the impression that no or few sandstones and conglomerates, being more resistant and less overgrown, are present.

Only in the valley of section 2 does the Prioro Formation show more sandstones, while in the upper part conglomerates and pebbly mudstones are also relatively abundant.

In sections 1, 3 and 4 successively, the quantity of coarse material decreases, and in sections 6 and 7 is even negligible. In section 11, however, there is once again much coarse material (this already begins in the area between sections 7 and 11), usually pebbly mudstones and conglomerates, and to a lesser degree also sandstones.

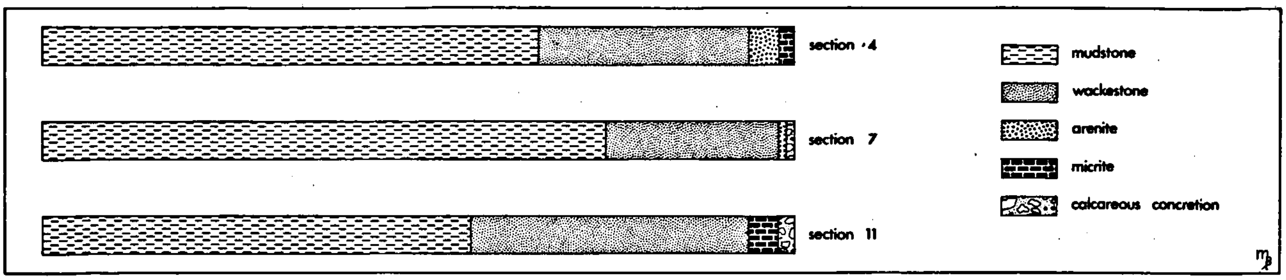

Fig. 7. Distribution of rock types in the Prioro Fm.

Pure limestones do not occur in this formation here, although there are some rather muddy micrites, in which slightly more carbonate than siliciclastic material is present, when studied microscopically. Sandstones with a carbonate cement also occur. This might indicate the presence of limestones elsewhere in the basin. As we shall see (II.3), some of these layers contain fossils that can be compared with those in the Pando Formation (IV.6). This probably indicates a situation comparable to the contemporaneous deposition of 'Kulm' and 'Kohlenkalk' in NW Europe (Paproth, 1969). The picture given by Weiler (1963) is also very similar to our situation.

The predominance of mudstones is expressed in the topography. This formation forms the lowest part of this area (1075 m). All larger hills in this area in which this formation is exposed have some more resistant material at their tops, small residues of conglomerates of the Ocejo Formation, or pebbly mudstones and small conglomerate lenses of the Prioro Formation itself.

II.3. PETROGRAPHY

From sections 4, 7 and 11 we investigated 187 samples in thin sections. Of two samples the shales were crumbling too much to allow the preparation of slides. Taking these two into account, the 189 samples (31 slides of pebbles are out of scope here) can be divided petrographically as follows:

micrites	: 2 %
calcareous concretions	: 1 %
mudstones	: 68 %
wackes	: 27 %
arenites	: 2 %

The percentages for the three sections separately are shown in Fig. 7. We see that 3 % is carbonate and 97 % siliciclastic material, mainly very fine-grained. One could say that the larger part of this formation consists of matrix. Grain-size measurements are not in agreement, however. We obtained the following results for some checked samples (in percentages):

sample no.	1	74	110	178	269	312
<25 microns (point-counting)	49.0	28.3	81.3	87.3	53.7	89.7
<32 microns (grain-size analysis)	18.3	19.8	15.3	35.2	17.0	19.5

These results distinctly show that these sediments are too hard to allow reliable grain-size analyses. For this reason all percentages were calculated from point-counting analysis. We obtained the following result:

1. abundant (more than 10 %):
This group only contains the matrix (70.7 %) and the quartz grains (13.1 %). Considering the matrix we must bear in mind that this is formed by many minerals and some organic matter. X-ray analysis revealed the presence of the following clay minerals: illite (being abundant) and chlorite (nearly always present), sometimes septachlorite. Irregular 14 Å mixed-layers, probably containing montmorillonite, are very rare.

Most of the slides contain 80–95 % of matrix (Fig. 8). The relatively few slides of the coarser samples are responsible for the much lower average value of 70.7 %.

2. normal (0.1–10 %):
This includes the group of authigenic minerals (9.2 %), the composition of which will be described in II.4. In addition, the group of rock fragments (3.4 %), muscovite (2.0 %) and biotite (0.9 %). On account of slide 160 (from the probably only shallow marine part of this formation as will be shown later), which contains 25.7 % of fossils (*Hemifusulina* sp.), the group of fossils also belongs to the normal constituents (0.2 %).

3. rare (less than 0.1 %):
Anatase, apatite, augite, chloritoid, chlorite, calcite, epidote, hornblende, opaque minerals (several), plagioclase (2 different kinds: one is angular albite, the other is a usually prismatic albite (?) which sometimes is slightly rounded), potash feldspar, rutile. staurolite, sillimanite, tourmaline and zircon.

4. negligible (less than 10 grains observed):
Antigorite, brookite, clinochlore, dolomite, kyanite, riebeckite, spinel, tremolite and vesuvianite.

Most of the minerals of groups 3 and 4 belong to the heavy minerals, probably because the latter are easily detected. The total amount of heavy minerals is 0.09 %, more than 95 % of which belong to the resistant species zircon, tourmaline, rutile and, to a lesser degree, epidote.

It is a pity that so few data are available on the sedimentary petrography of the older formations in the sur-

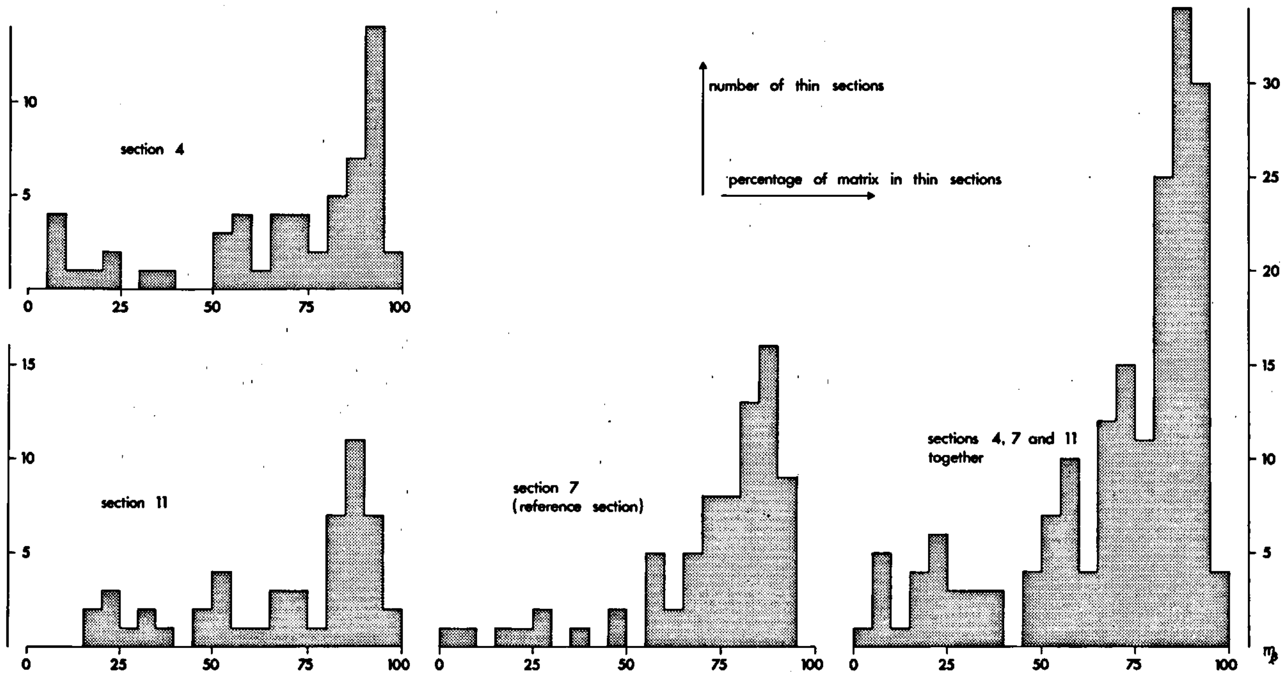

Fig. 8. Modal distribution of the matrix in the Prioro Fm.

roundings. Several minerals show such remarkable characteristics that they should be able to give reliable data on their provenance, thus facilitating a reconstruction of the palaeogeography. But it is possible to make conjectures concerning the source areas of some of these minerals. The Cambrian Herreria Formation is most probably the source of the quartz grains with zoned gas or fluid inclusions (Oele, 1964), possibly earlier derived from gneisses in Galicia (Comte, 1959). The Barrios Formation (Ordovician) may be responsible for some sillimanite grains (Oele, 1964), but another possibility will be shown below. Several minerals that can easily be detected in thin sections (e.g. staurolite, various colour varieties of the tourmaline) have never been described from older rocks, however.

There is also a problem concerning the presence of certain metamorphic minerals. Staurolite and kyanite, for instance, are formed during a degree of metamorphism unknown of sediments in the surroundings. Even the known parts of the late Precambrian (?) Mora Group have a lesser degree of metamorphism. Unless we assume that underneath the meseta (a vast region with a mainly Tertiary cover, that spreads out S of this mountain chain) more metamorphosed older Palaeozoic or even Precambrian rocks occur (de Sitter, 1962) (a speculation, that as yet cannot be based on facts), the nearest source area for these minerals would be Galicia, some 200 km or more to the W. Because of this distance, these minerals would most probably have arrived as rounded grains. The shape of the minerals (even a kyanite twin occurs!) makes this long transport improbable. However, Leguey & Rodriguez (1970) argued that this transport must have been possible: investigating heavy mineral samples from the N–S trending river valleys in the southern Cantabrian Mountains, they found that, from W to E, the quantities of staurolite, sillimanite, kyanite, andalusite and garnet decrease, while the grains of the more resistant species (zircon, rutile and tourmaline) become smaller and better rounded. This makes source rocks in the W the most probable. They also mention that chloritoid, staurolite and andalusite are frequent in the Lower Ordovician and Upper Gothlandian (Silurian) in the vicinity of the Galician massive, while sillimanite occurs in the Ollo de Sapo Formation, which is probably of Precambrian age, but in which the sillimanite is Hercynic (Upper Namurian, Capdevila, 1969). However, other formations in Galicia also contain sillimanite (Capdevila, 1969) and may have served as a source rock.

Although they suppose that the minerals found in the rivers were derived from Tertiary sediments, we believe that heavy minerals were supplied by the igneous rocks of Galicia as early as during the Carboniferous.

In this way the occurrence of most of the other minerals, both igneous and metamorphic, may also be explained. The possibility cannot be excluded, however, that all these sediments were derived from the larger Armorican massive elsewhere, including e.g. Brittany (Koning, pers. comm.; Ferm, in press).

Concerning the possibility of staurolite grains being transported over such long distances, it should be mentioned that it is not impossible that this mineral is much more resistant than is normally assumed (de Jong & Poortman, 1970). This mineral has even been found in Westphalian deposits more to the E (Pisuerga basin; Nossin, 1959).

For the same reason the occurrence in rather large quantities of biotite (or weathering products of biotite) is difficult to explain. It is true that this mineral occurs in igneous rocks, which are found as isolated small intrusives in several places, but it still remains a question whether these could yield such a large quantity. Another point is that these intrusives are considered to be of Westphalian D age, which is somewhat younger than this formation (II.7), but the reasons for this dating of the intrusives are not known to the present author. Perhaps they are a little older than is assumed.

As mentioned above, the Prioro Formation not only consists of mineral grains, but also contains a considerable amount (3.4 %) of rock fragments of all sizes. These fragments are usually angular and occur in the following percentages:

quartzite	: 27.6 %
phyllite	: 22.8 %
mudstone	: 22.2 %
shale	: 17.4 %
chert	: 5.0 %
sandstone	: 5.0 %

The remainder consists of limestone, clay galls and clay flakes.

Rock fragments gradually pass into pebbles (2–64 mm), which are the constituents of pebbly mudstones and conglomerates, but the quantity of pebbles cannot be expressed in percentages for want of a reliable method of investigation in this respect.

In the group of pebbles the quartzites are even more abundant than among the rock fragments: over 99 %. Sporadically some sandstone and limestone pebbles occur, the latter usually less rounded than the former. As well as some of the minerals mentioned above, there are a number of very characteristic quartzite pebbles which, however, could not be traced to the source rock owing to lack of knowledge of the older formations. We shall still indicate a tentative source rock, mainly based on observations of our own and on comments by students of our university working in this mountain chain on the older formations.

It is our intention to give more data on the pebble content in a future paper. Now it seems sufficient to mention the occurrence of two types of limestone, three types of sandstone, one type of mudstone and five types of quartzite pebbles. These were probably derived from, among others, Precambrian, Cambrian, Ordovician, Devonian and Carboniferous formations.

Although predominance of quartzite pebbles is a common feature in conglomerates (e.g. Williams, 1969; Cailleux, 1952) it is remarkable in this formation, since many other rock types are present at the León line (see geological map by Rupke and Helmig, 1965), which is supposed to be the area that supplied most material (II.8). Another interesting fact is that these quartzite pebbles are well-rounded or, at least, subrounded, although some of them are tectonically deformed (in the manner of Vargas et al., 1969). All these observations together lead to the conclusion that these pebbles have either been sorted and rounded in a strongly agitated environment (beach?) or are derived from an older conglomerate. There is, however, no evidence of the existence of such a conglomerate (the only thick older conglomerate known is the Curavacas conglomerate that, however, is at least partly time-equivalent to the Prioro Formation itself, and probably was not present in this area), so that we are inclined to believe that during the time of deposition of this formation a strongly agitated environment existed near the León line.

Carbonates are very scarce in this formation. They consist of micrites and calcareous concretions, which latter petrographically must also be called micrites, a normal feature in this kind of sediments (e.g. Tanaka, 1970). Only in one place (604) was a fossiliferous packstone found.

The concretions contain rather a lot of siliciclastic material (3.5–33 %) and may also contain dolomite rhombohedra (up to 47 %, slide 171).

The clastic carbonates usually contain much clay and iron oxides or hydroxides, and are poor in fossils, except for wackestone 604 (5.5 %). The average composition of the micrites is shown in Fig. 9.

It was already mentioned (II.2) that the proportion of coarser and finer material is not the same in all places. This remark was made for the macroscopic lithology there. It appears possibly to also make a division in the three sections based on the grain size of the samples as visible in the slides.

Five parts may be distinguished:

5. a coarse-grained part.
Section 4: samples 316–327 (mainly wackes, some mudstones, few arenites)
Section 7: samples 180–193 (wackes)
Section 11: sample 74 (wacke; here only the top sample)
4. a fine-grained part.
Section 4: samples 302–315 (mudstone)
Section 7: samples 161–179 (mudstone)
Section 11: samples 68– 73 (mudstone)
3. a relatively coarse-grained part.
Section 4: samples 300–301 (mudstone and wacke; a relatively thin level)

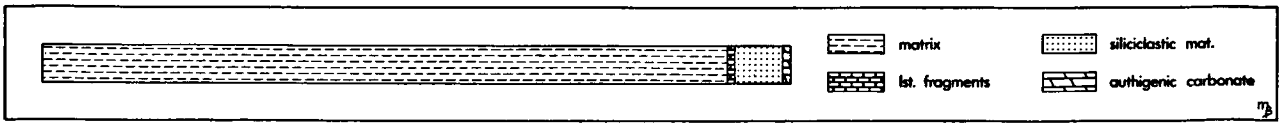

Fig. 9. Average composition of the limestones in the Prioro Fm.

Section 7: samples 145–160 (mudstones and wackes, some arenites)
Section 11: samples 42– 67 (mainly wackes, some mudstones)
2. a fine-grained part.
Section 4: samples 281–299 (mainly mudstones, some wackes)
Section 7: samples 119–144 (mudstones)
Section 11: samples 16– 38 (mudstones)
1. a relatively coarse-grained part.
Section 4: samples 268–280 (mainly wackes, some mudstones, few arenites)
Section 7: samples 110–118 (mainly mudstones, some wackes)
Section 11: samples 1– 15 (mainly wackes, some mudstones)

Although not equally distinct everywhere, this division into five parts, based on grain size, is possible in all three sections. Since, as will be shown further on, it is probable that the sediments in section 11 were supplied by another agent than those of sections 4 and 7, these changes in grain size must have been caused by a mechanism active in the whole region. The most probable assumption is that three periods of relatively rapid uplift in the hinterland alternated with two periods of slower (though still rather considerable) uplift.

In addition, some minerals do not occur in all parts of the sections. Sometimes they are only present in certain parts of the stratigraphic column, while their first appearance in the three sections occurs at apparently different stratigraphic levels. This is clear for hornblende and the angular albite and, to a lesser degree, for a special kind of clastic chlorite. In Fig. 10 it is clearly shown that these minerals appear considerably earlier in section 7 than in section 4. Because of the different supplying agent the picture in section 11 deviates. The conclusions that can be based on these mineral zones will be dealt with in II.8.

Here it is necessary to question whether this zoning is based on reliable observations. This problem was already mentioned in I.4. Here we shall give some examples, showing that our results are sufficiently reliable for our purpose.

First question: is the result of point-counting a reliable reproduction of the mineralogical composition of the handpiece? To solve this problem we made several thin sections of some handpieces. The results of 8 slides, taken from rock sample 651, an arenite with a calcite cement, are given in Table 1.

The differences between the percentages counted are rather small, so that it may be assumed that the percentages in one slide give a rather good impression of the petrographical composition of a rock sample (cf. Kalsbeek, 1969).

The second problem is whether a rock sample may be considered representative of the whole interval of 5 m from which it was taken. Since more samples were taken when clear differences in rock type could be observed in the field, this problem only refers to apparently more or less homogeneous intervals. To investigate this, samples were taken in such an interval. In section 11, for instance, sample 1 is the most resistant (coarsest?) and sample 2 the least resistant (finest?) piece, according to field observations. Both, however, look like very similar mudstones. Their composition, as calculated from point-counting, is shown in Table 2.

Although two samples cannot give more than an indication, the percentages found seem sufficiently similar to allow a rock sample to be considered a fair representative of its interval.

It should be noted that there appears to be a relationship between the percentages of some of the petrographic constituents distinguished here.

The most striking relationships are:

1. When the percentage of quartz increases, the percentage of the rock fragments also increases. This is shown

slide	matrix	quartz	rock fr.	fossils	heavy min.	authig. min.	remainder
651 a	6.7%	43.0 %	8.0 %	1.3 %	0.0 %	40.0 %	1.3 %
651b	5.0 %	46.7 %	7.0 %	2.7 %	0.3 %	35.3 %	3.0 %
651c	4.3 %	41.0 %	11.0 %	2.3 %	0.0 %	36.3 %	5.3 %
651d	5.0 %	43.7 %	11.3 %	3.7 %	0.3 %	35.7 %	3.0 %
651e	4.7 %	43.3 %	8.0 %	3.0 %	0.0 %	39.3 %	1.7 %
651f	3.0 %	46.0 %	11.7 %	1.0 %	0.3 %	35.0 %	3.0 %
651g	1.7 %	41.3 %	4.0 %	2.7 %	0.3 %	45.3 %	4.7 %
651h	3.3 %	46.0 %	5.7 %	2.3 %	0.0 %	37.3 %	5.3 %

Table 1.

no.	matrix	quartz	rock fr.	muscov.	biotite	heavy min.	auth. min.	remainder
1	49.0 %	21.7 %	1.3 %	2.7 %	3.7 %	1.3 %	20.3 %	0.0 %
2	54.3 %	19.3 %	0.7 %	3.3 %	3.3 %	0.3 %	17.7 %	1.0 %

Table 2.

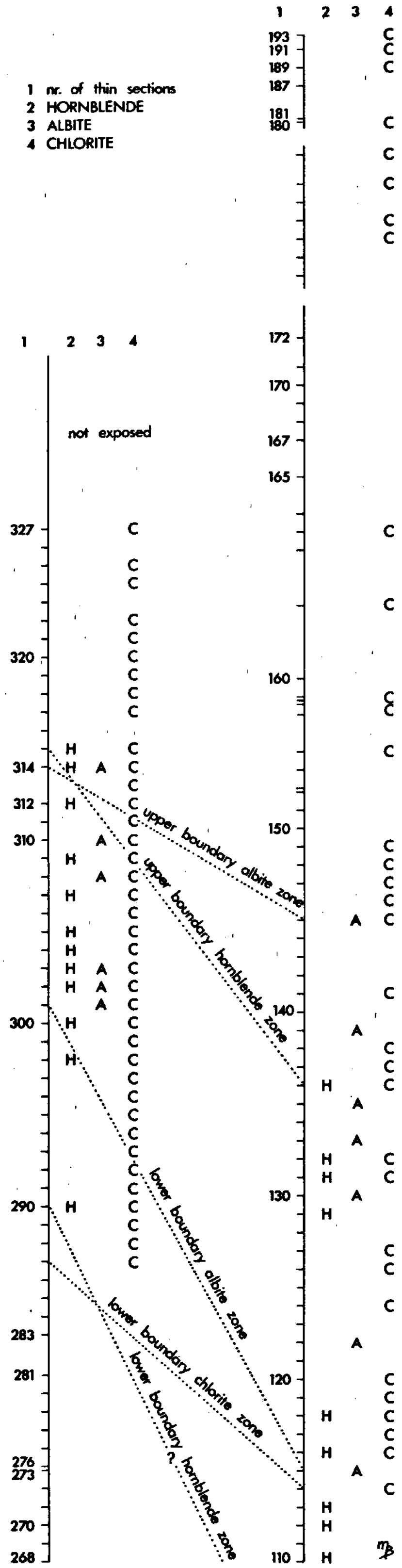

Fig. 10. Some mineral zones in the Prioro Fm. Sections 4 (left) and 7 (right).

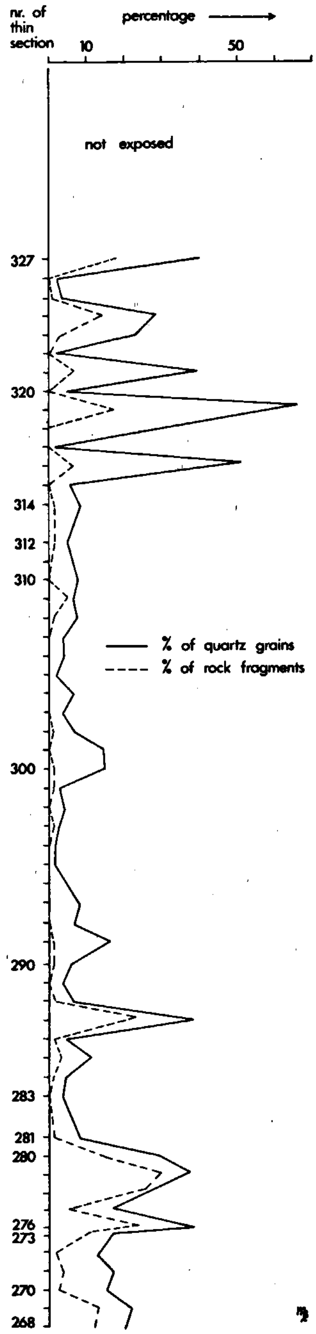

Fig. 11. Relationship between the percentages of quartz grains and rock fragments in section 4 of the Prioro Fm.

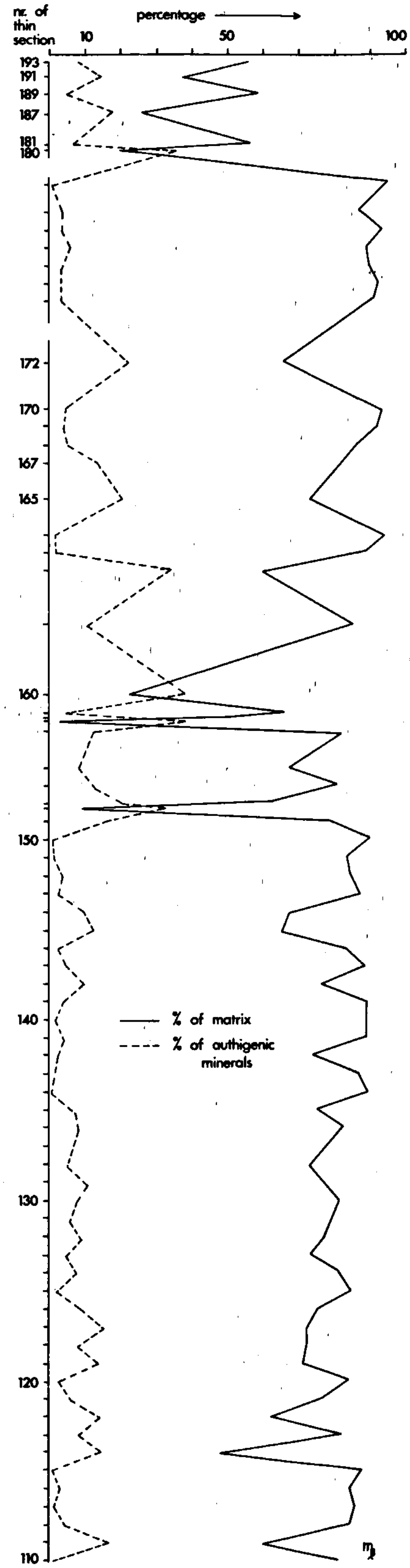

Fig. 12. Relationship between the percentages of matrix and authigenic minerals in section 7 of the Prioro Fm.

in Fig. 11 for section 4. It seems reasonable to explain this by the energy of the transporting medium, since these two components usually form the coarsest material.

2. There is a relationship between the percentages of muscovite and biotite, which can be explained either by a derivation from the same parent rocks or by similar hydrodynamic behaviour. Of these two minerals too few grains occur, however, to investigate which of the two assumptions is the most probable.

3. Although only a small number of heavy mineral grains were counted (0.09 %), they seem to be most frequent in slides with large quantities of quartz. Here, too, the transporting power may be responsible.

4. The percentage of authigenic minerals decreases when that of the matrix increases (Fig. 12). This relationship will be dealth with in II.4.

II.4. DIAGENESIS

The diagenetic phenomena belong to few types only and they are not of frequent occurrence either. This is most probably mainly caused by usually small grain size (68.5 % is mudstone), since with a decreasing quantity of matrix an increasing amount of diagenetic phenomena can nearly always be seen. This relationship is very clear in all three sections and is illustrated in Fig. 12 for section 7. It should be noted that in the percentage of authigenic minerals possible matrix-sized authigenic minerals are not included, since these are not recognizable as such. It is, however, probable that part of the matrix is authigenic, for this can at least be proved for some of the micas that are only slightly larger (30–40 microns). Other micas are definitely clastic. Many micas and quartz grains are orientated (Fig. 13). Since the angle between bedding plane and cleavage is usually very small (Chapter IX), it is often difficult to establish whether they were orientated during deposition or formed as a result of pressure. Both possibilities seem to be present, however, indicating that these sediments may

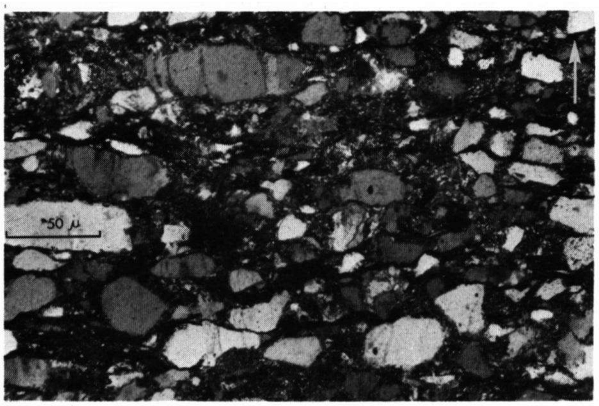

Fig. 13. Wacke with orientated grains. Note the angularity and low sphericity of the quartz. Thin section of slide 321. Nicols crossed.

already be called slightly metamorphic. The clay minerals present (II.3) are not in contradiction. In the following chapters we will see that a distinct difference exists in clay mineral content between the sediments of the Yuso and Cea Groups, probably originating from differences in diagenesis (or metamorphism).

The authigenic minerals consist for more than 90 % of iron minerals (mainly goethite and hematite). There is, however, a problem connected with this percentage to be dealt with later. The other authigenic minerals are chiefly quartz (usually in veins; sometimes as secondary quartz in slightly quartzitic arenites) and calcite (in veins; often a replacement of earlier crystallized quartz. In a few cases it was observed that this calcite in its turn has been partly replaced by quartz, resulting in authigenic quartz with a core of calcite).

Both the quartz and the calcite not only replace each other, but other minerals may also be affected. In this way muscovite may be replaced, but this may occur as well with more resistant minerals such as zircon (e.g. slide 1) and tourmaline (slide 157). Quartz (rarely) and calcite (in calcareous samples, e.g. 157 and 651) may also serve as a cement.

Apart from those mentioned above, there are only few authigenic minerals. This mainly concerns chlorite and muscovite. The muscovite in its turn has sometimes been replaced by microcrystalline chert (7). Since small quantities of muscovite frequently occur in chert fragments, it does not seem impossible that these cherts, classified as 'rock fragments', are at least partly authigenic replacements of muscovite clusters. Biotite may also be replaced by muscovite. More often, however, biotite is weathered to chlorite. All stages between pure biotite and chlorite were observed. Biotite may apparently disappear by hydratation. The colour changes during this process from rather dark brown (pleochroitic to dark yellow) to light yellow or even colourless, fully losing its pleochroism.

Most striking besides these is authigenic anatase, bright yellow, very small (10–30 microns) rhombohedra of which occur in many slides. It could not be determined from which of the minerals present the Ti required has been derived. The clastic anatase and rutile do show no signs of solution nor do they show secondary growth.

Authigenic pyrite can often be found in slides containing organic material, particularly within shells (cf. Tanaka, 1970). We therefore believe that a reducing environment existed (or possibly only micro-environments around decaying organisms). Fossil fragments show other replacements too, such as silicification (sometimes opalizing) and replacement by iron minerals (3, 61). These changes may result in a shell with a core of original calcite, around which a silicified zone can be found, surrounded by an outer wall of iron minerals. The silicification of fusulinids seems to proceed more easily along the walls than along the chambers, which have been filled with sparry calcite (Fig. 14). This might indicate that the calcite, driven out by the silica, was concentrated in the still empty chambers, where it crystallized as sparry calcite. Afterwards, part of the silicified outer part was replaced by the iron minerals.

The origin of the iron minerals seems to be rather recent. In some rock samples it was observed that in the part which had been exposed to the surface a zone exists apparently with a large quantity of iron minerals, while in a fresher part this percentage is considerably smaller. This indicates that the presence of the iron minerals is mainly caused by more or less recent weathering at the surface. Rain water penetrating deeper via the well-developed cleavage may be responsible for these minerals in parts that have not been directly exposed to the surface.

It should be noted that it was shown by means of point-counting that more than 90 % of the authigenic minerals consist of iron minerals. These percentages, however, were not supported by chemical analysis. The total iron content calculated from this analysis (thus including the quantity of iron in other minerals than the oxides and hydroxides) was usually much smaller than that obtained from the point-counting. This may be due to the fact that only a small quantity of the iron oxides and hydroxides is sufficient to hide other grains, so that point-counting gives too high values. For this reason, the percentages obtained from the laboratory appear to be more reliable. This is confirmed by the percentages themselves, since with the point-counting method the percentages were much higher than was to be expected. Twenty samples were checked in this way (Table 3).

sample	percentage of iron minerals	
number	by point-counting	by chemical analysis
2	17.7	7.73
42	27.7	7.53
43	23.0	6.40
48	21.7	11.55
61	7.0	6.36
73	2.3	6.19
74	6.0	5.74
110	1.0	5.94
130	8.3	6.71
150	2.0	7.98
166	5.7	16.90
172	23.0	9.49
178	4.3	7.68
189	4.7	6.13
269	10.3	5.67
279	28.0	4.83
289	2.7	5.72
312	3.0	6.46
316	29.0	7.61
661	24.0	3.49

Table 3.

It appears that in samples with few iron minerals a reasonable similarity exists between the point-counting and chemical analysis results, but that the percentage of iron minerals as found by point-counting seems to be exaggerated when the percentage itself is really high.

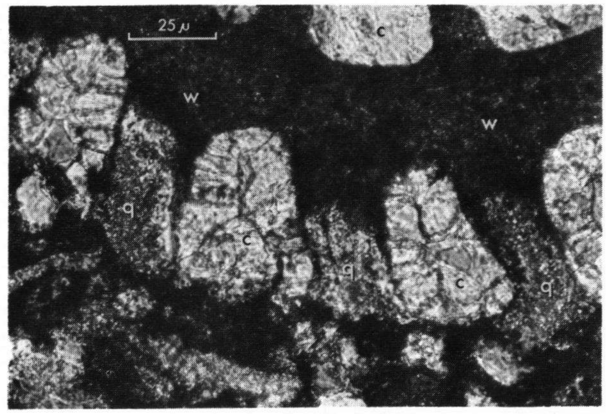

Fig. 14. Thin section of sample 160, showing part of a *Hemifusulina*. The calcitic wall (w) is replaced by authigenic quartz (q) at the outer rim, while the chambers are filled with sparry calcite (c).

As far as diagenesis is concerned, one more layer is important (35), for big concretions (up to about 20 cm) occur here with large authigenic pyrite cubes (1–2 mm). The ground mass (by colouring) turns out to consist for the most part of ankerite, while some siderite is also present.

Apart from all the authigenic minerals mentioned above, brookite, clinochlore and dolomite were also found sporadically.

II.5. SEDIMENTARY STRUCTURES

As stated in I.5, the lithological units will be interpreted on the basis of sedimentary structures and possible other data. For this formation it seems best to distinguish (field names) shales (a), sandstones (b) and pebbly mudstones and conglomerates (c).

II.5a. *Shales*

It was already stated (II.3) that micrites are very rare. They always show the same characteristics as the first type of shale that we shall describe here. Only their lime content is higher. For this reason we can refer to the description of the shales, as far as micrites are concerned.

The shales (petrographically mudstones and wackes) are usually very monotonous, with respect to the petrography as well as by the apparent lack of sedimentary structures. The scarcity of these structures is not a real fact, but is caused by the uniform grain size which conceals them. In favourable circumstances, however (rain water flow along bedding contacts), selective erosion may take place, showing structures that allow us to distinguish two types of shales.

II.5.a.1. *Graded shales.* – In this case the apparently homogeneous mud masses are composed of great numbers of graded beds with an average thickness of

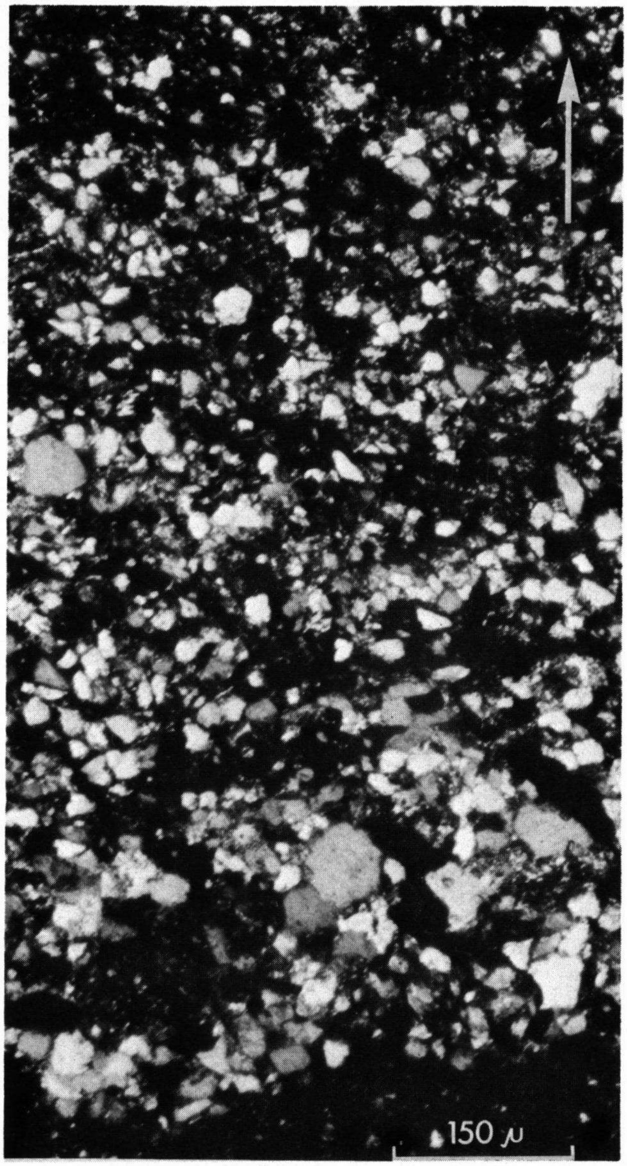

Fig. 15. Very thin graded layer, showing load casting at the base. Thin section of sample 155. Nicols crossed.

about 20 cm, with sharp and flat lower bedding planes (Fig. 6). The grading is usually so slight that no difference in grain size between base and top can be detected in the field, although the base is clearly more resistant. In thin section, however, it can be shown that the percentage of the matrix increases towards the top (in most cases with about 5–20 %, e.g. from 85 to 95 %), while the maximum grain size of the quartz decreases a little (e.g. from 60 to 40 microns). These characteristics indicate deposition by turbidity currents with a distribution grading, which indicates low-concentration currents (Allen, 1969). A low concentration here means that the suspended material forms less than 20 % by volume (Middleton, 1967).

The small differences in grain size explain why sedimentary structures can hardly be seen in the field. In the slides wavy lamination and scarce current ripples of very small dimensions may sometimes be visible by orientation of oblong grains. Very thin graded parts (450 microns) can also be shown. In slide 155 (Fig. 15) it can be demonstrated that these graded 'laminations' are not caused by whirling up and resettling of material deposited earlier, but by an allochthonous supply in suspension. At the base, which even in spite of its small thickness still shows erosion and loading, such a high concentration of heavy minerals occurs that this cannot be brought about by winnowing but only by exceptional supply. The erosion, too, proves that the material did not only settle from a suspension in quiet water.

On rock surfaces polished by water (cf. Walker, 1967b) a parallel lamination can nearly always be seen above which small current ripples are sometimes visible. Even though our best exposures were only a few metres long, this lamination could nowhere be traced over the whole exposure, so that the mode of deposition must have differed considerably from the one that caused the lamination traced over miles in a flysch basin, as found by Sujkowski (1957).

In relatively few cases the beds, once again, show a level with parallel laminations above the current ripples, with thinner laminae than nearer to the base. One rarely finds only the ripples with the parallel lamination above. The ripples tend to be climbing (ripple-drift cross-lamination), though the angle is usually small, indicating a high flow regimen (Walker, 1969). Nowhere does this flow regimen seem to have been sufficiently high, however, to cause structures that were interpreted by Walker (1967a) as antidunes, or dunes (Allen, 1970). Present structures can hardly ever be seen when the sediment is not polished.

Summarizing, these shales show the following properties:

1. thickness of the beds remains constant as far as they can be followed,
2. straight lower bedding planes,
3. distribution grading,
4. parallel lamination / small current ripples / very fine parallel lamination / or part of this sequence,
5. very fine-grained material, but texturally immature due to the presence of much larger grains,
6. compositionally very immature (II.3),
7. fossils absent (II.6), in sporadical cases present, often fragmentary,
8. very sporadical presence of groove and flute casts.

On the basis of these properties it seems justified to consider these deposits as originating from turbidity currents. Perhaps they should be called laminites (Lombard, 1963) rather than turbidites, because of their grain size and since parallel lamination is the most frequent structure. They may not be called distal, however, because of

Fig. 16. Contorted mud mass, containing a few quartzite pebbles (p). In the mud, a contorted parallel lamination (l) is sometimes still visible. Loc. 658.

the sole marks (though these are very scanty), the predominance of the a and b intervals (Bouma, 1962) and their intercalation between pebbly mudstones and conglomerates. The ABC-index (Walker, 1967b; Walker & Sutton, 1967) would therefore give a wrong picture here. The small number of sole marks can be explained by a relatively small erosional power, caused by too low a current velocity or a traction carpet (Dzulynski & Sanders, 1962), while it can also be explained by the impossibility to detect them, since no selective erosion has made them visible. Moreover, only very few lower bedding planes are exposed, and never more than a few decimetres.

The rather proximal deposition, the grain size (silt is the most common) and this sediment itself, together with those described below, give indications of a prodelta environment (Scruton, 1960; Lineback, 1968).

II.5.a.2. *Contorted and structureless shales.* – The second type of shale, too, is only recognizable by favourable weathering. These shales (or mudstones) then show a chaotic structure, in which contorted remains of older bedding planes (recognizable by their parallel lamination) can be detected (Fig. 16). This kind of sediment can pass both laterally and vertically into mud masses in which no structure has been left (cf. Lindsay, 1966). These sediments are interpreted as having been deposited by slumps and mudflows respectively (cf. Reading's (1970) facies type vii). It appears that they are more often present between the conglomeratic parts of this formation than between the turbidites or laminites described above. They may therefore be more proximal than these latter shales, which is also to be expected, since it is known (e.g. Bigarella et al., 1966; Dott, 1963; McBride, 1966; Morris, 1971) that slumps in distal direction may pass, via mudflows, into turbidity currents. Here this same relationship was found.

Both the shales from II.5.a.1 and II.5.a.2 are therefore interpreted as 'mass transported' sediments. Between these layers no material could be detected indicating an autochthonous sedimentation. This is not surprising, since a rough calculation (time span divided by number of beds) shows that in section 7, which is almost entirely composed of laminites, the average time interval between two turbidity currents is some 750 years, which is a fairly normal rate of sedimentation compared with the recent deep-sea sands, which are also interpreted by most authors as turbidites (e.g. Kuenen, 1964) and which show an average value of one turbidite in 1,000 years. The 750 years intervals in the Prioro Formation appear to have been too short to allow sufficient sedimentation to be detectable after compaction.

II.5.b. *Sandstones*

There is, roughly speaking, a difference between the sandstones in the lower and in the upper part of this formation. One type is more frequent in the lower part, two other types in the upper part.

II.5.b.1. *Graded sandstones.* – The older part is characterized by a number of sandstone layers within the shales (II.2). In this part the distance between two sandstone layers may vary considerably, but is usually between 50 cm and 10 m. The layers show all features of turbidites, including sole marks. Unfortunately, they are badly exposed in most cases, the best exposures being those along the road from Prioro to Tejerina. Where this road runs parallel to the strike, these sandstone layers can be followed over relatively large distances (some tens of metres). They belong to the best developed turbidites in this region, usually showing all Bouma's (1962) intervals as well as other characteristic properties. They may therefore be considered as coarser influxes alternating with finer-grained deposits, but their genesis is thought to be similar to that of the surrounding shale turbidites.

In the upper part of this formation this kind of sandstone is much rarer. Only in part of section 3 (see detailed section 1 in enclosure 3) are there sandstones with these same features.

II.5.b.2. *Structureless sandstones.* – In the younger part of this formation the sandstones are usually more difficult to interpret. This is partly caused by their scarcity, limiting the possibility of study. The most important finds of sandstones are in sections 2 and 4; those in section 3 were already dealt with in II.5.b.1. The exposures are small, so that no information could be obtained on lateral changes. As far as visible, these sandstones show an irregular appearance, while no sedimentary structures could be detected. This makes them quite different from the turbidites described above.

The absence of fossils, except for some wood fragments, and the absence of structures make it seem unlikely that they could be autochthonous sediments. The most reasonable interpretation seems a temporary influx of sand, transported downwards as a sandflow. In some of the sandstone layers so much matrix is present that a mudflow mechanism may also be considered a possible agent. Some sporadic pebbles may support this view. This interpretation explains both the often irregular appearance and the lack of fossils and structures.

II.5.b.3. *Sandy intervals.* – Only in one place in section 7 were a few metres of sediment found, which we consider to be an autochthonous, probably shallow marine deposit. It consists of shaly sandstones, alternating with a few sandy mudstones, while an arenite with a calcite cement is also present. Not only do several structures characteristic of a shallow marine environment such as abundant small channels, current ripples in various directions, wavy lamination, etc. suddenly appear, but at the same time there are also many fossils, chiefly fusulinids and brachiopods. Some pelecypods, crinoids and wood fragments are present, too. These fossils indicate a shallow marine environment. Although they have clearly been washed together, the distance of transport must have been neglibible, since even spines have in many cases not broken off.

Except for this place, only in one more locality is there a possibility (but no more than that) of a shallow marine origin. This is a level in section 7, at 40 m above the base of this section, where burrows have been found running through the sediment in all directions.

II.5.c. *Pebbly mudstones and conglomerates*

Pebbly mudstones frequently occur in this formation. Their appearance can vary from one single pebble or block (Fig. 17) in a shale layer several metres thick to a bed about 3 m thick, in which so many pebbles occur in the shale that they even touch each other in most places (section 1). It has already been stated (II.2) that they are more or less regionally concentrated: in the westernmost part of this area, along the road from Prioro to Tejerina, they are relatively rare, while in section 2 there is a sudden abundance. More to the E, in sections 1, 3, 4, 6 and 7 their quantity decreases rapidly, but gradually, so

Fig. 17. Large block (min. 223 x 75 x 46 cm) of Caliza de Montaña limestone in a pebbly mudstone or olistostrome. Loc. between sections 10 and 11.

that in sections 6 and 7 pebbly mudstones are already rare. E of section 7 once again frequent pebbly mudstones appear, which are still present in section 11 at the easternmost boundary.

It can also be observed that pebbly mudstones are more frequent in the uppermost part of this formation than at lower levels. In the older part they only occur in the southernmost part of our region, in which direction their number also seems to increase.

The mechanism of deposition of these sediments was investigated in great detail. Also because of the turbidites, slumps and mudflows around them, there can be no doubt that these masses were transported by mudflows (Crowell, 1957; Dott, 1963). Fragments of washed-in plants and shallow-marine fossils (mainly crinoids and brachiopods) are sporadically found in these masses, proving that they began to move in a shallow marine environment.

Although this cannot be proved for the bulk of the conglomerates (this name is used by us to indicate a sediment of pebbles within a sandstone matrix), it must be assumed that they were formed in a similar way. The difference, in our opinion, is only the sand/shale ratio in the material supplied. This is indicated by the presence of sediments, which form a transitional series between pebbly mudstones and conglomerates, since any percentage of sand or shale may be present in the matrix s.l. of these deposits. Another rather strong argument in favour of a similar origin is the presence of a gradual vertical transition from conglomerate into pebbly mudstone by grading of the matrix s.l. This grading has also been observed by other authors (e.g. Lindsay, 1966). The mechanism of deposition of these sediments was described earlier (van Loon, 1970). The interpretation was based largely on the characteristics of the pebbles (Cailleux, 1952), which change from bottom to top. While it was already previously known how pebbly mudstones originated (Crowell, 1957; Dott, 1963), it could now be established in more detail how these masses came to rest.

We also investigated whether there are differences between the pebbles of the pebbly mudstones at different stratigraphic levels. It appeared, however, that similar quartzite pebbles (to a lesser degree also sandstone and limestone pebbles) are always present, while their characteristics do not change regularly either. Neither are their differences in the petrographic composition of the matrix s.l. It is therefore impossible to find the stratigraphic level of a pebbly mudstone by investigating the petrography or pebble characteristics. These latter, however, may give some palaeogeographic information (II.3).

An estimate of the depth of the basin is difficult to make. We have seen, however (II.5.b.3), that a shallow part may have existed. Besides, assuming that the hinterland was situated near the León line (II.3, II.8), the basin will not have been very deep (though undoubtedly far below wave base) if we take into account a normal slope (0.5–3°) for these types of sedimentation (e.g. Kuenen, 1953; Anikouchine & Ling, 1963).

II.6. FOSSIL CONTENT

The Prioro Formation has for a long time been considered as barren (Brouwer & van Ginkel, 1964; Helmig, 1965). Little was therefore known about its age. To permit of a correct bio-stratigraphic dating, we paid much attention to its fossil content, which is indeed very low, although dating has been proved possible (van Loon, 1971). It appeared during our investigation that many species occur over the entire formation, although the number of specimens found is still rather limited. There are a few localities in which fossils seem relatively concentrated. In only 14 of the 188 slides of the three fully sampled sections (7 %) do fragments of fossils occur. Five of these slides belong to the shallow marine part in section 7 and two others belong to the few micrites present, so that it is clear that the other localities may really be called barren.

Nowhere were fossils found in growth position. Even in the shallow marine part they have been washed together. They are, however, less damaged here than elsewhere where they show many signs of transport. The lack of autochthonous fossils must be related to the high rate of sedimentation (II.5) and possibly a reducing environment, which prohibited life. Only a small number of tracks were found, showing that there must have been moments at which life was possible. Even in these levels, however, no autochthonous fossils were encountered which could indicate a relatively long period with little sedimentation and favourable conditions. Probably the tracks, as well as the burrows in section 7, were caused by animals able to live in a reducing environment (traces of pyrite can be found in nearly all slides, II.3).

Dating of the older part is still difficult. Only one identifiable fossil was found: a fragment of a washed-in:

terrestrial plant:
Mariopteris muricata (non von Schlotheim) Zeiller (591)

In the upper part much more fossils were found:

pelecypods:
Pernopecten carboniferum (Hind) Demanet (15, 113)
Pecten (Pseudamusium) sp. (308)
Edmondia aff. *arcuata* (Phillips) Demanet (308)
E. sp. (138, 698)
Annuliconcha interlineata (Meek & Worthen) (112)

Some of these are described and figured by van Amerom in: van Amerom et al., 1970.

brachiopods:
Productus cf. *carbonarius* de Koninck (698) (Pl. I, Fig. 1)
productids (663)
Reticulatia cf. *huecoensis* (King) (Coll. Winkler Prins)
Rugosochonetes acutus (Demanet) (556; cf.: 663)
Linoproductus sp. (Coll. Winkler Prins)
linoproductids (698)
Wellerella sp. (698)
Orthotetes sp. ex gr. *radiata* Fischer de Waldheim (651, 729)
spiriferids (651)
dicytyoclostids (240)

corals:
Lophophyllidium sp. (651)

fusulinids:
Schubertella ex gr. *kingi* Putrja (651)
S. sp. (604)
Hemifusulina sp. (160)
Fusulina sp. (604)
Beedeina sp. (604)
unidentifiable fusulinids (60, 61, 158, 160)

Sieswerda (1964b) found some specimens somewhat N of this area within a limestone lens in a conglomerate:
Beedeina bona (Rausser-Chernoussova) subsp. *lenaensis* van Ginkel
Staffella sp.
Pseudostaffella sp.
Schubertella sp.
Profusulinella sp.
Ozawainella sp.

Savage collected some samples from a limestone on the León line (his loc. M.V. 28) that yielded:
Staffella sp.

other forams:
Palaeotextularia sp. (651)
Bradyina sp. (651)
Ammodiscidae (651)
Calcitornella sp. or *Calcivertella* sp. (313)
Cornuspiridae (323)

Sieswerda (1964b) found the first three in his limestone lens, too, from which he described his fusulinids.

gasteropods:
several species not yet identified

trilobites:
one specimen not yet identified (663)

nautiloids:
one unidentifiable specimen (663)

crinoids:
many unidentifiable stems and ossicles (entire formation)

algae:
Dvinella comata Chvorova (651)
Uraloporella sieswerdai Rácz (651)
Ungdarella cf. *conservata* Korde (651)
U. sp. (604)
Ungdarella-Komia (604)
Komia abundans Korde (604)
K. sp. (604)
Anthracoporella sp. (604)
Archaeolithophyllum johnsoni Rácz (604)

land plants (washed in):
Linopteris obliqua (Bunbury) (663, 667; cf.: 663, 696)
L. obliqua var. *bunburyi* Bell (663)
L. neuropteroides (von Gutbier) (125; cf.: 696)
L. neuropteroides var. *minor* Potonié (663, 666, 667, 696, 698; cf.: 118, 130, 135, 138, 663, 667, 667, 696, 697)
L. cf. *neuropteroides* var. *linearis* Wagner (663)
L. subbrongniarti Grand'Eury (696; cf.: 125, 129, 664, 665, 696)
L. sp. (124, 125, 129, 308, 663, 664, 667, 696; ? sp.: 125, 665, 666, 696)
Neuropteris cf. *scheuchzeri* Hoffmann (134)
N. cf. *loshi* Brongniart (111)
cf. *N. rarinervis* Bunbury (663)
N. sp. (125, 663; ? sp.: 665)
Reticulopteris munsterifolia (Nemejc) (301)
R. sp. (301; ? sp.: 667)
? *Taeniopteris* sp. (696)
wood fragments are common over the entire formation

II.7. STRATIGRAPHIC INTERPRETATION

The older part of this formation only yielded a fragment of the plant *Mariopteris muricata,* which has a known range from middle Westphalian A – middle Westphalian C.

In the upper part, the plants indicate Westphalian C, most probably lower Westphalian C.

The brachiopods cannot yield very exclusive information. *Productus carbonarius* is known from upper Viséan – Westphalian B/C (Aegir marine band), while *Reticulatia huecoensis* indicates Lower Bashkirian – Kashirian, a range more or less similar to that of *Rugosochonetes acutus*. The brachiopods therefore all fall within the Lower Bashkirian – Westphalian B/C range. A Lower Moscovian age is the most probable (Winkler Prins, pers. comm.).

The fusulinids can possibly provide more exact information. The nearest relative of the encountered specimen of *Schubertella* will be *S. subkingi* (van Ginkel, pers. comm.), which in Russia has a range from Podolskian (possibly upper Kashirian) – lower Myachkovian (Rausser-Chernoussova et al., 1951). In the Cantabrian Mountains it has so far only been found in the Mesao Limestone Member of the overlying Pando Formation (van Ginkel, 1965), so that here we have a somewhat earlier appearance, which, however, may well coincide with the Russian range. The specimens of *Hemifusulina* belong to the most primitive forms of that genus (van Ginkel, pers. comm.), so that it indicates the basal part of the Kashirian. The specimen of *Fusulina* is also a primitive form.

The fusulinids found by Sieswerda (1964b) indicate lower Kashirian, but their locality cannot be correlated accurately with our area. The fusulinids from Savage's locality M.V. 28 indicate *Profusulinella* B or possibly even *Profusulinella* A subzone (probably Vereyan).

The algae possibly indicate *Fusulinella* B1 subzone, comparable to algal zone IV or the top part of zone III of Rácz (1965).

The pelecypods are known from the Aegir Marine Band (Westphalian B/C).

The other fossils do not give reliable stratigraphic data. It is therefore most probable that the Prioro Formation has an upper Westphalian B age at the base (*Mariopteris muricata*, fusulinids, brachiopods and pelecypods) and a Westphalian B/C age at the top (plants, brachiopods, algae, pelecypods).

II.8. PALAEOGEOGRAPHY

Palaeogeographic indications are scarce in this formation. There are a few sole marks in the turbidites. In the older part (in the S of our area) a south to north transport (if our tectonic reconstruction is correct) is indicated by the flutes, while the grooves show N–S orientations too.

In the upper part only one flute cast was found, roughly indicating a W to E transport, a direction also found for nearly all grooves (E to W or W to E). For these reasons we believe that the turbidites flowed eastwards. Only in two places (section 3) were N–S trending grooves found, while large wood fragments in a nearby locality also indicate this direction. This could be explained by a palaeogeographic picture in which a basin receives a sediment supply from the margins, which supply flows to the axis of the basin, and then rotates through 90° to follow this axis to the deepest point of the basin. The occurrence of a N–S trending channel (section 4) supports the idea that the material was derived from the northern and/or southern margin of the basin, and was transported along the axis towards the E (cf. Selley, 1968; Bailey, 1969; Walker, 1970; Contescu et al., 1966; Jipa, 1966).

Measurements of the current ripples lead to the same picture, since nearly all of them indicate transport directions to the S and E, with a small number indicating a transport to the N. These latter possibly indicate bottom currents (Hsu, 1964; Walker, 1970), active in the periods between the deposition of turbidites, an explanation supported by orientated fossils such as crinoids, which most probably rolled over the bottom. By means of photographs from recent sediments, the existence of such currents in deep water was proved (Heezen & Hollister, 1964).

It should, however, be noted that the number of ripple measurements is too small to draw fully reliable conclusions. More measurements were not possible, since the sediments are badly exposed, the ripples are scarce and cannot be seen on unpolished planes, while in places where ripples nevertheless can be measured the tectonic structure is not always sufficiently clear to allow a fully reliable reconstruction. All measurements indicated in Fig. 18, however, can be fully relied upon.

The assumption of a W to E transport is also supported by the grain-size distribution in the Prioro Formation. It

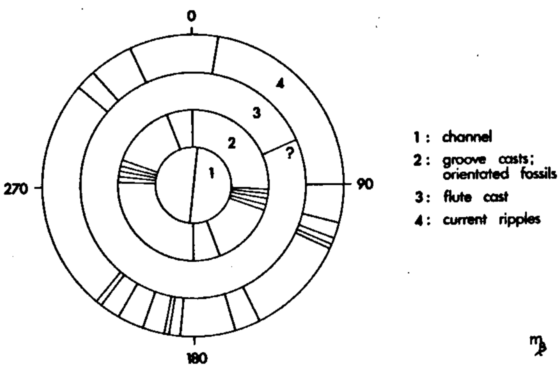

Fig. 18. Current directions in the Prioro Fm.

was mentioned (II.5.c) that the quantity of coarse material (sandstones and pebbles) gradually decreases in sections 2, 1, 3, 4, 6 and 7, i.e. in an eastward direction. This clearly indicates a W to E transport. Since much coarse material can also be found in sections 13 and 11 (mainly pebbly mudstones), it is necessary to assume that there were at least three points from which material was supplied. For the reasons mentioned above, these points must have been situated to the north of section 13 (or somewhat more westerly), north of section 2 and north of the region between sections 7 and 11.

As mentioned above, transport directions in the older part of this formation were only found in one place. In the southernmost part of the area studied, along the road from Prioro to Tejerina, in an isolated exposure with overturned layers (0/35), some turbidites were found one of which showed several N–S trending grooves and some flutes. When these layers are reconstructed into their most probable original position (taking into account an E–W trending fold axis, see Chapter IX), the flutes indicate a S to N transport. This can only be reasonably explained by assuming that the basinal axis was situated N of this locality. Since all other data indicated an axis S of Prioro, this axis must have run some 0–500 m S of Prioro.

There are some minerals which occur only at restricted stratigraphic levels (II.3). Considering sections 4 and 7, we see that in both sections a hornblende zone and a zone of angular albite can be dinstinguished (possibly also a clastic chlorite zone), which overlap for the larger part and are more or less parallel (Fig. 10). These zones are not easily detected, since these minerals occur in very small quantities (point-counting of more than 50,000 points yielded the following percentages: albite 0.001 %, hornblende 0.000 % and this chlorite 0.086 %). In a slide never more than a few of these grains are present, and these grains are often absent even in these zones (Fig. 10). The appearance of hornblende and this type of albite must indicate the erosion of a specific source rock in the hinterland. The fact that these minerals were found even less in section 11 than in sections 4 and 7 supports our assumption that the former section received its material from another point than the two latter sections.

In section 7 these mineral zones are thicker than in section 4. The part of the Prioro Formation above these zones is also thicker in section 7 than in section 4. The accumulation of more sediment in section 7 indicates a more rapid subsidence of the basin in that place. This is in accordance with the W to E transport directions found by means of ripples and flutes.

Considering the thicknesses of the Prioro Formation beneath these mineral zones, we see that in section 4 a sequence is present that cannot be found in section 7. Since both sections are limited at their bases by the Monte Viejo fault, this means that in section 7 a considerable part must be faulted down. The present base of section 4 must therefore be considered older than the present base of section 7.

II.9. CONCLUSIONS

The Prioro Formation, at least 670 m thick, consists entirely, except in one or possibly two places, of sediments transported 'en masse'. It closely resembles the German 'Flözleeres' (Dr. Paproth, pers. comm.). Slumps as well as mudflow and turbidity current mechanisms were active, probably on a delta-slope. This resulted in a rapid accumulation in a basin the E–W trending axis of which must be located somewhat S of Prioro. The sedimentation took place by supply mainly from the N, where at least three places can be designated from which the material was brought in. This sediment first followed a N–S direction along the delta slope, then bent into a W–E direction following the axis of the basin. There was some supply from the S, too, but considerably less than from the N. In the N a hinterland existed, which was influenced by the tectonic activity at the León line, and rose with irregular shocks, supplying much eroded material, that possibly indicates, by difference in grain size, three periods of rapid uplift, alternating with two periods of slower, but perhaps still rapid, uplift.

The almost completely siliciclastic material is very immature, both in textural and compositional respect, which is understandable on account of the small distance between the León line (where it was derived from) and the depositional site (probably varying between 0 and 7 km). Pebbles are supposed to derive from the older Palaeozoic formations, eroded at the León line.

The rapid sedimentation and reducing conditions prevented the development of an autochthonous fauna. In this almost barren formation sufficient fossils reworked from the contemporaneous shelf were found, however, to date this formation. Although there are some discrepancies between the various fossil groups, an upper Westphalian B age at the base and a Westphalian B/C age at the top seem most probable. A possible explanation of these discrepancies might be a diachronic character of this formation that is established on lithological properties only.

At one or two localities in the middle of this formation the deposits are interpreted as shallow marine, on the basis of fauna and sedimentary structures, indicating that a shallow environment of deposition is not incompatible with the mass transported sediments which surround them. These shallow parts were possibly the result of small, locally rapidly prograded deltaic lobes, which soon became eroded afterwards.

CHAPTER III

PANDO FORMATION: LOWER SANDSTONE MEMBER

III.1. INTRODUCTION

The name Pando Formation was derived from the Puerto de Pando (= Pando pass) on the road from Prioro to Pedrosa del Rey. Three members can be distinguished: Lower Sandstone Member, Mesao Limestone Member and Upper Sandstone Member. Together they compose the entire formation.

The oldest member, to be dealt with in this chapter, can well be observed in the S-flank. Its thickness varies considerably (enclosure 1). In the eastern bend of the syncline, too, this member can be found (section 11).

On the N-flank there are no outcrops of this member (except in one place) due to fault tectonics, unless this member there is developed in an entirely different facies not comparable to that of the S-flank. The only exception lies along the road from the Puerto de Pando to Pedrosa del Rey where a sequence occurs twice, which can best be interpreted as a partial tectonic repetition of this member.

It must be assumed that the Lower Sandstone Member in this northernmost part of the area studied, adjoining the León line, is considerably thicker than in the S-flank. Moreover, the transition into the overlying Mesao Limestone Member is very vague, because here rather thick limestones are already present in the Lower Sandstone Member, while in the Mesao Limestone Member thick sandstones still occur. Because of the scanty data obtainable from this area and the different aspects, this northern part will not be considered in the following paragraphs unless stated otherwise.

Three sections cutting through this member were sampled in detail:
section 4: 8 samples (this section is badly exposed)
section 7: 18 samples
section 11: 10 samples

These 36 samples together were used for the numerical petrographic data, while additional data were obtained from 14 more slides. The following localities belong to this member: 76–85, 194–211, 328–335, 554–556, 597–599, 602, 609, 612, 669–670, 700–702, 732–733 and 751.

III.2. LITHOLOGY

The frequent occurrence of sandstones is characteristic, as expressed in its name. They need not necessarily be pure sandstone, but can also be a sandy shale. Based on these lithological differences alone it is nearly always possible without any difficulty to draw a boundary between the Prioro and Pando Formations. Only in the valley of section 2, where the Prioro Formation is extremely sandy (especially in the top part), is the aid of fossils required. The appearance of many fossils is just as characteristic of this member as the presence of the sandstones (III.6).

The appearance of sandstone is usually gradual, but may be very sudden as in section 6 and, to a lesser degree, in section 7. It is possible that the sudden local change in lithology prompted previous authors (Brouwer & van Ginkel, 1964; Helmig, 1965; Boschma & van Staalduinen, 1968) to assume an unconformable contact between the Prioro and Pando Formations.

In spite of the large quantities of sandstones, the shales are still an important component, although their average grain size is much larger than in the Prioro Formation (compare II.3 with III.3) and they are often even sandy. The sand/shale ratio varies considerably in the various sections, and lies between approximately 4:1 and 2:3. Both the sandstones and the shales may be calcareous, although they are often decalcified probably under influence of recent weathering. Limestones are not unusual, although they are nearly always sandy. Pure limestones are very scarce.

Only one real conglomerate was found in this member, at a locality between sections 1 and 2. In other places pebbles may, however, occur in the sandstones as in section 6.

III.3. PETROGRAPHY

A microscopical examination of the 36 samples collected from sections 4, 7 and 11 shows distinct differences with respect to the Prioro Formation.

These samples are to be classified as follows:

1 = 3 % arenites
23 = 64 % wackes
3 = 8 % mudstones
4 = 11 % wackestones
4 = 11 % packstones
1 = 3 % grainstones

This means that 25 % is calciclastic and 75 % siliciclastic. For the three individual sections the lithology is

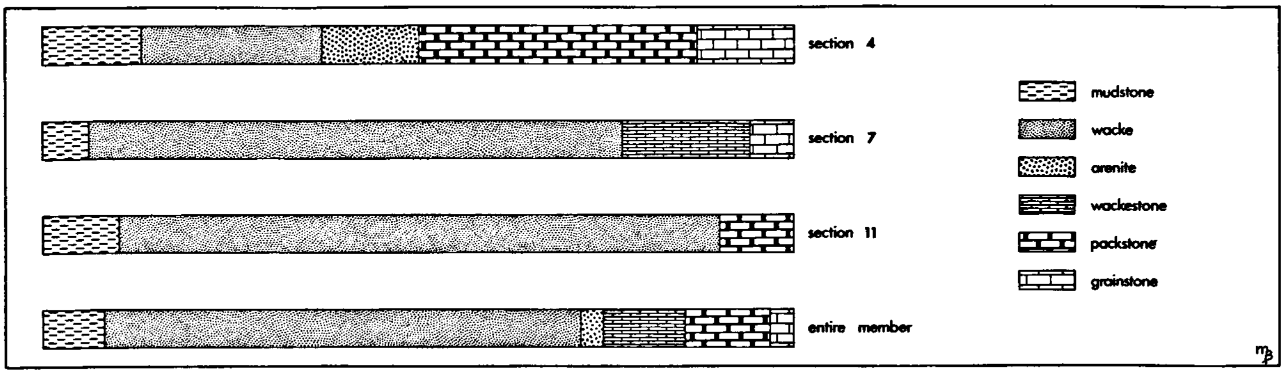

Fig. 19. Distribution of rock types in the Lower Sst. Mbr.

shown in Fig. 19. Sections 7 and 11 could be sampled almost completely, in section 4 some 40 m is not exposed. Field observations make it probable that the sediments not exposed are shales (mudstones), so that in reality the ratio between calciclastic and siliciclastic material will be approximately 1:4. These two types of sediments will be dealt with separately.

III.3.a. *Siliciclastic deposits*

The petrographic components of the 27 siliciclastic samples do not differ much from those in the Prioro Formation. The greatest difference is that their relative ratios have changed. Most important is the decrease of the matrix content, which can already be observed in the field by the fewer shales. It is mainly the quartz and the fossil fragments, which are much more common. There is also a considerable change in the group of the rock fragments, which are predominantly limestones in this member.

The mineralogical composition of these siliciclastic samples is, in greater detail:

1. abundant (more than 10 %):
The matrix is still the most important constituent (45.4 %). X-ray analysis showed the presence of the clay minerals illite (abundant) and chlorite (a normal constituent). The presence of irregular 14 Å mixed-layers is rare. The clay minerals are therefore very similar to those of the Prioro Formation.

The modal distribution of the matrix (Fig. 20) is distinctly different from that in the Prioro Formation.

Quartz grains comprise 31.5 %, while the whole group of authigenic minerals accounts for 15.6 %. The components of this group will be dealt with in III.4.

2. normal (0.1–10 %):
biotite (1.3 %) and muscovite (1.1 %) have about the same percentages as in the Prioro Formation, but the fossil content has increased to 0.6 %. The amounts of rock fragments (3.2 %) do not differ much. They can be divided into:

limestone	: 60.6 %
chert	: 18.0 %
phyllite	: 7.3 %
mudstone	: 6.6 %
shale	: 4.0 %
clay flakes	: 1.8 %
quartzite	: 1.7 %

The remainder consists of sandstone, opal and clay galls.

3. rare (less than 0.1 %):
albite, anatase, cassiterite, chlorite, epidote, hematite, hornblende, opaque minerals (several), plagioclase (the prismatic ? albite), rutile, staurolite, tourmaline, volcanic glass and zircon.

4. negligible (less than 10 grains observed):
apatite, augite, chloritoid, clinochlore, corundum, enstatite, gruenerite, kyanite (only replaced by calcite), pyrochlore, sillimanite, sphene, stilpnomelane.

Most of the minerals of groups 3 and 4 again belong to the heavy minerals, which in total once again constitute 0.09 %. Here too, the most resistant species are responsible for this percentage, only the quantity of rutile has decreased somewhat.

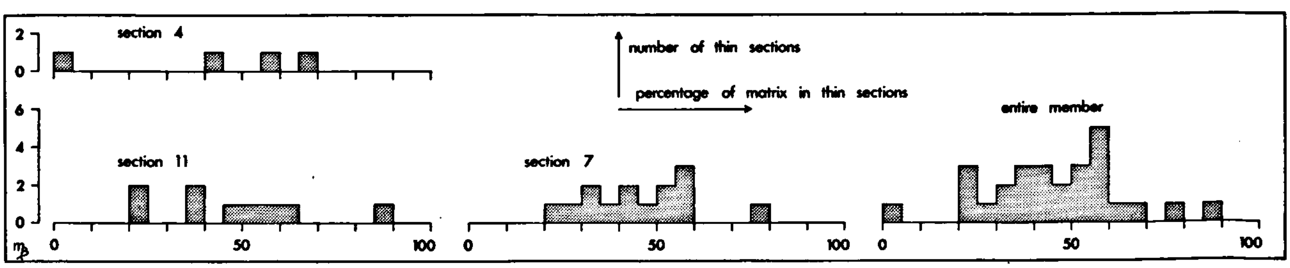

Fig. 20. Modal distribution of the matrix in the Lower Sandstone Member.

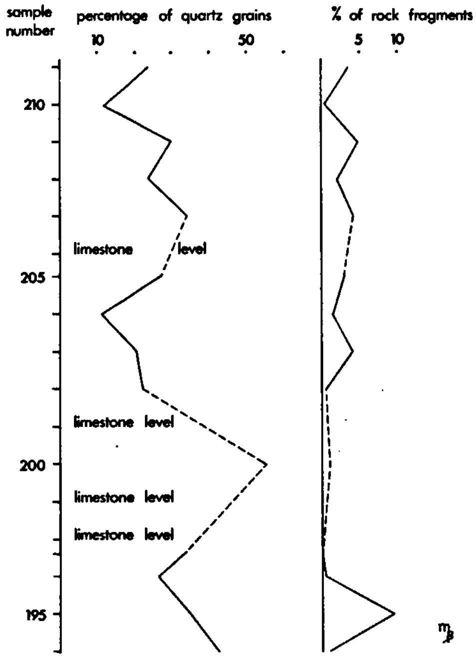

Fig. 21. Relationship between the percentages of quartz grains and rock fragments in section 7 of the Lower Sst. Mbr.

Most of the minerals present were also found in the Prioro Formation (II.3.a). New, among others, is the volcanic glass, probably basic because of its relief. This is present in sample 78 in relatively large quantities, indicating volcanic activity, although tuff layers were found nowhere. These are probably too thin to be recognized or have been washed away.

It is likely that most of these minerals were also primarily derived from Galician rocks. The roundness of the most resistant species indicates more than one sedimentary cycle, although prismatic crystals are also present.

The pebbles were not examined in thin sections. During the field work only quartzite pebbles were observed, which looked quite the same as those of the Prioro Formation.

Between the petrographic components distinguished, the same relationships could be established as those found in II.3. Presumably due to the smaller number of slides examined, the relationship between the heavy minerals and the quartz, and between the muscovite and biotite is even less clear. But the resemblance between the percentages of quartz and rock fragments is rather obvious (Fig. 21), while the relationship between the percentages of matrix and authigenic minerals will be shown in the following paragraph.

III.3.b. *Limestones*

Calcareous concretions or boundstones were not found in this member, all limestones are therefore calciclastic. Just as the siliciclastic sediments in this member, they are coarser than those of the Prioro Formation. Micrites are even absent, indicating currents that could prevent sedimentation of the finest particles or wash them away afterwards. There was still a supply of siliciclastic material at the time that limestone sedimentation was predominant.

The average petrographic composition of the three types of limestone present (grainstone, packstone and wackestone) is shown in Fig. 22.

It is interesting that the three sections sampled show differences in grain size of the limestones: those of section 4 are coarsest, those of section 7 intermediate and those of section 11 finest. This possibly means that the largest development of limestones (by building biogenetic banks?) occurred in the W, from where the material was transported towards the E.

III.4. DIAGENESIS

The relatively small quantity of matrix in the siliciclastic sediments, as compared with that of the Prioro Formation (45.4 and 70.7 %, respectively), is accompanied by an increase in the authigenic minerals (15.6 as against 9.2 %).

The following authigenic minerals were found:

quartz as secondary growth in samples which show quartzitization,
iron minerals, i.a. lepidocrocite (see below),
pyrite, concentrated around fossils, in small quantities nearly everywhere,
chert and opal, in silicified fossils,
chlorite, as a weathering product of biotite and as radial aggregates like in slide 83,
calcite, in veins, as a cement and as a replacement (e.g. of kyanite in slides 329 and 669),

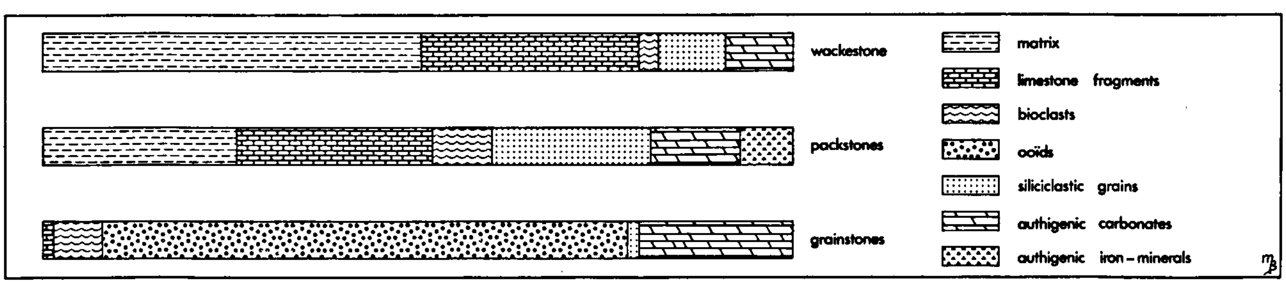

Fig. 22. Average petrographic composition of the limestone types in the Lower Sandstone Member.

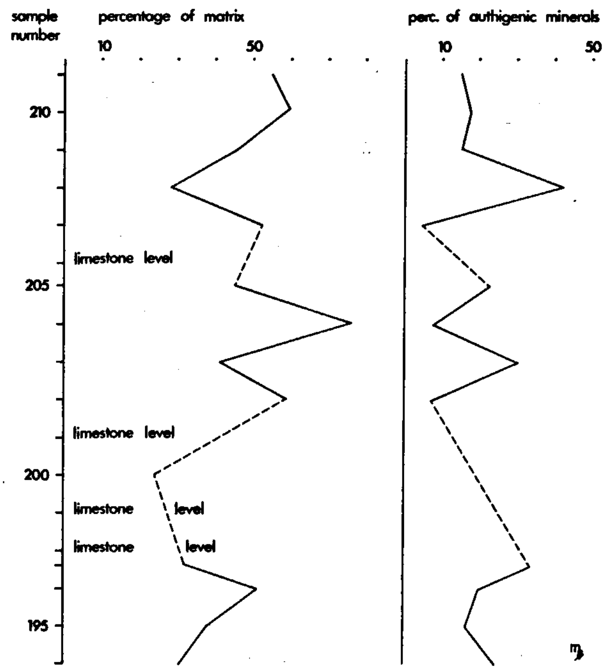

Fig. 23. Relationship between the percentages of matrix and authigenic minerals in section 7 of the Lower Sandstone Member.

anatase, as small bright yellow rhombohedra as in the Prioro Formation,
dolomite, most probably as a replacement of calcite, and siderite, here a rare occurrence.

Only few of these minerals occur in considerable quantities:

iron minerals	: 92.7 %
chlorite	: 4.2 %
calcite	: 2.7 %
quartz	: 0.3 %

Just as in the Prioro Formation, nearly all authigenic minerals therefore seem to be iron minerals (goethite, hematite and much amorphous material). It appears that here, too, the percentage of these minerals is inversely proportional to the percentage of matrix. For section 7 this is shown in Fig. 23. Since the percentage of the iron minerals was found by point-counting, it will in reality be smaller (compare II.4).

One of the most characteristic phenomena is that fossils have nearly always been opalized, in contrast to the Prioro Formation where they are usually preserved as calcite, although opalizing also occurs. In some cases the fossils are replaced by authigenic quartz or chert, while the original calcite has very rarely been preserved. A secondary replacement of the silicified fossils by authigenic calcite is even rarer. In the fossils of primary calcite other changes may occur. In slide 83 a fragment of a brachiopod is present, in which two zones occur (parallel to the fibrous structure) of silicification and of iron minerals. Fossils may also be completely replaced by iron minerals.

Slide 80 is one of the very rare slides examined from our entire area, that show a strongly pleochroitic iron mineral, probably lepidocrocite, in microscopically clearly visible quantities (approx. 400 microns).

Another rare feature among the iron minerals can be seen in slide 210, where strange interference colours, very similar to those of staurolite, are observed.

III.5. SEDIMENTARY STRUCTURES

The first appearance of a regular supply of sand is coupled with a total change in sedimentary environment. While in the Prioro Formation hardly any indications could be found of a shallow environment and autochthonous sediments, these predominate in the Lower Sandstone Member, as in the entire Pando Formation. This interpretation is based on the following evidence.

The sandstones are usually strongly channelling, with a width to depth ratio that may vary considerably. If the sandstones overlie each other, they often channel and the width to depth ratio is high. When they overlie shales they nearly always channel, while the width to depth ratio is lower, but may vary more than in the first case. Sometimes the sandstones show loading in the shales. In all channels clay galls can often be found, but real pebbles are much scarcer. When present they have always been concentrated in the lowermost part of the channel. Their rare occurrence must be a matter of supply, since in all three members of this formation pebbles are scarce.

Both the sandstones and the mudstones show abundant sedimentary structures, which, however, do not show much variation. Apart from the loading and channelling mentioned above, we are almost exclusively confronted with wavy and parallel lamination, and current ripples which usually have much larger dimensions than those in the Prioro Formation, but are still small-scale. Only in one locality in section 11 could wave ripples be demonstrated with certainty, proving the shallow character of this member. That the environment must have been marine is proved beyond doubt by the fossils (III.6). These data indicate a delta-front environment.

It is worth noting that the only well-developed conglomerate (III.2) shows a distinct imbrication, indicating a south to north movement of the water. Possibly it represents Reading's (1970) facies viii. The conclusions from this observation will be dealt with in III.8.

Besides all these distinctly autochthonous sediments some allochthonous sediments are also present. In section 4 several graded layers still occur, which show many features of turbidites. One layer even shows the whole sequence of Bouma (1962). In this layer rather many fossil fragments occur at the base, while clay galls are also present, indicating erosive power of what is interpreted as a shallow marine turbidity current.

Somewhat higher up in this section a sandstone is present with a slump structure (cf. Oomkens, 1967). This clear indication of mass transport in a shallow marine

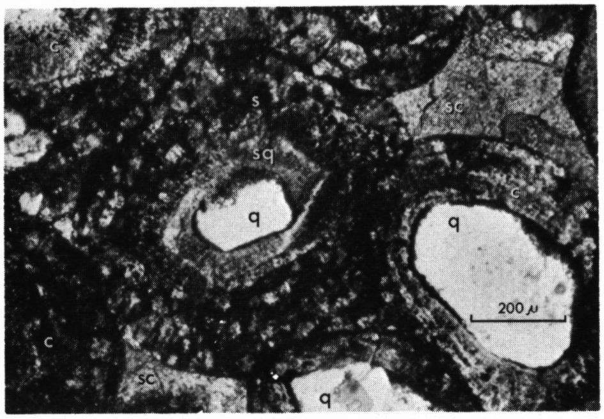

Fig. 24. Thin section of sample 334. Oolite with sometimes silicified (s), originally calcitic (c) ooids. Pore space filled with sparry calcite (sc). Cores of the ooids consist of quartz grains (q), sometimes surrounded by authigenic quartz forming euhedral crystals (sq).

environment supports the possibility, mentioned above, of turbidites in this environment.

Some of the limestones also indicate a very shallow environment, for the grainstones (334, 670) are real oolites (Figs. 22 and 24).

III.6. FOSSIL CONTENT

The base of the Pando Formation is characterized by the sudden abundance of fossils. This is such a striking phenomenon that it can serve for determining the boundary between the two formations. This is a favourable circumstance for those sections in which it is difficult to draw a boundary on lithological grounds, since the appearance of sandstone is gradual, as in sections 4 and 11. The fossils occur both in the siliciclastic and calciclastic sediments. Their abundance may be expressed by their frequency in the slides: in 25 of the 36 slides fossils were found (= 69 %; compare with the Prioro Formation: 7 %).

In a few cases fossils were found in an apparent growth position, in many cases fragile parts of fossils such as spines were found not to have broken off the fossils (mainly brachiopods) encountered in the shales. It therefore seems reasonable to think of a quiet and muddy environment. Moreover, all fossils are only known from a shallow marine environment, except for the plant fragments, which were washed in.

In this member the following fossils were found:

pelecypods:
Pecten (Pseudamusium) medium (Herrick) sensu Fedotov (612)
Anthraconeilo sp. (335)
Annuliconcha interlineata (Meek & Worthen) (612)

brachiopods:
Dictyoclostus ? aegiranus Böger & Fiebig (700, 733) (Pl. I, Fig. 2)
Orthotetes sp. ex gr. *radiata* Fischer de Waldheim (76)
O.? sp. (612, 700)
Schizophoria sp. (612, 751; ? sp.: 732) (Pl. I, Fig. 4)
Productus cf. *carbonarius* de Koninck (700)
productids (612, 732, 733)
Linoproductus latiplanus Ivanov (Coll. Winkler Prins)
linoproductids (76, 612)
Levipustula cf. *breimeri* Winkler Prins (751)
Orthotichia sp. (612)
Orulgania ? sp. (612)
Karavankina cf. *rakuszi* Winkler Prins (751)
K. sp. (612) (Pl. I, Fig. 3)
Reticulatia cf. *huecoensis* (King) (76)
Brachythyrina sp. (76)
Kozlowskia cf. *aberbaidenensis* (Ramsbottom) (612)
Crurithyris sp. (612)
Choristites sp. (700, 751)
Martinia sp. (612; ? sp.: 612,700)
Rhipidomella sp. (612)
Zaissania sp. (612; ? sp.: 700)

fusulinids:
Hemifusulina sp. (554, 700; ? sp.: 198, 201, 329)
Ozawainella cf. *angulata* (Colani) (206)
O. sp. (201)
Pseudostaffella sp. (670)
Schubertella sp. (670)

other forams:
Endothyra sp. (197)
Tetrataxis sp. (670)

crinoids:
calyx of dicyclic Inadunata (732) (Pl. II, Fig. 1)
stems and ossicles abundant throughout this member

land plants (washed-in):
Linopteris cf. *neuropteroides* von Gutbier (204)
L. neuropteroides var. *minor* Potonié (204; cf.: 204)
L. *cf. subbrongniarti* Grand'Eury (204)
L. sp. (204; ? sp.: 204)
Lepidophloios sp. (732)
? *Neuropteris* sp. (204)
wood fragments are abundant throughout this member.

It thus appears that the number of species is not extremely large. One must, however, bear in mind that this member is only a rather thin one (enclosure 1), and that there are no more than a few fossil localities. In these localities, the richest of which are 612 for shells and 204 for plants, the number of species is usually limited, although there are abundant individuals. Among the brachiopods, for instance, the genera *Schizophoria* and *Orthotetes* are represented by so many specimens that they can be considered characteristic of this member, although they occur in the other stratigraphic levels as well.

Towards the top of this member we usually find a decrease in the fossil abundance, in some cases (sections 6 and 7) fossils even become scarce at the top.

III.7. STRATIGRAPHIC INTERPRETATION

The brachiopods indicate lower Podolskian or uppermost Kashirian. *Linoproductus latiplanus*, however, is only known from the Vereyan of Russia, but since it is associated with the other brachiopods mentioned in III.6, it must now be assumed to range into the Kashirian or even into the lower Upper Moscovian.

The few plants indicate Westphalian C, a more exact dating is impossible.

Among the fusulinids the specimens of *Hemifusulina*, found in slide 700, are still primitive forms, although clearly less primitive than those found in the Prioro Formation (van Ginkel, pers. comm.), where they indicate the base of the Kashirian. Since they are more primitive than those of the overlying Mesao Limestone Member, dated by van Ginkel (1965) as upper Lower Moscovian or lowermost Upper Moscovian, they must be placed somewhere in the Kashirian. Most probably they must be dated as *Profusulinella* B subzone/*Fusulinella* A subzone (van Ginkel, pers. comm.).

Since only these few stratigraphic data are available, the law of superposition must help us. Bearing in mind that this member must be younger than the Prioro Formation and somewhat older than the well-dated Mesao Limestone Member (IV.7), a lower Westphalian C (not in contradiction with the possible ranges of the groups mentioned above) age is considered most likely for this member.

III.8. PALAEOGEOGRAPHY

Because of the very limited thickness of this member (max. 100 m in section 6) only few indications for a palaeogeographical reconstruction can be found. The most interesting is the conglomerate between sections 1 and 2. It shows a fairly well-developed imbrication, which indicates a movement of the water from SSW to NNE. Considering that this conglomerate lies in between clearly shallow marine sediments, and that it does not look like a fan that was invaded by the sea, it must be assumed that this conglomerate was also deposited in a shallow marine or littoral environment. The only recent conglomerates with the same properties are found along coasts where the action of waves, more or less perpendicular to these coasts, causes a shoreward movement of water resulting in this imbrication, which cannot be disturbed by the waves coming back since the power of their backwash is not sufficient. For the Lower Sandstone Member this interpretation leads to a coast line, which has a WNW–ESE direction.

Ripple measurements give a picture of currents in all directions, with frequencies, however, to the NW and to

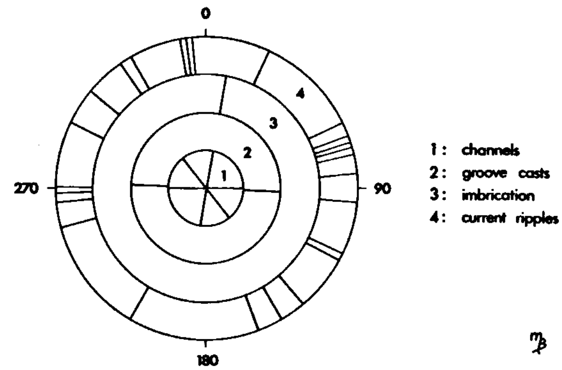

Fig. 25. Current directions in the Lower Sandstone Member.

the SE (Fig. 25). These might indicate tidal influence. If the tidal currents were directed perpendicular to the coast, it would mean that the coast line trended more or less NE–SW, which is not in full accordance with the interpretation based on the imbricated conglomerate. Actually, the coast line may have been irregular and the tidal currents might have had varying directions.

At one place groove casts were found, indicating an E to W or a W to E transport, which is the same direction found for groove casts in the Prioro Formation. W to E transport in the basin is supported by the various types of limestone in sections 4, 7 and 11 (III.3.b).

As in the Prioro Formation, some places can be indicated from which the supply was greatest. This is clearly the case in section 6. While in section 4, at only a small distance W of this section, sandstone is still rather scanty and the transition from the Prioro Formation gradual, section 6 shows a very sudden transition with a high sand content. In section 7, somewhat to the E, the sand content decreases, until in section 11 it is again fairly unimportant. Based on these observations a source of supply may be assumed to lie in the vicinity of section 6. For similar reasons this can be said of section 2 in which very thick sandstone layers (up to 2 m) occur. This point was also found in the Prioro Formation as a source of sand supply.

The overall picture is therefore still one of supply from the N, with a transport to the S into the basin, and a transport to the E along a coast.

III.9 CONCLUSIONS

The Lower Sandstone Member is characterized by a sudden occupation by a fauna and a large increase in the sand content, which latter may be abrupt or gradual. These sandstones, as well as the shales in between, contain many structures indicating a shallow delta-front environment, while the limestones and the fauna indicate that this shallow environment was marine.

As is shown by the grain size of the material and by the washed-in plants, the coast may not have been far away. Palaeogeographic data indicate an E–W or NE–SW trending coast line. There were a few localities from which the material, eroded in the hinterland which still possessed a considerable relief, was supplied.

Between these autochthonous sediments some allochthonous sediments are present, too. These mass-transported deposits are, however, fairly unimportant and could not prevent an abundant fauna from living on the bottom. The number of species is, however, still limited, which hampers dating. Aided by the law of superposition, a lower Westphalian C age is considered most probable.

CHAPTER IV

PANDO FORMATION: MESAO LIMESTONE MEMBER

IV.1. INTRODUCTION

This is the only part of this formation which has already been distinguished as a separate member by previous authors. However, they did not indicate by which criteria they distinguished it, so that the levels which they considered to be the lower and upper boundary are not known. Of course, the presence of limestones was the reason of their decision, especially as the limestones can be recognized from great distances in places where they are well developed. This easy recognition is partly the result of their being well exposed, while the mudstones in between are usually covered by vegetation. A gradual increase in carbonate at the base and a sometimes gradual decrease at the top makes it difficult to draw indisputable boundaries for this member. In the opinion of the author the lower boundary can best be drawn where the first distinct limestone occurs at the base of a succession in which limestones are very frequent. This is pretty well possible in the N-flank. In the S-flank, however, there are parts in which limestones are scarce, or in which even no pure limestone occurs. Here the lower boundary was drawn where the sandy – sometimes somewhat calcareous – mudstones of the Lower Sandstone Member pass into the first – always present – very calcareous mudstone of a calcareous interval.

The upper boundary was chosen on identical grounds: when the last limestone in a succession containing much limestone disappears, while the quantity of sandstone (of the Upper Sandstone Member) strongly increases or, where no limestones were formed, at the transition of mainly calcareous into mainly sandy mudstones. This boundary can be drawn relatively easily.

These boundaries are therefore strongly dependent on the local facies, expressed in the supply of siliciclastics and the deposition of carbonates, and are therefore far from time-equivalent. This results in considerable changes in thickness of this member in the sections measured. It must be borne in mind that by this procedure deposits can be formed at the same time, which in one place will be considered to belong to the Mesao Limestone Member, in another place to belong to the Lower or Upper Sandstone Member.

The name of this member was derived from the Cueto Mesao (Mesao Peak) near the Puerto de Pando. This peak, however, owes its shape to the resistant cover of conglomerates of the Ocejo Formation.

From (parts of) five sections sampled in detail the samples were used for numerical petrographic data. These sections are:

section 4: 17 samples
section 7: 41 samples
section 10: 26 samples
section 11: 24 samples
section 13: 42 samples

In total these are 150 samples. Additional data were provided by 45 thin sections. The following localities belong to this member: 86–109, 212–252, 336–352, 484–526, 557–588, 593, 595–596, 601, 605, 607, 611, 613–617, 626–633, 648–650, 653–654, 671–674, 686–693, 703–710, 712–720, 722–724, 734–748, 759–760 and 762.

IV.2. LITHOLOGY

Limestones are, of course, the most characteristic part of this member. In the field it is usually impossible to see whether these limestones are biogenic or clastic. We shall therefore deal with them later on (IV.3). The limestones alternate with calcareous mudstones, which are sometimes sandy. The limestones are somewhat muddy, or contain thin mudstone layers (Figs. 26–27). They may vary considerably in total thickness, but never exceed 50 m in this region. In places where they reach thicknesses of some tens of metres they cause hills because of their relatively high resistance to erosion. The hills in the NE part of this region are formed in this way.

Sandstones are very rare; only N of the Puerto de Pando are they important. Conglomerates are absent, except for two rather small conglomeratic banks at the base of section 13. Pebbles, however, can still be found in many places in the limestones, usually single, sometimes in small groups. But they are a rare occurrence. In the mudstones they are even much rarer.

The transitions between the various kinds of lithology are usually very gradual, both laterally and vertically, but abrupt changes also occur.

In horizontal sense we can see that the limestones in the N-flank are much better developed than those in the S-flank, while the biogenic limestones in the western part of our region are much more important than in the eastern part.

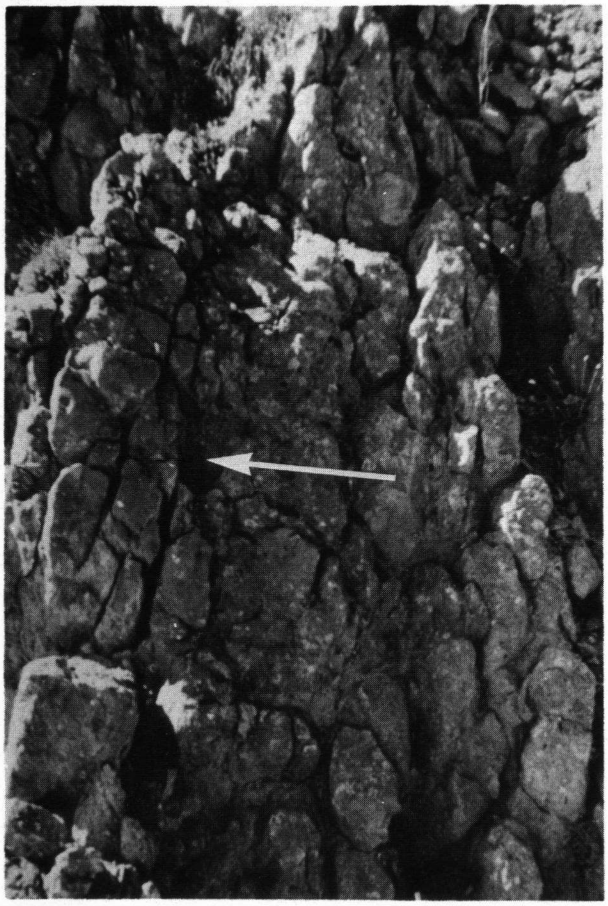

Fig. 26. Normal appearance of the limestones in the Mesao Limestone Member. Note the fairly irregular bedding planes. Loc. 690.

Fig. 27. Nodular limestone. Bedding plane hardly visible. Loc. 687.

IV.3. PETROGRAPHY

Rock samples were taken at stratigraphic distances of 5 m in sections 4 (whole member, but badly exposed), 7 (completely sampled), 10 (except from a small part at the base of this member), 11 (top of this member absent) and 13 (probably small parts of both base and top absent). These sections yielded (see also Fig. 28) 8 (= 47 %), 27 (= 66 %), 20 (= 77 %), 7 (= 29 %) and 21 (= 50 %) limestone samples, respectively.

In this member limestones for the first time constitute a majority (83 out of 150 samples = 55 %). Even supposing that the parts in section 4 not exposed consist of mudstones (which is very probable), the percentage of the limestones is still more than 50 % for the whole member. However, this division in limestones and siliciclastic sediments was made by microscopical examination. In the field many rocks were considered to be mudstones, which microscopically appeared to consist of impure micrites.

Once again we shall deal with these different groups separately.

IV.3.a. *Siliciclastic sediments*

These can be classified in the following manner (see also Fig. 28):

3 = 2 % arenites
31 = 21 % wackes
33 = 22 % mudstones

In total 67 = 45 % siliciclastic sediments.

Compared with the Lower Sandstone Member the wacke/mudstone ratio is much lower. Probably the hinterland supplied much less coarse material, possibly due to a slower uplift.

The petrographic components are quite the same as in the older deposits. The relative ratios of these components, however, differ considerably from those in the Lower Sandstone Member, but all in all resemble those of the Prioro Formation (see II.3). The most important difference is that the percentage of fossils here is more than 11 times the percentage in the Prioro Formation. Beyond doubt, this is the result of the erosion of the very fossiliferous limestones between these siliciclastic deposits in this member (see IV.3.b and IV.6).

When the percentages of the components are inves-

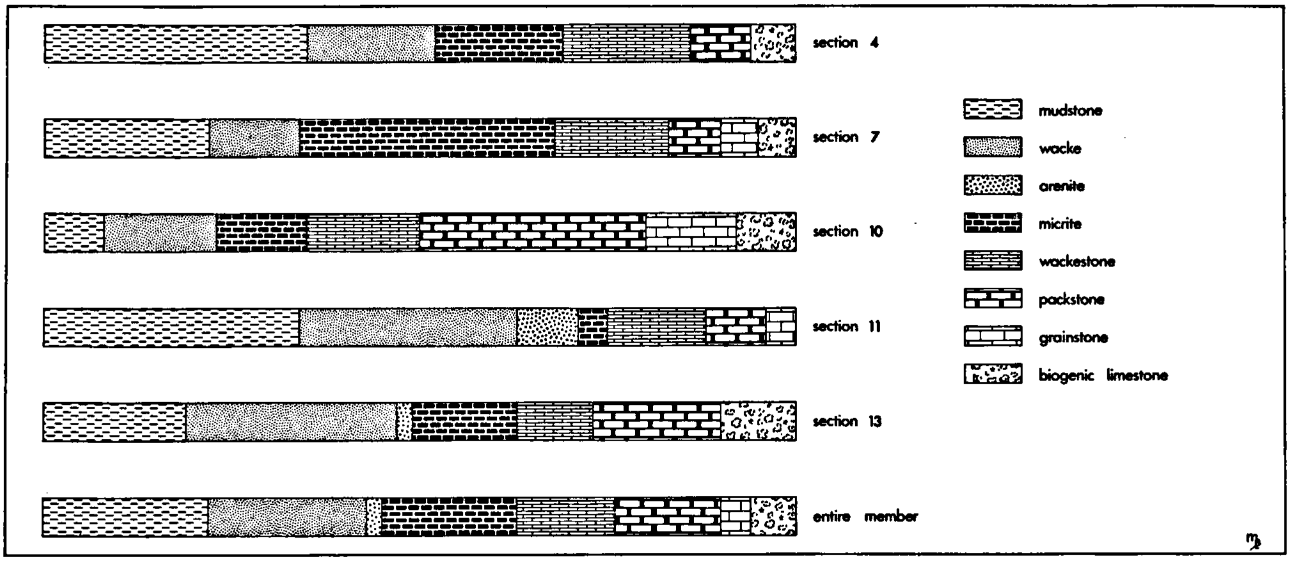

Fig. 28. Distribution of rock types in the Mesao Limestone Member.

tigated in greater detail it appears, however, that there are considerable differences as compared with the Prioro Formation, especially as far as the distribution over the slides is concerned. This is very clear for, for instance, the percentage of matrix. Although this is 69.9 % here and 70.7 % in the Prioro Formation, Fig. 29 shows that there is no predominance of slides with 80–95 % of matrix such as in the Prioro Formation, but that there is a considerable distribution.

Point-counting of 64 slides (samples 94, 102 and 493 yielded no slides) yielded the following results:

1. abundant (more than 10 %):
As mentioned above, matrix constitutes by far the largest group with 69.9 %. The following clay minerals were identified by X-ray analysis: illite (abundant) with somewhat less chlorite.

Quartz grains constitute 19.4 %.

2. normal (0.1–10 %):
Muscovite (1.0 %) and biotite (1.3 %) do not differ much from the amounts in the Lower Sandstone Member, while fossils now constitute 2.0 %. The authigenic minerals have decreased to 8.2 %. Their components will be dealt with in IV.4.

The last group belonging to this class is the group of the rock fragments, increased to 4.1 %. These fragments consist of:

limestone	: 84.5 %
chert	: 4.8 %
phyllite	: 3.0 %
shale	: 2.1 %
mudstone	: 2.1 %
clay galls	: 2.0 %
clay flakes	: 0.6 %
quartzite	: 0.4 %

The remainder consists mainly of sandstone and opal.

As was to be expected in this limestone member, the percentage of limestone fragments has increased even further. These fragments consist mainly of microspar and spar, although fragments of nearly all other types of limestones in this member can also sporadically be found.

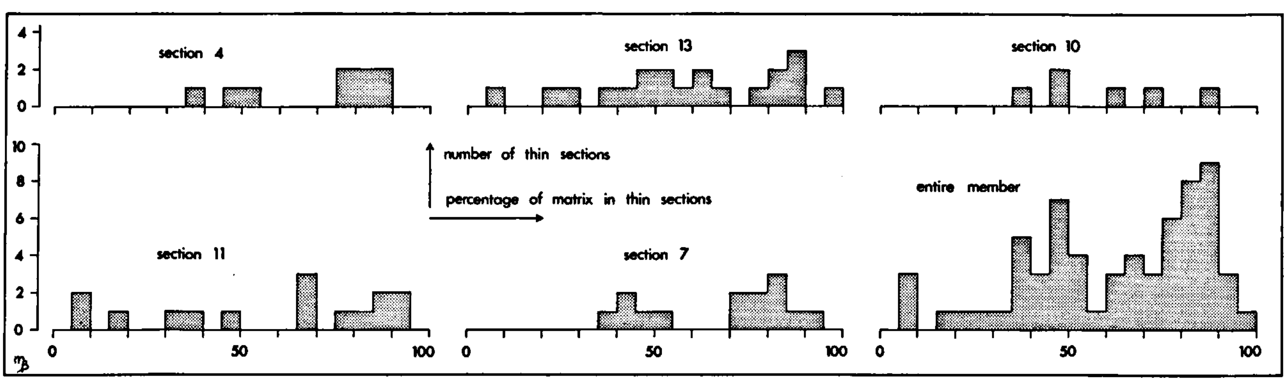

Fig. 29. Modal distribution of the matrix in the Mesao Limestone Member.

The relatively high percentages of clay galls and flakes, indicating considerable erosion of the not yet completely consolidated mudstones, are remarkable. Because of the frequent differences in grain size in the mudstones, their plastic deformation is often clear microscopically.

3. rare (less than 0.1 %):

Albite, anatase, apatite, augite, cassiterite, chlorite, epidote, hornblende, iron minerals (hematite ?), opaque minerals (several), plagioclase (the prismatic ? albite), potassium feldspar, rutile, tourmaline and zircon.

4. negligible (less than 10 grains observed):

Actinolite, andalusite (?), brookite, chloritoid, clinochlore, corundum, diaspore, dolomite, hypersthene, kyanite (?), olivine, orthite, phlogopite, spinel and zeolite.

The quantity of the heavy minerals is somewhat less (0.07 %) than in the older deposits (0.09 %). Once again, the bulk consists of the usual most resistant minerals, among which especially tourmaline is relatively abundant, showing many varieties in colour.

It seems unnecessary to assume other source rocks for the minerals present than we did before (II.3 and III.3). Only the mineral diaspore is difficult to explain, but it may originate from volcanic rocks.

The presence of olivine, a rapidly weathering mineral (Pettijohn, 1941; Milner, 1962; Ollier, 1969), is typical. The only explanation for its presence is rapid burying within non-permeable sediments (mudstones).

Pebbles are scarce in this member (IV.2), and only few of them have been examined. They all belong to one of the quartzites mentioned in II.3, except for three limestone pebbles. Since these latter in thin section do not resemble those of the older Palaeozoic formations which still crop out in the vicinity (i.e. León line N of Tejerina) of the finding place of these pebbles (section 13), the present author is inclined to consider these limestone pebbles as intrabasinal fragments.

Among the finer-grained components the relationship between the percentages of quartz and rock fragments, which run parallel in the older deposits, has almost completely disappeared. This must be due to the high percentage of limestone fragments derived from intrabasinal erosion, by which the original relationship between the quartz grains and the rock fragments supplied by the hinterland was masked. This effect will even be greater, because the limestones were situated between the mudstones like 'islands', on account of which fact the supply of limestone was much less local than that of the siliciclastic sediments, which are dependent on the supply by rivers. This assumption is the more probable, since in section 4 the original relationship between quartz and rock fragments is still observable, in section 11 somewhat less, while it has completely disappeared in the other sections. Sections 11 and 4 are precisely the ones with the fewest limestones (29 and 47 % respectively, for section 4 even as few as approximately 30 %, if the parts not exposed are mudstones, see IV.2).

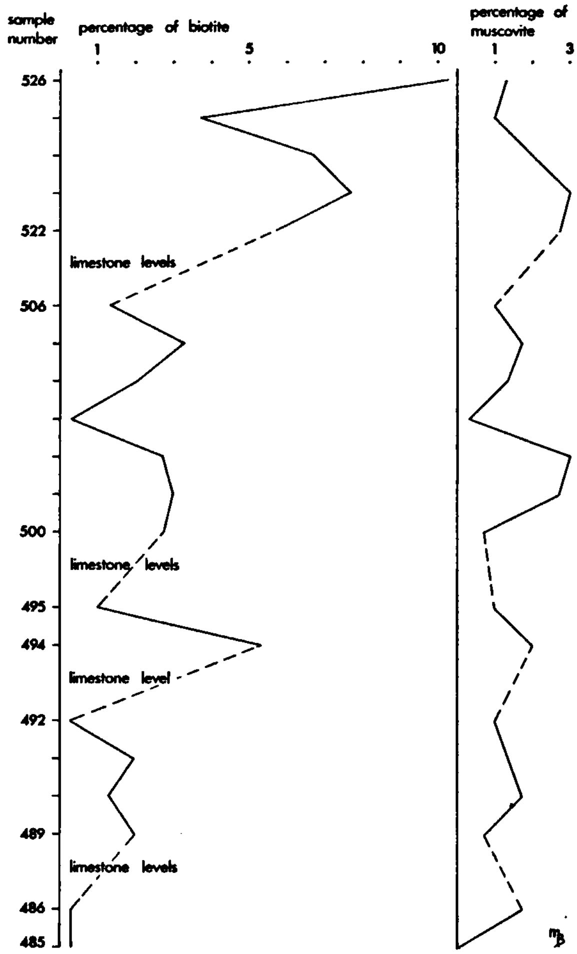

Fig. 30. Relationship between the percentages of biotite and muscovite in section 13 of the Mesao Limestone Member.

The relationship between the percentages of muscovite and biotite that run parallel is, however, very clear. This is illustrated for section 13 (Fig. 30). Only in the lowermost part do these percentages show no interdependence.

IV.3.b. *Limestones*

The limestones constitute 55 % of this member (see above). They can be grouped as follows:

6 = 4 % grainstones
21 = 14 % packstones
20 = 13 % wackestones
27 = 18 % micrites
9 = 6 % boundstones

The matrix is always recrystallized, and usually should be called a microspar or even a pseudospar (sensu Folk, 1965). Since most of the larger grains are also recrystallized, it is often very difficult to interpret these limestones. It is, however, possible to state reliable differen-

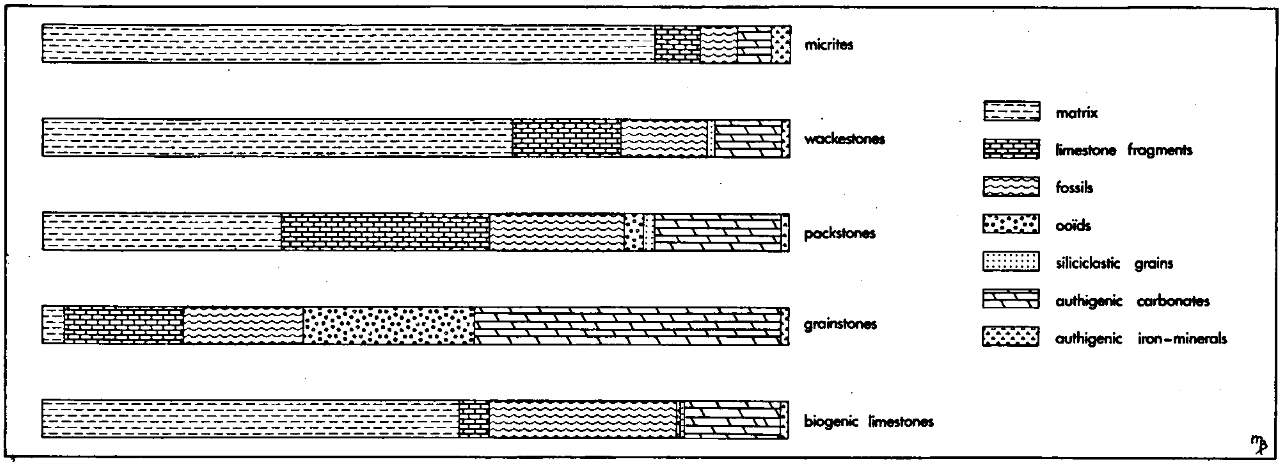

Fig. 31. Average petrographic composition of the limestone types in the Mesao Limestone Member.

ces between the limestones of the five sections sampled (see Fig. 28).

In spite of these differences, these sections show a number of similarities the most important of which are: no or only a few grainstones, and no or only a few boundstones. The first observation is in accordance with what we saw in IV.3.a: there are only a few coarse sediments. The second may be due either to erosion of the original biogenetic banks to calciclastic material or to non-recognition. This latter possibility was recently emphasized by de Meyer (1971), but he believes (pers. comm.) that the former possibility is mainly responsible here.

It is interesting that the limestones may contain widely varying amounts of siliciclastic grains (0–20 %), but that these grains most frequently occur in the limestones at the base of the sections, and become less important toward the top.

The average total composition of the five types of limestone is shown in Fig. 31.

IV.4. DIAGENESIS

As far as the siliciclastic sediments are concerned, the percentages of matrix and of authigenic minerals increase and decrease, respectively, showing a mirror image as compared with the transition between the Prioro Formation and the Lower Sandstone Member. These changing percentages do not only bear upon this member as a whole, but we see once more that in the slides matrix and authigenic minerals are inversely proportional, as is shown for section 7 (Fig. 32), though somewhat less clear than in the older deposits.

The following authigenic minerals were encountered:

iron minerals (goethite, hematite, lepidocrocite were shown by X-ray analysis, amorphous material is also present), usually as vague stains, probably originating from weathered biotite, sometimes as replaced fossils, and often as small cubes which are pseudomorphs after pyrite;

pyrite itself, varying from 'drusy' to rather coarse crystalline, often found especially in limestones, probably a consequence of a reducing micro-environment around a decaying animal;

calcite, predominantly as veins and as replacements,

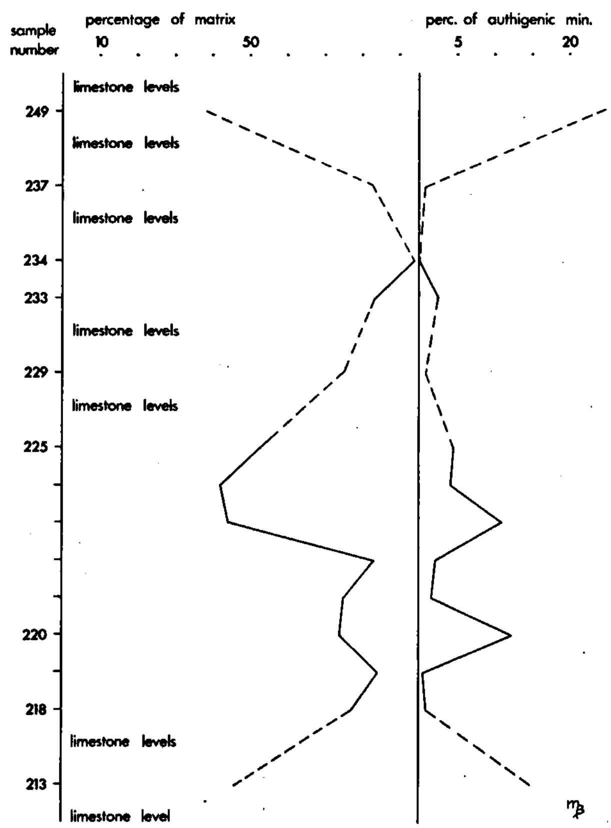

Fig. 32. Relationship between the percentages of matrix and authigenic minerals in section 7 of the Mesao Limestone Member.

Fig. 33. Thin section of sample 97. Fossil is broken due to compaction of the surrounding mud.

usually of quartz, sometimes of other minerals such as muscovite and zircon;

chlorite, usually as a weathering product of biotite, sometimes as radial aggregates;

opal, as opalization of fossils (rare in this member);

quartz, as a replacement in fossils or of mineral grains as calcite and dolomite, as veins and as secondary quartz in quartzitic layers, only a few of which, however, occur, e.g. 99 and 100 in section 11, although the coarser sandstones are often slightly quartzitic. In a few limestones a number of euhedral quartz crystals are present, such as recently found in sebkhas;

muscovite;

tourmaline: this mineral is often attacked by calcite, but in other places secondary tourmaline may be formed (100, 492);

rutile, as small crystals (99);

dolomite, as a frequent replacement of calcite, sometimes in the form of rhombohedra, up to approx. 250 microns;

zeolite, probably natrolite or thomsonite, as the filling of a crack;

siderite, as small rhombohedra.

In the siliciclastic sediments only few of these minerals occur in significant quantities:

iron minerals	: 94.1 %
calcite	: 4.5 %
chlorite	: 0.6 %
quartz	: 0.6 %

The quantity of iron minerals is therefore once more responsible for the percentage of authigenic minerals in these sediments.

Compaction is an early diagenetic process in the mudstones, possibly partly responsible for the mass transport – still occurring in this member too – (see IV.5), because differences in relief might have been created between these mudstones and the less compactible limestones. This compaction can be seen in slides by compression and breaking of fossils (Fig. 33) and by folding of small dikes (e.g. the zeolite dike in slide 104) (Fig. 34). The latter proves that this zeolite must also be very early diagenetic.

Fig. 34. Thin section of sample 104. Crack was filled with zeolite (z), compressed due to compaction.

In the limestones many different diagenetic features occur, the most important of which are:

a. dolomitization. In the slides dolomite was distinguished from calcite by staining methods (Friedman, 1959; Dickson, 1966). The dolomitization occurred irregularly. The limestones can vary from nearly pure calcite to nearly pure dolomite. It was not possible to find any regular change, either in time or in space. Dolomitization appears to start from irregularly distributed cores, which often do not differ from their surroundings when examined by means of a microscope. But stylolites also often appear to be zones from which dolomitization starts (Fig. 35). This would mean that dolomitization is, at least in part, a relatively late diagenetic process, occurring later than stylolitization. In literature (e.g. Dapples, 1967), however, dolomitization is usually considered to be early diagenetic and pre-lithification.

When the whole ground mass of a limestone has been dolomitized, fossil fragments, especially brachiopods and

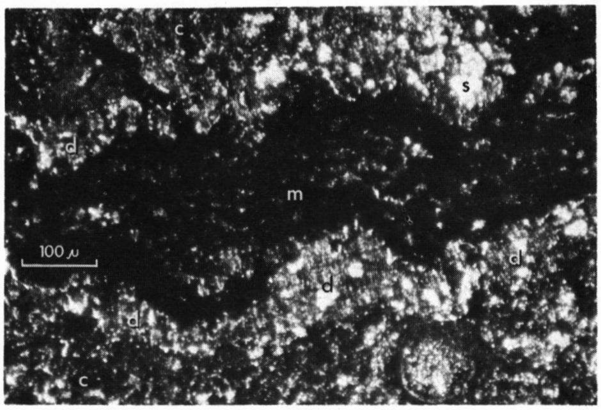

Fig. 35. Thin section of sample 241. Concentration of siliciclastic mud in a stylolite (m). At both sides of the stylolite authigenic dolomite (d) has been formed. Most of the limestone, however, still consists of calcite (c) with some siliciclastic material (s).

pelecypods, often still consist of calcite. This shows that dolomitization is realized most easily when the particles are small, irrespective of whether these have themselves been recrystallized or not.

b. pressure solution phenomena. Both well-developed, strongly dentated stylolites and just beginning, still nearly flat ones are observed. From the degree of stylolitization it can be calculated that large parts of the limestones (up to approx. 25 %) have been dissolved. It is not known where the dissolved material was transported to, but the calcareous matter in the mudstones is possibly partly derived from these limestones. The residue that remains in the limestones consists mainly of clay minerals (Fig. 35), often with a considerable quantity of iron minerals and some pyrite. Since the limestones are often 'dirty' and contain much clay, these residual zones can be very thick (214).

c. birdseye structures. These are places with a sparitic filling within an area of micrite, microspar or pseudospar (see IV.4.f). These patches show a variation in length of approx. 100 microns to some mm. It is possible to distinguish two different types:
1. more or less rounded patches with grains of sparite of about equal size. We presume that these have been formed by recrystallization, starting from a core such as a fossil.
2. rounded or irregular patches, sometimes connected by 'veins' filled with sparite which becomes coarser from the outside to the inside. Sometimes the transition into the surrounding ground mass is gradual. These birdseyes originated, as may be assumed on account of the grain-size distribution, from crystallization in cavities, which, as is shown by the 'veins', were at least partly connected with each other. De Coo and Deelman (Univ. of Leiden) carried out a number of experiments, in which water was added to a dry lime mud, a situation which will occasionally occur in supratidal lime mud environments. When the water entered the mud the air was pressed out of the pores, causing structures just like those described above. They also studied recent supratidal mud environments, where they found the same structures. By these experiments the author is also convinced that these structures may indicate a supratidal environment, although other possibilities, such as worm tracks etc., cannot be fully excluded. Shinn (1968) also supposed birdseyes to be a possible indication of a supratidal environment.

d. veins. In most cases these consist of calcite. It was observed that they break through stylolites (212, 577), but also that stylolites cut them into parts that no longer touch. This proves that veins were formed during a long time range, from before until after stylolitization.

In one slide both pure calcite and nearly pure dolomite veins may occur. Typically, the dolomite veins more often cut through the calcite veins (226, 572) than the reverse (243). Since the dolomite veins must have originated from calcitic veins (partly dolomitized veins (243) also suggest this), the following explanation appears to be the only possibility: calcite veins were formed during a long period (see above); at a certain moment dolomitization began, and for reasons not yet understood the youngest calcite veins, which sometimes cut through the older ones, were replaced first; when dolomitization stopped, before all veins were replaced, this led to dolomite veins cutting through the calcitic veins. In a following phase calcite veins may again be formed which cut through both dolomite and older calcite veins. Since dolomitization appears to be a rather late diagenetic feature (IV.4.a), it follows that there is a difference between the 'older' and 'younger' calcite veins.

e. silicification. In many slides fossils or parts of fossils appear to have been silicified. They are usually fusulinids and other foraminifers, while echinoderms (especially crinoids) and bryozoans also seem to be silicified relatively easily. The silica required for this process can be derived from the often considerable quantity of siliciclastic material (IV.3.b) or from dissolved *Glomospirae* (a siliceous foraminifer) of which still a large number is present (IV.6).

In the field silicified fossils, mainly brachiopods and corals, can be found. This may indicate a fore-reef environment (Dapples, 1967).

f. recrystallization. This is one of the most important changes that have taken place in the limestones. Both the large particles and the ground mass have often been recrystallized. For this reason the fossils in the slides often could not be identified.

The ground mass, presumably micrite in most cases, has nearly always been recrystallized into a microspar or even a pseudospar (sensu Folk, 1965). Folk mentions that this is relatively rare, only occurring in limestones

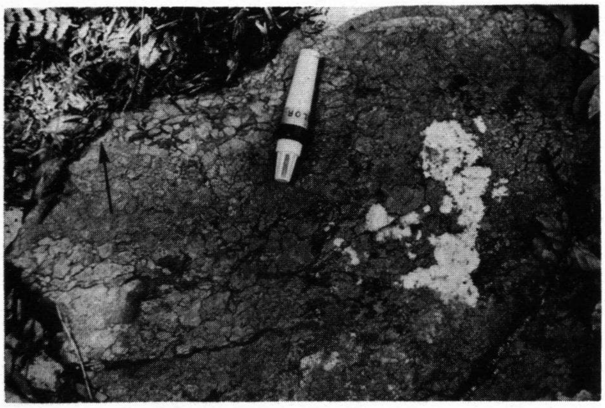

Fig. 36. Debris of a biogenetic bank. The limestone fragments contain abundant algae. Their angularity indicates a short transport. Loc. between sections 10 and 11.

containing shale. Our observations are in full agreement with this remark. In a few cases recrystallization of the ground mass led to very large calcite crystals measuring up to several cm. This recrystallization of the ground mass has nearly always caused all sedimentary structures to disappear.

IV.5. SEDIMENTARY STRUCTURES

It is worth noting that the Mesao Limestone Member must have been deposited, over its entire thickness (max. 200–250 m), in a very shallow marine environment. The sediments were often affected by wave action (Reading's, 1970, facies iv), which is shown by the occurrence of oolites over the entire stratigraphic range of this member. Wave ripples were also observed in a siliciclastic portion of section 11.

The oolites are not the only limestones which indicate a shallow environment. Part of the limestones are biogenic (IV.3.b), and can best be described as biogenetic banks (Baars, 1963). In these banks many fossils occur, mainly algae, which indicate a very shallow environment (Johnson, 1961). In several places at the ends of these biogenetic banks angular fragments occur (Fig. 36) which, because of their resemblance to the biogenetic bank, are interpreted as talus material, also indicating a shallow (wave action), possibly even intertidal environment. This material changes laterally into finer-grained clastic limestones, which interfinger with, or gradually pass into, the surrounding mudstones. These latter may be derived either from the eroded hinterland (according to Veevers, 1969, a slow uplift of the source area results in carbonates when the sea floor subsides slowly, and in siltstones and shales when there is a rapid subsidence), or from clay accumulation as a result of tidal action (van Straaten & Kuenen, 1958).

There are also limestones which, by the occurrence of birdseyes, indicate a supratidal environment (IV.4). The frequent presence of stromatactis (e.g. 605) is another argument in favour of subaerial exposure, and the occurrence of broken up algal mats, resulting in an intraformational breccia (Fig. 37), is also in favour of this assumption (Roehl, 1967).

Although sedimentary structures are difficult to distinguish in limestones, especially when recrystallized, it is possible in several localities where the limestones contain much mud to observe parallel lamination (sometimes slightly wavy) and very sporadically current ripples as well. Usually the limestones do not contain channels, but in section 10 the contacts with the mudstones are usually strongly erosive.

The absence of much siliciclastic material, the sorting and the hardly rounded fossil fragments are indications of the quietwater character of most of the limestones (Plumley et al., 1962). In spite of this generally quiet environment mass transported limestones also occur in this member. They show much variation:

a. There is one mudstone level near the base of this member containing a number of limestone masses with a

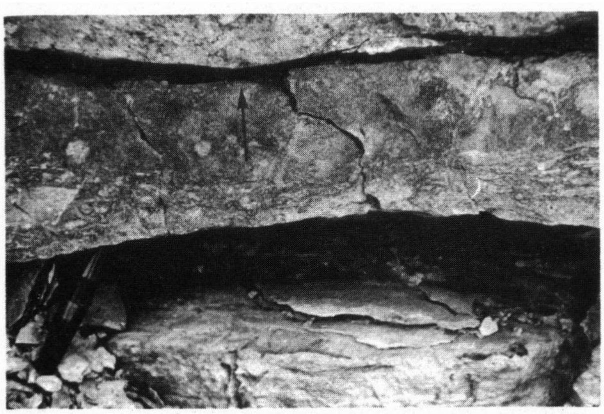

Fig. 37. Limestone bank with broken-up algal mats. This 'intraformational breccia' indicates subaerial exposure (Roehl, 1967). Loc. 235.

Fig. 38. Limestone olistolite or 'Gleitklippe', lying within mudstone, partly pushed away. Some 500 m W of loc. 611.

maximum size of a few tens of metres, which must have slid off in a manner transitional between olistolites (Badoux, 1967) and slumping (Fig. 38). Since these limestone blocks lie within a mudstone that is apparently autochthonous, they cannot have slid down in an olistostrome, but as a single block. In such a case Görler & Reuter (1968) prefer the term 'Gleitklippe'. The large size of these blocks is not remarkable, since this has frequently been observed in both ancient and (sub)recent sediments (Renz et al., 1955; Plafker et al., 1969; Moore et al., 1970). There is a striking resemblance to the blocks described by McBride (1966).

All blocks appear to have derived from the same layer that was broken up; the largest ones have been slightly rolled up. Because of their deformation these latter indicate an approx. N to S transport (cf. Ksiazkiewicz, 1958). Although the contacts with the surrounding mudstone are often not or badly exposed, it could be confirmed that this mudstone has been partly pushed away and now curves around the limestone (cf. van de Graaff, 1971). That this is no tectonic feature is proved by the layers below and above this mudstone not having been disturbed. The fact that a limestone layer is broken up, but that some parts could be deformed during their transport, proves that the limestone had reached a stage of considerable but not of full consolidation.

b. Slumping can be clearly demonstrated in a few cases. The degree of contortion can change laterally in a layer. Especially when a high percentage of clay is present, the conditions for slumping seem favourable as well as when lithological differences exist, such as between rather small pieces (approx. ½–1½ m) of biogenetic limestones within a calcareous mudstone or micrite. This can be shown for level 249 in section 7. Even in the slide it can be seen that the ground mass has been contorted, while the large limestone particles were too consolidated and broke when the stress became too great (Fig. 39).

Probably slumping also occurs in some of the oolites. Although the grain size should be rather uniform – and usually it is – so that sedimentary structures are hardly visible, there are, however, some oolites that show vague irregular structures. In slides they prove to contain a considerable quantity of clay and silt, which is, of course, unusual for sediments which are supposed to have been formed in agitated water. Besides the sedimentary structures this quantity of fine material may indicate short sliding over a muddy bottom (van der Meer Mohr, pers. comm.).

c. Convolute lamination was observed in a limestone layer that showed no other sedimentary structures (Fig. 40). Although this layer could not be traced in the field because of the vegetation, it is most probably the same as a limestone layer, some 200 m further on, in which slump structures occur. It therefore appears that these two structures may be related.

d. As noted earlier (IV.2), there are still some quartzite pebbles in this member. Where a single pebble occurs in a coarse clastic limestone or in a biogenetic bank, it may be assumed that it was the depositional site of an occasional coarse supply of siliciclastics. However, there are also very badly sorted clastic limestones with a high percentage of clay, in which several quartzite pebbles (up to 15 cm in diameter, section 9) and patches of mudstone (mudstone pebbles? , possibly parts of an eroded mud bottom) are situated. Since it is hard to believe that both the pebbles and the clay particles were supplied and deposited at the same time as a result of particle by particle transport, these sediments were possibly deposited by a lime mudflow. We did not succeed in finding conclusive evidence either in favour of or against such a mode of transport. But because other mass transported limestones are present in this member, it must have been possible, at least theoretically.

e. Turbidites are most probably present as well. In any case, there are graded limestones (Fig. 41) which often show a parallel lamination, sometimes small current

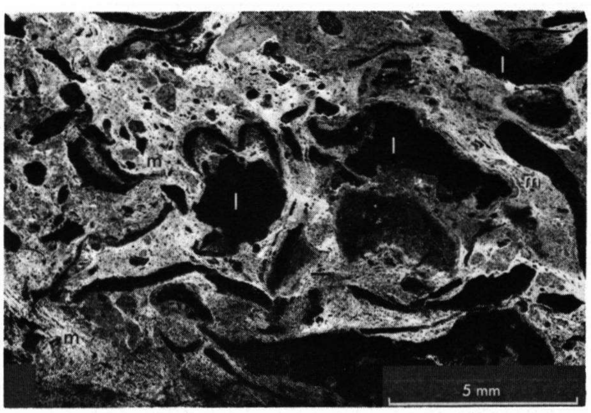

Fig. 39. Negative print of thin section of sample 249. In this slump level the consolidated limestone fragments (l) are broken, while the siliciclastic mud matrix (m) is contorted.

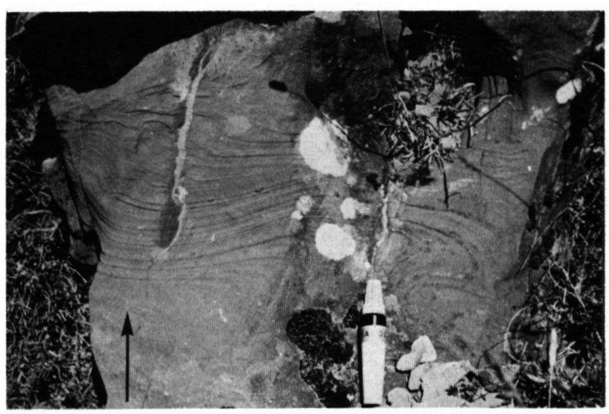

Fig. 40. Limestone with convolute lamination, visible by an alternation of mainly calcareous and less calcareous laminae. This limestone probably passes laterally into a slump. Loc. 561.

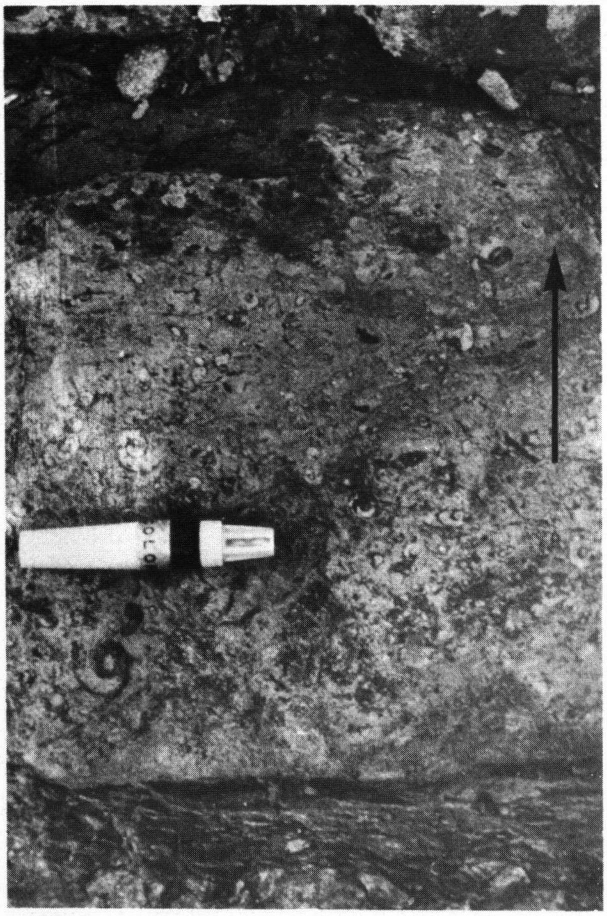

Fig. 41. Graded limestone turbidite. Grading caused by decreasing size of fossils and increasing amount of siliciclastic mud towards the top. Base shows load casting, but is flat as compared with layers of Figs. 26 and 27. Loc. 714.

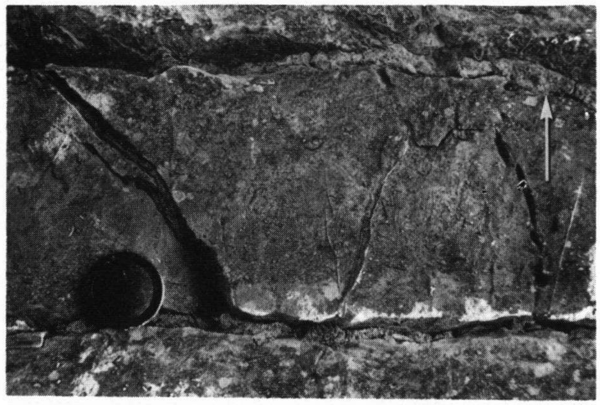

Fig. 42. Limestone turbidite with a pre-phase (Meischner, 1964) at the base. Same exposure as Fig. 41.

Fig. 43. Limestone lens formed by a biogenetic bank in the basal part and by an oolite with only few fossil fragments at the top. Intercalated between fossiliferous mudstones. Loc. 230.

ripples or vague structures (convolute lamination?) above these structures, and in sporadic cases another zone of fine parallel lamination in the top part, the uppermost part of which gradually passes into a calcareous mudstone without structures. Similar limestones were described by Thomson & Thomasson (1969). This picture also resembles that of Bouma's (1962) turbidite sequence, passing at the top into an autochthonous sediment. There are a few layers in which below the graded lower part, usually containing fossils, yet another zone is to be found mainly of clay and silt-sized material, which, however, belongs to the graded layer above (Fig. 42). This was also observed in Germany by Meischner (1964) in Carboniferous limestones. Meischner called this part the pre-phase. In the author's opinion the explanation is not yet satisfactory.

In spite of all these properties it is difficult to prove that these sediments are really turbidites. The largest exposures in which such layers occur are approx. 5 m long, so that it could hardly be checked how these layers change laterally. In the exposure itself, however, the thickness of these layers does not change, while the lower bedding planes are sharp and flatter than those of other limestones, although load casting in underlying mudstones may cause a rather irregular bedding plane. The degree of exposure prevents the study of the lower bedding planes themselves, so that it is unknwon whether sole marks are present or not.

Because all these deposits lie between sediments that are beyond doubt shallow marine, it must be assumed that here, as in the Lower Sandstone Member, mass transport took place in a very shallow environment (cf. Sturani, 1969).

It should be mentioned, however, that most limestones do not show sliding. The biogenetic banks pass both laterally and vertically into mudstones. Although sometimes an oolitic bank may have served as a barrier (Fig. 43), the biogenetic banks were usually only protected by their own debris (Fig. 36). This low degree of protection is possibly the reason why so few biogenetic banks have been conserved, and why nearly all limestones are clastic. The clastic limestones generally have a limited extension as well, preventing the establishment of a correlation in the S-flank, even between two neighbouring sections. This limited lateral extension, when compared with the thickness of these banks, supports the idea that very local biogenetic banks supplied considerable quantities of calciclastic material. Only in the eastern part of the syncline do limestone layers exist which can be traced over several hundreds of metres.

The mudstones are always calcareous. The percentage of carbonate can become so high that there is a gradual transition into a limestone. This amount of carbonate for the most part belongs to the matrix, although calcitic fossils may also be important. Since the percentage of clay is normally high, sedimentary structures are often difficult to see on account of the small grain-size variation. Where the mudstones contain some more sand, however, they usually show much more structures, especially parallel lamination, somewhat less wavy lamination. Current ripples are rarer, possibly because they are more difficult to detect. They are always small-scale ripples.

The mudstone layers have irregular thicknesses, probably partly because of having to adapt themselves to the limestone banks when compaction begins. They often wedge out completely, and when they are sandy they channel into the underlying mudstones. We interpret these sediments as muddy shoals between the limestones, while the more sandy mudstones could indicate the places along which most of the transport took place. This should also explain the channelling.

As was the case with the limestones, there are also a number of mass transported mudstones. As far as could be seen, only slumping is present (339), and even that feature is rare.

IV.6. FOSSIL CONTENT

This member is extremely rich in fossils, both individuals and species. The clastic limestones especially contain many echinoderms and bryozoans, while in the biogenetic limestones algae are the most important. The percentage of fossils is very high (IV.3), especially in these limestones. In spite of a somewhat lower percentage (2.0 %), the mudstones yielded more fossils for identification because here they could be detached much more easily. The number of faunal elements in the mudstones is also much larger, since they contain the species from the eroded limestones and the species which lived in the mudstones themselves. All fossil groups encountered in this area are present in the mudstones of this member. Most probably many fossils in situ occur, such as brachiopods (e.g. *Rugosochonetes acutus*) of which all spines have been preserved. Many others showing two opened valves still connected to each other may have been transported over a small distance only.

Differences in environment existed, for in one slide (e.g. 693) both quiet-water ostracods and turbulent-water specimens may occur (Michel, pers. comm.). Since both types have been well preserved, transport cannot have taken place over a great distance. These two environments must therefore have been very close together, a conclusion that was also drawn from the various kinds of lithology (IV.5). The ostracods, which appear to be very useful environmental indicators (Bless, 1970), unfortunately could not be studied in detail. The differences in environment are also expressed by the differences in the percentage of samples that contain fossils in the five sections. The most reliable way to obtain comparable results was a microscopical examination of slides of these samples. The results are shown in Table 4.

section	number of slides examined			percentage of slides		
	with fossils	without fossils		with fossils	without fossils	
4	12	5	17	70	30	100
7	39	2	41	95	5	100
10	26	0	26	100	0	100
11	18	4	22	81	19	100
13	23	19	42	54	46	100
whole member	118	30	148	79	21	100

Table 4.

section	number of siliciclastic slides			percentage of sil. slides		
	with fossils	without fossils		with fossils	without fossils	
4	4	5	9	45	55	100
7	12	2	14	85	15	100
10	6	0	6	100	0	100
11	11	4	15	73	27	100
13	2	19	21	10	90	100
whole member	35	30	65	53	47	100

Table 5.

This picture, however, is somewhat distorted since fossils were encountered in all slides of limestones: section 4: 8 slides; 7: 27; 10: 20; 11: 7 and section 13: 21. The differences are therefore caused by the fossil content in the siliciclastic sediments and the limestone/siliciclastics ratio. The results of the siliciclastic parts are shown in Table 5.

Section 13 therefore appears to be poorest in fossils as far as the siliciclastic sediments are concerned (10 %), section 10 the most fossiliferous (100 %).

Many of the fossils have already been previously recorded (van Loon, 1971). In the following lists they are mentioned together with later finds.

IV.6.a. *Limestones*

From the limestones or small shaly intercalations in the limestones the following fossils were collected:

brachiopods:
Meekella sp. (497)
spiriferids (692)

pelecypods:
Pterinopecten sp. (631)

corals:
Rotiphyllum sp. (746)
Pseudozaphrentoides sp. (689)
Chaetetes sp. (514, 709)
cf. *Carcinophyllum* (*Axolithophyllum*) sp. (232)
cyathopsids (690)
syringoporids (235, 690)

Somewhere in section 5 Sieswerda (1964a) also found: *Syringopora* sp.

fusulinids:
Beedeina sp. (87, 92, 227, 737)
Beedeina sp. or *Fusulina* sp. (242, 509, 574; ? : 107, 108)
Beedeina sp. or *Dagmarella* sp. (519)
? *Eofusulina* sp. (92, 95)
Millerella sp. (511, 519, 562, 690, 717, 748)
Ozawainella sp. (87, 95, 244, 613)
? *Putrella* sp. (613)
Pseudostaffella sp. (92, 95, 226, 242, 243, 246, 509, 511, 519, 557, 574, 690, 708, 737, 748; ? sp.: 717)
Profusulinella sp. (87, 557, 737; ? sp.: 562)
Schubertella ex gr. *obscura* Lee & Chen (557)
S. sp. (613, 632, 692, 693, 705, 708; ? sp.: 562)
Staffellinae (92, 95, 212, 737)

Van Ginkel (1965) found the following fusulinids in his locality L 11, which is our locality 692:
Staffella cf. *pseudophaeroidea* Dutkevitch
Parastaffella sp.
Pseudostaffella ex gr. *parasphaeroidea* (Lee & Chen)
Schubertella cf. *subkingi* Putrya
Hemifusulina ex gr. *moelleri* Rauser-Chernoussova
Fusiella cf. *praecursor* Rauser-Chernoussova

foraminifers (siliceous):
cf. *Glomospira* sp. (throughout this member)
cf. *Glomospirella* sp. (idem, but rare)

foraminifers (other):
Ammodiscidae (throughout this member)
Indothyridae (idem)
Cornuspiridae (idem)
many other not identified foraminifers (idem)

crinoids:
Amphoracrinitidae or Actinocrinitidae (calyx) (235)
stems and ossicles abundant all over this member

sponges:
Amblysiphonella barroisi Steinmann (215, 226, 516; ? : 710)
spicules are relatively rare, but occur throughout this member

conodonts:
Idiognathodus sp. (715; possibly reworked)
Streptognathodus sp. or *Gnathodus* sp. (715)
Streptognathodus sp. (714)
Hindeodella sp. (712)

algae:
Epimastopora rolloensis Rácz (93; cf.: 557)
E. bodoniensis Rácz (92)
E. sp. (336, 588, 613, 708)
Girvanella sp. (566)
cf. *Ungdarella conservata* Korde (521)
Ungdarella-Komia (212, 348, 509, 570)
Komia abundans Korde (244, 511, 515, 587, 588)
K. sp. (92, 93, 348, 349)
Archaeolithophyllum johnsoni Rácz (108, 217, 232, 248, 351, 497, 498, 507, 508, 513, 519, 563, 564, 565, 578, 585; ? : 228)
Anthracoporella sp. (496, 517, 520; ? : 227, 251, 578)
Dvinella comata Chvorova (515)
Donezella sp. (588)
dasycladaceans (92, 216, 562, 572, 578, 717)
codiaceans (243, 578, 588, 708)

In some cases boring Algae were found in crinoids (cf. Sadler, 1970).

Sieswerda (1964a) found in this member:
Uraloporella sp.
Petschoria sp.
Girvanella sp.
Komia sp.

land plants (washed-in):
Linopteris neuropteroides (von Gutbier) (250)
L. cf. *neuropteroides* var. *minor* Potonié (250)
L. cf. *neuropteroides* var. *linearis* Wagner (704)
L. obliqua (Bunbury) (232)
L. subbrongniarti Grand'Eury (250; cf.: 252)
L. sp. (250, 690; ? sp.: 250, 692)

IV.6.b. *Siliciclastic sediments*

The sandstones yielded only few fossils, but the mudstones are very fossiliferous. There is no difference in fauna, for all species found in sandstones were also found in the mudstones.

From these sediments the following fossils were collected:

brachiopods:
Brachythyrina cf. *strangwaysi* (de Verneuil) (Coll. Winkler Prins)
B. sp. (742)
Crurithyris sp. (742)
Choristites sp. (742)
Dictyoclostus ? aegiranus Böger & Fiebig (742)
Fluctuaria undata (Defrance) (Coll. Winkler Prins)
Karavankina aff. *dobsinensis* (Rakusz) (102)
K. rakuszi Winkler Prins (686)
K. cf. *paraelegans* Sarycheva (Coll. Winkler Prins)
K. sp. (593, 742)
Martinia sp. (742; ? : 523)
Productus cf. *carbonarius* de Koninck (742)
Rugosochonetes acutus (Demanet) (742, 747; cf.: 742) (Pl. I, Fig. 5)
R. skipseyi (Currie) (Coll. Winkler Prins)
R. sp. (593)
Schizophoria sp. (614, 742; ? : 614, 688, 742)
Chonetinella sp. (Coll. Winkler Prins)
Antiquatonia sp. (742; ? : 742)
Linoproductus latiplanus Ivanov (Coll. Winkler Prins)
L. cf. *magnispinus* Dunbar & Condra (Coll. Winkler Prins)
Kozlowskia aberbaidenensis (Ramsbottom) (Coll. Winkler Prins)
K. pusilla (Schellwien) (Coll. Winkler Prins)
Reticulatia huecoensis (King) (Coll. Winkler Prins)
'Horridonia' sp. ex gr. *incisa* (Schellwien) (Coll. Winkler Prins)
Levipustula breimeri Winkler Prins (Coll. Winkler Prins)
Orthotetes sp. ex gr. *radiata* Fischer de Waldheim (742)
Rhipidomella sp. (Coll. Winkler Prins)
marginiferids (691, 742)
productids (653, 738, 747)
rhynchonellids (691)
spiriferids (614, 741, 742)
chonetids (688)

pelecypods:
Annuliconcha interlineata (Meek & Worthen) (742)
Allorisma sp. (102)
Anthraconeilo sp. (102)
Aviculopecten delepini Demanet (102)
Crenipecten foerstii Herrick (102)
Edmondia aff. *arcuata* (Phillips) Demanet (102)
E. aff. *gibbosa* McCoy (738)
E. sp. (251)
Grammatodon cf. *sangamonensis* (Worthen) (100).
? *G.* sp. (102)
Myalina verneuilli (McCoy) Hind (102)
Pecten (Pseudamusium) ufensis (Tschernyschev) Fedotov (102)
P. (P.) purvesi (Demanet) (102, 106)
P. sp. (102)

corals:
Lophophyllidium sp. (738)
Zaphrentites sp. (686, 741, 742)
Carcinophyllum (? Axolithophyllum) sp. (613)

sponges:
Amblysiphonella barroisi Steinmann (614)
Cystauletes mammilosus King (601)

trilobites:
Brachymetopus cf. *ouralicus* (de Verneuil) (688)
Ditomopyge aff. *granulata* (Weber) (688, 742, 747)
? *Paladin* cf. *mucronatus* (McCoy) (617, 686, 688)
? *P.* sp. (691)

goniatites:
? *Pseudoparalegoceras* sp. (614)
cf. *Politoceras politum* (Shumard) (614)
Gonioloboceras welleri Smith (Coll. Dr. Gandl)

nautiloids:
aff. *Metacoceras* sp. (686)

crinoids:
cf. *Talanterocrinus* sp. *Synerocrinus* sp. and *Amphicrinus* sp. (calyx) (691) (Pl. II, Fig. 2)
dicyclic Inadunata (calyx) (759)
synerocrinids (calyx) (691)
unidentifiable calyces (102, 614, 733)
stems and ossicles abundant throughout this member

fusulinids:
? *Fusulina* sp. (488)

forams (siliceous):
cf. *Glomospira* sp. (entire member) (Pl. I, Fig. 10)

forams (other):
Ammodiscidae (all over this member)
many other unidentified forams (idem)

algae:
Ungdarella sp. (574)
Ungdarella-Komia (574)
Archaeolithophyllum johnsoni Rácz (574)
A. johnsoni or *Anthracoporella* sp. (249)
dasycladaceans (748)

land plants (washed-in):
? *Cordaianthus* sp. (686)
Alethopteris cf. *davreuxi* (Brongniart) Goeppert (653)
Linopteris neuropteroides (von Gutbier) (691; cf.: 614, 617, 742)
L. neuropteroides var. *minor* Potonié (610, 617, 691, 707, 713, 742; cf.: 601, 615, 617, 686, 691, 707, 724, 742)
L. neuropteroides var. *linearis* Wagner (742; cf.: 742)
L. cf. *obliqua* (Bunbury) (617, 691, 741, 742)
L. obliqua var. *bunburyi* Bell (617, 688, 691, 742; cf.: 615)
L. subbrongniarti Grand'Eury (601; cf.: 601, 617, 686, 688, 742)
L. sp. (221, 593, 601, 617, 654, 686, 691, 722, 742; ? : 220, 742)
Lobatopteris (Pecopteris) waltoni (Corsin) Wagner (672)
Neuropteris cf. *peyerimhoffi* P. Bertrand and cf. *piesbergensis* Gothan (719)
N. sp. (617, 742; ? : 617, 653, 692)
Paripteris ? gigantea Sternberg (617)
Palmatopteris furcata (Brongniart) (742) (Pl. II, Fig. 4)
P. sp. (686)
? *Pterophyllum* sp. (742)
Reticulopteris munsteri (Eichwald) Gothan (713; cf.: 688, 742)
Sphenopteris polyphylla Lindley & Hutton (611)
Calamites sp. (742)
Sphenophyllum sp. (742)

From the rather large number of leaves washed into these sediments, the conclusion can be drawn that the coast line was not far away. In section 13 there are even some very thin coaly layers, which indicates an even larger supply of plant material. This might have been in the close vicinity of a river mouth, a conclusion that is also drawn from the grain-size distribution (IV.8). As can be seen from the list above (compare with enclosure 1), the other sections in which most plant fossils are found are sections 5, 8 and 11.

IV.7. STRATIGRAPHIC INTERPRETATION

It was previously observed (van Ginkel, 1965; van Loon, 1971) that the fossils from this member give contradictory results in dating. Most of the fossils considered, however, in combination most probably indicate a middle Westphalian C age or uppermost Lower Moscovian – lower Upper Moscovian. In detail the following datings were made:

IV.7.a. *Fusulinids*
To obtain the most detailed dating possible, this member was subdivided into three stratigraphic parts. The lower part indicated top *Fusulinella* A subzone, the middle part top *Fusulinella* A or base *Fusulinella* B1 subzone and the upper part *Fusulinella* B1 subzone. Because of this gradual change in age these datings appear to be very reliable.

IV.7.b. *Brachiopods*
Since brachiopods are present which range from Lower Carboniferous – Podolskian (e.g. *Fluctuaria undata*) as well as brachiopods that are typical Upper Moscovian representatives (e.g. *Kozlowskia* spp., *Karavankina* spp.), this group indicates a Podolskian age.

IV.7.c. *Trilobites*
Considering the known ranges of the species found, Namurian A or B would be the only possible dating (van

Loon, 1971). It should, however, be mentioned that too little is nowadays known of the Upper Carboniferous trilobites to rely on their stratigraphic values. A good example is *Paladin mucronatus*, a species that in Poland was considered as a guide fossil for the marine band Franciszka (X) of the marginal beds from the Namurian A until, in 1961, Bojkowski found this species in many more stratigraphic levels, all of which, however, still in the Namurian A.

Gandl (pers. comm.), who originally identified the specimens from our region as *P. mucronatus* (see van Loon, 1971), now has indications that these specimens are homeomorphous to a *Ditomopyge*. For that reason they have now been recorded as ? *P.* cf. *mucronatus*. Winkler Prins (1968, p. 69) also stated that *P.* is synonymous with *Weberides*, known from the entire Pennsylvanian in America.

Since the other trilobites have also been designated in terms of aff. and cf., it is clear that the dating as Namurian (*B. ouralicus* has a known range from Viséan (!) – Namurian according to Osmólska (1968), and Tournaisian (!) – lower Namurian according to Hahn & Hahn (1969) who doubt whether the Asturian specimens belong to this species) on the basis of our trilobites has no real stratigraphic value. In our opinion it will be necessary to collect much more material in Spain in order to make a statistical analysis of the three species present, especially because all three genera present have the same generic ancestor: *Phillipsia* (Weller, 1937).

IV.7.d. *Goniatites*

These indicate lower Westphalian C (*Politoceras politum*), but the stratigraphic value of this group, too, is negligible. Wagner-Gentis (in press) mentions *G. welleri* and *Pseudoparalegoceras* sp. as Westphalian C, too.

IV.7.e. *Sponges*

The sphinctozoan sponges indicate Upper Moscovian by the presence of *Cystauletes mammilosus* (van de Graaff, 1969).

IV.7.f. *Algae*

The algae from three stratigraphic parts were examined, as were the fusulinids. They clearly show an evolution, but the resulting datings are not in accordance with the other results. The lower part was dated as lower algal zone III of Rácz (1965), the middle part as zone IH or IV and the upper part as zone IV. According to these datings, the Mesao Limestone Member was deposited during a time span that, roughly, can be correlated with middle Westphalian A – middle Westphalian D. Why the algae yield such a diverging result has not as yet been explained.

IV.7.g. *Land plants (washed-in)*

Several species are known from NW Europe. It seems reliable to date the flora as Westphalian C.

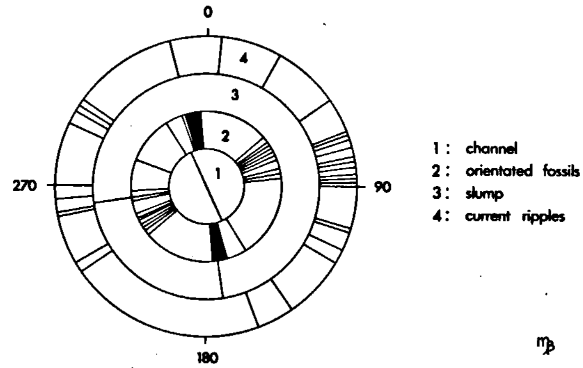

Fig. 44. Current directions in the Mesao Limestone Member.

IV.8. PALAEOGEOGRAPHY

Transport directions could be measured from several kinds of sedimentary structures in this member (Fig. 44).

Many fossil concentrations occur in the mudstones with clearly orientated fossils. We assume this to have been caused by current action, thus indicating a line of transport. Usually it concerns crinoid stems and, to a lesser degree, plant debris. Both show two distinctly preferred directions, roughly perpendicular to each other, indicating transport in NNW–SSE and WSW–ENE directions, possibly one of these perpendicular to the coast line (slope? ; tidal influence?), the other parallel to it due to the action of long shore currents.

Measurements of current ripples give rise to a somewhat more varied picture, but mainly indicate a transport towards the E, with a smaller component pointing to the W. These measurements are supported by the axis of the contortion in the head of a slump, indicating a WSW to ENE transport.

Not in accordance with these measurements are the limestones in the olistolite level mentioned above (IV.5), which probably slid down in southward direction, the same direction which is more or less indicated by a big channel (NNW–SSE).

A hinterland in the N is also probable when we take into account the very sandy facies of this member N of the Puerto de Pando. The pebbles in this member are also found, for by far the most part, in the N-flank. It can further be mentioned that the coarsest siliciclastic parts lie in sections 10, 11 and 13.

Together these observations give the impression of a coast line in the northern part or somewhat N of our area of study (probably along the León line) from which limestone olistolites, and possibly the turbidites too, slid down to the S along the steepest slope. Material that reached the axis of the basin curved to the ENE along this axis. This picture is a confirmation of the conclusions drawn from the measurements in the older deposits (II.8 and III.8). The distribution of the grain sizes also supports this view, indicating a supply from points that were situated N of sections 13, 5 and 10 + 11. This is in accordance with the concentrations of plant material (IV.6) in sections 13, 5, 8 and 11. These points of supply are therefore almost the same as in the older deposits (II.8 and III.8).

The occurrence of mass-transported limestones can be understood if it is borne in mind that the Mesao Limestone Member may have a maximum thickness of 200–250 m, but has been deposited entirely in very shallow water. There must have been a subsidence of the basin, just sufficient to keep up with the sedimentation. This subsidence caused a flexuring at the basinal margin, so that at some moment the slope became too steep for the deposited sediments that slumped into the basin, possibly sometimes passing into a mudflow or turbidite. Even the almost completely consolidated limestones could be broken up. It is not impossible that topographic differences, caused by the differential compaction of limestones and mudstones, accelerated this process.

From the basinal margin in the N siliciclastic material was thus introduced near sections 13, 5 and 10 + 11. This mainly fine-grained material came into a very shallow coastal plain, often influenced by wave action, in which a number of locally limited biogenetic banks developed that supplied large quantities of calciclastics, making the carbonates more important than the siliciclastic sediments. These latter were most probably supplied in small quantities, as is indicated by the possibility of biogenetic banks developing. In the siliciclastic sediments sufficient measurements could, however, be made, to see that the material was mainly transported along the coast, usually to the E, sometimes to the W, probably by small irregularities in the topography of the basin bottom, either by irregular subsidence or by differences in relief due to the occurrence of biogenetic banks.

Possibly on account of fluctuations in the sea level the limestones were often in a supratidal position, so that birdseye structures could develop, while authigenic quartz crystals give indications that even sebkhas originated, probably only locally.

IV.9. CONCLUSIONS

The Mesao Limestone Member consists, for somewhat more than one half, of usually impure limestones, passing laterally and vertically into calcareous mudstones. There are only few biogenetic banks, the only biogenic limestones present, that, however, supplied much calciclastic material. These banks were situated in a shallow coastal plain to which a relatively small quantity of siliciclastic material was supplied from the hinterland. This caused an interruption in the prograding of the deltaic complex, developed by the sedimentation of the older deposits. The shallow character of the mudstones formed by this material is apparent from their sedimentary structures, while the oolites prove this for the limestones. This is supported by the fossils present in large quantities both in the limestones and in the mudstones. The large number of species indicates a middle Westphalian C or uppermost Lower Moscovian – lower Upper Moscovian age, although some fossil groups indicate diverging ages.

Since the 200–250 m thick member was entirely deposited in a very shallow sea, the basin must have subsided over the same 200–250 m in relatively short time. This unstable picture is confirmed by the occurrence of mass-transported sediments that indicate a coast line in the N, a steepest slope in a N to S direction and a longest axis of the basin trending WSW–ENE, while the basin became deeper towards the E, although irregularities existed, forcing sediments to be transported to the W. Probably partly due to the presence of limestones, the topography was somewhat more complicated than during deposition of the Prioro Formation or Lower Sandstone Member.

The very rare occurrence of authigenic euhedral quartz crystals in the limestones, which might indicate a sebkha environment for which, however, an arid climate is required, is remarkable. This assumption is in contradiction with the vegetation found and indications of the climate of the Upper Carboniferous elsewhere in Spain.

CHAPTER V

PANDO FORMATION: UPPER SANDSTONE MEMBER

V.1. INTRODUCTION

All sediments situated above the Mesao Limestone Member, but below the unconformable Cea Group, belong to this member. These sediments lie in the core of the syncline, which is covered by vegetation, probably because of the abundant water circulation in the rocks, facilitated by the large number of faults and joints. This causes the degree of exposure to be sufficient in only a few places. The only sections that could be studied lie in the valley from Prioro to the Puerto de Pando (section 7), and along the road between the same points (section 4). In both cases only the S-flank is exposed rather well. Nowhere could such a section be found in the N-flank.

The original thickness of this member is unknown, since the axis of the syncline, of which this member fills the core, plunges westward, where the sediments become covered unconformably by the Cea Group (Fig. 69). Section 4 shows a thickness of about 165 m, which is a minimum. The movement of several faults could not,

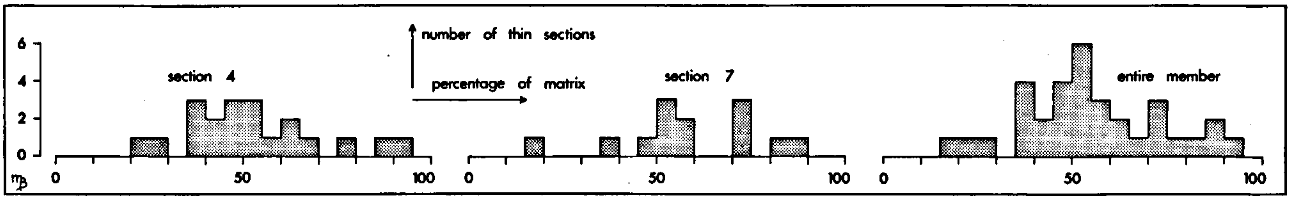

Fig. 45. Modal distribution of the matrix in the Upper Sandstone Member.

however, be established, so that this thickness cannot be more than an estimate.

The two sections mentioned above (4 and 7) were completely sampled. They yielded 21 and 14 samples, respectively, the slides of which were used for the numerical data in the petrographic part (V.3). In addition, 12 slides were studied for additional data. The following samples belong to this member: 253–266, 353–373, 600, 606, 618, 655–656, 675–685, 694–695 and 721.

In nearly all respects (sedimentary structures, modal distribution of the petrographic components, faunal elements, relief features, etc.) there is a striking similarity to the Lower Sandstone Member of the same formation. Only the smaller quantity of limestone is distinctly different.

V.2. LITHOLOGY

The amount of limestones decreases rather abruptly at the transition from the Mesao Limestone Member to the Upper Sandstone Member, allowing the upper boundary of the former to be drawn much more easily than its lower boundary. At the same time, the siliciclastic material becomes much coarser, so that many, though still dirty, sandstones (V.3) were formed, alternating with sandy mudstones; subsequently, however, the proportion of sand increases so that towards the top the sandstones become thicker and more numerous. Even in the uppermost part, however, the sediments never become sufficiently coarse to be called conglomerates.

The almost complete absence of limestones (V.3) is striking, especially after the abundance in the Mesao Limestone Member. The few limestones are all clastic. Nowhere were biogenic limestones found. The sandstones and mudstones are also only sporadically somewhat calcareous. There are some parts which have been decalcified by dissolution of fossils (usually echinoderm fragments), but here, too, the amount of carbonate was never considerable.

V.3. PETROGRAPHY

Microscopical examination of the thin sections made from the 35 samples from sections 4 and 7 reveals many similarities to the Lower Sandstone Member. Although the percentages of the petrographic components are not always the same (matrix and biotite, for instance, are more abundant here, and quartz and authigenic minerals are scarcer), the modal distribution of matrix (Fig. 45) and quartz seems to be quite identical (compare with Fig. 20).

Based on the petrographic composition the samples can be divided as shown in Fig. 46. The small quantity of limestones and the total absence of arenites which are always present, though scarce, in the older deposits, are striking.

V.3.a. *Siliciclastic sediments*

Pointcounting of the 33 samples yielded the following frequencies of the petrographic components:

1. abundant (more than 10 %):

Matrix (55.3 %) is still by far the most important component. By means of X-ray analysis the clay minerals were shown to be only abundant illite with very little chlorite. Irregular 14 Å mixed-layers may rarely be present.

Quartz constitutes 25.5 %, while the group of authigenic minerals constitutes 11.8 %. They will be dealt with in V.4.

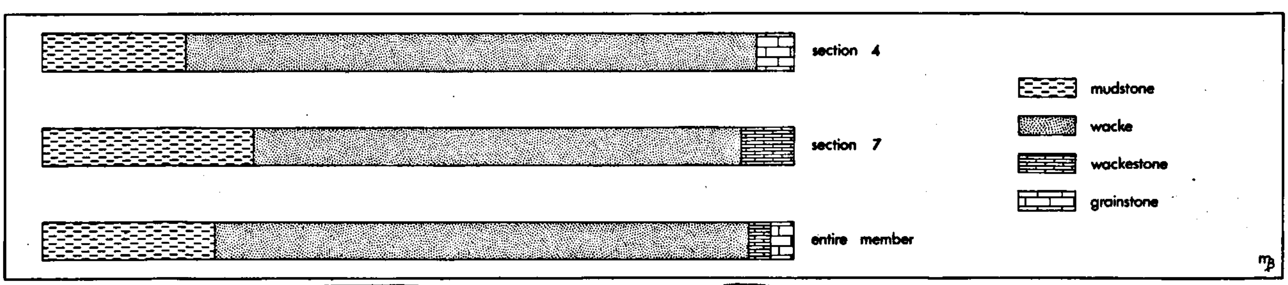

Fig. 46. Distribution of rock types in the Upper Sandstone Member.

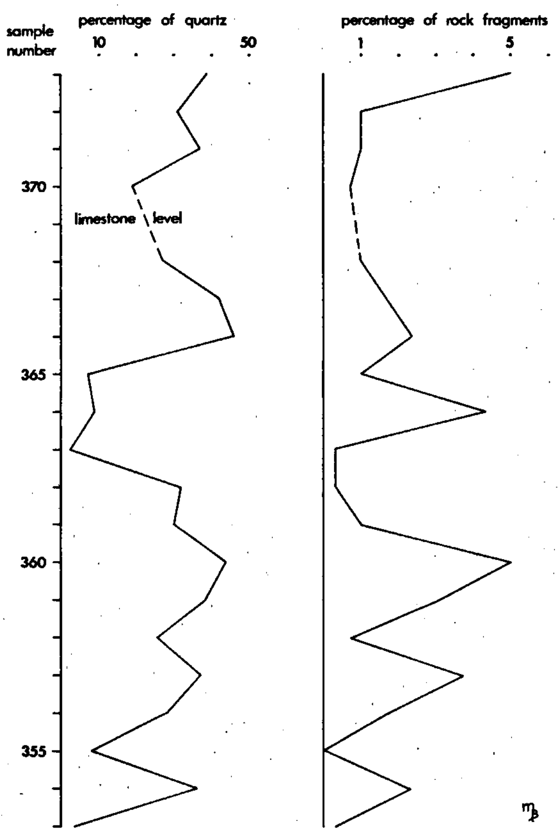

Fig. 47. Relationship between the percentages of quartz grains and rock fragments in section 4 of the Upper Sandstone Member.

2. normal (0.1–10 %):
Biotite has increased considerably (3.6 %) as compared with the older deposits, while muscovite (1.1 %) remains fairly constant. Fossils have decreased to 0.6 %, which is the same as in the Lower Sandstone Member, but only 1/3 of the percentage in the Mesao Limestone Member.

Rock fragments have decreased to 2.0 %. This group contains the following components:

quartzite	: 29 %
phyllite	: 22 %
chert	: 16 %
mudstone	: 11 %
shale	: 6 %
limestone	: 6 %
opal	: 3 %
clay flakes	: 3 %
clay galls	: 2 %

The remainder mainly consists of sandstone. The decrease in the percentage of limestone fragments is striking when compared with that in the Lower Sandstone and Mesao Limestone Members (III.3.a and IV.3.a). This low percentage again improves the visibility of the relationship between the percentages of quartz and rock fragments, masked in the Mesao Limestone Member by the high percentage of intrabasinal limestone fragments (Fig. 47).

3. rare (less than 0.1 %):
Albite, anatase, apatite, cassiterite, chlorite, chloritoid, epidote, iron minerals (all hematite?), opaque minerals (several), plagioclase (the prismatic ?albite), rutile, tourmaline and zircon.
4. negligible (less than 10 grains observed):
Anthophyllite, augite, brookite, corundum, fluorite, kyanite, protolithionite, sillimanite, spinel and volcanic basic (?) glass.

The heavy minerals are much scarcer than in the older deposits: only 0.03 %. Among these tourmaline, especially the brown and green varieties, is by far the most important. The absence of hornblende is conspicuous, since in the older deposits it appeared to be coupled with the angular albite which is indeed present here, even in larger quantities than in the other members of this formation.

The prismatic plagioclase present in the Prioro Formation in relatively large quantities (0.014 %), but almost entirely absent in the lower two members of the Pando Formation, is again relatively frequent (0.02 %), too, nearly always in the form of poorly rounded grains. Because of the apparently rather low resistance to attrition (in the Prioro Formation a considerable percentage is rounded) they must have been derived from a nearby source. Apart from broken specimens, the size of these grains is very constant: approx. 250 x 30 x 30 microns to approx. 200 x 25 x 25 microns.

The relatively small percentage of rock fragments in this member may have been caused by the phyllite falling into matrix-sized pieces, as could many times be observed. Another reason may be the difficult recognition of phyllite, shale and mudstone in the matrix-sized deposits, or of quartzite fragments in wackes that are slightly quartzitic.

The number of mineral species is rather small, as was to be expected from the small number of thin sections (only 33). There are no minerals that give new indications concerning source rocks. The presence of volcanic glass may again indicate renewed volcanic activity. This glass was only found in the Lower and Upper Sandstone Members of the Pando Formation, while it is absent in the Mesao Limestone Member and in the Prioro Formation. This means that it seems to be connected with the coarser sediments. Apparently tectonically active periods resulted in volcanism and coarse sediments, the latter probably being supplied by a more strongly uplifted hinterland.

V.3.b. *Limestones*
Only two limestone levels were found in this member: 369 in section 4, and 257 in section 7. As mentioned above, these are a grainstone and a wackestone, respectively, with petrographic compositions as shown in Fig. 48.

Just as in the Lower Sandstone Member, micrites are absent, indicating a relatively agitated environment, which is in accordance with the siliciclastic sediments

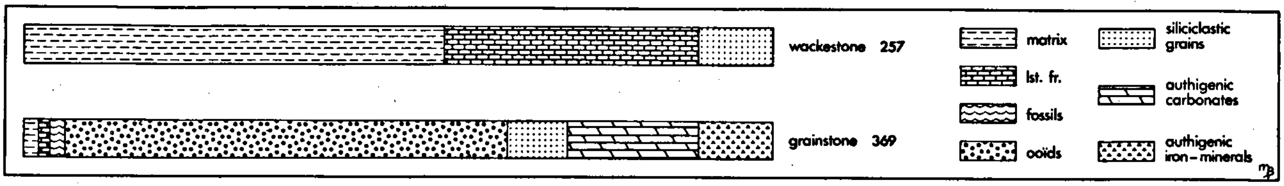

Fig. 48. Petrographic composition of the limestones in the Upper Sandstone Member.

(V.3.a), the sedimentary structures (V.5) and the fossils (V.6).

It should be noted that the limestone in the W (369) is coarser than that in the E (257), possibly again indicating a W to E transport (compare with III.3.b and V.8).

V.4. DIAGENESIS

The relatively low percentage of matrix (55.3 %) is once again linked to a rather high percentage of authigenic minerals (11.8 %).

The following minerals belong to this group:

iron minerals (mainly hematite and wustite, also some goethite and lepidocrocite; amorphous material present in unknown quantities): found as rust-coloured stains (possibly originating from weathered biotites); as small cubes of hematite in pseudomorphs after pyrite; as flat crystals which sometimes push away the surrounding material; as ferruginous fossils (mainly the outermost rim, preserved when the rest of the fossil was dissolved, proving that the introduction of iron occurred, at least partly, before decalcification); and as the filling of cracks;

chlorite: as a weathering product of biotite;

muscovite: as 'booklets' or small needles;

quartz: as veins (sometimes as chert; the occurrence of veins in which the coarser crystals lie against the outer rim is typical); as a fibrous rim (chalcedone?) around grains of other mineralogical composition; as cement or small local growth on quartz grains in quartzitic wackes; and as a replacement of other minerals such as calcite and tourmaline (362) or of fossils (mainly bryozoans) which, however, is a rare feature in this member. In this manner fossils are often replaced by silica in the form of opal;

calcite: as cement in the few limestones; as veins; as total or partial replacement of minerals such as quartz. It can also be found as a recrystallization of fossils (e.g. 364) or of the calcareous matrix, resulting in a microspar (257, compare IV.4);

pyrite: always drusy, especially in the vicinity of fossils. Coarse crystals were not found.

The proportions of these minerals are as follows:

iron minerals	: 97.8 %
quartz	: 1.5 %
chlorite	: 0.4 %
muscovite	: 0.2 %
remainder	: 0.1 %

The occurrence of this high percentage of iron minerals most probably results from the good permeability in these relatively coarse sediments, hence also from the low percentage of matrix (Fig. 49), although this inverse proportionality between these components seems less conspicuous than in the older deposits. This is possibly due to weathering before the deposition of the unconformable Cea Group. This might also be the cause of the decalcification found in many places (see sections of enclosure 1).

V.5. SEDIMENTARY STRUCTURES

The structures indicating a shallow delta-front environment over the entire thickness of this member are again characteristic. In section 4 we find at the base

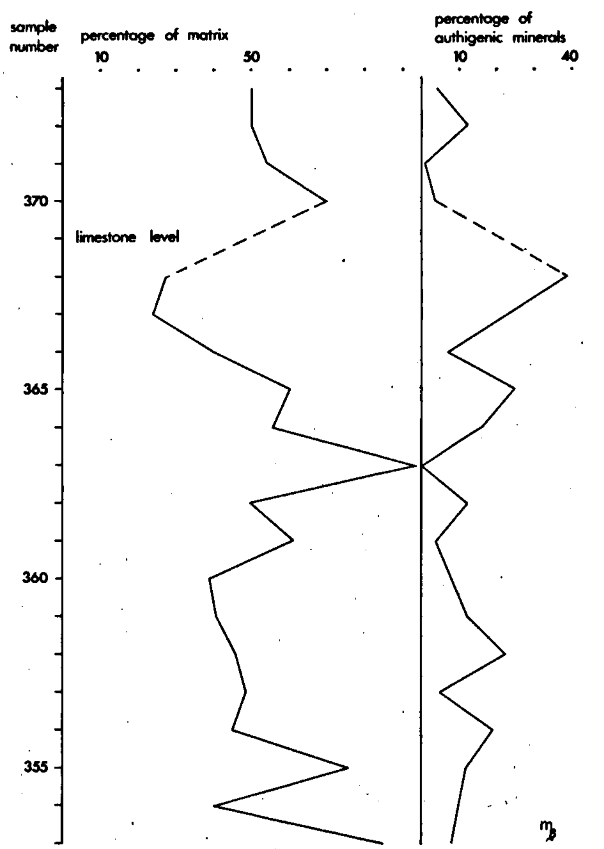

Fig. 49. Relationship between the percentages of matrix and authigenic minerals in section 4 of the Upper Sandstone Member.

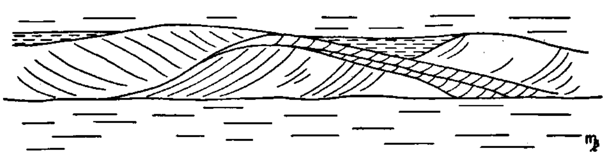

Fig. 50. Current ripples in opposite directions in loc. 695. After field sketch.

some levels with wave ripples, occasionally with an interference pattern, and almost at the top the oolite previously mentioned (369). In between we mainly find channelling sandstones, sometimes with a large-scale cross-bedding (a few decimetres) at the base, and current ripples at the top. Parallel lamination, however, is by far the most common structure (Scruton, 1960). The current ripples indicate transport in all directions, in one layer often even in opposite directions (Fig. 50). This might indicate tidal influence.

Very sporadically (section 1) carbonaceous layers occur. The plant material was always washed in, and does not indicate subaerial exposure. However, the shore-line must have been quite near, especially in the upper part, as is indicated by the grain size, the amount of plant material, the poorly rounded albite(?) prisms (V.3), etc.

Mass-transported sediments are very rare. We observed the following cases:

a. Graded beds: These occur relatively often, sometimes with parallel lamination and current ripples. The grading (the top parts of these layers show a decrease in average grain size of 9.0 to 33.1 % with respect to the basal parts) is caused by a regular decrease in the average grain size of quartz grains, and an increase in the percentage of matrix. In most cases it is not necessary, however, to assume that these layers were deposited by turbidity currents.

The occurrence of true turbidites, however, can be clearly demonstrated at the base of this member in section 5. The turbidites in locality 606 show all Bouma's divisions (see detailed section 4 of enclosure 3).

b. Ball-like structures: In several places (sections 4, 7 and 8) sandstone balls are found. Since these lie in the middle of other sandstone deposits, it is improbable that they are pseudo-nodules. We interpret them as the heads of slumps.

c. Sole-marked beds: In section 1 a sandstone level can be found (600), showing both flute and groove cast at the base. The exposure was too small to allow clear interpretation of the mechanism of deposition of these sandstones. It is possible that turbidity currents were the cause, but there are no distinct indications.

Since the entire member was deposited in very shallow water (turbidites do not necessarily constitute a contradiction (III.5 and IV.5)), the rate of sedimentation must have kept pace with the subsidence of the basin. The fact that the siliciclastic material is much coarser than in the Mesao Limestone Member leads us to assume that the hinterland was uplifted much more rapidly. This gave rise to a larger supply of siliciclastic material that made the environment unfavourable for biogenic limestones. The clastic limestones present were possibly derived from previously formed limestones that became eroded (stage of 'cannibalism' at the end of the basin development, compare, for instance, Lovell, 1969). Since these limestones, too, are very rare, we assume that the limestones of the Mesao Limestone Member were covered so quickly with a large quantity of siliciclastic material that after a very short period erosion was no longer possible.

V.6. FOSSIL CONTENT

In this member fossils are much scarcer than in the Mesao Limestone Member, both in the field and in the slides (V.3). In the few places where a calcareous sediment is encountered, however, a suddenly abundant fauna can sometimes be found. Among a few other less spectacular localities, this refers to a locality a few metres thick along the road from Prioro to the Puerto de Pando (section 4) near kilometre stone 7 (loc. 680, including loc. 364–367), where the following fossil groups can be found together: brachiopods, pelecypods, gasteropods, crinoids (stems, ossicles and some calyces), echinoids, algae, bryozoans, corals (solitary), goniatites (rare), trilobites, fusulinids (rare), siliceous and calcareous forams and ostracods. Tracks are also present in large numbers.

Many of the fossils from this member have previously been mentioned (van Loon, 1971). A more detailed list follows here:

brachiopods:
Avonia (Quasiavonia) echidniformis (Chao) (680)
Antiquatonia sp. (681; ? : 680)
Kozlowskia ex gr. *pusilla* (Schwellwien) (260)
? *K.* sp. (680)
Linoproductus neffedieui de Verneuil (683)
L. sp. (680)
linoproductids (680)
Karavankina rakuszi Winkler Prins (680)
K. sp. (680)
Meekella eximia (von Eichwald) (263, 266)
? *Martinia* sp. (681)
Orthotetes sp. ex gr. *radiata* Fischer de Waldheim (263, 655, 680, 681)
? *O.* sp. (260, 655)
productids (680)
Rugosochonetes cf. *acutus* (Demanet) (680)
R. sp. (680)
Schizophoria sp. (260, 680; ? : 680, 681, 682)
Brachythyrina cf. *strangwaysi* (de Verneuil) (680)
spiriferids (680, 683)

Possibly from this member (loc. 680?), de Alvarado et al. (1942) sampled:
Chonetes sp.
Spirifer bisulcatus Sow.
S. cf. *tornacensis* Kon.
Productus rugatus Phill.

Probably from this member of the Pando Formation, Winkler Prins (1968 and pers. comm.) collected the following brachiopods not found by the present author:
Karavankina aff. *dobsinensis* (Rakusz)
K. cf. *paraelegans* Sarycheva
Linoproductus cf. *magnispinus* Dunbar & Condra
Zaissania aff. *zaissanica* Sokolskaya
Hustedia aff. *remota* (von Eichwald)
Juresania cf. *kalitvaensis* (Likharev)
Levipustula cf. *breimeri* Winkler Prins
? *Fluctuaria undata* (Defrance)
Globosochonetes aff. *waldschmidti* (Paeckelmann)
'Horridonia' sp. ex gr. *incisa* (Schellwien)
Cancrinella sp.
Rhipidomella sp.
Chonetinella sp.
echinoconchids
marginiferids
rhynchonellids

pelecypods:
Grammatodon sp. (353)
Paleoneilo cf. *sharmani* (Etheridge jun.) Demanet (370)
Pecten (Pseudamusium) medium (Herrick) sensu Fedotov (263)
Schizodus? sp.

corals:
Palaeacis sp. (680)
cf. *Zaphrentites* sp. (680)

trilobites:
Ditomopyge sp. (680)
Paladin cf. *shunnerensis* (King) (680)
P. sp. (695)

De Alvarado et al. (1942) mentioned:
Phillipsia eichwaldi Fisch.

goniatites:
Pseudoparalegoceras sp. (364, 680)

Van Ginkel (1965, p. 209) reported a goniatite from the Mesao Limestone Member that was probably found in a part that we consider to be Upper Sandstone Member. It was identified by Kullmann (see also Kullmann, 1962, p. 106–107) as:
Pseudoparalegoceras cf. *russiense* (Tzvet)

crinoids:
calyx of a dicyclic inadunate crinoid (732) (Pl. II, Fig. 3)
stems and ossicles throughout this member

fusulinids:
Hemifusulina sp. (682)
Parastaffella sp. (364)
Millerella sp. (695)
Fusulinella sp. (695)
Fusiella sp. (rather close to *Profusulinella librovitchi* (Dutkevitch)) (682)

Probably from the base of this member in section 5 (we called this part 'transitional beds' in van Loon, 1971) Wagner (1962, p. 3382) collected:
Fusulina cylindrica Fischer (var. ? *hispanica* Gübler) (det. F.T. Barr). These samples were later examined by G. Schmerber (in Wagner, 1966b, p. 25), who identified them as:
Pseudotriticites fusulinoides Putrja
Dutkevitchella böcki Moeller
Hemifusulina moelleri Rauser

These identifications were commented upon by van Ginkel (in van Loon, 1971).

forams (siliceous):
cf. *Glomospira* sp. (364)
cf. *Glomospirella* sp. (364)

forams (other):
Endothyra sp. (695)
? Ammodiscidae (364)
other unidentified specimens (257, 364, 369, 682, 695; ? : 370)

land plants (washed-in):
Linopteris cf. *neuropteroides* (von Gutbier) (721)
L. cf. *neuropteroides* var. *linearis* Wagner (695)
L. cf. *subbrongniarti* Grand'Eury (695)
L. sp. (695, 721)
Neuropteris sp. (721)

Wagner (1962) reported, from the same locality in which he found the fusulinids:
Linopteris neuropteroides var. *minor* Potonié.

V.7. STRATIGRAPHIC INTERPRETATION

As in the Mesao Limestone Member, there are a number of discrepancies between the datings based on the various groups of fossils. This has been dealt with previously (van Loon, 1971). On account of the relative paucity of species (except for the brachiopods), the dating of most groups cannot, however, be very accurate.

The brachiopods indicate uppermost Podolskian or lower Myachkovian, the latter possibility being more probable than the former. This is indicated by species such as *Juresania kalitvaensis*, *Zaissania zaissanica* and *Hustedia remota*, although specimens in hand were identified in terms of aff. and cf.

The trilobite *Paladin shunnerensis* is only known in one locality in the Lower Namurian of Yorkshire (England). For this reason the stratigraphic value must be considered very small.

The goniatite *Pseudoparalegoceras russiense* indicates the goniatite zone G 1 (Kullmann, 1962), which can be considered equivalent to uppermost Namurian. However, Kullmann (pers. comm.) now considers a longer range (Namurian C – Westphalian C) possible for this species, and Westphalian C (or perhaps B) the most probable in this case.

The plants do not indicate more than a general upper Westphalian age, although lowermost Stephanian cannot be entirely excluded.

According to Schmerber, the fusulinids collected by Wagner (1962, 1966b) indicate Podolskian or Myachkovian. Van Ginkel (in van Loon, 1971) dates them as upper Myachkovian or even Kasimovian, but doubts whether these identifications are correct. He himself collected samples from the top of the Mesao Limestone Member (his loc. L 11, our loc. 692; compare with IV.6), probably only a few metres below Wagner's locality, and the fusulinids from these samples were dated as Upper Kashirian or lower Podolskian (van Ginkel, 1965; IV.7).

Our fusulinids indicate the *Fusulinella* zone, most probably the *Fusulinella* B1 subzone (lower Podolskian).

In spite of these contradictions, a Podolskian – lower Myachkovian age is considered most probable, mainly on the basis of the rich brachiopod fauna and the fusulinids. This age is considered to be more or less equivalent to upper Westphalian C – lower or middle Westphalian D, which is in good accordance with the dating of the plants.

V.8. PALAEOGEOGRAPHY

Apart from the flute and groove casts mentioned above (V.5), no orientated sole marks were observed. These casts indicate a WSW to ENE transport, as in the Mesao Limestone Member.

Current ripples point in all directions with two maxima, the larger of which indicates a transport to the SE, the slightly smaller one a transport to the NW. These two directions often occur in one single layer (Fig. 50), and can be explained as the results of tidal currents, more or less perpendicular to a SW–NE trending shore line. A large channel, running NNW–SSE, fits well into

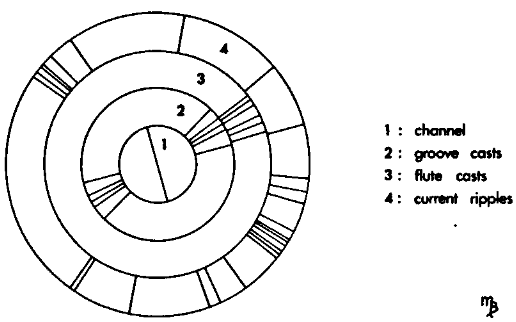

Fig. 51. Current directions in the Upper Sandstone Member.

this picture (Fig. 51). The flutes and grooves might indicate a longshore current parallel to the coast in a NE direction. The grain-size differences in the limestones (V.3.b) possibly also indicate a W to E transport.

The grain sizes of the siliciclastic sediments indicate that the coast probably was not far away, a suggestion that is supported by the abundant plant material, and even by the presence of poorly rounded plagioclase prisms (V.3).

The coarsest sections are 1 and 4. A large quantity of material was probably supplied in the vicinity of these sections. This is the more probable for section 1, because here a number of small coaly layers are found. E of section 7 it was not possible to obtain sufficient data for detecting possible other points of supply, neither was this possible W of section 1.

Since there is a coarsening upwards sequence it must be assumed that the coast line came closer. It looks as though here a final phase has been reached of a deltaic development during which, from the time of deposition of the Prioro Formation onward, gradually more sand was supplied (considering the Mesao Limestone Member as a period of stagnation). At the same time there is a gradual decrease in the influence of mass transport, while plant fragments and sedimentary structures, indicating shallow water, become more frequent. In the top of this member, sandstone banks of 80 cm are already found and the mudstone is scarce. This development would probably have rapidly led to subaerial exposure by further progress of the delta. Since the top part was eroded before the deposition of the unconformable Ocejo Formation, it could not be established whether this continental stage was reached before the uplifting started.

V.9. CONCLUSIONS

Probably due to a stronger uplift of the hinterland, the supply of coarse siliciclastic material suddenly increased, making the environment unfavourable for biogenic limestones. The lack of these almost prevented the formation of clastic limestones as well. The fauna also changed, again resembling that of the Lower Sandstone Member, and also indicating a very shallow marine environment. The exact dating is difficult, but a Podolskian to lower Myachkovian age is considered likely.

The shallow delta-front environment is also indicated by the sedimentary structures and by some sporadic oolites. Indications of mass transport are very scarce, but ripples give good indications of the directions of transport, two major groups of which can be distinguished: one perpendicular to the approx. NE–SW trending shore line as a result of tidal currents, the other parallel to the coast due to a longshore transport in NE direction.

In this period a deltaic development came to an end, although a non-marine stage was not reached (or was eroded).

CHAPTER VI

OCEJO FORMATION

VI.1. INTRODUCTION

The Ocejo Formation, named after the village of Ocejo de la Peña, some 8.5 km SW of Tejerina, constitutes the lower part of what was called the Cea Group by Koopmans (1962). With an angular unconformity (Fig. 52), it lies upon older sediments, in the area studied mainly the Prioro and Pando Formations, but N of Tejerina various strongly folded older Palaeozoic formations. The base of the Ocejo Formation is therefore sharply defined, and can easily be traced in the field, because it is formed by a resistant, sharply based and often erosive conglomerate, the thickness of which varies considerably.

The upper boundary is formed by the contact between this mainly or entirely continental formation and the marine Barranquito Member of the Tejerina Formation (Chapter VII). Compared with the stratigraphic division of the Cea Formation by Helmig (1965) (= Koopman's and our Cea Group), it thus includes the Carrión Member, the Villacorta Beds of the Prado Member and the lowermost part of the Prado Member above these latter beds.

Two sections (12 and 13) of the Ocejo Formation were sampled. For practical reasons the manner of sampling differed from that in the other deposits under study and is dealt with in VI.3. From these sections 35 and 12 samples, respectively, were used for petrographic analysis, and 37 in total for examination of the pebbles. Thin sections of 19 other samples were studied for additional data.

The following localities belong to this formation: 374–436, 527–543, 619–625, 634, 639–647, 657, 758, 761 and 763.

Fig. 52. Contact between the Pando and Ocejo Formations. Shales of the Pando Fm. (PF) nearly vertical (as shown by parallel lamination), erosive basal conglomerate of the Ocejo Fm. (OF) nearly horizontal. Loc. 763.

VI.2. LITHOLOGY

A detailed section (our section 12) was described by Wagner et al. (1969). Helmig (1965) also gives an extensive description of the lithology. Our own observations are shown in the sections of enclosure 2.

The larger part of this formation is characterized by the occurrence of conglomerates that may differ considerably from each other, or even from one place to another within one layer, as far as composition is concerned. Both polymict conglomerates (the majority, e.g. the basal conglomerate in the N-flank) and oligomict conglomerates (e.g. some quartzite conglomerates in the middle of the sequence), or locally even monomict conglomerates (especially in the S-flank) occur. There is a tendency for polymict conglomerates to be more common in the N-flank than in the S-flank. We shall discuss this in greater detail in VI.3. Based partly on the alternation of limestone and quartzite conglomerates, Helmig (1965) drew a lithological boundary (Carrión/Prado Member) at the level of the first limestone conglomerate. The first and second limestone conglomerates, including the material in between, were named Villacorta Beds by Helmig. Since these two first limestone conglomerates may occur at very different stratigraphic levels, such a unit would have lower and upper boundaries suddenly changing in an upward direction. These Villacorta Beds are therefore of no value as a marker horizon, as was considered by Helmig.

The large quantity of conglomerates makes this formation the most resistant of the units dealt with in this thesis. This fact is expressed in the topography, where the conglomerates form the highest ridges (max. 1640 m on the Pico de la Teja), usually without any vegetation. Even some hills in the surroundings of Prioro, consisting of shales of the Prioro Formation, owe their conservation to a small cover of these conglomerates.

The conglomerates alternate with finer-grained material, which latter tends to have a cover of vegetation, especially between two nearby conglomerates. The intervening sediments consist of sandstones, mudstones and coal layers (sometimes with a seatearth), usually in this stratigraphic order and repeated several times in many localities, before another conglomerate begins.

A few metres below the top of this formation the conglomerates stop abruptly and only the sandstone, mudstone and coal sequences are found.

VI.3. PETROGRAPHY

Sections 12 and 13 were sampled in a different manner to those of the older deposits, since the conglomerates tend to be rather monotonous, while the sediment layers in between are often not exposed. For this reason one

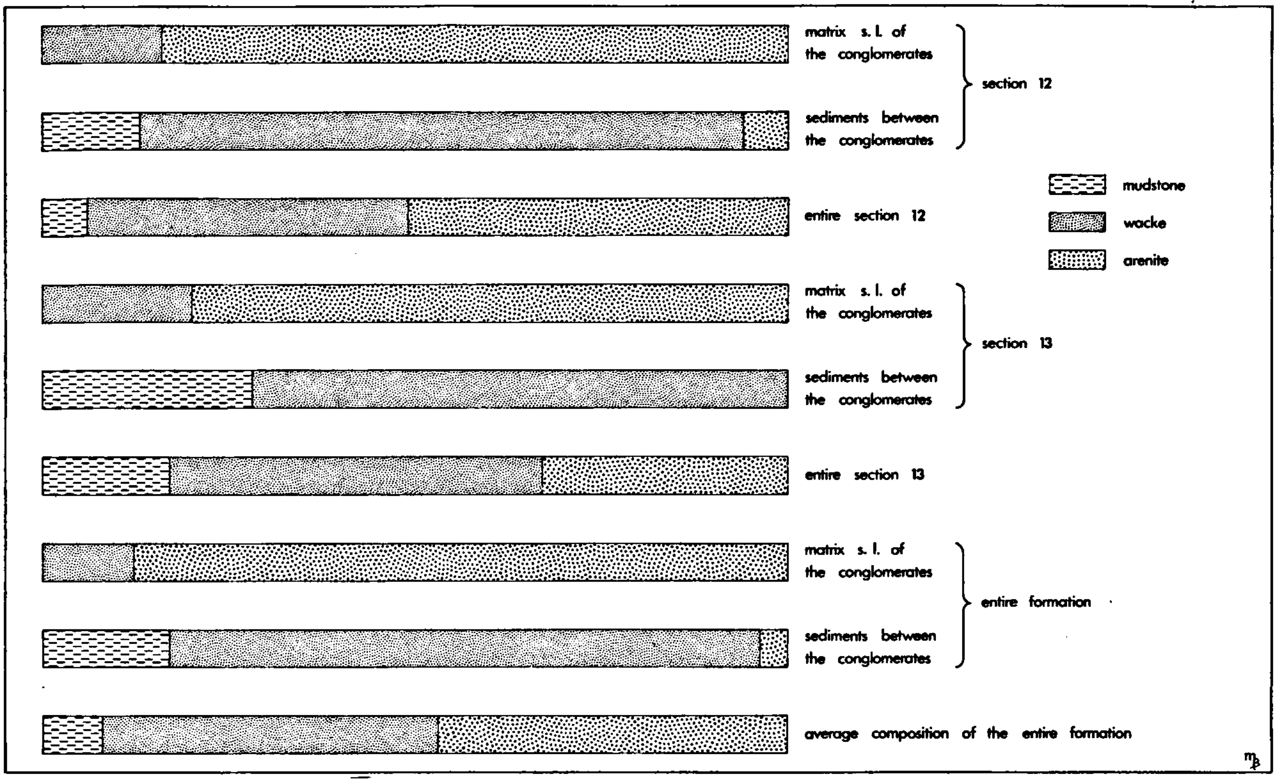

Fig. 53. Distribution of rock types in the Ocejo Formation.

sample was taken from the matrix s.l. of each conglomerate layer where possible (pebbles were also collected, but are not used here; a short description is given below), whereas samples from the sediment layers in between could be taken only sporadically (see enclosure 2).

It appears that considerable differences exist between the degrees of textural maturity of the matrix s.l. of the conglomerates and the sediments in between. These differences also exist between the two sections sampled. This is expressed by the petrographic names of these samples, as shown in Fig. 53.

The petrographic composition may also vary considerably.

For the 47 samples examined altogether, the following values were calculated:

1. abundant (more than 10 %):
Quartz (33.4 %), rock fragments (27.2 %), matrix (20.3 %) and the group of authigenic minerals (17.3 %) to be dealt with in VI.4.

Compared with the older deposits we find considerable differences. In the first place the strong decrease in the percentage of matrix, that no longer even constitutes the bulk. The modal distribution of the matrix-sized material is shown in Fig. 54. The clay minerals encountered are abundant illite and kaolinite, with minor amounts of chlorite and 14 Å mixed-layers, possibly with a few other rare components.

The rather strong increase in the group of authigenic minerals is also striking, but most spectacular is the enormous increase in the quantity of rock fragments. Although these are present in considerable numbers in the sediments between the conglomerates, the high average percentages are especially due to the matrix s.l. of the conglomerates as will presently be shown. Hence a gradual transition exists between the matrix s.l. and the pebbles of the conglomerates. The rock fragments may be grouped as follows:

limestone : 55.3 % (laminated micrite, recrystallized fossiliferous micrite, microspar, recrystallized fossiliferous wackestone, recrystallized pelletiferous packstone, pelletiferous grainstone and crystalline carbonate)
quartzite : 38.7 %
coal flakes : 1.8 %
phyllite : 1.5 %
chert : 1.4 %
mudstone : 0.4 %
shale : 0.4 %

The remainder consists mainly of sandstone and opal, with rare clay flakes and clay galls.

2. normal (0.1–10 %):
Biotite (0.4 %) and muscovite (0.2 %). These minerals constitute no more than about 1/10 of the percentages in which they occur in the Upper Sandstone Member of the Pando Formation (V.3). This low percentage may be due to the high content of plant material (compare with

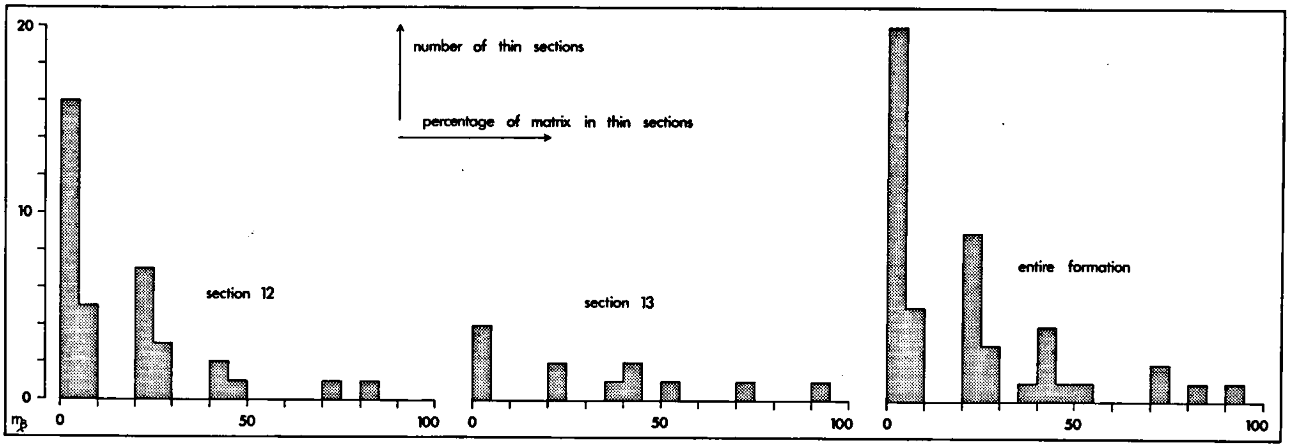

Fig. 54. Modal distribution of the matrix in the Ocejo Formation.

quantity of coal flakes and plant fossils) (Timofeev & Bogolyubova, in press).
3. rare (less than 0.1 %):
anatase, apatite, cassiterite, chlorite, epidote, fossils, iron minerals (hematite?), opaque minerals (several), plagioclase (the prismatic ? albite), potassium feldspar, rutile, tourmaline and zircon.
4. negligible (less than 10 grains observed):
augite, brookite, chloritoid, corundum and dolomite.

The total quantity of the heavy minerals is 0.06 %. It is most conspicuous that, besides tourmaline, cassiterite plays a dominant role. It was most probably derived from broken-up quartzites, since quartzite pebbles were encountered containing many cassiterite grains.

It is also striking that the prismatic ? albite, which, as a matter of fact, is somewhat more rounded here than in the older deposits, was found in only two levels (414–419 in section 12 and 533–534 in section 13), though in both levels in relatively large quantities. This might provide a possibility for correlating these two sections.

If we consider the petrographic composition of the sediments in this formation not as a single entity, as we did above, but distinguish the sediments according to their sections and their mode of occurrence (conglomerate or not), we obtain a much more informative picture.

There are many differences: the matrix s.l. of the conglomerates contains much less matrix-sized material and much more authigenic minerals than the other sediments, while the rock fragments, too, play a more important role. The micas, however, are less frequent here than in the other sediments. These observations indicate stronger erosion in the hinterland, a coarser supply and a higher flow regime during the periods of formation of the conglomerates (see also VI.5).

Comparing sections 12 and 13, situated in the N-flank and S-flank respectively, we find much more matrix in 13 than in 12. This has been influenced somewhat by the relative quantities of sampled conglomerates and other sediments in both sections, but it is also expressed in both kinds of sediment separately. As was to be expected (compare with II.3, III.3, IV.3 and V.3), the percentages of authigenic minerals are higher in 12 than they are in 13. For a large part, this refers to the cement in the conglomerates: in section 12 a carbonate cement is present in 16 out of 19 (= 84 %) conglomeratic samples, while in section 13 this is the case in 1 out of 5 (= 20 %). This also results in the circumstance that among the authigenic minerals calcite is more important in section 12 (58.9 %) than in section 13 (20.4 %) (see also VI.4). At the same time we find that in section 12 the limestone pebbles dominate, while they are relatively rare in section 13. The limestone fragments show the same tendency: in section 12 they constitute 72.5 % of the rock fragments, in section 13 only 12.4 %. On the contrary, micas are more important in section 13 (0.98 %) than in section 12 (0.37 %).

All these observations are petrographic indications of a transport with a N to S component, because (1) section 12 is coarser than section 13, (2) section 12 contains more unstable minerals (carbonates) than section 13 and (3) easily transported minerals (micas) are more frequent in section 13 than in section 12.

We examined 37 pebbles in thin section. It was possible to distinguish many rock types: sandstones and quartzites, both sometimes calcareous, and many limestone types (e.g. a griotte). According to our own observations in the Cantabrian Mountains, and according to several students of Leiden University, these pebbles may derive from the Mora Group (Precambrian), the Herreria and Oville Formations (Cambrian), the Barrios Formation (Ordovician), the San Pedro Formation (Silurian/Devonian), the La Vid, Santa Lucia and Portilla Formations (Devonian) and the Caliza de Montaña Formation (Lower Carboniferous), while for several pebbles no source can be given with any certainty.

VI.4. DIAGENESIS

With 17.3 % the authigenic minerals in this formation constitute a considerable proportion (VI.3), but they are not distributed homogeneously over this formation, as will be shown below.

The following authigenic minerals were encountered:

calcite: this is present mainly as the cement in the conglomerates; further as veins; as secondary calcite around limestone fragments; and as a replacement (e.g. of quartz; Fig. 55);

dolomite: as small rhombohedra in the limestone cement of the conglomerates and as local dolomitization of this cement;

quartz: as a cement in some of the conglomerates, but more often in arenitic sandstones; also as veins and replacement (e.g. of calcite and tourmaline);

iron minerals: usually as stains; sometimes as replacement (Fig. 55) or as veins, while some red sandstones occur in which the grains are coated with an iron oxide or hydroxide layer that in this way may locally serve as a cement. These sandstones often show greenish spots which look like burrowing. According to Kroonenberg (1971), this indicates a position within the phreatic zone;

chlorite: the rather few biotite grains present have been changed into chlorite less frequently than in the older deposits. This apparent low frequency may be due, however, to alteration of the chlorite as a result of the plant material present (Timofeev & Bogolyubova, in press). Radial aggregates occur, but most of the authigenic chlorite present can be found in association with fine-grained authigenic quartz (often in veins); in a few cases chlorite seems to replace quartz (Fig. 56);

muscovite: rare in this formation, but present as radial aggregates and as a weathering product of biotite;

chert: in veins, and once as a cement (527). It can here be shown to be a replacement of an originally calcite cement (cf. Orme & Ford, 1970);

pyrite: probably for want of zoo-fossils, this is rarer here than in the older deposits; it was only found as small spots with very fine drusy pyrite.

Only four of these minerals are present in quantities larger than 0.1 %, as is shown in Fig. 57. This figure distinctly shows the decreasing importance of the iron minerals as compared with the older deposits: in these latter their average percentages were always more than 90!

This formation is also the first in which cementation of sediments plays an important role. This mainly concerns the carbonate cement. Sediments containing no or little calcite are often quartzitic, but secondary quartz is rare. In all these sediments pressure solution is observed which, depending on the quantity of carbonate, leads either to suture contacts between the quartz grains or to stylolitization.

Where originally calcareous sediments have been completely or partially decalcified, a rare phenomenon in this formation, a relatively large quantity of iron minerals is present which indicates that the limestone, at least partially, served as a source. The 'stains' of iron minerals are never intersected by veins, while the few veins of these iron minerals present intersect all other types of veins. This proves that the precipitation of the authigenic iron minerals is caused by a relatively late diagenetic process.

VI.5. SEDIMENTARY STRUCTURES

Fluvial cycles are characteristic of this formation (Fig. 58). The cycles are composed of the following four units:

4. Coal layer, usually thin. A seatearth sometimes occurs underneath. The coal or coaly layer contains often abundant iron minerals (400).
3. Mudstone, often sandy, usually with a gradual transi-

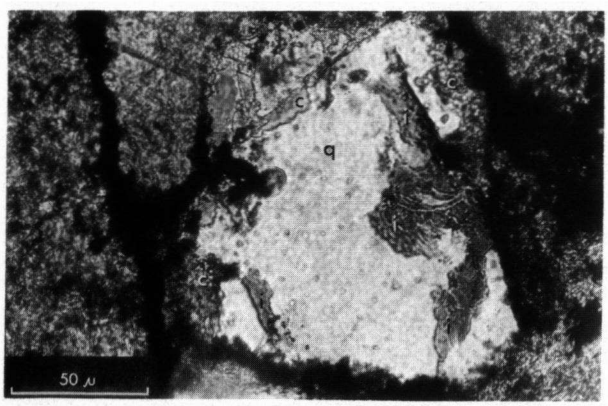

Fig. 55. Thin section of sample 421. Quartz grain (q) partly replaced by an iron mineral (i), partly by calcite (c). Nicols crossed.

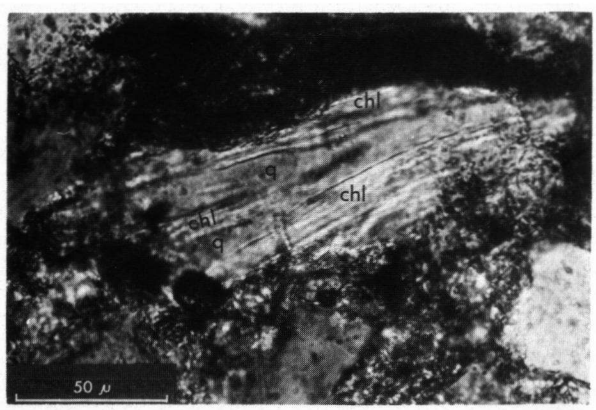

Fig. 56. Thin section of sample 423. Grain consisting by half of quartz (q) and by half of chlorite (chl.) Nicols crossed.

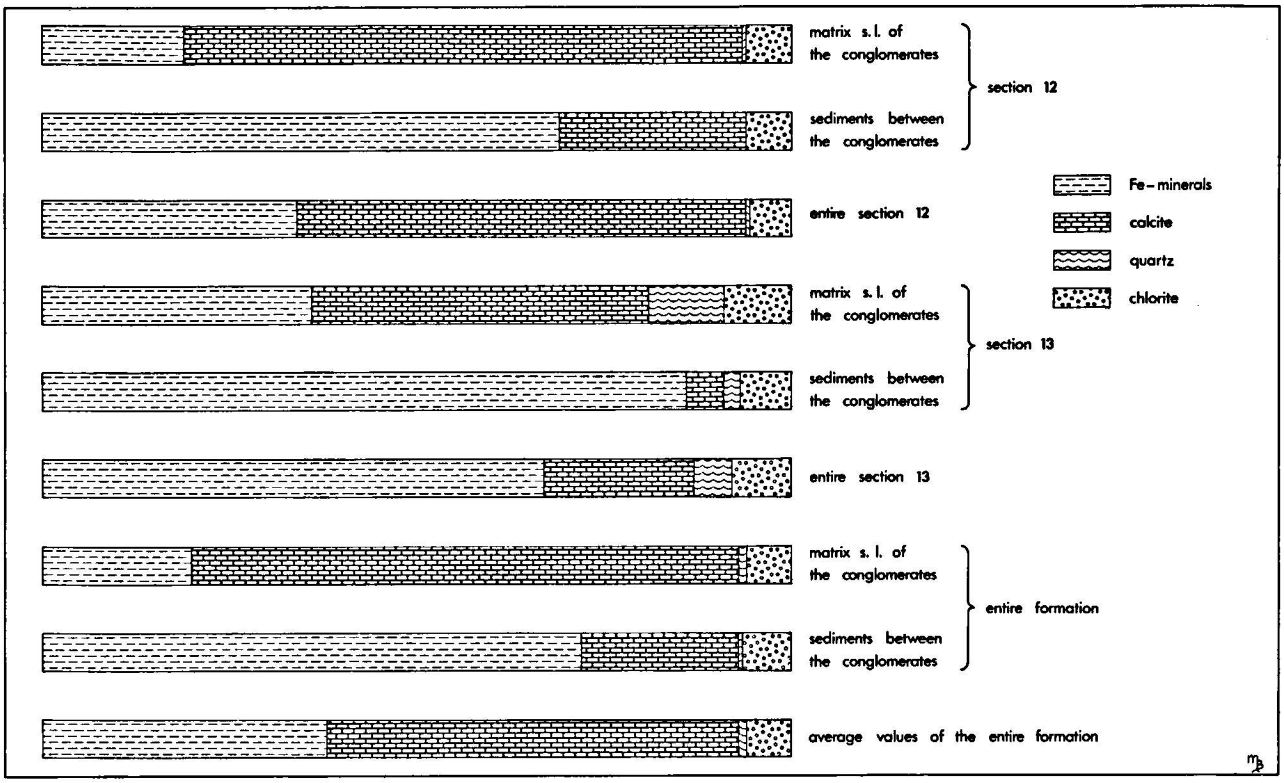

Fig. 57. Relative frequence of authigenic minerals in the Ocejo Formation.

tion from the underlying sandstone. The most common structure is parallel lamination, sometimes wavy lamination as well. Current ripples are much rarer. Much plant debris, sometimes still identifiable pinnules.

2. Sandstone. In most cases rather coarse-grained, with average grain sizes fining upwards. Parallel lamination and current ripples or cross-bedding can nearly always be found. Plant material is common, but rarer than sub 3 and always restricted to debris. The boundary with the underlying conglomerate tends to be rather abrupt.

1. Conglomerate. Usually overlies the underlying finer-grained material in a slightly channelling manner. In most cases the pebbles touch and lie more or less parallel to the bedding plane, sometimes with some local imbrication (Fig. 59). The pebbles are nearly always reasonably rounded, but angular fragments also occur, mainly in the N-flank, especially in the basal conglomerate. If we ignore distinctly broken quartzite pebbles, these angular fragments mainly concern less resistant rocks (limestones and sandstones). The maxi-

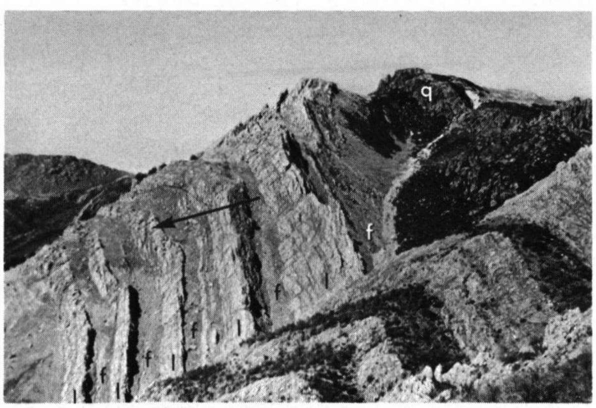

Fig. 58. Part of section 12, showing quartzite (q) and limestone (l) conglomerates with fluvial cycles (f) in between. Belonging entirely to the Ocejo Fm., except for the left (Barranquito Mbr., Tejerina Fm.).

Fig. 59. Detail of limestone conglomerate. Resistant quartzite pebbles (sometimes imbricated) show a higher relief. Loc. 619 (basal conglomerate in section 12).

Fig. 60. Transition from conglomerate into finer-grained fluvial material. Only a few conglomeratic strings are left. Mega cross-bedding in the coarse-grained sandstone. Loc. between sections 1 and 4.

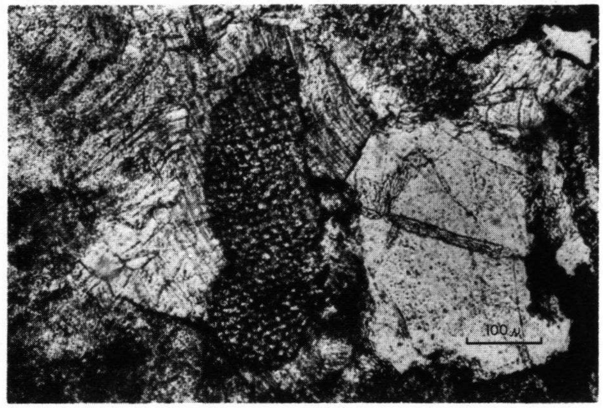

Fig. 61. Thin section of sample 412. Echinoid fragment in the matrix s.l. of a conglomerate.

mum size of the pebbles depends on the rock type. Considering one of these (quartzite is the most suitable), we find that in the conglomerates grading may occasionally be present. It is sometimes even possible to distinguish several 'subunits' in one single conglomeratic bank by this grading. Sandstone intercalations may occur especially in the top of the conglomerates or of a 'subunit'. These sandstones usually show cross-bedding. The presence of pebbles with a size of several cm in the foresets indicates a very strong current. Previous authors (e.g. Wagner et al., 1969; de Jong, 1971) even mention torrential currents. In a few cases the transition into sandstone unit 2 is not abrupt but very gradual, showing more and more sandy intercalations, until only a few conglomeratic strings are left (Fig. 60). The conglomerate layers, that may be up to several tens of metres thick, can be traced laterally for hundreds of metres, but usually subsequently wedge out, although there are levels that can be traced over our entire area.

In most cases the cycles as described above are incomplete (see detailed section 6 of enclosure 3), especially as far as the conglomeratic units are concerned which, however, on account of their thickness and resistance, are the most spectacular where present. One cycle, without a conglomerate at the base, was studied in somewhat greater detail:

a sandstone bank (460) 405–210 cm thick lies, with an erosive contact, upon the underlying mudstones. At the base small pebbles of up to 12 mm are present, as well as a few washed-in stem remains (cf. *Calamites* sp.). Rare wood fragments occur in all levels of this decalcified bank, while the sedimentary structures present, viz. current ripples of small-scale size and parallel lamination, become more frequent in upward direction. This part passes into a finer-grained layer approx. 350 cm thick that is still sandy at the base (461) (petrographically a wacke, somewhat finer-grained than wacke 460), but that is silty at the top (463) (petrographically a coarse mudstone), where ferruginous concretions, current ripples, parallel lamination and unidentifiable washed-in wood fragments are common. In the middle of this irregularly graded layer another small channel is present which has a size of approx. 200 x 16 cm. Petrographically it is a very coarse wacke (462).

Above the silty part (463) follows a seatearth 20 cm thick (464), with 45–50 cm coal above it. A new sequence, starting with a fine-grained sandstone (465), rests erosively upon this coal layer.

Where the conglomeratic part is absent the sandstone usually shows an erosive base; clay galls and other coarse material may be found in the lower part. The coal layer is also absent in most cases. This may be due to erosion during the next cycle of deposition, but is probably more often caused by non-deposition as a result of the following cycle beginning before sufficient vegetation could develop.

On the basis of these observations we consider this sequence as fluvial. The conglomerate layers, however, are not a channel lag deposit, since they show many differing characteristics: they tend to be too thick and too extensive laterally, they contain too many too angular pebbles, and possibly show too little imbrication. Considering also the polymict character of these conglomerates (to the S they become less polymict, VI.3), an origin as alluvial fan built up by sheet floods seems the most probable. The shape of many quartzite pebbles that look like fragments, bounded by joints and subsequently somewhat rounded (Bodenhausen, Univ. of Leiden, pers. comm.), is in agreement with this. The variation in current direction, as shown by the imbrication in these conglomerates (VI.8), may also indicate this origin. These conglomerates belong to Reading's (1970) facies type (i)a.

In the conglomerates remarkable quantities of marine fossils (Fig. 61) sometimes occur in the matrix s.l. (a few per slide). Since marine fossils are also present in the limestone pebbles and limestone fragments (VI.3), the fossils may also be rock fragments. It is, however, also possible (see below and VII.3, 5), that these alluvial fans extended into the sea or came quite near to it, allowing a marine influence during periods of minor transgression or even during heavy storms. This matter has not, however, been investigated.

As was mentioned in VI.2, there are polymict, oligomict and even (possibly only locally) monomict conglomerates. Since these can suddenly alternate, at least two supplying media should be considered. Because the limestone conglomerates are never pure, but some quartzite conglomerates are, one of these two media must have supplied exclusively quartzite pebbles, the other exclusively limestone pebbles or both limestone and quartzite pebbles with minor quantities of other rock types. There is yet another reason for assuming these two different supplying media: the roundness of the pebbles. The quartzite pebbles tend to be better rounded, although they are more resistant. This suggests either a reworking of a pre-existing conglomerate or a first local rounding with transport to the depositional site afterwards. Reworking is improbable, since only the Westphalian Curavacas conglomerate could yield so many pebbles. In II.3 we already mentioned, however, that this conglomerate most probably did not extend to this area. The only other possible source of pebbles is the Prioro Formation which, however, did not contain enough pebbles to give rise to the considerable amounts of conglomerates under discussion here. We must therefore assume, as we did for the Prioro Formation, that fragments of an eroded quartzite were rounded locally in rather agitated water. Since this does not seem the case for the other pebbles, this indicates at least two supplying media.

The explanation by Helmig (1965) of the occurrence of conglomerates here, viz. strong tectonic activity, seems very plausible. Because of the degree of rounding and the lithological composition of the pebbles we assume that a repeatedly active fault zone (the León line) must have caused an escarpment. Since the older Palaeozoic formations had been intricately folded (see map by Helmig and Rupke, 1965), possibly already in the Famennian (Wagner, 1970) and again during the Palentian (or Curavacas) and Leonian phases (Chapter IX), many of these formations became eroded along this escarpment at the same time, resulting in the great variation in pebbles. This erosion of the escarpment and the deposition of the material led to a profile of equilibrium, expressed by the formation of alluvial fans that wedged out to the S. This explanation of the conglomerates deviates considerably from that by de Jong (1971).

In quieter times meandering rivers, sometimes preceded by braided ones, developed on these fans in rather flat plains with vegetation (Reading's, 1970, facies (i)b, (i)c and (ii)), as appears from the sediments between the conglomerates, which show cycles ending in coal layers (sometimes with a seatearth). One mudstone level with a few quartzite pebbles might indicate a lacustrine environment or a crevasse splay deposit.

In the uppermost part of this formation no more conglomerates occur. It is reasonable to suppose that some degree of tectonic quiet was established.

The present author is not convinced of the assumption by Helmig (1965) and by Wagner et al. (1969) that this entire formation is continental. There are places which, in our opinion, could be shallow marine or littoral as well. Apart from the possibility that the sea sometimes influenced the conglomerates (see above), this in the first place concerns a locally strongly quartzitic sandstone bank (639, 642) that can be traced a few metres above the basal conglomerate in the entire S-flank (except S of Tejerina due to faulting) and in the entire N-flank (except for the westernmost part). Although it still contains not very resistant rock fragments, it is well sorted and clean, and shows no other structures than parallel lamination and very rare small current ripples usually with hardly inclined foresets. Sometimes mudstone intercalations can be found, a few mm to a few cm in thickness, which are hardly ever exposed. Based on these observations it seems more plausible that this bank was deposited behind a barrier in an environment sometimes agitated by wave action, and subsequently in a very quiet environment, possibly at low tide, rather than that it formed part of a fluvial sequence. Since no fossils were found, a definite proof of a marine character cannot, however, be given.

Higher in the sequence, especially in the E part of our area (though also present in the SW), similar quartzitic laminated sandstones occur, interpreted as beaches, that can even be more mature, both compositionally and texturally. An example is sample 645, that consists of the following components, as calculated by means of point-counting:

quartz grains	: 74.7 %
authigenic quartz	: 24.3 %
authigenic chlorite	: 0.3 %
matrix	: 0.7 % only

If in these cases marine or littoral sediments are indeed present this is very important for the palaeogeographic picture. The fact that only few winnowed sediments are present might indicate a dominant supply by longshore currents (Walker & Harms, 1971). Because we did not spend much time on these layers, further study will be required in the future to solve this problem.

VI.6. FOSSIL CONTENT

The flora of section 12 was studied in detail by Wagner (in: Wagner et al., 1969). He mentions a large number of species from this member, obtained from localities that cannot always be compared with our localities. For this reason we give Wagner's locality numbers, preceded by a 'W', and refer to his publication for an exact location. In this formation Wagner found:

Linopteris cf. *elongata* Zeiller (W 1806)
L. neuropteroides von Gutbier (above W 1833–35, W 1184)
L. neuropteroides var. *linearis* Wagner (W 1184, W 1822, W 1823)
L. sp. (above W 1822)

Neuropteris scheuchzeri Hoffmann (W 1181 (= some 3.5 km more to the E), W 1184, W 1757–1804, W 1806, W 1822, above W 1822, W 1833–35, above W 1833–35)
N. ovata Hoffmann (W 1184, W 1757–1804, W 1805, above W 1833–1835)
Mixoneura brittannica (von Gutbier) (W 1757–1804)
M. sp. (W 1181, W 1796)
Callipteridium cf. *armasi* (Zeiller) (W 1796)
C. (Praecallipteridium) jongmansi (P. Bertrand) (W 1183, W 1757–1804)
Pecopteris unita Brongniart (W 1184, W 1757–1804, W 1833–35)
P. ocejensis Wagner (W 1181)
P. dentata Brongniart (W 1757–1804)
P. monyi Zeiller (W 1184)
P. cf. *rarinervosa* Corsin (W 1184)
P. hemitelioides Brongniart (W 1184, W 1822)
P. sp. (W 1184, W 1805, above W 1833–35)
Sphenopteris ovalis von Gutbier (W 1833–35)
Sph. cf. *macilenta* Lindley & Hutton (W 1181)
Sph. (Oligocarpia) gutbieri Goeppert (W 1184)
Sph. cf. *nummularia* von Gutbier (W 1184)
Sph. sp. (W 1757–1804)
Annularia stellata (von Schlotheim) (W 1181, W 1184, W 1757–1804, W 1822, W 1833–35)
A. sphenophylloides (Zenker) (W 1184, W 1757–1804, W 1822, above W 1822, W 1833–35, above W 1833–35)
Sphenophyllum emarginatum Brongniart (W 1181, W 1184, W 1757–1804, W 1833–35)
Sph. sp. (above W 1833–35)
Calamostachys tuberculata Sternberg (W 1757–1804, W 1822)
Alethopteris missouriensis D. White (W 1184, W 1757–1804, above W 1833–35)
A. grandinioides Kessler (W 1181, W 1184)
A. grandinioides Kessler var. *grandinioides* (W 1757–1804)
A. grandinioides var. *subzeilleri* Wagner (W 1757–1804)
A. cf. *lesquereuxi* Wagner (W 1757–1804)
A. lesquereuxi var. *cerverae* Wagner (W 1184)
A. bohemica Franke (W 1184)
A. kanisi Wagner (W 1822)
A. sp. (above W 1822, W 1833–35)
Odontopteris cantabrica Wagner (W 1184)
Mariopteris cf. *rotunda* Huth (W 1184)
cf. *M. nervosa* (Brongniart) (W 1184)
M. sp. (W 1833–35)
Pseudomariopteris ribeyroni (Zeiller) (W 1184)
Dicksonites pluckeneti (von Schlotheim) (W 1184)
Polymorphopteris polymorpha (Brongniart) (W 1184)
P. gothani (Guthörl) (W 1184)
Lepidodendron cf. *scutatum* Lesquereux (W 1184)
L. sp. (W 1757–1804, W 1833–35)
Cordaites sp. (abundant in many levels)
Lobatopteris sp. (W 1833–35)
Artisia sp. (W 1757–1804)

Helmig (1965) also gives lists of plants found by him and other students of Leiden University. The locations are again given in their original form, preceded by 'H' (Henkes), 'HM' (Helmig) or 'S' (Savage). It was not possible to identify all their localities with certainty. Most probably the following comparisons can be made:

H 356 = (probably) HM 1532 = some 20 m N of our loc. 657
H 499 = some 100 m stratigraphically above our loc. 657 (= top Ocejo Fm.)
H 593 = some 60 m N of our loc. 657
H 594 = some 75 m N of our loc. 657
H 875 = (probably) our loc. 429
H 876 = (possibly) W 1833–35 = (possibly) between our loc. 409 and 410

Helmig (1965) mentions:

Neuropteris scheuchzeri Hoffmann (H 356, HM 1532)
cf. *Lobatopteris alloiopteroides* Wagner (H 356)
L. sp. (H 593)
Sphenophyllum emarginatum Brongniart (H 356; cf.: S 71)
Alethopteris grandini Brongniart (HM 1532)
A. kanisi Wagner (H 499) (listed as *A. kamissi* Wagner)
A. sp. (H 356, H 875)
Callipteridium gigas (von Gutbier) (H 499, H 593, H 594, H 876)
C. pteridium Zeiller (H 593, H 594, H 876)
cf. *C.* sp. (S 71)
Annularia stellata (von Schlotheim) (H 594, H 876, HM 1532, S 71)
A. sphenophylloides Zenker (H 876)
Lepidodendron geinitzii (?) (H 593) (this species is not known to the present author)
L. sp. (H 356, H 593)
Sublepidodendron lycopodioides (Sternberg) Nathorst (S 71)
Asterophyllites equisetiformis von Schlotheim (S 71)
Pecopteris dentata Brongniart (H 594)
P. sp. (H 594, H 875, HM 1532, S 71)
cf. *Astherotheca* aff. *cyathea-arborescens* (S 80)
Odontopteris cf. *obtusa* Brongniart (H 499)
O. cf. *reichi* von Gutbier and cf. *brardi* Brongniart (H 499)
O. sp. (H 499)
Sigillariophyllum sp. (S 71)
cf. *Calamites* sp. (S 80)
Cordaites sp. (H 876)

The present author did not collect many plants because of the knowledge already available. We found:

Calamites carinatus Sternberg (657)
C. cf. *suckowi* Brongniart (657)
C. sp. (763)
Astherotheca sp. (539)
Pecopteris unita Brongniart (642)
P. cf. *ocejensis* Wagner (429)
cf. *P. (Polymorphopteris) polymorpha* Brongniart (642)

P. sp. (642)
P. (Astherotheca?) sp. (429)
Astherotheca sp. (539)
Calamostachys tuberculata Sternberg (657)
Astherophyllites? sp. (429)
Callipteridium (Praecallipteridium) armasi (Zeiller) Wagner (429, 539) (Pl. II, Figs. 9–9a)
C.? sp. (763)
Sphenopteris cf. *gutbieri* Goeppert (657)
Sph. sp. ex gr. *obtusiloba-nummularia* (429)
Cordiates sp. (642)
Linopteris neuropteroides var. *linearis* Wagner (539)
cf. *Reticulopteris munsteri* (Eichwald) Gothan (539)
Annularia stellata (von Schlotheim) Wood (429)
A. sphenophylloides (Zenker) von Gutbier (429)
Neuropteris scheuchzeri Hoffmann (539)
Lepidodendron sp. (763)
Alethopteris ambigua Lesquereux (539, 657) (Pl. II, Fig. 13)
A. grandinioides Kessler var. *grandinioides* (763)
A. cf. *kanisi* Wagner (763)
A. sp. (763)

It is remarkable that all these four investigators found some species not found by the others. These differences cannot be caused by differences in identification, as all these fossils, except those of Savage (det. van Amerom, Heerlen), were identified by Wagner (Sheffield). This therefore indicates a very rich flora; which is as yet incompletely known. Since Wagner et al. (1969) made exhaustive collections in our section 12 we only can explain this by assuming large differences in flora from place to place.

The marine fossils mentioned in VI.3 and VI.5, the origin of which is not yet understood (rock fragments or marine influence) concern:

crinoids (stems and/or ossicles) (392, 421, 426)
echinoderms in general (391, 407, 411, 412, 416, 417, 421, 422, 425, 428, 647, 657) (Fig. 61)
bryozoans (391, 392, 412, 416, 420, 421, 422)
brachiopods (416, 421)
calcispheres (426)

VI.7. STRATIGRAPHIC INTERPRETATION

Wagner et al. (1969) dealt with the stratigraphy of this formation in detail, based on the floral content. Better than we can do they argued that, on account of the evolution of the flora present, this formation must be considered as uppermost Westphalian D in the lower part, and possibly somewhat younger in the upper part. Up to locality W 1184 only Westphalian elements are found, whereas above this point younger forms also occur, although the assemblages are still distinctly older than basal Stephanian A. Wagner et al. located this part in the Cantabrian, a stage proposed by Wagner (1966a, 1969), with a proposed lower boundary of the Cantabrian in locality W 1184.

During the latest session of the Subcommission on Carboniferous Stratigraphy (Krefeld, August 1971), however, the lower boundary of the Cantabrian was placed at the base of the Lores Limestone (upper limestone of the Corisa Formation) in a section in the Casavegas syncline in the province of Palencia. According to Wagner (pers. comm.), the Lores Limestone is probably time equivalent to the Barranquito Member of the Tejerina Formation in our area. Since the base of the Barranquito Member is the same as the top of the Ocejo Formation, this implicates that the entire Ocejo Formation should be considered as uppermost Westphalian D.

The problems concerning the Cantabrian stage will be dealt with in VII.7.

The boundary between Westphalian D and Cantabrian in any case differs from Helmig's (1965) boundary between the Carrión and Prado Members. Helmig's division is therefore neither based on lithology (VI.2) nor on valid bio-stratigraphic arguments. For these reasons we did not follow his division of the Cea Group (or Formation), but that by Wagner et al. (1969).

VI.8. PALAEOGEOGRAPHY

The coarse-grained material of the Ocejo Formation indicates an emerging hinterland or a subsiding basin. Because of the thickness of this formation, both are probable. The rapid thinning to the ESE (see also map by Helmig and Rupke, 1965) indicates a transport in this direction. Other data also suggest this: the decreasing amount of less resistant pebbles and rock fragments (VI.3), and of authigenic carbonates (VI.4) in this direction, while the imbrication in the conglomerates also indicates this transport direction (Fig. 34).

The depositional environments are quite different from those in the older deposits: from an escarpment in the N, originating from tectonic activity (León line) which was more intensive in the W than in the E, large alluvial fans extended towards the ESE. At the base of these and, during periods of relative tectonic rest, upon these fans a plain was situated in which rivers flowed, mainly meandering, sometimes braided. The current ripples in their deposits indicate a transport more or less parallel to the feet of these alluvial fans (Fig. 62).

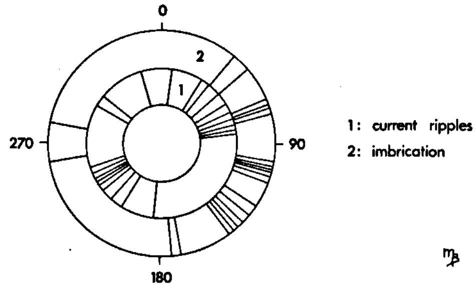

Fig. 62. Current directions in the Ocejo Formation.

This rather flat plain probably gradually passed into a littoral environment with beaches and barriers, and possibly even into a fully marine environment. The sea transgressed (if indeed it transgressed) our area from the SE to the NW, as is shown by the location of the possibly marine or littoral sediments.

Although this picture is quite different from that of the older deposits it can be explained by assuming a coast line that was situated only a few km further to the S or SE.

VI.9. CONCLUSIONS

After a rather short period (as shown by the datings) of – only locally distinct – folding and uplifting the deposition of the Ocejo Formation began on a rather flat eroded surface. Although the basal conglomerate is erosive one cannot speak of valley fills as was done by Wagner et al. (1969).

The sediments mainly consist of an alternation of alluvial fans and sequences of meandering rivers. The plant material that is present in large quantities and that even led to the occurrence of coal layers indicates an age of uppermost Westphalian D.

The thickness (max. here 571 m, and 583 m according to Wagner et al., 1969) and the coarse and often angular material indicate a rising hinterland and a subsiding basin. At most times the subsidence of the basin seems to have kept pace with the filling up. Small irregularities led to possibly littoral and marine sediments by a sea transgressing our area from the SE. A real marine character could not, however, be established with certainty.

This sequence can be considered as a logical continuation (top sets) of the older deltaic development.

CHAPTER VII

TEJERINA FORMATION: BARRANQUITO MEMBER

VII.1. INTRODUCTION

Since the Barranquito Member is defined as the marine band in the local development of the Cea Group, situated just above the strongly conglomeratic part, the lower and upper boundaries are theoretically well defined. There may, however, be some difficulties in tracing them exactly, on account of the entirely or almost entirely barren parts (VII.6) and the lithology which is usually identical to the surrounding deposits (VII.2, 3). The limited thickness (always less than 50 m), however, precludes important mistakes during mapping when using scale 1:25,000, for the thickness has always to be exaggerated on the map.

The low resistance of the sediments is the cause of much vegetation and consequently few exposures. In the E bend of the syncline and in many parts in the NE of the Cea area this member could not be found. In the former case this may be due to faulting, in the latter, however, it is the result of the low degree of exposure.

This formation is named after the village of Tejerina in the W part of our area, while this member is named after the small brook (Spanish: barranquito) that runs from the Pico de la Teja along Tejerina to the S. Along this brook runs a path, which forms the only well-exposed section through this member (section 12). Six samples were collected from this section, if possible every 5 m, for petrographic examination. This was also done in section 13 that yielded only three samples. Additional petrographic data were obtained from three more slides. Our localities 437–442, 544–546, 635–636 and 752–757 belong to this member. The small vertical extension of this member and the poor degree of exposure are responsible for this small number of samples and localities.

VII.2. LITHOLOGY

The lithology can differ considerably from place to place. The finer material (silt and clay) nearly always dominates. Only in section 12 (compare with Wagner et al., 1969, Fig. 2) is there a distinct alternation between mainly sandy and mainly muddy parts, but even there this member is considerably less resistant than the surrounding deposits. Only one other locality was found in which the material consists of sandstone (usually decalcified). The only reason for considering this exposure to belong to the marine band, however, is the occurrence of an imprint (3 mm) of a crinoid stem. Localities where the deposit is still calcareous are scarce, and always concern silty deposits. Silt is the most frequent grain size: in some localities nearly all material is silt.

From NW to SE the material becomes finer-grained. Whether this is a gradual change could not be established because of the large unexposed parts. In section 13 (S-flank) we still find some fairly clean sandstones, however.

Although this marine band is very useful for correlation purposes, we cannot call it a distinct marker bed on account of the varying lithology (usually identical to the sediments of the surrounding deposits), the lack of marine fossils in several places and the dense vegetation. Helmig (1965) did not even mention the existence of this marine band.

VII.3. PETROGRAPHY

On the basis of the samples collected it is impossible to obtain a reliable picture of the petrography of this member. The reason is that only section 12 could be sampled in a reasonable manner (VII.1) and that this section differs considerably from the others by a much coarser lithology (VII.2). Samples from a part that will be interpreted as a littoral deposit (VII.5) were the only to be taken in the other section sampled. This littoral part, however, is also one of the few other coarse localities. In other places only a number of isolated samples could be taken that cannot, of course, give an idea of the petrography of the entire member in that place.

It will be understood that the numerical data on the petrography of this member are not reliable. Nevertheless, we shall give the results of our examination:

section 12: 6 samples: 5 wackes
1 mudstone
section 13: 3 samples: 1 arenite
1 wacke
1 mudstone

This result indicates that mainly wackes are present with some minor amounts of arenites and mudstones. This is in contradiction with the impression in the field that most sediments are mudstones.

The various petrographic components in the total of the 9 slides occur in the following percentages:

1. abundant (more than 10 %):
Matrix (40.8 %). Just as in the marine deposits of the Prioro and Pando Formations, and in contrast to those in the fluvial Ocejo Formation, matrix is the most important component. The modal distribution, shown in Fig. 63, more resembles that in the other marine strata than that in the Ocejo Formation. By means of X-ray analysis the clay minerals were identified as mainly illite and kaolinite. Some chlorite, septachlorite and 14 Å mixed-layers, probably containing montmorillonite, are also present.

Quartz (36.6 %) is also an important constituent, while the authigenic minerals (14.8 %) will be dealt with in VII.4.

2. normal (0.1–10 %):
Biotite (0.6 %), muscovite (0.2 %) and the group of rock fragments (6.5 %) belong to this class. The latter group is composed of various rock types in the following proportions:

limestone	: 59.9 %
coal flakes	: 14.9 %
mudstone	: 7.9 %
quartzite	: 5.0 %
phyllite	: 5.0 %
chert	: 3.7 %
opal	: 1.6 %
shale	: 0.5 %
sandstone	: 0.5 %

The remainder consists mainly of clay flakes.

3. rare or negligible (less than 0.1 %):
Anatase, apatite, brookite, cassiterite, chlorite, clinochlore, dolomite, epidote, fossils, iron minerals (? hematite), opaque minerals (several), rutile, tourmaline and zircon.

The total quantity of heavy minerals is 0.03 %. Among them cassiterite, as in the Ocejo Formation, is still very important.

Comparing these percentages with those in the older deposits, it is striking that they do not resemble the percentages of the Ocejo Formation, but in many respects those of the older marine sediments, especially those of the Lower Sandstone Member of the Pando Formation (III.3). Yet there are important differences, e.g. the percentage of fossils which is much smaller here. This is partly caused, however, by the fact that sections 12 and 13 are the poorest in fossils of the Barranquito Member. Higher percentages must be found elsewhere: in a thin section of sample 753 we calculated 14.3 % of fossils.

The large number of coal flakes among the rock fragments is a striking feature. This must be due to erosion of parts of the Ocejo Formation. This assumption is supported by our observation that coal flakes are most common in the coarser sections (12 and 13).

The small number of thin sections available prevents a reliable investigation of the relationships that may exist between the various petrographic components. The same trends found in the older deposits are present, but in our opinion it is dangerous to draw conclusions.

VII.4. DIAGENESIS

In the few thin sections that could be examined only a limited number of diagenetic features were observed, mainly authigenic minerals. These are:

iron minerals (goethite, hematite, wustite and amorphous material, as shown by X-ray analysis): as vague spots;

chlorite: usually as small flakes, sometimes as radial aggregates and rarely as an alteration product of biotite.

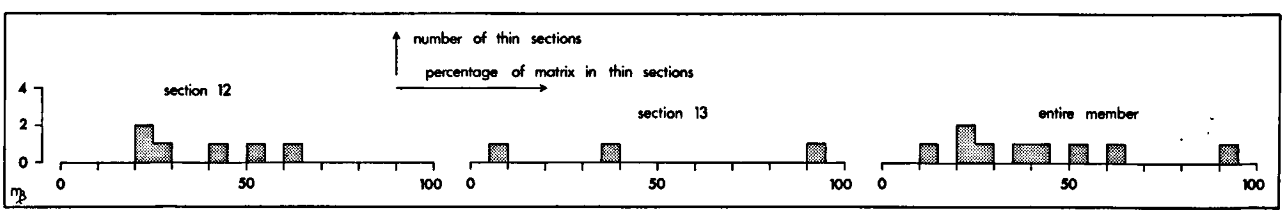

Fig. 63. Modal distribution of the matrix in the Barranquito Member.

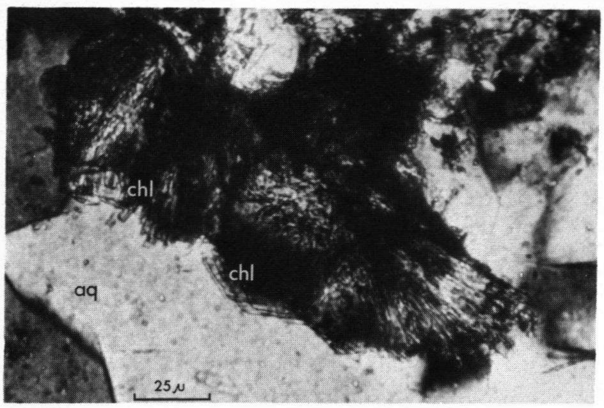

Fig. 64. Thin section of sample 544. Authigenic quartz (aq) partly replaced by chlorite (chl). Nicols crossed.

In a few cases it replaces quartz (Fig. 64)! ;
anatase: as bright yellow rhombohedra and as relatively large groups of crystals that have themselves been altered;
quartz: as secondary rims; as veins; and possibly as a replacement (see below);
pyrite: usually fine-grained, sometimes concentrated around fossil remains;
calcite: as a replacement of quartz.

The tourmaline that is often strongly attacked, as in slide 635, and of which in such a case only a few parts have been preserved is remarkable. The space created by the disappearance of the tourmaline is now occupied by quartz, but it is not clear whether this mineral was the original replacing mineral.

Only four authigenic minerals are present in significant quantities:

iron minerals	: 63.8 %
quartz	: 15.4 %
calcite	: 12.6 %
chlorite	: 8.0 % (often together with authigenic quartz)

The percentage of iron minerals is relatively low. It is much lower than in all other marine deposits under study, but considerably higher than in the continental Ocejo Formation. This relatively low percentage may, for a large part, be due to the presence of the authigenic quartz. In contrast to what is found in the older deposits a considerable part of the sandstones is slightly quartzitic, while secondary quartz, forming a beginning of a cement, is much more abundant than it previously was.

VII.5. SEDIMENTARY STRUCTURES

On account of the low degree of exposure and the limited thickness of this member, only relatively few observations could be made. As far as observed, the structures always indicate a shallow environment, while the fossils indicate a shallow marine environment (VII.6). The sediments tend to be distinctly, but irregularly layered. This can be well observed in the exposure just W of our locality 757, where the material consists mainly of silt. In the coarser part (section 12) most layers are irregular, too, and even some channels can be encountered. There is one part, however, that is much more regular (detailed section 8, enclosure 3).

The number of different sedimentary structures is limited: parallel and wavy lamination are common, while small-scale current ripples are relatively frequent. In sporadic cases a flaser-linsen structure was found. A shallow environment seems the most logical interpretation. The circumstance that hardly any channels were found, except in section 12, must be due to the almost complete absence of strong currents, which is in accordance with our grain-size distribution.

Since it is hardly possible lithologically (VII.2) and petrographically (VII.3) to distinguish this member from the surrounding fluvial sediments, we must assume that the influence of the sea was limited. The general absence of winnowed littoral deposits (these are only present in section 13) between this marine band and the continental underlying and overlying sediments indicates that no well-developed, long-existing beach was present. This implies that the continental (marsh?) deposits must have changed gradually into shallow marine deposits, and that the boundary between them was rather vague, possibly strongly dependent on the tide. This indicates a near-coast character of the Barranquito Member, an assumption supported by the frequent plant remains (VII.6) and coal flakes (VII.3).

VII.6. FOSSIL CONTENT

Compared with the shallow marine sediments of the Pando Formation (III.6, IV.6 and V.6), only a limited number of fossil groups occurs here. The most important are: brachiopods, pelecypods, gasteropods and crinoids. Ostracods, cephalopods, forams, algae(?) and washed-in land plants occur much less frequently.

Although exceptions exist, one may say that the more fine-grained and the more calcareous a locality is, the more fossiliferous it will be. The most fossiliferous localities are numbers 753, 754, 756 and 757. In all these localities fossils are abundant, but the number of species is extremely small.

Just as the sedimentary structures (VII.5), the fossils indicate a very shallow marine environment. The washed-in plants indicate that the coast was not far away.

The following fossils were found:

brachiopods:
Chaoiella sp. (756)
Juresania sp. (752) (Pl. II, Fig. 6)
Linoproductus sp. (756) (Pl. II, Figs. 7–8)
linoproductids (756)

Wagner et al. (1969) mention, from the top part in section 12:
Lingula sp.

pelecypods:
not yet identified (752, 753, 754, 755, 756, 757)

gastropods:
unidentified (752, 755, 756) (Pl. II, Fig. 5)

nautiloids:
one unidentifiable specimen (755)

crinoids:
stems and ossicles in nearly every locality, but rather scarce

land plants (washed-in):
Callipteridium? sp. (752)
Sphenopteris sp. (757)

VII.7. STRATIGRAPHIC INTERPRETATION

Since the Barranquito Member most probably is time-equivalent to the Lores Limestone (Wagner, pers. comm.) which is proposed as the lower boundary of the Cantabrian, the Barranquito Member must be considered to have a lowermost Cantabrian age.

Since Cantabrian deposits were only recognized as such a few years ago, the data on the flora and fauna are still rather scanty. A better knowledge of these fossils would, however, be of no help since the species we found are long ranging species, which makes an exact dating impossible.

There have been attempts to correlate the Cantabrian with the Russian stages (e.g. Wagner & Winkler Prins, 1970; Winkler Prins, in press): the Westphalian/Cantabrian boundary is supposed to roughly correspond with the Myachkovian/Kasimovian boundary. On the basis of fusulinids, some limestones now considered as middle Cantabrian were attributed a Kasimovian age by van Ginkel (1965).

It seems worth while to mention the large number of publications on the Cantabrian, often partly based on observations in our section 12, N of Tejerina. These usually contain arguments in favour of introducing a Cantabrian stage.

Already in 1964 and 1965, Wagner mentioned deposits in NE León (e.g. N of Tejerina) and NW Palencia (both in the Cantabrian Mountains) in which typical Westphalian and Stephanian elements occur together, causing these transitional floras to be considered younger than uppermost Westphalian on the one hand, but older than the lowermost Stephanian on the other. This, of course, is directly related to the definitions of both Westphalian D and Stephanian A.

The type area of the former is the Saar-Lorraine basin, where the Faisceau de Steinbesch forms the uppermost part of the Westphalian D. This is covered, with an angular unconformity (locally a disconformity), by the Holz conglomerate, considered there as the base of the Stephanian A. Although this conglomerate is erosive, the Westphalian D is thought to be complete or nearly complete in the central Lorraine area (Pruvost, 1934; Corsin, 1952). Although several plant species continue from the Westphalian to above the Holz conglomerate, many other new appearances and several disappearances cause a 'Florensprung', a sudden change in the flora (Corsin, 1952; Wagner, 1964; Bouroz et al., 1970; for the Saar basin: Germer et al., 1968), indicating a locally important stratigraphic hiatus (Corsin, 1952; Wagner, 1964, Bouroz et al., 1970) between the Faisceau de Steinbesch and the Holz conglomerate. Laveine (pers. comm.), however, believes that in some places the hiatus may be of minor importance. In any case, more to the W (bore-hole La Houve 2) the Holz conglomerate is absent, giving rise to the assumption that no hiatus is present there (Alpern et al., 1971; Corsin et al., 1968). In the Saar area itself, however, it is known that even in places where the Holz conglomerate rests apparently concordantly upon the older sediments, a probably important hiatus is present (Weingardt, in press). Without further evidence, it may therefore not be concluded that no hiatus exists in bore-hole La Houve 2.

The type area of the Stephanian A lies near St. Etienne in the Loire basin (Assise de Rive-de-Gier), where it rests upon basement rocks, so that no transition occurs from the Westphalian. Unfortunately, relatively few layers rich in plants are present, while the flora seems relatively poor in species. This flora is, however, as yet rather poorly known (Wagner, 1969). Another problem arises from the isolated location, which prevents a reasonable lateral extension (Wagner, 1966a). Nevertheless, it is possible to compare other deposits with this Assise de Rive-de-Gier, such as those in the French Cevennes (Bouroz et al., 1970) and the Carboneros Beds near the village of Barruelo in the NW of the Spanish province of Palencia (Wagner, 1966d).

In both cases it refers to deposits lying in an uninterrupted sequence, but the underlying sediments are younger than Westphalian D when the flora is compared with that of the type area (Faisceau de Steinbesch in the Saar-Lorraine basin). These very sediments are considered to be of Cantabrian age. It is even possible to draw a boundary in these series between the Cantabrian and the Stephanian A (Wagner, 1966d; Bouroz et al., 1970). That these Cantabrian sediments cannot be considered as equivalents of either the Westphalian or the Stephanian with a locally diverging flora is shown by the evolution in the flora found in the Cantabrian sediments. Corsin & Corsin (1971) do not agree, however, and consider the Cantabrian as partly Westphalian D, partly Stephanian A (see below).

In the Cantabrian Mountains a number of sections can be found that partially overlap. Correlation between these sections is not only possible with the aid of the flora, but especially by means of marine intercalations (such as the Barranquito Member of the Tejerina Formation under study). The oldest of these sections is our section 12, the youngest the one near Barruelo (see above), and between them a number of sections from the Valderrueda basin (Palencia and León). These sections together show a floral evolution, starting in the Westphalian D and passing via transitional beds (Cantabrian) into the Stephanian A. The boundary between the Westphalian D and the Cantabrian was originally drawn by Wagner (1966a, c) and Wagner et al. (1969) in our section 12; the boundary between the Cantabrian and the Stephanian A was drawn by Wagner (1966d) (see above).

The most important evolutionary developments that can be followed in these sections are described by Wagner (1969). They concern, for instance:

Alethopteris grandinioides Kessler (typical of the Westphalian) → *A. grandinioides* var. *subzeilleri* Wagner → *A. zeilleri* Ragot (typical of Stephanian A–C);

Callipteridium (Praecallipteridium) jongmansi (P. Bertrand) → *C. (Eucallipteridium) striatum* Wagner (typical Stephanian element);

Neuropteris ovata Hoffmann var. *ovata* (typical upper Westphalian) → *N. ovata* var. *grand'euryi* Wagner (Stephanian A–B).

The simultaneous occurrence of typical Westphalian elements (e.g. *Neuropteris scheuchzeri, Mariopteris* spp., *Alethopteris* *lonchitifolia*) and typical Stephanian elements (e.g. *Odontopteris reichi, Callipteridium pteridium*) also indicates transitional beds between the Westphalian D and the Stephanian A (Wagner, 1964).

Some authors (e.g. Corsin & Corsin, 1971) do not agree. They give some reasons to consider the lower part of the Cantabrian as upper Westphalian D and the upper part as lower Stephanian A. In their opinion, the boundary between Westphalian and Stephanian lies somewhere in Wagner's middle Cantabrian. This boundary is also said to be important for palynological reasons. It should be mentioned, however, that this boundary lies stratigraphically below the classic one (Holz conglomerate); in the Saar-Lorraine area this boundary would lie at the base of the Tritteling conglomerate (Corsin et al., 1968).

In the Cevennes an evolution, comparable to that in the Cantabrian Mountains, can be found, although there the base of the Cantabrian may be absent (Bouroz et al., 1970).

Because of the observed gradual evolution via transitional beds it is, in our opinion, acceptable that these fill up a time gap between the Westphalian D and Stephanian A (one must bear in mind that neither in the type area of the Westphalian D nor in that of the Stephanian A an uninterrupted transition exists between these stages). In this case it is correct to introduce a new stage (Cantabrian), although there are some arguments against this (Bode, in press) for reasons of definitions. However, these very definitions can easily be changed, as was done during the session of the S.C.C.S. of the I.U.G.S. (Krefeld, August, 1971).

The evolutionary developments show that, unless the floral evolution took place more rapidly in the Cantabrian than in the Westphalian and Stephanian (but there is no reason for assuming this), the Cantabrian lasted a considerable time (Wagner, 1969). A lower, middle and upper Cantabrian can even possibly be distinguished (Wagner, 1969; Bouroz et al., 1970). Wagner et al. (1969) even proposed a part of our section 12 (enclosure 2) as the stratotype section of the lower Cantabrian, in which the contact between the Westphalian D and the Cantabrian is fixed. Now, however, this boundary is placed elsewhere (see above).

At the session of the Subcommission on Carboniferous Stratigraphy (S.C.C.S.) of the International Union of Geological Sciences (I.U.G.S.) (Krefeld, 1971), a proposal to suggest the introduction of a Cantabrian stage at the forthcoming International Geological Congress in Montreal was accepted. This proposal was tabled by the S.C.C.S. that, in order to obtain first hand information, organized an excursion to the Cantabrian Mountains in September 1970.

The Cantabrian is considered by the S.C.C.S. as the lowermost part of the Stephanian s.l. In our opinion, however, it would be more practical and more correct (for reasons of definitions made earlier) to consider this Cantabrian stage as the uppermost part of the Westphalian (cf. Bode, in press), although evolutionary reasons, in the opinion of some authors (e.g. Rotay, in press), are in favour of attributing the Cantabrian to the Stephanian.

VII.8. PALAEOGEOGRAPHY

Without apparent lithological changes, the Barranquito Member follows the Ocejo Formation and the transition into the Corriello Member is equally inconspicuous. Only the occurrence of marine fossils instead of coal layers and seatearths characterizes this member. This indicates a very small change in circumstances, which, however, was sufficient to allow a continental facies to pass into a marine facies.

It would seem that the sea level rose or – more probably – the basin subsided just sufficiently to allow the sea to transgress the Ocejo Formation. Most probably the basin subsided at the same rate as before, but the hinterland rose less or not at all. This might explain the sudden absence, here and in the top part of the Ocejo Formation, of conglomerates. Less erosion yielded less material, resulting in a slower filling of the basin, thus allowing the sea to transgress.

Some indications were found of the direction of the incoming sea: because of the coarseness of the material in the NW, diminishing towards the SE, a supply of clastics from approx. NW seems most probable. This should imply a sea transgressing from approx. SE in a NW direction. The current ripples (only a few were measured) might indicates a transport from the land into the sea (N to S) and along the coast (E to W and W to E) (Fig. 65). This agrees quite well with the picture given above.

It also agrees with what we know of the Ocejo Formation, where the supply came from the WNW (VI.8), possibly with a sea that came from the SE. The difference with the Ocejo Formation is that there is now no doubt about the marine character of the sediments (only in the NW may some fluvial intercalations occur, cf. Wagner et al., 1969), because the sea at any rate transgressed further or stayed longer.

VII.9. CONCLUSIONS

A decrease in the uplift of the hinterland in the NW is thought to have caused a decrease in sediment supply and hence a transgression from the SE to the NW. The sea came in during a relatively short period at the very beginning of the Cantabrian, but always remained shallow, as may be concluded both from the fossils and from the sedimentary structures.

The fauna, sometimes rich in specimens, is poor in species, indicating either an unfavourable environment or a regression before a rich fauna could settle. Since a fauna may become rich very rapidly under favourable circumstances, the first possibility seems the more probable. This is in accordance with the brachiopod species present, that may live under rather unfavourable conditions (Winkler Prins, pers. comm.). We presume that the Barranquito Member was deposited in a fully marine environment in the SE and in semi-separated parts of the sea, probably in bays between lobes of the delta, in the NW part. The possibly fluvial sediments present there (VII.5) may also be explained in this way.

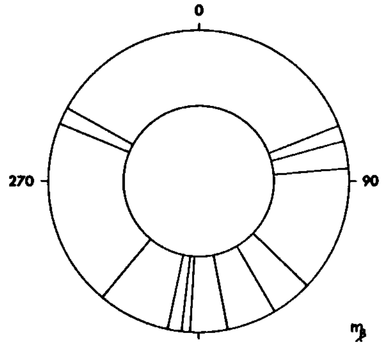

Fig. 65. Current directions in the Barranquito Member.

CHAPTER VIII

TEJERINA FORMATION: CORRIELLO MEMBER

VIII.1. INTRODUCTION

All sediments lying above the marine Barranquito Member and belonging to the Cea Group are here considered to belong to the Corriello Member of the Tejerina Formation. This member therefore completely belongs to Helmig's (1965) Prado Member. Just as in the Ocejo Formation, it appears that the thickness may vary considerably, the S-flank being much thinner than the N-flank (see sections in enclosure 2).

The Corriello brook, after which this member was named, is a W–E running tributary of de Río de Tejerina and debouches in this river just S of the village of Tejerina.

In this member two sections were sampled in detail: 12 and 13. Where possible, a sample was taken every 5 m. In section 12 this yielded 33 thin sections for petrographic examination and 7 in section 13. Ten pebbles that seemed to represent all the different types, were collected from the uppermost conglomerate of section 12 for thin section analysis. The top part of this member, possibly some 300–350 m, is unexposed or nearly unexposed for which reason no samples were taken.

The following localities belong to this member: 443–483, 547–553 and 637–638.

VIII.2. LITHOLOGY

This member is lithologically completely identical to the uppermost part of the Ocejo Formation, consisting as it does of alternating sandstones, mudstones and coal layers, in principle in this order. Only few conglomerate banks occur, again always with sandstone intercalations, while some single pebbles may occasionally occur in the sandstones.

The sandstones may vary from very coarse to very fine-grained, often containing a considerable amount of finer material (VIII.3). Really clean, winnowed sandstones were nowhere encountered. The 'shales' can vary from nearly pure clayey banks to very sandy mudstone layers.

Beneath the coal layers a seatearth is often present, according to our observations much more frequently than in the Ocejo Formation. In other cases the organic material from the coal layers may have been washed in and may be mixed with much siliciclastic material.

The lack of conglomerates makes the sediments in this member relatively non-resistant to erosion, causing an undulating relief.

VIII.3. PETROGRAPHY

In many respects this member resembles the sediments of the Ocejo Formation that lie between the conglomerates. This resemblance concerns the lithology (VIII.2) and also, as we shall see, the sedimentary structures, the fossils, the depositional environment and the palaeogeography. The petrography, too, shows considerable similarities.

Section 12 could be sampled very well, though only in the lower part, but section 13 yielded much less samples, which, due to the vegetation, only refer to the more resistant parts. For that reason the results obtained from this section must be considered less reliable than those from section 12.

Sections 12 and 13 yielded 33 and 7 samples, respectively, that can be classified as shown in Fig. 66.

The petrographic composition of the 40 samples examined shows the following average picture:

1. abundant (more than 10 %):
Quartz grains are again the most important (37.4 %), but matrix is also well represented (32.7 %). The modal distribution of the latter is shown in Fig. 67 and closely resembles that in the Ocejo Formation (Fig. 54). The matrix usually consists of siliciclastic material, but may contain limestone, or even be mainly composed of calciclastic material as in the micrite 549. The clay minerals in the matrix are mainly kaolinite and illite. 14 Å mixed-layers are also a normal constituent, but chlorite is rather rare.

The group of authigenic minerals is also abundant (15.1 %) and will be dealt with in VIII.4, while the rock

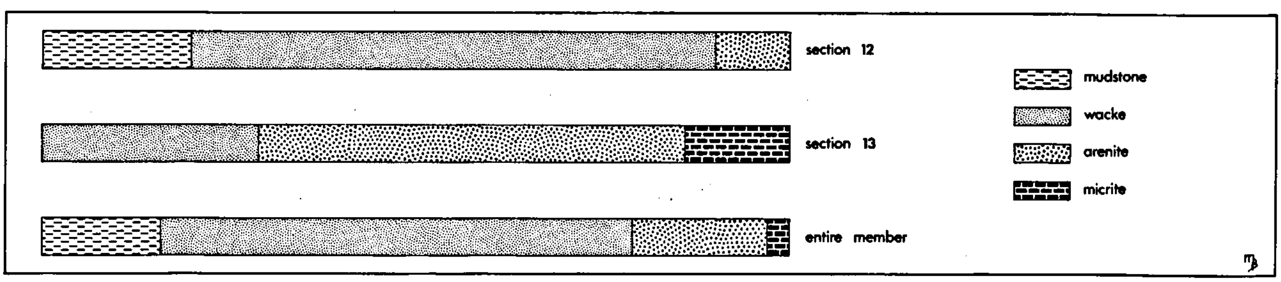

Fig. 66. Distribution of rock types in the Corriello Member.

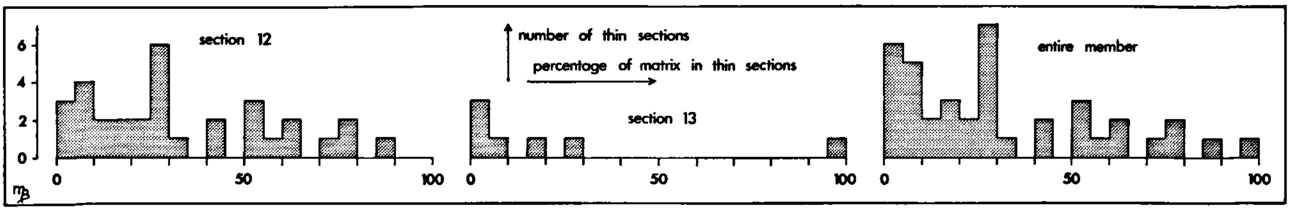

Fig. 67. Modal distribution of the matrix in the Corriello Member.

fragments constitute 13.2 %. These latter occur in the following proportions:

limestone	: 86.7 %
coal flakes	: 4.5 %
quartzite	: 3.8 %
mudstone	: 1.3 %
chert	: 1.2 %
phyllite	: 1.1 %
sandstone	: 0.5 %
shale	: 0.2 %
opal	: 0.2 %

2. normal (0.1–10 %):
Once again, only the minerals biotite (0.4 %) and muscovite (0.1 %) belong to this group (compare with VI.3).
3. rare (less than 0.1 %):
Anatase, cassiterite, chlorite, epidote, fossils, iron minerals (? hematite), opaque minerals (several), rutile, tourmaline and zircon.
4. negligible (less than 10 grains observed):
Apatite, chloritoid, clinochlore, corundum, fluorite, hornblende, periclase, plagioclase (one specimen of the ? albite prisms described in II.3 occurs here in slide 547) and zoisite.

The total quantity of heavy minerals is only 0.01 %, which is considerably less than in the older deposits. It is remarkable that cassiterite is still rather frequent at the base, but becomes rarer towards the top, finally even disappearaning almost completely. Since this tendency is not shown by the other heavy minerals, we assume this to be caused by the end of the erosion of an older sedimentary unit, possibly the Oville quartzite (see VI.3). The roundness of the grains excludes a direct derivation from pegmatites or similar rocks, neither are these rocks known in the vicinity.

All pebbles examined were sampled from the uppermost conglomeratic bank in section 12. There are 4 types of quartzites, a micrite, a fossiliferous micrite, a fossiliferous wackestone and a fossiliferous grainstone.

VIII.4. DIAGENESIS

The authigenic features in this member closely resemble those of the Ocejo Formation to which we may refer (VI.4).

The authigenic minerals in the Corriello Member are:
iron minerals (X-ray analysis shows lepidocrocite, wustite, magnetite, hematite and ilmenite, while amorphous material may also be present): as vague spots, sometimes as veins and a few times as a replacement of fossils;
anatase: as small bright yellow rhombohedra, but more often as altered masses;
pyrite: mainly as very fine-grained material, sometimes a little coarser;
chlorite: as a rare alteration product of biotite, but more frequently as authigenic flakes associated with authigenic quartz;
muscovite: as radial aggregates and probably as a rare colourless alteration product of biotite;
quartz: as secondary rims, rarely as veins;
calcite: as a replacement of quartz (rather rare), of other minerals such as tourmaline (very rare), as a cement and sometimes as veins;
antigorite: as a replacement of quartz.

These minerals occur in the following proportions:
section 12:

iron minerals	: 92.5 %
calcite	: 4.3 %
chlorite	: 2.1 %
quartz	: 0.9 %

section 13:

iron minerals	: 52.0 %
calcite	: 39.0 %
chlorite	: 7.1 %
quartz	: 1.8 %

total:

iron minerals	: 87.6 %
calcite	: 8.5 %
chlorite	: 2.7 %
quartz	: 1.0 %

Pressure solution, resulting in stylolitization in the calcareous sediments, was observed a few times (470, 550, 552, 553). The occurrence of compaction can be shown in slide 453, in which veins containing iron minerals (hematite ?) are not visibly compressed where parallel to the bedding plane, whereas similar veins perpendicular to the bedding plane show curves indicating compression.

VIII.5. SEDIMENTARY STRUCTURES

For the most part we can refer to our observations in the Ocejo Formation (VI.5), as far as the fluvial cycles are concerned. From the sedimentary structures, which for the most part are fully identical to those in the Ocejo Formation, and especially from the vertical alternations it appears that once again the sequence consists of fluvial cycles that, however, have often been preserved or developed only partially. Since the relatively frequent occurrence of coal layers beneath which a seatearth is found (VIII.2) indicates that erosion played a minor role, we assume the incompleteness of most of the cycles to be due to incomplete development.

We have the impression that cycles with a seatearth are more frequent in the lower part of this member than in the Ocejo Formation. In the top part of this member a reversal seems to take place. According to Read & Dean (in press), this should mean that the basin subsided more rapidly during the beginning of this member than at the end.

In the fluvial sequences sediments were encountered from nearly all possible sub-facies (Oomkens, 1967), such as coarse lag deposits, small and large channel fills, natural levees (coarse sandstone, probably homogenized mainly by burrowing, although the growth of plants may also be responsible) and point-bar deposits. This picture of meandering rivers gives the impression that a period of rest was achieved during which the hinterland supplied only little coarse material. Only in sporadic cases (VIII.2) was sufficient coarse material supplied to form conglomerates, possibly again as the result of some late tectonic movements along the León line. On account of the important lateral extent (many km) of these thin conglomeratic layers (less than 2 m thick), the polymict character and the angularity of the pebbles, it seems probable that these layers were deposited by sheet floods. These few layers therefore seem identical in origin to the numerous conglomerates in the Ocejo Formation.

In the Corriello Member no indications were found of winnowing in a littoral environment, nor do sediments influenced or reworked by marine conditions appear to be present.

VIII.6. FOSSIL CONTENT

For an exhaustive description of the flora we can once again refer to Wagner et al. (1969). In that paper the following plants from this member are recorded (reproduced here with Wagner's locality numbers):

Neuropteris ovata Hoffmann (W 1794–1861, W 1800–07), W 1808, W 1809, W 1815–17, W 1818–25, W 1829, W 1830–31, W 1832, W 1837–38, W 1840, above W 1840)
N. scheuchzeri Hoffmann (W 1800–07, W 1818–25, W 1826, W 1830–31, W 1836, W 1840, above W 1840)
N. sp. (cf. *praedentata* Gothan) (W 1818–25)
N. sp. (W 1818–25)
Mixoneura raymondi (Zeiller) (W 1794–1861, W 1818–25)
Linopteris neuropteroides (von Gutbier) (W 1836; cf.: W 1832)
L. cf. *brongniarti* (von Gutbier) (W 1815–17)
L. cf. *obliqua* (Bunbury) (W 1826, above W 1840)
L. sp. (W 1794–1861, W 1808, W 1818–25, W 1830–31, W 1837–38, W 1840)
Odontopteris cantabrica Wagner (W 1794–1861, W 1818–25)
O. sp. (W 1830–31)
Callipteridium jongmansi (P. Bertrand) (W 1794–1861, W 1818–25, W 1830–31)
C. sp. (W 1800–07, W 1808, W 1837–38, W 1840)
Alethopteris missouriensis D. White (W 1794–1861, W 1800–07)
A. lesquereuxi Wagner (W 1794–1861, W 1818–25, above W 1840)
A. grandinioides Kessler (W 1800–07, W 1830–31)
A. grandinioides Kessler var. *grandinioides* (W 1794–1861, W 1815–17, W 1818–25)
A. grandinioides var. *subzeilleri* Wagner (W 1815–17, W 1818–25, W 1839)
A. bohemica Franke (W 1794–1861)
A. ambigua Lesquereux (W 1818–25)
A. kanisi Wagner (W 1818–25)
A. zeilleri Ragot (above W 1840)
Sphenopteris cf. *rotundiloba* Nemejc (W 1830–31)
Sph. nov. sp.? (aff. *nummularia* von Gutbier) (W 1818–25)
Sph. sp. (W 1794–1861, W 1808, W 1815–17, W 1818–25, W 1836, W 1837–38)
Dicksonites pluckeneti (von Schlotheim) (W 1826, W 1830–31, W 1837–38, above W 1840; cf.: W 1832)
? *D.* sp. (W 1794–1861)
Lobatopteris vestita (Lesquereux) (above W 1840)
L. sp. (W 1794–1861)
Polymorphopteris polymorpha (Brongniart) (W 1794–1861, W 1800–07, W 1808, W 1809, W 1818–25, W 1829, W 1830–31, W 1832, W 1837–38, above W 1840)
P. sp. (W 1808)
Sphenophyllum emarginatum Brongniart (W 1794–1861, W 1808, W 1826, W 1830–31, W 1832, W 1836, W 1837–38, above W 1840)
Pecopteris unita Brongniart (W 1794–1861, W 1815–17, W 1818–25, W 1826, W 1840, above W 1840)
P. dentata Brongniart (W 1794–1861)
P. monyi Zeiller (W 1794–1861)
P. bredova Germar (W 1794–1861)
P. punctata Corsin (W 1794–1861)
P. hemitelioides Brongniart (W 1815–17, W 1837–38, above W 1840)
P. gothani (Guthörl) (W 1818–25)
P. acuta Brongniart (W 1837–38, W 1840)
P. sp. (W 1800–07, W 1818–25, W 1830–31, W 1837–38, W 1840, above W 1840)
Annularia sphenophylloides (Zenker) (W 1794–1861, W 1800–07, W 1815–17, W 1818–25, W 1830–31, W 1840)
A. stellata (von Schlotheim) (W 1794–1861, W 1800–07, W 1808, W 1818–25, W 1829, W 1830–31, W 1840, above W 1840)

Calamostachys tuberculata Sternberg (W 1794–1861, W 1818–25, above W 1840)
Palaeostachya sp. (W 1794–1861, W 1818–25, W 1830–31, W 1840)
Lepidodendron wortheni Lesquereux (W 1794–1861)
L. cf. *scutatum* Lesquereux (W 1809)
L. sp. (W 1794–1861, W 1800–07, W 1818–25, above W 1840)
Lepidostrobus sp. (W 1794–1861)
Lepidophyllum sp. (W 1818–25)
Calamites sp. (W 1800–07)
Cyclopteris fimbriata Lesquereux (W 1818–25)
C. sp. (W 1818–25)
Cordaites sp. (W 1818–25)
Alloiopteris cf. *cristata* (von Gutbier) (W 1826)
A. sp. (W 1837–38)
? *Potoniea* sp. (W 1832)

Helmig (1965) also mentioned some plants from this member. In the part of his Prado Member that can be correlated with our Corriello Member the following plants were found (original localities are given):

Callipteridium pteridium Zeiller (H 870)
Alethopteris cf. *grandinioides* Kessler (H 870)
Neuropteris scheuchzeri Hoffmann (H 870)
Linopteris? cf. *neuropteroides* Potonié (H 870)
Pecopteris sp. (H 870)

From some 600 m SW of Tejerina (just outside the area under study now, and lacking a locality number by Helmig) he lists (det. van Amerom):

Odontopteris minor-zeilleri (H. Potonié)
Linopteris obliqua (Bunbury)
Alethopteris grandini Brongniart
Sphenopteris sp.
Pecopteris polymorpha Brongniart
Calamostachys sp.

Because of all plant material previously collected, the present author sampled only a few specimens. He found:
Annularia sphenophylloides (Zenker) (638)
A. stellata (von Schlotheim) (638)
A. jongmansi Walton (638)
A. sp. (551)
Cordaites sp. (638)
Acitheca polymorpha (Brongniart) (637)
Lobatopteris sp. (638)
Polymorphopteris cisti (Brongniart) Wagner (549, 638)
cf. *P. polymorpha* (Brongniart) Wagner (549)
P.? sp. (549)
Linopteris sp. (551)
Astherotheca? sp. (549)
Pecopteris cf. *dentata* Brongniart (551)
P. fruct.? (551)

Just as in the Ocejo Formation, all collectors found some species not found by the others. According to the previous reasoning (VI.6) this indicates a flora that differs from place to place.

Beside the flora, both Wagner et al. (1969) and the present author found a non-marine fauna. This consists of small non-marine bivalves, and is richest just above the marine Barranquito Member. In the remainder of the Corriello Member these bivalves are extremely rare. We found:
? *Anthraconaia* aff. *pruvosti* (Chernyshev) (juvenile form?) (443) (Pl. II, Fig. 12)
Calver (in: Wagner et al., 1969) described and figured:
Anthraconaia aff. *pruvosti* (Chernyshev) (W 1798, W 1861)
A. sp. (W 1820).

VIII.7. STRATIGRAPHIC INTERPRETATION

Helmig (1965) considered these deposits to be basal Stephanian A. Wagner (1962, 1964) originally also interpreted the floras of this member as belonging to Stephanian A. After the recognition of some contradictions, which resulted in the proposal in favour of establishing a Cantabrian stage (see VII.7), he placed these sediments in the lower Cantabrian (Wagner et al., 1969). For a detailed argumentation we refer to this latter paper.

The few fossil plants sampled by the present author do not provide any evidence against a dating as lower Cantabrian.

VIII.8. PALAEOGEOGRAPHY

During this interval land appears to have definitely replaced the sea. No indications have at any rate been found of a marine influence or even of littoral deposits.

Most probably the land prograded in a more or less SE direction, just as in the Ocejo Formation. This is indicated, for instance, by the thinning of this member in this direction.

Since few conglomerates are present. the relief in the nearby hinterland will for the most part have disappeared, although so much finer-grained material was still supplied (the max. thickness of this member is approx. 550–600 m) that there must have been considerable erosion in the hinterland.

The almost exclusive presence of sequences indicating meandering rivers shows that the depositional area must have been rather flat. A mature relief had probably developed, with a broad river valley in which the length of the meandering river was mainly determined by the length of the parts perpendicular to the main direction. For if we assume (the decreasing thickness towards the SE is in favour of this assumption, and no arguments could be found against it) that the hinterland was situated in the NW and the sea in the SE (compare with VI.8 and VII.8), the main direction of the river(s) must have been from NW to SE. Ripple measurements in the

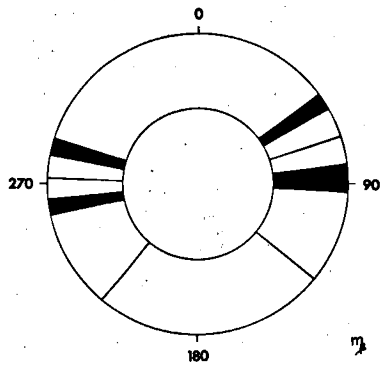

Fig. 68. Current directions in the Corriello Member.

fluvial deposits, however, only show transport directions more or less perpendicular to this main direction (Fig. 68). In our opinion this is only possible by large meanders in a broad and mature river valley.

VIII.9. CONCLUSIONS

The Corriello Member which has completely developed in a fluvial facies, thins rapidly towards the SE. Because of the considerable thickness and the mature relief in the depositional area, it seems probable that the basin subsided slowly and gradually, keeping pace with the sedimentary filling-up.

Meandering rivers led to the deposition of typical fluvial cycles that often did not have sufficient time to develop completely. Nevertheless, many coal layers occur, often with a seatearth. The abundant plant material that can be found in the top parts of the cycles indicates a lower Cantabrian age.

With this fully continental phase the regressive deltaic development (Scruton, 1960; Visher, 1965) that began with the deposition of the Prioro Formation (delta-slope deposits) seems to have reached its final stage, not only in a sedimentary, but also in a geotectonic sense (Timofeev, in press).

CHAPTER IX

STRUCTURAL HISTORY

The structure of the area studied has already previously been described by Rupke (1965) and Helmig (1965). We refer to their publications and will only give some general information and a few corrections and details.

Helmig and Rupke distinguished two unconformities: one between the Prioro and Pando Formations and one between the Pando and Ocejo Formations. According to Helmig and Rupke, these unconformities separate the Ruesga (≈ Namurian), Yuso (≈ Westphalian) and Cea (≈ Stephanian) Groups from each other. Since the present author was able to work in more detail, he was in a position to conclude that, on grounds of palaeontology, sedimentology and tectonics no unconformity exists here between the Prioro and Pando Formations, contrary to the assertions of many others (e.g. de Sitter, 1962; Martinez Alvarez & Torres Alonso, 1967; Boschma & van Staalduinen, 1968). Both formations, of a Westphalian age and belonging to the Yuso Group, have normal contacts here.

The sediments of the Pando and Ocejo Formations, however, are indeed separated by an angular unconformity which is quite clear on the map (Fig. 69), but which can only rarely be observed in the field. The map also shows that the Yuso sediments must already have been slightly folded when sedimentation of the Cea Group began. During and after deposition of these latter sediments both groups were folded further as a result of the continued activity of the same synclinal structure. This resulted, roughly, in a large E–W trending syncline with an axis plunging towards the W.

This picture became complicated due to a very large number of rather small to very small fold structures, while a considerable faulting activity took place too. Both folds and faults seem to be strongly dependent on the differences in competence of the rocks.

The León line, which here, according to Rupke (1965), represents the Las Salas anticlinal structure, and which can also be traced in the anticlinal structure that forms the northern boundary of our area, is the most important fault. The opinion of Marcos (1968), who does not consider this fault exceptionally important, is opposed by the fact that the León line can be traced over a very great distance (de Sitter, 1962); the importance of this fault was also recognized by us. Especially because this fault zone was active during the deposition of the sediments under study (according to Marcos, 1968, at least from upper Westphalian to Stephanian B–C), its influence as a facies boundary must have been great, a conclusion also arrived at by Helmig (1965) and Rupke (1965) on the basis of their field observations. This influence is distinctly shown by the fact that the area S of this line (the so-called Leonides) was elevated during Westphalian times and by being eroded supplied the material to the subsiding basin (the so-called Asturides) that was situated in the N between this fault zone and the Asturian block (Wagner, 1970 and in press). In the Leonides only a few locally limited depressions existed in which Westphalian sediments were deposited (Rupke, 1965). Elsewhere in the Leonides the Westphalian is absent, while it occurs everywhere N of this line, usually reaching a thickness of several km (e.g. Feys et al., in press). One of the few exceptions in the Leonides is the area studied. This depression was limited in the W by the Las Salas 'high' and in the N by the

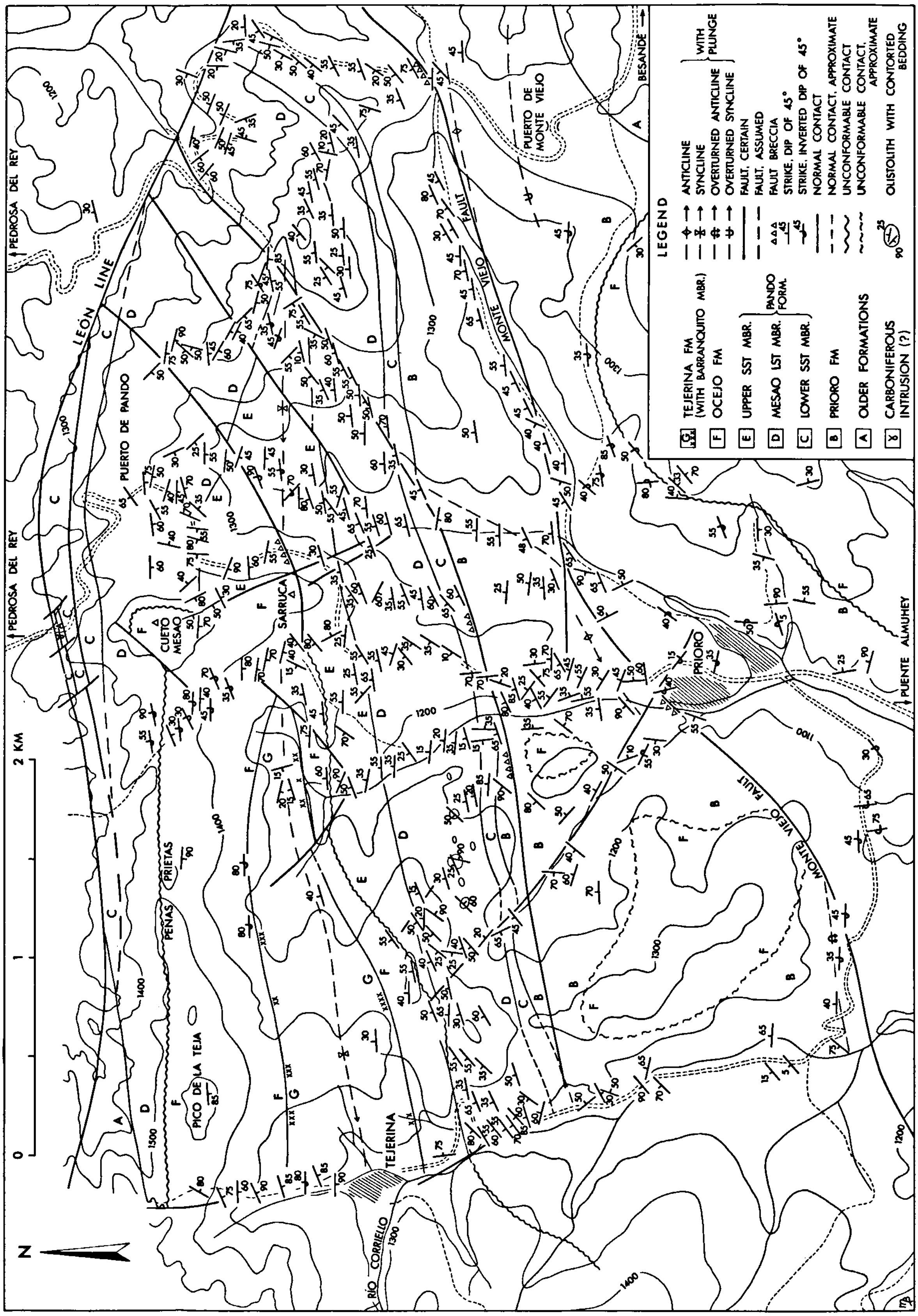

Fig. 69. Structural map of the Prioro-Tejerina area. Only major faults are indicated.

rising blocks at the León line. As was suggested in the preceding chapters, these two areas probably yielded most of the material that filled up this basin. It is unknown where the boundary of this depression was situated in the S and E, because it is hidden by the cover of younger sediments or below upthrusts. It seems very improbable that the basin extended more than a few km in those directions (compare with Rupke, 1965, Fig. 16). It is therefore not unlikely that those southern and eastern areas, too, were responsible for the supply of some material to the Prioro-Tejerina basin. Since it is not impossible that in the S, beneath the Tertiary cover, the old metamorphic axial zone of the Hercynic Cantabrian Mountain chain is present (de Sitter, 1962), this could explain the occurrence in our sediments of metamorphic minerals (II–VIII.3). The relatively limited attrition of these minerals is easier to explain in this way than by assuming a transport from Galicia.

The anticlinal structure which runs parallel in the S also assumes the form of a fault line, the Monte Viejo fault which comes next in importance to the León line. It practically forms the southern boundary of the area investigated, since exposures are extremely scarce S of this fault. With some reserve, Helmig (1965, p. 130) considered it to be an upthrust. Our observations are in agreement with this: the folding gave rise to a fault approximately in the core of the anticline, and the same pressure responsible for the forming of the E–W trending anticlinal and synclinal structures pushed the N flank of this anticline somewhat over the southern flank. This is shown by two small sandy outcrops belonging to the Lower Sandstone Member of the Pando Formation or to the uppermost part of the Prioro Formation, found in the direct vicinity of the unconformable Ocejo Formation in the S.

Approximately one km N of the Monte Viejo fault, a topographic depression runs more or less parallel to this fault. For some parts a fault character could be proved. Together with numerous smaller faults it seems to form part of a fault system running approx. ENE–WSW (Fig. 69; only the more important faults could be shown in this fig.).

Another fault system, the number of faults of which is also very large but which has resulted in faults with less important displacement, runs more or less NW–SE. Together with the system mentioned above they seem to be the result of a N–S pressure, which may be the same pressure responsible for the E–W trending fold structures (see above).

The faults occur preferably in places where there is a marked difference in competence of the rocks. This is clearly shown at the Prioro/Pando Formation transition (especially where the latter starts abruptly (III.2) with thick sandstones), at the vertical and lateral changes of limestone into mudstone in the Mesao Limestone Member of the Pando Formation and at the unconformable contact between the Pando and Ocejo Formations. The most important faults (e.g. the Monte Viejo fault) appear to be connected with non-resistant zones (sometimes with a fault breccia), that have led to a topographic depression. This makes the river pattern very similar to the fault pattern (Fig. 69). Most probably the many depressions in which no fault could be established (e.g. on account of vegetation or uniform lithology), but which are orientated according to the fault systems mentioned above, may be interpreted as similar fault zones.

The occurrence of these fault systems in both the Yuso and the Cea Groups shows that these faults are post-Cea, or were formed during a time-span from at least Westphalian B to Stephanian s.l. (Cantabrian) (compare with dating of the León line by Marcos, 1968, see above). In the Yuso Group many more faults occur than in the Cea Group. This might indicate the latter possibility, although the fact that the Yuso sediments are less competent than the Cea sediments may have played an important role. Further evidence in favour of fault movements having already begun in Yuso times is, however, constituted by the presence of the many apparently syntectonic sediments (especially in the Prioro Formation) and the occurrence of volcanic minerals in the Pando Formation (III.3, V.3) that seem to derive from contemporaneous eruptions.

This tectonic activity during the deposition of the Prioro Formation is probably related to the latest part of the Curavacas phase (Kanis, 1956) of the Hercynian orogenesis, which might be time-equivalent (at least partially) with the Prioro Formation. Towards the E the influence of this folding phase always becomes greater (Nederlof, 1959; Koopmans, 1962; Brouwer & van Ginkel, 1964; van Veen, 1965; Savage, 1967; Boschma, 1968; van de Graaff, 1971; Maas, in prep.), which results in the above-mentioned unconformity between the Ruesga and Yuso Groups. W of our area no indications have been found of this unconformity (Sjerp, 1966; Evers, 1967). Our area may therefore be considered a transitional zone. The Curavacas folding phase has been compared with the Sudetic phase (de Sitter, 1960, 1962). Wagner & Wagner-Gentis (1963) introduced the term 'Palentian folding phase' as a synonym.

The folding phase responsible for the angular unconformity between the Yuso and Cea Groups is the Leonian phase (Wagner, 1962, 1963a, b), which can be compared with the Asturian phase (de Sitter, 1960, 1962). This phase, running diachronically through the mountain chain, just as the others, can be dated here as upper Westphalian D by means of the fossil content in the underlying and overlying sediments.

The Cea Group has also been strongly folded: the N flank of the syncline is even usually overturned. There must therefore have been at least one more folding phase. According to de Sitter (1960), this was not an Alpine but a pre-Triassic movement, and he compares it with the Saalic phase of the Hercynian orogenesis. Helmig (1965), however, states (p. 143) that 'the deformation of the Cea formation has apparently been a long-lasting intermittent process, which occurred in three stages: 1. a synsedimentary stage, 2. a post-

Stephanian and pre-Cretaceous stage and 3. a middle Tertiary stage.'

During the folding both cleavage and joints were formed, especially in the fine-grained sediments. These phenomena were not investigated by us. It could, however, be established that contrary to former assumptions (Henkes, 1959?) the Prioro Formation does not exclusively show a cleavage parallel to the bedding. An angle between cleavage and bedding often occurs. In the Pando Formation, too, these two possibilities are present. The conception of the previous authors may have been influenced by the lithology: in the Prioro Formation, mainly consisting of shales (or mudstones), the bedding plane is often very difficult to find, while the cleavage is responsible for a pseudo-bedding. Besides, the angle between bedding and cleavage tends to be small (less than 15°). In the Pando Formation, however, the different lithology facilitates the recognition of the bedding plane, while the angle between cleavage and bedding plane (influenced by the competence of the rocks) is usually larger than in the Prioro Formation. In mudstones, however, the angle is usually still rather small (10–15°), e.g. near the Puerto de Monte Viejo.

Joints are rather rare, but in the E part of our area they occur more frequently. There we are confronted with AC joints, which in uniform parts may also be mistaken for bedding planes.

In many thin sections phenomena can be observed that indicate deformation, such as small faults (e.g. 503) with micas that are bent along these faults, and small graben (e.g. 505).

CHAPTER X

CONCLUSIONS

The sediments of the Prioro-Tejerina area form one thick regressive sequence constituting as they do a prograding deltaic complex. The lithological units distinguished in the preceding chapters can be considered to be various phases in this development, which will be summarized here.

A period of strong tectonic activity (Curavacas phase?) during the deposition of the Prioro Formation at the transition between Westphalian B and C supplied a large amount of material into a small subsiding basin, situated as a depression between the rising blocks of the so-called Leonides. The axis of the basin ran WNW–ESE. The material was mainly supplied laterally by various modes of transport 'en masse', nearly exclusively from the northern border where a rising zone existed along an active fault line (the León line). The sediments supplied are extremely immature, consisting mainly of fine-grained material, especially mudstones. Grain-size differences in these mudstones may indicate a few phases of more and less tectonic activity.

The sediments were partly deposited upon a delta slope, partly at the base of this slope by a transport parallel to the basinal axis. In both cases the water must have been deep (below wave base), although rather near-shore. The rapid sedimentation resulted in an environment that was extremely unfavourable to life, since, apart from some trails, no autochthonous fossils are present. Allochthonous fossils are also very scarce, often only occurring as fragments.

The accumulation of sediment (by enlargement of the deltaic complex) took place more rapidly than the subsidence of the basin, so that this latter became shallower. This is expressed by the sedimentary structures and the fossils of the Lower Sandstone Member of the Pando Formation (probably lower Westphalian C). The sediments consist of alternating sandstones and mudstones, often with a rich fauna. Like the structures and the grain size, this indicates a shallow (sometimes above wave base) near-coast environment.

A succeeding period of relative quietness (Mesao Limestone Member, middle Westphalian C) was possibly caused by a decrease in or an interruption of the uplift in the hinterland, resulting in the supply of only little and fine-grained material (mudstones). This allowed the development of biogenetic banks in the very shallow shelf zone. Abrasion, probably caused by wave action, yielded large quantities of calciclastic material deposited as clastic limestones. Around the limestones the siliciclastic material was deposited. Both in the limestones and in the mudstones abundant fossils indicate a favourable and very shallow environment, rich in oxygen. The occurrence of algal limestones over a thickness of 200–250 m indicates a slow subsidence of the basin with which the rate of sedimentation kept pace.

A new period of stronger tectonic activity (accompanied by a supply of volcanic minerals) brought about an increase in the supply of coarser material. The formation of biogenic limestones became impossible due to the increased supply. Only a few clastic limestones were still deposited. The prograding of the delta is shown by a gradually increasing sandstone/shale ratio towards the top, by the presence of more channels in the sandstones and by the increasing thickness of the sandstones towards the top in the Upper Sandstone Member of the Pando Formation (probably upper Westphalian C to middle Westphalian D). The fossils, abundant especially in the calcareous parts, indicate a very near-shore and shallow marine environment.

Possibly as a result of the prograding deltaic complex, the basinal axis gradually turned from WNW–ESE in the Prioro Formation via W–E to SW–NE in the Upper Sandstone Member.

A period of strong tectonic activity followed (Leonian, i.e. Asturian phase) causing the uplift and tilting of this area, combined with a slight folding. This took place in the upper Westphalian D. The tectonic pressure during this phase is probably responsible for the differences in clay-mineral content in the sediments below and above the resulting unconformity.

Above the angular unconformity deposition of mainly continental sediments (piedmont deposits and fining-upwards fluvial cycles) initially took place, between which a few littoral and possibly also some marine reworked sediments occur (Ocejo Formation, uppermost Westphalian D). Transport took place from NW to SE, which indicates that the situation of the basin had hardly changed: only the shore-line had moved somewhat towards the S.

At the transition between the Westphalian D and the Cantabrian a small transgression occurred. Some tens of metres of shallow marine sediments were deposited (Barranquito Member of the Tejerina Formation). The grain-size distribution indicates a supply from NW to SE. A regression followed, still in the lower Cantabrian. The sediments of the Corriello Member of the Tejerina Formation are almost exclusively composed of fining-upwards fluvial cycles, and possibly a few piedmont deposits. Transport was directed more from N to S. With this phase the deltaic development had reached its final stage.

Subsequently all sediments were again folded, probably during various phases.

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