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SEDIMENTOLOGICAL ASPECTS OF FOUR LOWER-PALEOZOIC FORMATIONS IN THE NORTHERN PART OF THE PROVINCE OF LEON (SPAIN)

E. OELE *)

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INHOUD

*) Geological survey of the Netherlands, Haarlem.

Samples mentioned in the text can be found in collections of the „Rijksmuseum van Geologie en Mineralogie" at Leiden, where also information as to their location in the sections given can be obtained.

ABSTRACT

This paper deals with the sedimentary structures and sedimentary petrography of the four lowermost formations of the Paleozoic as developed in the Northern part of the Province of León (Cantabrian Mountains, Spain). Three of the four formations have a detrital character, and one consists of dolomites and limestones. Mineralogically, the detrital formations are mature. Consequently the differences are small, but diagnostic. The source rocks will have been non-sedimentary.

The *Herreria Sandstone Formation* is the oldest formation. Only its upper 200 metres are described here. This part consists of medium-grained quartzites with intercalations of shales, coarse quartzites, and conglomeratic beds.

The detrital quartzes contain various kinds of inclusions and are often composite. Microcline, the common feldspar, is often kaolinized. Both minerals are secondarily enlarged. The source of the secondary quartz is duscussed; this quartz is held to have been supplied partially, and precipitated, form formational waters. The latter have the tendency to increase salinity, which lowers the silica solubility.

The layers show predominantly a parallel lamination, but cross-lamination occurs as well. In two parts of the sequence the layers are wedge-shaped.

The depositional environment is assumed to have been shallow, near the shore, with fluviatile influences.

The *Lancara Dolomite Formation* can be subdivided into Dolomite s.l. and Griotte. The *Lancara Dolomite* s.l. contains dolomites, limestones, oolitic limestones, and breccias. The diagenetic process of grain growth transformed the original detrital texture of the limestones and dolomites. Dolomitization is assumed to have been postdepositional. Recrystallization due to mechanical stresses occurs as well.

The oolitic limestones too are built up of various types of calcite in a textural sense. The time-relations between these types is discussed. These limestones contain authigenic quartzes, indicating high salinity of the environment.

The *Lancara Griotte* consists of nodular limestones and shale layers with limestone nodules. The limestones are detrital in origin. The origin of the griotte is discussed: it is attributed to solutional processes.

The depositional environment of the Lancara Dolomites s.l. is thought to be comparable to the recent Bahama Bank deposits. That of the Griotte is less distinct, but must have been shallow neritic. The red colour of the griotte may point to a warm, humid climate.

The *Oville Sandstone Formation* is characterized by its clayey nature, high lime content, and the authigenic mineral glauconite. The micas show replacement by carbonates, a relatively unknown process. The origin and source of the glauconite is dealt with: cryptocrystalline aggregates are thought to have initially been clay, while the crystalline glauconites are altered micas.

Of special interest are the slump structures. Since they are the result of a thixotropic behaviour of the sediments some rheological principles are briefly reviewed. It is also stated that internal slumping and convolute laminations are related in the sense that both are expressions of a false-body thixotropic state of the sediment. Such a state is to be expected within a certain range of moisture content: internal slumping occurs at the lowest values, convolute lamination at the highest values of the range. However, convolute lamination is observed more commonly in turbidity deposits because such deposits settle at higher rates than other sediments, consequently their moisture contents must have been higher.

In this thin-bedded complex, parallel lamination dominates but small-scale cross-lamination is also present. Other sedimentary structures observed are load casts, pseudo-nodules and "Linsen" structures.

The depositional environment is held to have been deltaic i.e. the formation represents a chain of deltas.

The *Barrios Quartzite Formation* consists of quartzites with few shale beds and locally a conglomerate. The quartzes are limpid and do not contain inclusions. Composite grains are scarce. Feldspars are not kaolinized, only sericitized. The occurrence of the mica phengite is diagnostic.

Most of the beds are wedge-shaped, which gives the formation a special appearance. Most beds have an slightly inclined lamination.

Like the Oville deposits the Barrios sands are held to be deposits of a deltaic environment.

INTRODUCTION

During the past few years the section for structural geology of the University of Leiden under the direction of Professor L. U. de Sitter, has been mapping the southern slope of the Cantabrian Mountains in Spain. A structural map has already been published (de Sitter, 1962). During the mapping, various sedimentary structures observed in the Lower-Paleozoic rocks attracted the attention. It seemed interesting to investigate some of the Lower-Paleozoic formations, and their sedimentary structures in particular, from a sedimentological point of view. However, soon after the investigation started it became clear that, although structures such as ripple marks, slump balls, and so forth are present, their number is insufficient to determine the current directions or location of shore-lines. On the other hand, the petrography of these formations, which have such different aspects in the field, offered many possibilities.

Since no study of the petrography of consolidated sediments had previously been made at the Geological Institute of the University of Leiden, the methods used and the results reached in this study are preliminary in many respects. This paper therefore represents first results, the more so because the investigation cannot yet be regarded as terminated. It deals with the four lower formations of the Paleozoic rocks. Sections of each of these formations were measured in the Esla region, but unfortunately the basal formation, the Herreria Sandstone, is not exposed there and had to be investigated in the Curueño Valley.

Acknowledgements

The work presented here was accomplished thanks to the assistance of many persons. I am greatly indebted to Prof. Dr. A. J. Pannekoek for his reading of the manuscript and critical remarks as well as for his assistance in the field, to Dr. P. Hartman for his interpretation of the x-ray photographs of the clay minerals and carbonates, and to Dr. P. M. J. Ypma for the identification of some of the opaque minerals. Fruitful dicussions with Dr. J. M. Mabesoone (now at Recife), Mr. H. Koning, Mr. P. Floor, Mr. P. J. C. Nagtegaal and others were of great help in completing the work. I also wish to thank Dr. F. Kalsbeek (now at Aarhus) very much for his stimulating help during the first stages of the investigation, and Dr. D. Boschma, Mr. J. Rupke, and Mr. H. A. van Adrichem Boogaert, from whose knowledge I greatly profitted in the field. Mrs. I. Seeger corrected the English text with great care for the subject.

I am also indebted to Dr. J. Lips of the Radiological Department of the Leiden University Hospital for his cooperation in making the x-ray photographs of sedimentary rocks. The x-ray photographs of the clay minerals were made by Mr. A. Verhoorn. The drawings, which form such a substantial part of the work, were executed by Mr. J. Bult. The photographs, also indispensable to the text, were made by Mr. J. Hoogendoorn. Miss Angeles Dekker was willing to type a great part of the manuscript in her spare time, for which I am much obliged. The thin sections essential to the study were made by Mr. C. J. van Leeuwen and Mr. M. Deyn.

The inhabitants of Valdoré made my stay in the field a great pleasure. I want especially to thank the family of Angel Fernandez, the García family, and the family of Pedro Lopez for their hospitality and their great cordiality.

CHAPTER I

GEOLOGICAL SETTING

Stratigraphy and age

The stratigraphic division proposed by Comte (1959) has been followed. The terms used for the successive formations are given in Table I. Due to the paucity of fossils, the division is primarily lithostratigraphic. Although the same holds for other divisions in use (Lotze & Sdzuy, 1961), it is easy to compare the divisions with each other.

TABLE I Stratigraphic Table

	Formigoso Shales	Silurian	
	Barrios Quartzite	Ordovician	
	Oville Sandstone	---------	
	Lancara Limestone: Griotte / Dolomite	Cambrian	
	Herreria Sandstone		
	Precambrian		

Since the formations are poor in fossils, their age is difficult to determine. No fossils have been found in the Herreria Sandstone, except for some tracks which are being studied by Seilacher. Its age is held to be Cambrian because of its conformable position underneath the dated Lancara Formation. The Lancara Dolomites have also failed to yield fossils, the red limestones on top of the Dolomites containing only indeterminable fragments. A rich fauna has been found in the Lancara Griotte, however; among these fossils the Trilobites are important.

The fossils reveal a Middle-Cambrian age. Trilobites have also been observed in the Oville Formation. The base of the Oville Formation is known to be still Acadian. Since the top of the Oville contains no fossils, the boundary with the Ordovician cannot be drawn with certainty.

The Barrios quartzites do not contain fossils either. A bottom track ascribed to *Cruziana* has been found. The Barrios Formation is overlain by Silurian shales of the Formigoso Formation.

Structural geology

Recently, various papers dealing with the structural geology of the southern part of the Cantabrian Mountains have been published by de Sitter (1959, 1961, 1962). Their main lines may be recapitulated here.

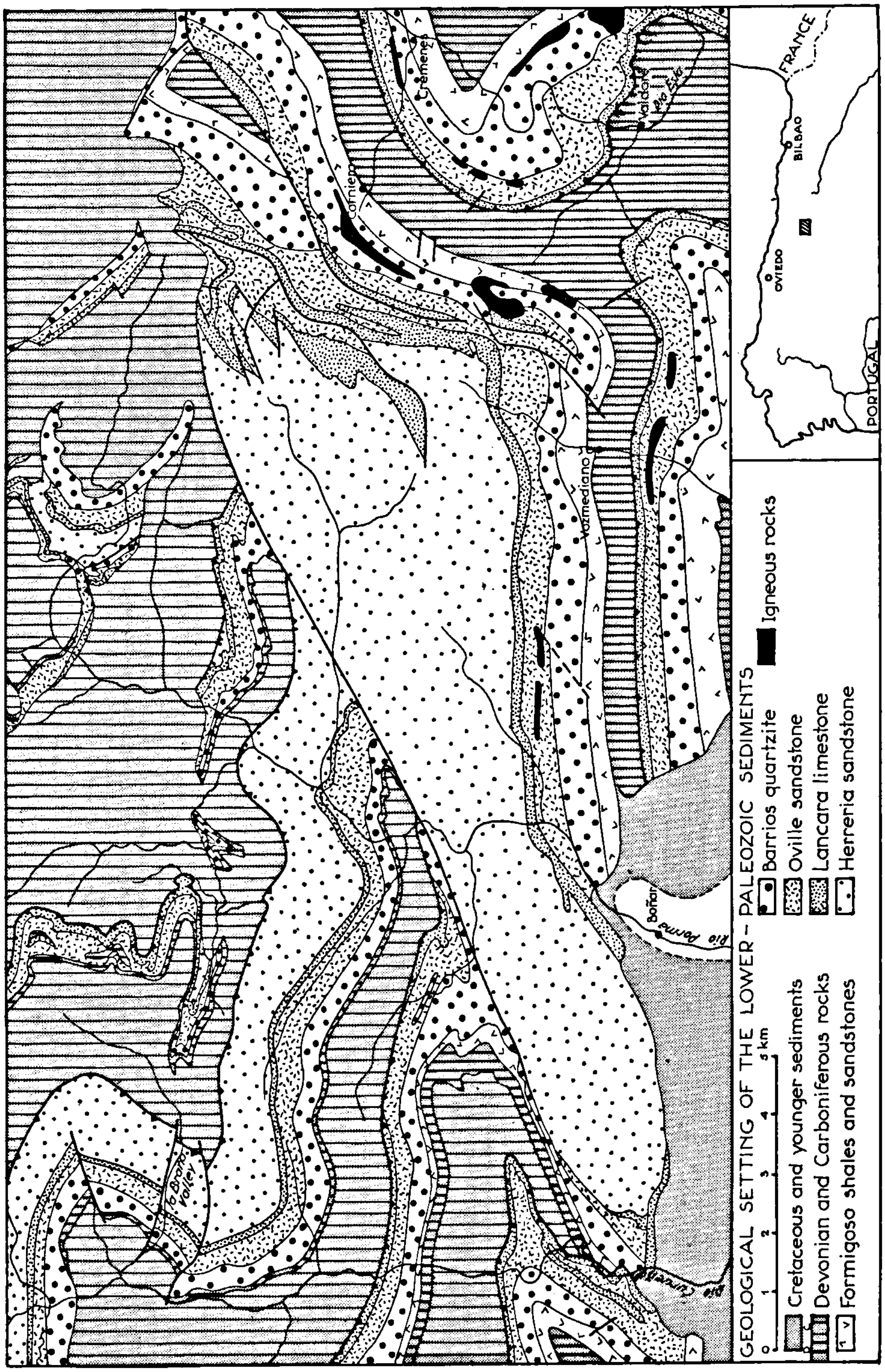


Fig. 1. Geological setting of the Lower-Paleozoic in the area investigated.

Structurally, the southern slope of the Cantabrian Mountains can be divided into two longitudinal units (Fig. 1): the Asturides in the North and the Leonides in the South, only the latter being important to our subject. The structural history of the Leonides can be sketched as follows.

After a Precambrian diastrophism, the first tectonic movement during the Paleozoic is a tilting and subsequent erosion of the northern part of the Leonides just before the Famennian. The main movement, however, took place just before or during early-Westphalian times (Sudetic phase), while later phases of the Hercynian folding resulted in a refolding of the previously-created structures. The Alpine orogeny placed the rocks in their final position.

In the Bernesga-Esla region the main orogenetic phase created various thrustsheets: in the western part, five successive thrustsheets can be distinguished (the most northerly one being termed Bodon), in the E. only one thrustsheet (the Esla nappe) is present. The sheets comprise all the formations ranging from Middle-Cambrian up to Visean or Namurian. The thrustplane coincides with the base of the Lancara griotte, but occasionally it is located stratigraphically lower in the Lancara Limestone, as is the case with the Esla thrustplane in some places.

Intrusive rocks

Throughout the whole area under consideration here, igneous rocks crop out. They are mostly lenticular bodies, but in the Barrios Formation tuffaceous matter and lapilli have also been found. The lapilli measure about 20 cm, but even larger ones have been observed. Most remarkable is the absence of a contact metamorphic zone around the igneous bodies.

These rocks have been described by Comte (1959) and I can only subscribe to his description. The rocks have an ophitic texture. The dominant mineral is lath-shaped plagioclase (labradorite); chlorite, which is an alteration product of olivine, and augite are also present.

METHODS OF INVESTIGATION

Measurement and the presentation of sections

Sections of formations are usually presented in rather small-scale drawings, which make it necessary to omit various details. To overcome this disadvantage, here parts of each section are presented in detail, use having been made of the graphic log of Bouma & Nota (1961) (Fig. 2). Using the graphic log, the student is on the one hand forced to look for the various properties as listed in the columns of the graph and will not so easily overlook special features, while on the other hand a good impression of the habitus of the rocks is obtained at a glance.

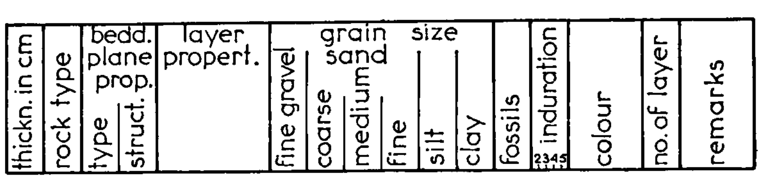


Fig. 2. Graphic log as proposed by Bouma & Nota (1961).

Since some of the important properties of sediments are not listed (e.g. sorting; mineralogy and texture; roundness of the grains), environmental conclusions cannot be drawn from such graphs except perhaps when very special environments are involved (e.g. cyclothems). In addition, transporting agents can only exceptionally be read from these graphs (e.g. turbidity current deposits). Some of the properties recorded in the graphic log will be briefly reviewed, the others speak for themselves.

Bedding contacts. For the sake of completeness, the kind of bedding contacts is mentioned. Under bedding contacts is understood the contacts between bedding planes with an actual parting. It is, of course, only an assumption that such bedding planes delimit depositional units. However, partings may logically be expected where the properties of the sediments show a change, and therefore it may be assumed that the partings mark off sedimentary units.

Since unconsolidated sediments usually lack partings between the beds, the successive beds can be discerned only by means of a macroscopically observable change in grain size or change in sedimentary structure such as, for instance, cross-bedding in the sense of McKee & Weir (1953). But differences in grain size are not always trustworthy. In consolidated sediments it is often observed that successive beds show a parting and yet are identical as to grain size, while in other instances the reverse holds and one bed may contain pebbles at the base and fine sand at the top, without a gradual transition between the two extremes.

Two aspects of the contact plane can be distinguished: the sharpness and the form. Following Bouma (1962), the sharpness of the plane is subdivided into: very sharp, sharp, distinct, transitional within 0.5 cm, and transitional over more than 0.5 cm. The form is flat, undulating, or irregular. The classes, of course, lack absolute class limits, which makes the handling of the subdivision rather subjective.

It follows that the sharpness should not be deduced from the difference in grain size as suggested by Bouma. By doing so, we reduce the information which can be obtained in the field, and we do not know how much information is lost. As long as the precise value of contacts and bedding planes is uncertain, it is better to describe the contact planes as they are observed in the field. Moreover, the change in grain size is recorded in a special column, and the "type of bedding" column would therefore give exactly the same information, although the latter column gives additional data as to the form of the contact plane.

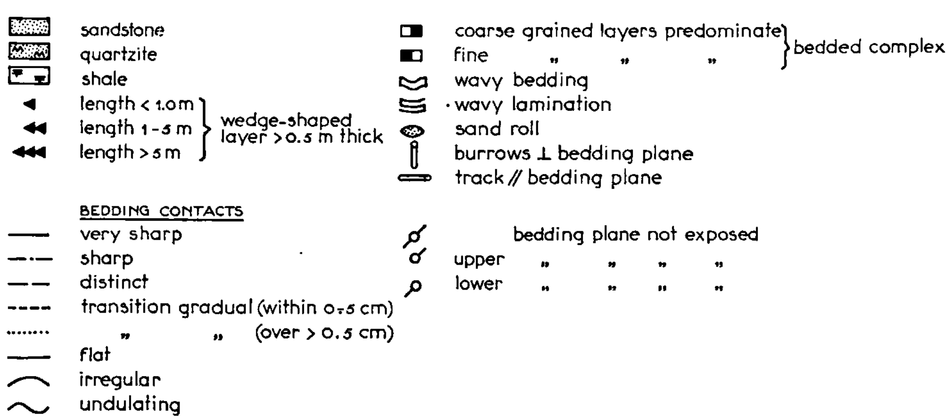


Fig. 3. Symbols used to designate properties of the rocks in the graphic log.

Fig. 3 represents the symbols used to designate the properties of the contact plane. It is noted that sharpness and form are presented by the same symbol, i.e. the sharpness symbol is drawn straight, bent, or s-shaped, indicating the plane is flat, irregular, and undulating, respectively.

Induration. Following Bouma, the rate of induration is expressed by the numbers 1 to 5, indicating: 1. loose sediment 2. slightly consolidated (grains can be detached by fingernail) 3. moderately hard (grains can be detached by knife) 4. hard (fractures pass between the grains) 5. weakly metamorphic (fractures pass through the grains).

Colour. The colour of the rocks was determined by means of the colour chart of Cailleux & Taylor (1952). Since the chart is not satisfactory for estimating greyish colours, the method was only applied to reddish and yellowish sediments. Cailleux & Taylor give also the corresponding numbers of the Munsel Rock Colour Chart. Because the latter chart is more widely known, the figures of this chart have been used.

Grain-size distribution

The grain size was determined by the ribbon-counting method first proposed by Kalsbeek and later advocated by van der Plas (1962). Since such counting methods are very time-consuming, without, in our case, adding much information concerning the depositional environment or transporting agent, only a few samples were measured. One curve of each formation is presented to give the reader an idea of the grain-size distribution of the respective formations. The diameters are grouped in classes whose boundaries nearly coincide with the Udden $\sqrt{2}$-scale (1400, 1190, 850, 600, 420, 300, 210, 150, 105, 75, and 50 microns). The classes are plotted along the x-axis. Of each sample, 250 grains were measured.

Determination of carbonates

Since the Lancara Formation consists of limestones and dolomites, a rapid method to determine the nature of the carbonate was required. Recently Warne (1962) introduced a staining method by which the calcite is ultimately coloured red, dolomite remaining colourless, and ankerite is staining purple. However, when this method was applied, most of the rock chips appeared to be stained purple. To test the results, x-ray photographs were made of some of the stained specimens. All purple-coloured specimens appeared to be dolomite identical with the colourless specimens, the red-stained carbonate being calcite. As none of the specimens consisted of ankerite according to Warne's definition, it was assumed that even extremely small quantities of iron affected the staining method.

Notwithstanding this disadvantage, the method was used, the red-stained carbonate being determined as calcite, the colourless or purple-stained carbonate as dolomite.

Directional measurements

Orientation of bedding structures. To measure directions of ripple marks and the axes of slump balls, the tilt compensator of Bouma (1962) was used. The directions mentioned have therefore been corrected as to tilting due to the folding.

Orientation of long axes of grains. The apparent orientation of grains was examined according to the method used by Bouma (1962). Most of the samples chosen show a more or less distinct orientation of the grains. It is noted that they often have a "double maximum" (Schwarzacher, 1951). It is assumed that the long axes tend to be oriented parallel to the current direction, as experimentally shown by Dapples & Rominger (1945). The current directions therefore correspond to the maxima involved.

Dr. Kalsbeek tested one of the samples of the Oville Formation, having been analyzed for its grain orientation, also for an orientation of the c-axes of the quartz grains. These crystallographic axes, however, show a random orientation and do not coincide with the dimensional long axes.

Determination of small-scale sedimentary structures

Recently, Hamblin (1962 a) demonstrated that hidden sedimentary structures are revealed by x-ray photographs. Such pictures of rock slices, several centimeters thick, display small-scale structures, if present. With the cooperation of the Radiological Department of the Leiden University Hospital the method was applied to some samples, the photographs being made by Dr. J. Lips and his assistants. Fig. 12 presents one of the samples. The cross-bedding is clearly visible, but still better results might have been obtained if a thinner slice had been used. This slice was rather thick (6 cm).

The etching method, also developed by Hamblin (1962 b), was first applied to the specimen of which the x-ray photograph is given. Unfortunately, the results were not satisfactory, which can be ascribed to the very slight variations in the mineralogical composition of the successive laminae.

CLASSIFICATIONS AND DEFINITIONS

Classification of sandstones

General remarks. About two years ago the present author, in an internal report, propagated the use of the sandstone classification, proposed by Füchtbauer (1959) (Fig. 4). This classification was chosen and is still preferred for reasons which will be dealt with below. Füchtbauer plotted several analyses of various rocks in his classification triangle supporting his proposition. Recently, deVries Klein (1963) and Huckenholz (1963 a, b) critically reviewed *in extenso* most of the sandstone classifications hitherto proposed, and Huckenholz made it clear that the arkose of the type locality in Auvergne (France) falls in the graywacke field of Füchtbauer's classification, an inconvenience this classification appears to have in common with most of the other classifications. After having defined the term sandstone, the choice of the parameters will be discussed.

Definition. A sandstone is a lithified aggregate of detrital mineral grains or rock fragments having a median grain size of between 50 and 2000 μ and containing less than 50 % of chemically precipitated minerals.

Choice of the parameters. Pettijohn (1948) stated that the properties chosen as parameters of a sandstone classification should be informative as to the genesis of the sandstones, because knowledge of the genesis is the ultimate aim of research. Unfortunately, the choice of such parameters seems to be difficult, as can be concluded from the great number of genetic classifications that have been proposed but which are all unsatisfactory. One aspect of the genesis of the sandstone is rarely revealed by one only property, so we cannot expect to determine its whole history in three or four parameters.

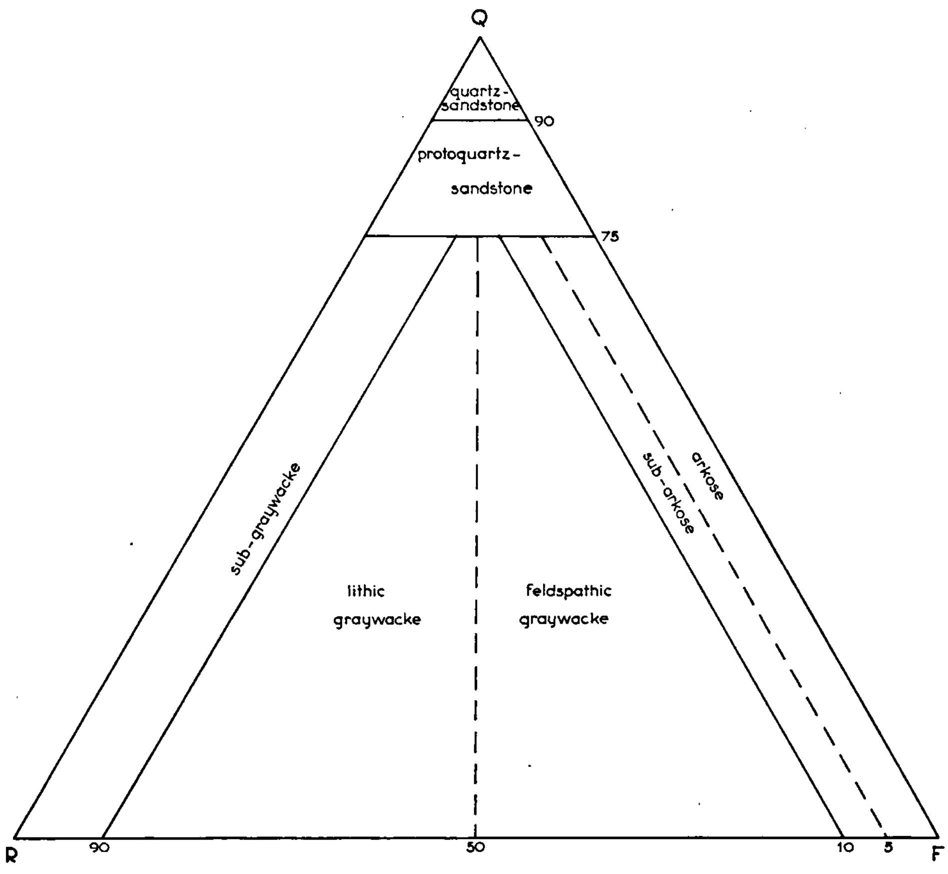


Fig. 4. Classification of sandstones after Füchtbauer (1959), with addition of the subgraywacke group.

Certain properties are too easily overgeneralized. For instance, a high feldspar percentage has been ascribed to a source area composed of plutonic rather than metamorphic rocks. However, for Surinam Bakker & Müller (1957) described fluviatile sediments poor in feldspars although the sediments were derived from granitic rocks. On the other hand, Pliocene sediments derived from the metamorphic Canigou Massif are known to be very rich in feldspars (Oele, Sluiter and Pannekoek, 1963). It should be noted that feldspars may give much information on the source rocks. Tobi (1961) demonstrated a relationship between the law according to which plagioclases are twinned and the kind of rocks in which these plagioclases occur. He supposed the relationship to be useful for sedimentologists as a tool for determining the source rocks.

Other genetic classifications have been proposed, for instance, by Folk (1954), Packham (1954), and Crook (1960), but all these classifications have some disadvantages. Folk's classification is meant to reflect the source rocks; fragments of composite quartz are therefore divided into those derived from metamorphic and those derived from plutonic rocks, but this division seems somewhat arbitrary. Moreover, deVries Klein has pointed out that the quartzes, despite recycling, do not lose their source identity.

Both Packham and Crook used sedimentary structures to classify the sandstones. According to their structures, sandstones are divided into graywackes (deep water deposits from turbidity currents) and arkoses (shallow water deposits). But various other authors have shown that not all graywackes are necessarily turbidity deposits and that they are not necessarily deep water deposits.

On the basis of the foregoing, it appears preferable to make use of a descriptive classification. Parameters other than the mineralogical ones are inappropriate for such a classification. Because our knowledge about sedimentary structures is still preliminary, they cannot be used to classify the sandstones. Moreover, many sandstones lack such structures. Objection can also be offered to the use of grain-size distribution and grain roundness for this purpose. In dealing with lithified sediments, both these characteristics are measured in thin sections, resulting in data that cannot be expressed in exact values, making comparison almost impossible. It seems more adequate to base the classification on the mineralogical composition as is done in petrology. The other properties cannot be ignored, however, and they must agree with the name chosen on mineralogical grounds.

The main detrital constituents of sandstones are quartz, feldspars, rock fragments, and phyllosilicates. Especially quartz and feldspar occur in such amounts that they have been chosen as parameters in most of the existing classifications. The amounts of the other two constituents vary strongly. According to Huckenholz, rock fragments may constitute up to 63 percent of the sandstones, the percentage of the phyllosilicates depending on the percentage of clay minerals present.

Füchtbauer (1959) chooses quartz, feldspar, and rock fragments as parameters, while Huckenholz, modifying the classification of Krynine (1948), used quartz, feldspars, and phyllosilicates. Although the arkoses of the type locality are located in their proper field of Huckenholz's classification, while in Füchtbauer's system these arkoses are located in the graywacke, Füchtbauer's classification is preferred for the following reasons.

Since, in the first place, muscovite and biotite occur in all kind of sediments, and in varying amounts, it is not conceivable that they could be useful in the classification. Moreover, Huckenholz adds to these micas others such as illite and other hydrous micas, which implies that the phyllosilicate parameter consists of muscovite and biotite plus micaceous matrix. This parameter seems too complex to be useful. The opposite holds for rock fragments in Huckenholz's classification; if counted according to their constituting components, there is no indication of how many fragments were present or of what percentage of the respective parameters represents single crystals. In both cases useful information on the primary constituents is lost. Counting cementing silica as quartz has of course the same result.

The objections made to Huckenholz's classification appear to me serious enough to prefer Füchtbauer's system, despite the wrong location of the arkose of the type locality.

It must also be admitted that every classification is only approximate. Sorting and roundness of the grains are of great importance, and in the end they may be conclusive as to the name to be applied to the rocks involved.

In Füchtbauer's system clay contents and cement contents are circumscribed as follows: 10—25 % argillaceous etc.; 25—50 % very clayey. It is noted that quartz-cemented quartzsandstones are termed quartzites as is commonly done.

Thickness of the strata

For the thickness of the strata the terminology of McKee & Weir (1953) has been used. The terms are given in Table II.

TABLE II Terminology of thickness of strata

Terms	Thckness	Terms	Thickness
very thick-bedded	> 120 cm	laminated	0.2—1 cm
thick-bedded	60—120 cm	thinly laminated	> 0.2 cm
thin-bedded	5— 60 cm		
very thin-bedded	1— 5 cm		

The term "metadepositional"

In dealing with slump structures of the Oville Formation, the term metadepositional has been used. This term is meant to designate the stage after a sedimentary bed has settled and before renewed sedimentation takes place. The present author proposed the term to Mr. P. J. C. Nagtegaal, who introduced it in a recent paper (Nagtegaal, 1963).

CHAPTER II

HERRERIA SANDSTONE

INTRODUCTION

The term, introduced by Comte (1959), is somewhat misleading because it is applied to a group consisting mainly of quartzites. Yet some of the beds are poor in cementing material and can indeed be termed sandstones. In the Esla region, the Herreria is found only W. of this river. The best exposure, however, is at La Braña, where a small branch of the Curueño River cuts the upper 200 meters of the formation almost perpendicularly to the strike. The base of the formation is not reached in this profile, the lower part being exposed only in the region of the Luna River which falls outside the scope of this study. The section at La Braña is shown in Fig. 5 a and b.

The transition of the Herreria Formation into the Lancara Formation is poorly exposed in the La Braña section, but can be observed in the Porma Valley and is shown in Table III as measured by Lotze & Sdzuy.

TABLE III Transition of Herreria Formation into Lancara Formation near Boñar (after Lotze & Sdzuy, 1961).

Top:	Lancara Limestone Formation
7 m	not exposed.
20 m	red, thick-bedded, cross-laminated, micaceous sandstones with sandy calcareous beds intercalated.
2 m	dark red, sandy shales with organic tracks.
1 m	cross-laminated, red quartzite.
3 m	alternation of red shales, thin sandstone beds with organic tracks and thicker beds with ripple marks.
10 m	not exposed.
Base:	Herreria Sandstone Formation.

The sediments range from silty shales to conglomeratic quartzites. The bulk of the sediments consists of medium-grained quartzites. The shale beds and conglomerates occur throughout the whole formation, but are irregularly distributed. Lotze & Sdzuy (1961) could subdivide the formation into 3 sequences, but such a division cannot be made in the La Braña section.

The conglomerates contain small pebbles of quartzite, up to 5 cm in diameter, although occasionally the diameter may reach values as high as 1 dm. Comte mentions the occurrence of rhyolitic pebbles near Boñar, but I have not observed such rock fragments. These pebbles are sub-rounded to rounded. They are irregularly distributed in the layers, although lamination is sometimes visible. This will be described in dealing with the stratification.

SECTION OF THE UPPER PART OF HERRERIA SANDSTONE AT LA BRAÑA.
PROFILE IS INTERRUPTED, WHERE SECTION IS NOT EXPOSED.

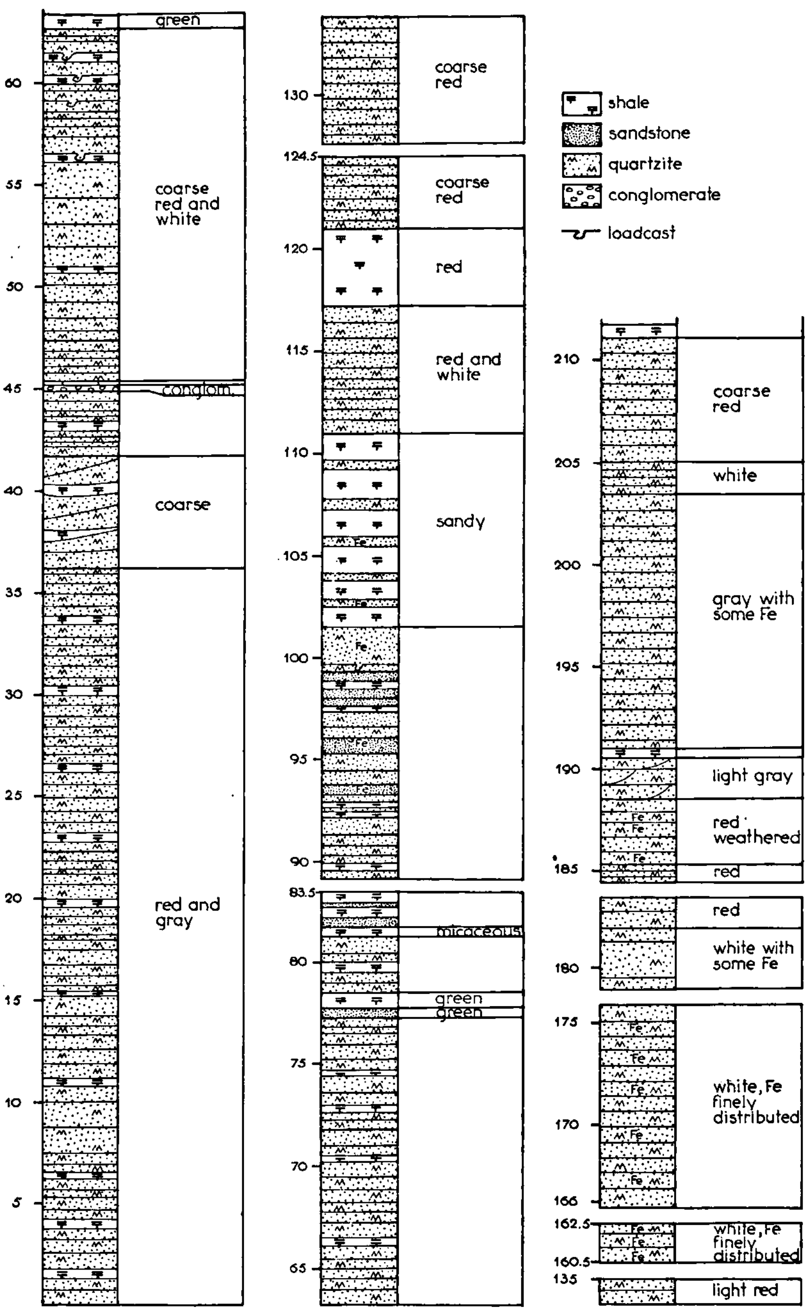


Fig. 5a.

DETAILED SECTIONS OF HERRERIA SANDSTONE
A corresponding to height 186.25 - 189.0 m of sect. given in Fig. 5a
B " " " " 78.80 - 79.81 " " " " " "
C " " " " 62.70 - 62.90 " " " " " "

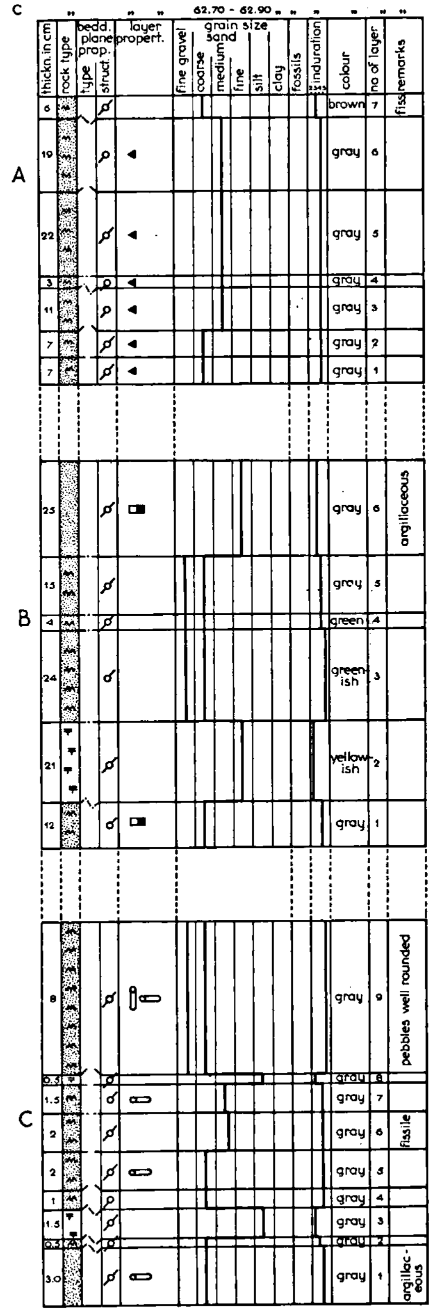


Fig. 5b.

MINERALOGICAL COMPOSITION

The mineralogical composition of the quartzites is comparable to that of the matrix of the above-mentioned conglomerates. Compositional data are given in Table IV. Because of their composition, the rocks must be classed as quartzsandstones.

TABLE IV Mineralogical composition of quartzites of Herreria Formation in percent by Volume

Spec.	Quartz	Sec. q	Feldsp.	Rock fragm.	Musc.	Biot.	Opaque	Clay
204	71.2	9.6	9.4	2.2	tr	tr	1.6	6.0
203	83.5	4.8	3.8	3.1	tr	tr	0.2	4.6
202	66.1	2.0	2.8	1.5	0.2	0.2	2.4	24.6
200	73.5	11.0	1.2	2.5	0.6	1.4	0.8	8.8
199	56.5	6.2	3.5	3.9	tr	1.0	20.4	8.6
198	91.1	3.6	0.4	1.7	0.2	tr	1.0	2.0
197	86.4	4.4	1.2	3.4	0.2	0.8	1.0	2.4
196	76.6	9.6	2.2	3.8	0.6	tr	3.0	4.2
195	88.6	7.0	tr	2.0	tr	tr	1.0	1.4
194	89.7	3.0	0.8	2.3	tr	0.6	0.8	2.8
193	87.8	4.2	0.8	3.2	tr	tr	3.6	0.4

Quartz

The chief component is quartz, comprising about 90 % of constituents Q + F + R. The grains are rounded to sub-rounded, although some sub-angular grains are also encountered and even the originally hexagonal form can occasionally be observed. The quartz shows undulating extinction.

The grains themselves are frequently composite. Composite grains have internal contacts that are elongate or composed of many small quartzes with irregular boundaries.

The quartzes often contain inclusions in the form of rounded tourmalines and zircons, apatites or short grains of biotite, and in one instance even a feldspar. Tiny liquid or gaseous inclusions show parallel lineations, in some cases two sets of lineations crossing each other.

Secondary overgrowths. Most of the grains have a secondary overgrowth. This overgrowth is in optical continuity, but the primary or detrital grains are visible, due to well-developed dust-rings. The overgrowths are not always limpid; they may contain inclusions of iron oxide *. Their surface, as visible in thin section, sometimes even surpasses that of the original grains.

Dust-rings. The dust-rings mentioned above are produced by two different kinds of material: a dark opaque matter, probably manganese (or perhaps organic matter), which is the most common, and clay.

* The overgrowths should be clear according to Gilbert (1958), but, as stated below precipitation of quartz and dust-ring material took place at the same time.

Feldspars

The feldspar in only one case reaches a percentage of 11 %, taking Q + F + R at 100 %; in all other cases the content varies between 4 % and less than 1 %. It is chiefly microcline; few grains of plagioclase are found. The grains are present in fresh as well as in altered condition. The alteration product is mainly kaolinite which derived from the microcline, sericite being the product of the plagioclases. In the sediment we do not see transitional specimens between the clear, non-kaolinized microclines and the kaolinized ones. Grains limpid only in the centre were not observed, but grains kaolinized only in the centre were seen. These observations lead to the conclusion that the kaolinization antedated the diagenetic stage and that for the same reasons an origin due to weathering in the source area may be excluded. Consequently, the kaolinized microclines must have derived from igneous or metamorphic rocks.

Hematite was found especially in the kaolinized microclines. It is present in irregular patches, but is also distributed throughout the kaolinite as submicroscopical grains. It seems reasonable to assume that the hematite was introduced by circulation of water in the diagenetic stage. Commonly, precipitation appears to have taken place along cleavage and twinning planes.

Many microclines have secondary overgrowths giving rise to rhombohedral outlines. Dust-rings as clearly visible as those around the detrital quartzes have not been developed. The difference between the overgrowths and the detrital feldspars is mainly that the former are not turbid or contain only some hematite grains, while the latter can be kaolinized. Another criterion for distinguishing between the two is the frequently-occurring optical discontinuity. Fig. 6 gives an example of a feldspar with secondary overgrowth.

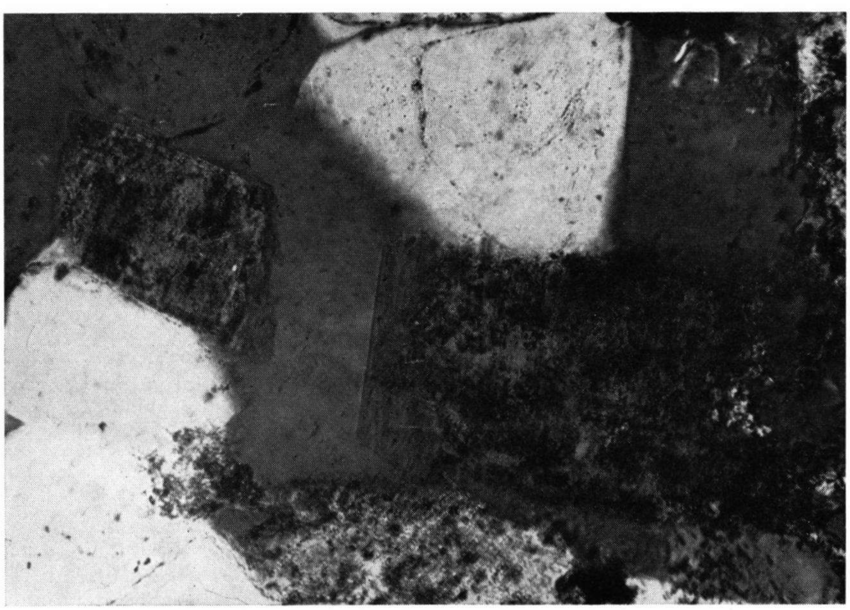


Fig. 6. Feldspars with secondary overgrowths. crossed nicols. 200 ×.

The overgrowth here extinguishes simultaneously with the adjoining quartz overgrowth, but this may be accidental.

Most of the feldspars also have a rim of chloritic material. In such cases the outlines of the feldspars themselves are far more irregular than those of the feldspars without the clay rims. It is likely that the clay was formed by alteration of the feldspars during diagenesis.

The feldspars are less rounded than the quartz grains. Euhedral grains are not uncommon, but they usually show worn edges. In one case quartz inclusions were observed in microcline.

Micas

Among the detrital constituents, the micas are present only in small amounts, generally not more than 1 %. Biotite as well as muscovite occur, but both can have altered due to sericitization. The length of the flakes varies in the main between 100 and 350 μ, which is not very long in comparison with the observed diameters of the quartz and feldspar grains. The biotites are pleochroic from light to dark green, but also light to dark brown. The micas frequently contain ferrous material, which is sometimes nicely arranged along the basal cleavage. The flakes terminate in a sheaf-like fashion, but they may also have rounded edges. I want to make special mention of some biotite flakes with a longitudinal axis perpendicular to the basal cleavage: in this respect they resemble authigenic mica, but here such biotite is not of an authigenic origin, firstly, because the same mineral has been encountered with rounded edges, and secondly, because they are not in a bleached condition. The last argument weighs the more heavily because other biotites show bleaching features.

Rock fragments

These too, form part of the mineralogical content. They are grains, and even pebbles, of quartzitic composition, characterized by internal suturing or elongate contacts. Other rock fragments are well-rounded pebbles presumed to have consisted originally of clay. In fact, they are aggregates of extremely small grains which due to their size cannot be determined with the microscope. A few chert grains are also seen.

Binding material

The matrix consists mainly of sericite in the finer-grained beds, and the coarser quartzites also contain chlorite. Both sometimes form spherulites, suggesting an authigenic origin.

The cement is quartz, with some feldspar and hematite. In a few cases the hematite becomes very important as cementing material (see Table IV, 199), so that we may call such quartzites iron quartzsandstones. Only a few occur.

TEXTURE

Grain contacts

The fabric is strongly dominated by the secondary overgrowths. The contacts of the secondary overgrowths with each other or with the original quartzes can be straight, slightly curved, zig-zag, or slightly sutured. The contacts of the original quartzes with each other may also be tangential. Fig. 7 shows a beautiful example of several modes of contact: straight, slightly curved, zig-zag, and slightly sutured. The slightly curved and zig-zag contacts can be readily explained by the meeting of overgrowths, starting on primary grains, whose crystallographic axes lie unoriented in the sediment, thus giving a random position of the crystallographic axes of one grain with respect to the others. The presence of such contacts does not imply a deformation of planes that were once straight.

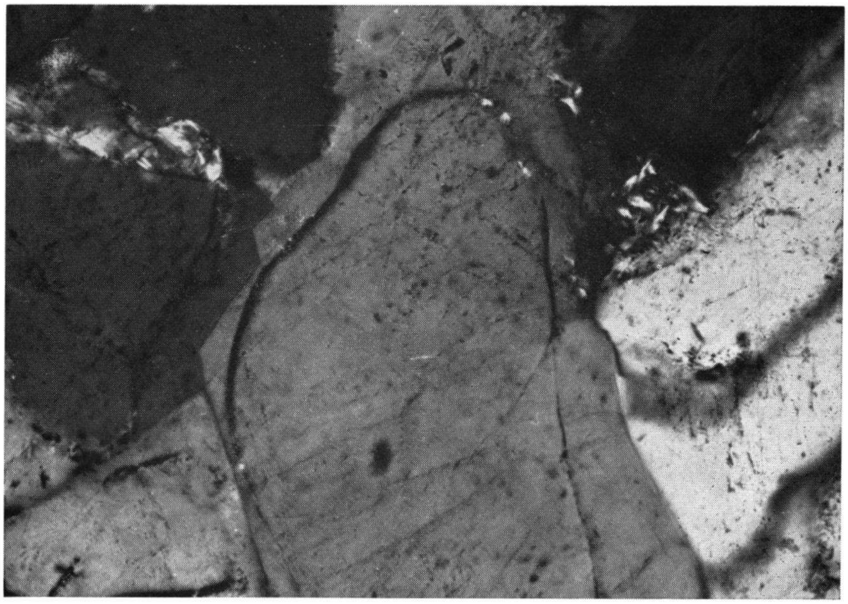


Fig. 7. Various types of grain contacts of quartzes. crossed nicols. 200 ×.

In several instances there can be no doubt that the overgrowth tends to develop outlines in accordance with the crystallographic faces. Although Gilbert (1949) denied such a tendency, recent measurements done by Basumallick (1962) have revealed its existence.

The simplicity of the grain contacts reveals simple pore filling (Taylor, 1950) rather than pressure solution. The few suturing contacts are probably due to local stresses. Their scarcity makes it impossible to term them pressolved beds, although locally pressure solution did occur. It may be noted that the Herreria sandstones were covered by sediments ranging up to about 2700 metres in thickness.

The contacts between the quartzes and the micas are mainly straight. The micas sometimes show a sheaf-like termination at both ends, giving the impression of mushrooming (Taylor, 1950) and resulting from being pressed together in the middle and finding release at both edges (Fig. 8).

Moment of origin of secondary quartz and dust-rings

As we have already seen a dust-ring may be built up of either manganese (or organic matter) or clay. The first precipitated as tiny crystals dispersed in a zone around

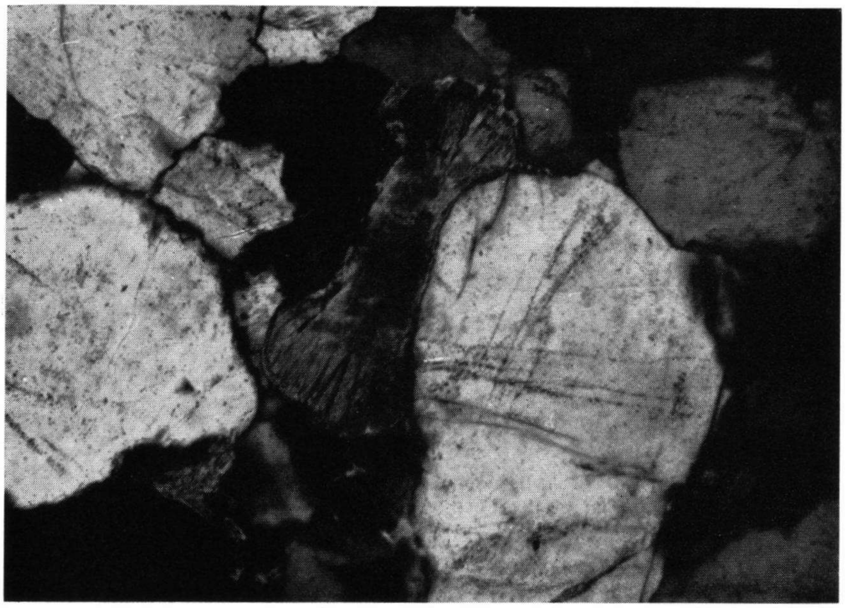


Fig. 8. Mica showing features resembling mushrooming. crossed nicols. 200 ×.

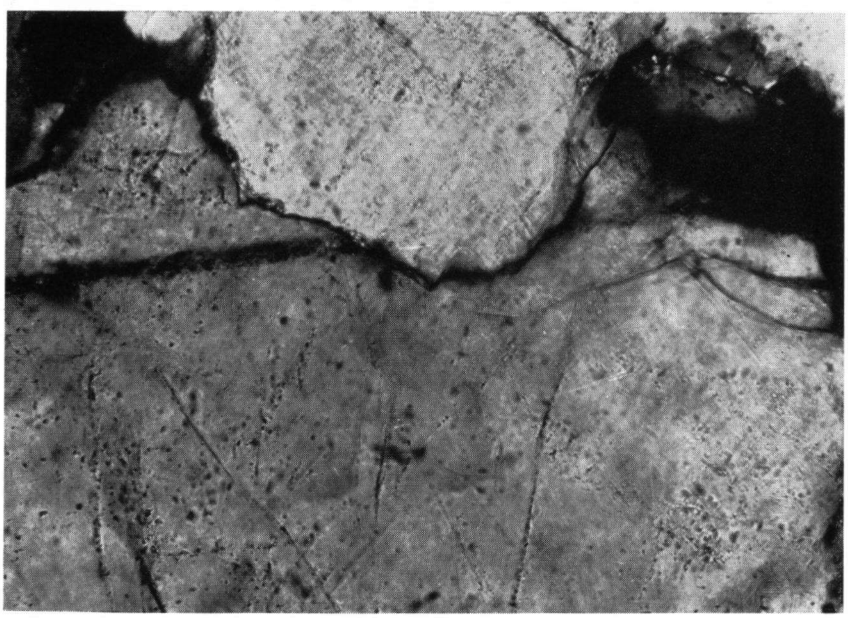


Fig. 9. Pitting of quartz due to pressure solution. crossed nicols. 200 ×.

the primary grains rather than forming a closed rim by juxtaposition. In fact, the manganese started precipitating at the same moment as the secondary quartz but stopped earlier.

Closer investigation of the dust-rings in their relation to the filling of the pits makes it possible to approximate the time of precipitation of the secondary quartz and the dust-rings. Generally, in the pits we observe clay overlying the dust-rings, although the latter is present here in minor amounts. The clay is assumed to be detrital in origin. There are only two stages in which settling of mechanically-supplied clay would be possible: the depositional and the early burial stages. If settling occurred in the former, we must consider the clay now present as residual, that is to say a remnant of what was winnowed out; if settling dates from the latter stage, it was normal sedimentation in almost standing water. Unfortunately, there is no further evidence to indicate which of the two stages we are dealing with. From the reduction of the amount of dust-ring material underlying the clay and the presence of particles of it in the clay itself, the conclusion can be drawn that not much time elapsed between the forming of the dust-ring and settling of the clay.

With respect to the origin of the pitted appearance of the quartz grains, the moment of precipitation of the manganese is also important. Since most of the pits show filling with manganese and clay, we know that their formation antedates at least the early-burial stage. They thus originated in the depositional stage or earlier. This is not in accordance with the supposition made by Giles (1932) that pitting is more often due to interpenetration of quartzes and their overgrowths, but does not necessarily exclude the occurrence of this kind of pitting. An instructive example of the origin of the pits by pressure solution is given in Fig. 9.

The grains in the Herreria sandstone presumably became pitted by a solution process as indicated by Kuenen and Perdok (1962). Although secondary overgrowth may be absent, where clay is present on the grain's surface this is not generally the case. I could not find any indications of replacement of clayey material by (secondary) quartz as suggested by Heald (1956a).

Source of the secondary silica

The source of the quartz building the cement of so many sandstones is still being debated. Pettijohn (1957, p. 656) gives a summary of the existing theories and Siever (1959) also discusses the matter to some extent, subjecting the hitherto accepted hypotheses to a critical review.

Intrastratal solution and subsequent precipitation is commonly held to be the main source. Waldschmidt (1941) and later Heald (1950) regarded the silica released by interpenetration of the quartz grains and pressure solution to be of great importance in the formation of the cement. Although in the above-mentioned paper Siever even doubts the action of the process of pressure solution in general, solution due to interpenetration of the grains must be rejected as far as the Herreria sandstone is concerned, because it is not observed sufficiently frequently. The grain contacts indicate that only small amounts of quartz have been dissolved at those sites.

Another hypothesis is that of Goldstein (1948), who supposes that the small grains dissolve and that the dissolved silica precipitates onto the larger grains. It is impossible, however, to establish the presence in the past of grains that are now absent. It seems to me that the influence of such dissolution must have been limited here, because it is so unlikely that more than 10 percent of the volume of the quartzites which is occupied by secondary quartz originated from solution of small grains and subsequent precipitation on larger grains.

Indirect precipitation from normal seawater has been advocated by Siever (1957). This process is mainly a biochemical one. Silica-secıeting organisms such as diatoms withdraw silica from the seawater and build their skeletons of amorphous silica. After burial, the skeletal remains dissolve rather easily and this matter precipitates again as secondary overgrowths and small quartz crystals. Some objections can be made to this hypothesis. Even for late-Paleozoic deposits the presence of diatom-like forms is hypothetical (Siever, 1959), and this will hold the more strongly for the early-Paleozoic sediments. Besides, in a more recent paper Siever (1962) records the presence of undissolved diatom skeletons, while the sediment water is undersaturated with regard to amorphous silica. Apparently, solution requires special chemical conditions. Although silica-secreting organisms may play a role in concentrating silica, a quartz cement in sandstones will originate only occasionally via the medium of those organisms.

Mineralogical alteration has also been considered as a possible source of the silica. Siever (1962) and Towe (1962) both mention the liberation of silica due to alteration of montmorillonite into illite or muscovite. The small amounts of clay and muscovite in the Herreria sandstone make it impossible to explain the origin of the secondary quartz in this way.

Direct precipitation. Siever (1959) also suggests direct precipitation from the formational waters. During diagenesis, the formational waters show increasing salinity. Since NaCl appears to influence the solubility of quartz (see also page 44), this increase would cause the precipitation of quartz. However, the cause of the increasing salinity is a difficulty here. De Sitter (1947) and von Engelhardt (1961) propose the action of clay beds as ionic filters, but the shales in the Herreria sandstone are mainly silt and cannot have played such a role.

Supply. A last suggestion, also from Siever (1959), is a supply. Weathering, chiefly chemical, in the source area supersaturated the rivers with respect to silica. Entrapped in the sediment, the excess silica was precipitated. Todd (1963) assumes such a source for the quartz cementing the Tensleep sandstone. He finds an argument favouring this theory in the decreasing amount of cement with increasing distance from the source area. It must be admitted that he also attributes an influence to aeolian activity: tiny splinters (100—1000 Å) split off from the quartzes and were blown into the waters, where they easily dissolved, due to their small size, raising the amount of solved silica. There are no indications to justify the assumption of aeolian activity in the source area of the Herreria sandstone, but an oversaturation of fluviatile waters remains possible.

The supply, however, requires special climatic conditions in the source area. The hypothetical chemical weathering could be expected in a tropical humid climate, but no evidence of this except the reddish colour of the deposits can be found. The conditions, however, under such climatic influences must have been very different because of the absence of a vegetation cover in the early-Paleozoic. The amounts of clay are rather low, but it is possible that clay was winnowed out from such near-shore sediments.

None of these hypotheses is satisfying. If the last two appear to me to be the most reasonable with regard to the Herreria sandstone, this is nevertheless based on rather weak arguments, being only that both increasing salinity in formational waters and supersaturated surface waters have indeed been observed.

Grain-size distribution

It appears to be difficult to get a picture of the grain-size distribution Not all of the quartz grains are surrounded by dust-rings, and as a result the diameter of these detrital grains cannot be measured. Such grains are of course not to be used in the counting procedure. When the number of grains discarded for this reason becomes too large, the result of the measurements must show discrepancies with reality. Hence, only a few beds give useful data. Fig. 10 represents the grain-size distribution of one of these beds. Although the sorting cannot be expressed in the coefficient of Trask, it is at once clear that the degree of sorting must be very good.

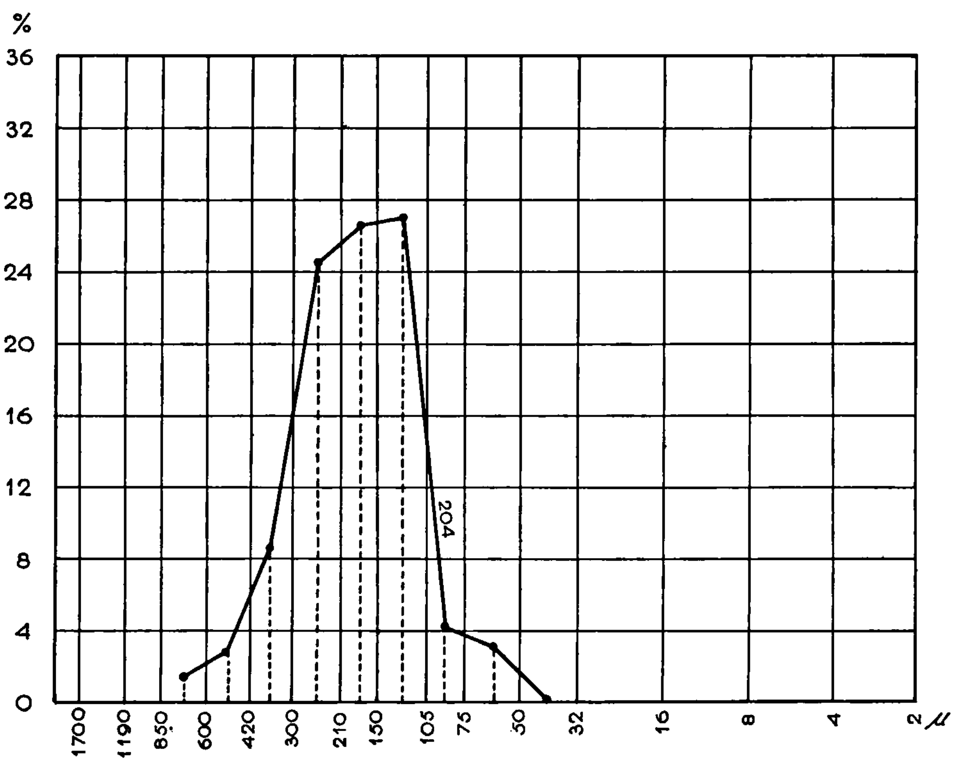


Fig. 10. Grain-size distribution of an average quartzite of the Herreria formation.

Grain orientation

Fig. 11 shows that the long axes of the grains have a distinct preferred orientation. The four examples given show two directions. Although the measurements made on only a few specimens cannot serve to determine the depositional environment, the changes in orientation revealed by these four examples are in good agreement with the environment as suggested for the Herreria Formation.

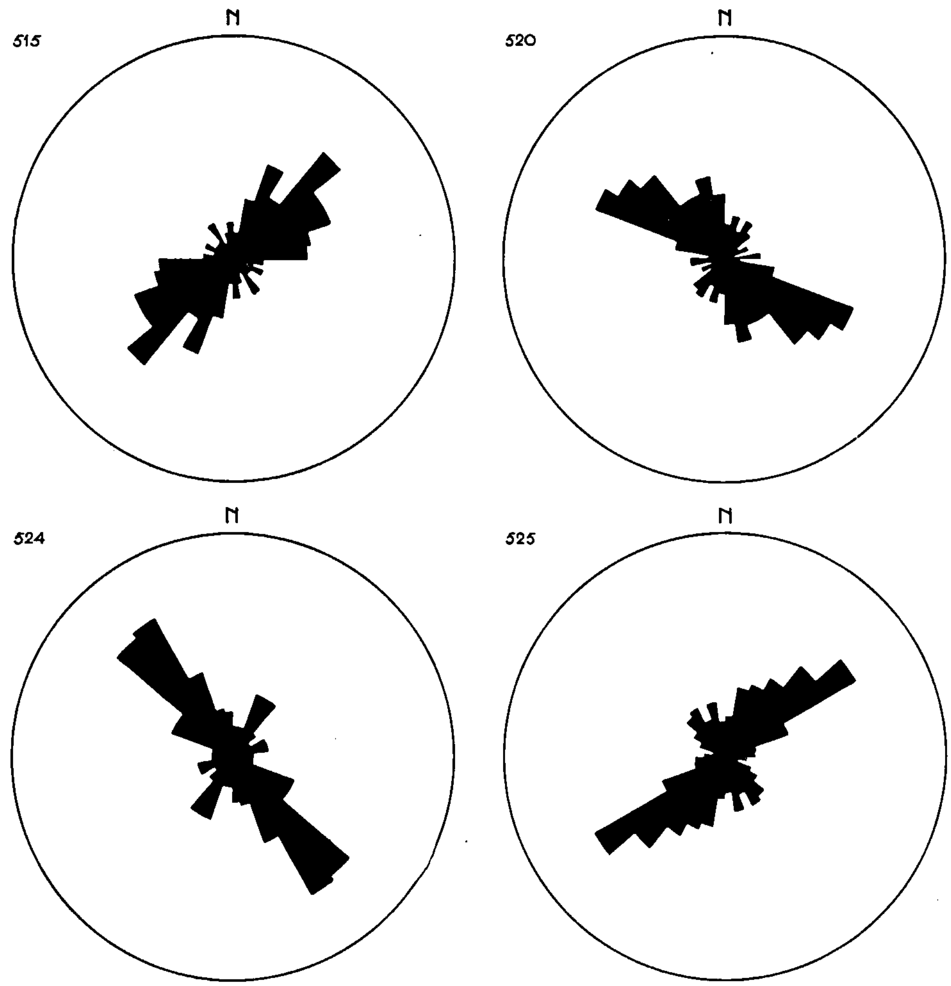


Fig. 11. Orientation of the long axes of grains.

LAYER-PROPERTIES

Bedding contacts

The contacts of the layers are mostly sharp or distinct. A more gradual transition between the layers is exceptional. The form of the contact planes is in general undulating or irregular, but sometimes even flat.

Thickness

The thickness of the beds varies considerably. The quartzitic layers are on the average 20—30 cm thick, but the values vary between 1 and 50 cm and in one

Fig. 12. Quartzite lens showing internal cross-lamination. X-ray photograph. About half natural size.

case even a 2.10 m bed was found. The conglomeratic beds in some places reach considerable thickness (up to 4.00 m). The shale beds are ordinarily either a few centimetres or about half a metre thick. Several of them measure more than a metre, but in such cases thin beds of quartzite are intercalated. The whole complex is medium to thin-bedded.

Stratification

The stratification is on the whole of the parallel type. In the La Braña profile, however, a very rapid change of thickness in a lateral direction was observed in two series of layers. Whithin a metre the thickness of the beds may increase from 0 to about 50 cm. The beds in this way build an interfingering pattern. Most of the beds have convex upper bedding planes. The beds appear as homogeneous; no internal lamination could be detected. Some thin, shaly layers are intercalated. Because of their form, I am inclined to consider these beds to have been sand bars, which are in fact megaripples.

Along the road to Oville some peculiar deviations of the normal stratification occur. In this locality I observed a shale bed containing quartzite lenses with rather flat bottom planes and convex top planes (Fig. 12). These are more or less plano-convex lenses with a height of about 10 cm, a long axis of 20 cm, and some 15 cm wide. In the field I explained their presence as being remnants of beds that had been removed by erosion. After making a polished section the internal structure was still not clear, but after making an x-ray photograph according to the method introduced by Hamblin (1962a), some of the structure could be detected. This structure suggests that the lenses are indeed remainders of ripple marks.

Another irregularity in the parallel stratification is visible a few metres above the shale beds containing the quartzite lenses just mentioned. Fig. 13 shows a

Fig. 13. Fore-set beds in the Herreria Sandstone along the road to Oville.

photograph of the layering. Here we are dealing with a cross-bedded series. Due to synsedimentary erosion and some minor folds, the picture is not immediately clear.

Although the formation is a well-bedded complex, it is difficult to see any internal lamination of the beds. In thin sections the beds appear to contain small-scale current bedding features.

These features indicate only variations in the direction, not in the velocity, of the current, because the grain size in the laminae differs but slightly.

In some of the conglomeratic beds, lamination is clearly visible. This lamination is produced by a kind of grading which is in some cases also the texture of a whole bed. Such a bed or lamina has small quartz pebbles at the bottom which decrease in size upwards. As a result, the upper part consists of coarse sand. The decrease is not as gradual as is the case in true graded bedding settled from turbidity currents; there are jumps in the decrease of the grain size. In the sandy part of the laminae, isolated pebbles also occur. In addition, the laminae sometimes show a change in grain size in a lateral direction; the sandy merging into coarser material (Fig. 14). The lamination itself is inclined, corresponding to the foresets of megaripple lamination.

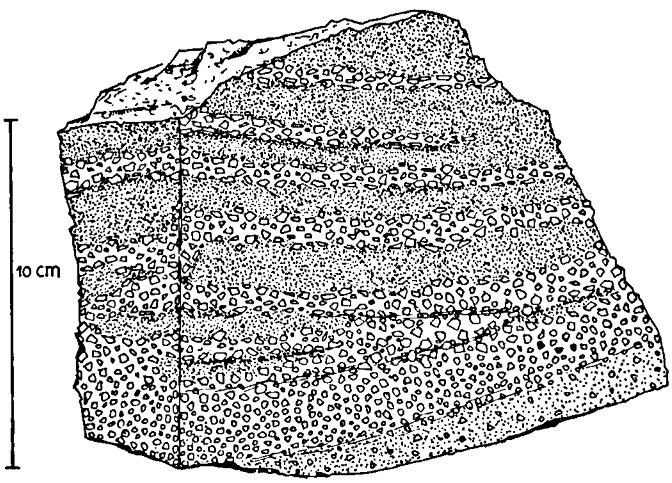


Fig. 14. "Graded" conglomeratic layers in the Herreria Sandstone.

The vertical grading indicates the fluctuating character of the current that formed the ripple pattern. The lateral variety in grain size and bedding irregularities are explained by small changes of current direction with subsequent erosion and renewed deposition. These fluctuations were not so great as to change the whole ripple pattern, but they did influence the size of the particles which settled and came to rest on the bottom.

Colour

The colour of the sediment is rather characteristic. Most of the coarser deposits are red or white, but a few are green. These last, however, contain a high percentage of clay (Table IV, 202) and in this respect differ from the other quartzites. The shales have a red or greenish colour. The green colour of the shales is different from that of the quartzites. The clayey material in the green quartzites is chloritic; the shales, to the contrary, lack chlorite. The origin of the green colour of the shales might be explained by presence of Fe·· in illite, as has been stated by Rasumowa (1960). The red colour can be ascribed to the presence of hematite.

Clay pebbles

Clay pebbles occur throughout the formation. They measure about 0.5 to 1.0 cm and are well rounded.

BEDDING-PLANE STRUCTURES

Load casts

Load casts were observed several times. Fig. 15 shows a load cast with a flat bottom. The latter is due to the small thickness of the clay layer underlying the sandstone bed. In the load cast an impression is visible, caused by a pebble that rested on the bottom of the clay bed. Often load casts are asymmetrical, which can be explained by assuming a lateral flowage of the sand during its downward movement. A relation between the size of the load casts and the thickness the sandstone bed from which they originated could not be found: all the observed load casts are comparatively small because the clay beds in which they sank were thin.

Burrows and tracks

Burrows and tracks are also found, with a diameter of about 3 mm.

Fig. 15. Impression of gravel in the bottom of a load cast. About natural size.

THE SOURCE AREA AND DEPOSITIONAL ENVIRONMENT

Source area

The maturity of the sediment actually complicates a reconstruction of the source area. The minerals occurring in the sediments are only of the most ordinary kind, and not even the inclusions in the quartzes are indicative in this respect. The only diagnostic mineral present is the blue tourmaline. According to Krynine (1948), this mineral was initially derived from pegmatites. Because it is described as one of the minerals most resistant to mechanical wear, the roundness must be the result of long-distance transport or polycyclicity. As is discussed below (page 89), it is held that the source rocks were non-sedimentary.

The climate in the source area must have been warm and at least seasonally humid, as can be concluded from the red bed colour of the sequence (see also page 00).

Depositional environment

The coarseness of the sediments indicate either a marine (but near the shore) deposition or a fluviatile deposition. Several arguments, however, contradict a fluviatile environment: the high degree of sorting, the scarcity of fluviatile structures, and the great extent of the deposits. Consequently, it is concluded that the environment must have been marine. The cross-bedding and the delta-like foreset beds, as well as the observed grading, reflect a deposition in rough, shallow water, perhaps even influenced by fluviatile currents. The coarse-grained layers would then represent sand from which the clay was winnowed out. The preservation of the red colour, which implies oxidized waters, also strengthens this supposition. It is generally assumed that load-casts are indicative of rapid accumulation. Summarizing, we may say that the environment must have been shallow neritic near the shore.

CHAPTER III

LANCARA LIMESTONE

INTRODUCTION

This formation is well exposed in the Esla and Porma regions, although in the Esla region, due to tectonic movements, only the uppermost part is present. In the Porma valley near the village of Vegamián, the formation measures 115 meters, but the total thickness varies considerably (De Sitter & Zwart, 1957 p. 18) elsewhere.

The Lancara Formation can be divided into two parts: the Dolomites s.l. and the so-called Griotte. Great differences in texture make both easy to discern in the field. Because of their special nature and the way in which the griotte forms the transition into the younger Oville Formation, these two parts will be dealt with separately.

Fig. 16 shows a section through the formation near the village of Valdoré in the Esla valley. This section is given even though it lacks the lowermost part, because the transition into the Oville Formation is much better developed and exposed here than elsewhere.

LANCARA DOLOMITE S.L.

The Lancara dolomite s.l. comprises several types of carbonate rocks, among them breccias, oolitic limestones, and dolomites. The basal part of the formation, as seen in an outcrop near Vegamián (some km north of Boñar) consists of shaly dolomite beds measuring a few centimeters each and thin-bedded dolomite beds. The younger layers are thicker and constitute a thick-bedded complex. Fresh fracture planes show the gray colour of the rocks, but on weathered surfaces it is yellow-white. The uppermost beds are reddish, due to hematite finely distributed throughout the rocks. These beds are also somewhat coarser (medium to coarse-grained) than the older ones (fine to medium-grained).

The red bed sequence, measuring about 10 m, consists mainly of calcite, but dolomites are still intercalated. The limestone texture corresponds to the texture of the limestones in the griotte part and will therefore not be dealt with separately. It is noted, however, that the griotte limestones contain more organic and clastic fragments, most of the former limestones having a more crystalline appearance due to grain growth.

Dolomites

The dolomites predominate, x-ray diffraction patterns revealing the high degree of purity of their chemical composition. The texture, however, is not very uniform, and thin sections make it possible to distinguish three different types. The first type of dolomites comprises those in which organic remains such as shell debris are still visible due to a somewhat coarser grain size than that of the surrounding dolomite. Besides the difference in grain size, the dolomite of the shell fragments

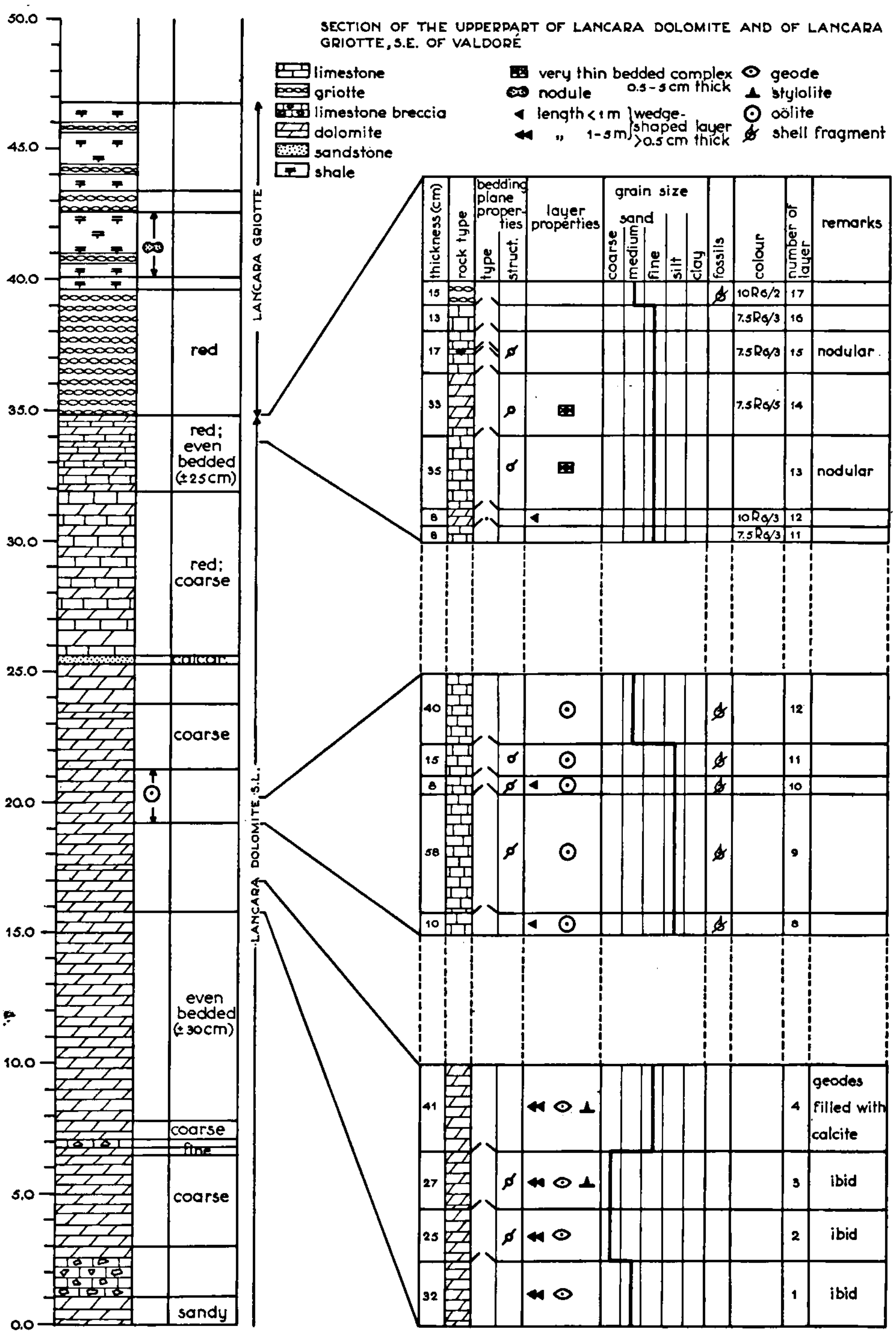

SECTION OF THE UPPERPART OF LANCARA DOLOMITE AND OF LANCARA GRIOTTE, S.E. OF VALDORÉ
limestone
griotte
limestone breccia
dolomite
sandstone
shale
very thin bedded complex 0.5-5 cm thick
nodule
length < 1 m
" 1-5 m
wedge-shaped layer >0.5 cm thick
geode
stylolite
oölite
shell fragment
LANCARA GRIOTTE
LANCARA DOLOMITE, S.L.
red
red; even bedded (±25 cm)
red; coarse
coarse
even bedded (±30 cm)
fine
sandy
thickness (cm)
rock type
bedding plane properties
type
struct.
layer properties
grain size
sand
coarse
medium
fine
silt
clay
fossils
colour
number of layer
remarks
nodular
geodes filled with calcite
ibid
10R6/2
7.5R6/3
7.5R6/5
10R6/3
50.0
45.0
40.0
35.0
30.0
25.0
20.0
15.0
10.0
5.0
0.0

Fig. 16.

is limpid, whereas the surrounding dolomite is turbid. The second type is the most frequent, and in thin section exhibits a mosaic structure built up of an alternation of coarser and finer patches (fig. 17). Although the pattern of the two different grain sizes is recognizable, the range of each is rather wide so that the ranges may have an overlap, as can be seen from the values in Table V. The boundary between the patches is seldom sharp and always highly irregular, producing embayments

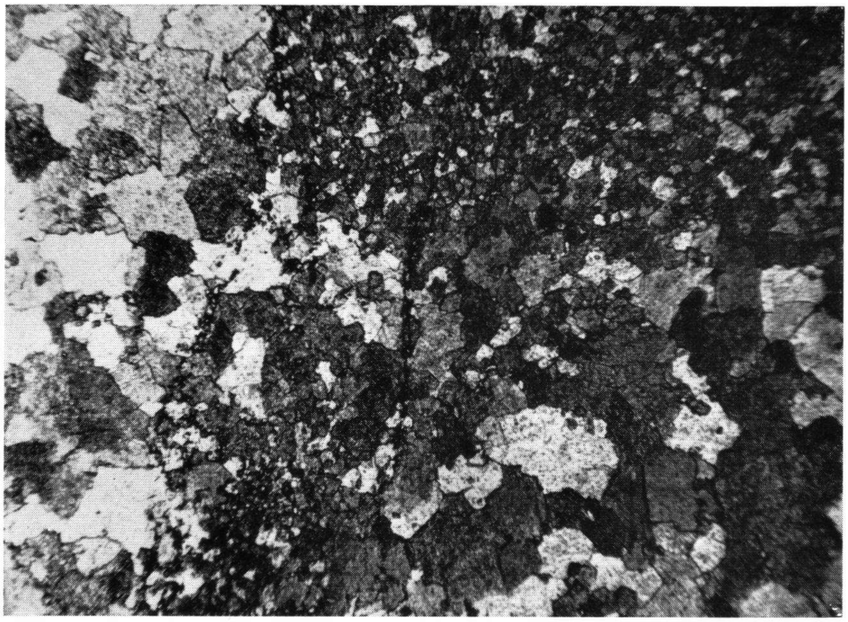


Fig. 17. Alternation of coarser and finer patches in dolomite. crossed nicols. 30 ×

TABLE V Some examples showing the twofold character of the grain-size of most of the dolomites, in microns.

hand-specimen	490	491	492	493	494	495
fine-grained mosaic	30—50	—120	100—150	±30	60—100	30—90
coarse-grained mosaic	150—500	150—500	200—500	100—300	150—300	100—300

of the one in the other. Either the coarse-grained or the fine-grained part can be dominant; the extreme case, being a mosaic consisting of only one of these parts, represents in textural respect the third type. This last type may also show a great spread. So we may list the various types as follows, after which the relations between them will be discussed:

1. Mosaic texture with ghost structures of fossils.
2. Mosaic texture with two ranges of grain size.
3a. Mosaic texture of either coarse or fine grains, but uneven-grained.
 b. Idem, but even-grained.

ad Type 1. The first type represents the simplest case. It is clear that dolomitization took place after sedimentation because the shells must have been built up of calcite or aragonite. (The same therefore holds for the beds without fossils, as will be seen below.) It is interesting that grains similar to those constituting the organic remains also build other forms: rhombohedral (Fig. 18) and rounded triangular forms can be observed. Yet the only explanation seems to be that they, too, must be considered organic in origin and that they are merely cleaved parts of such large-sized fossils as trilobites.

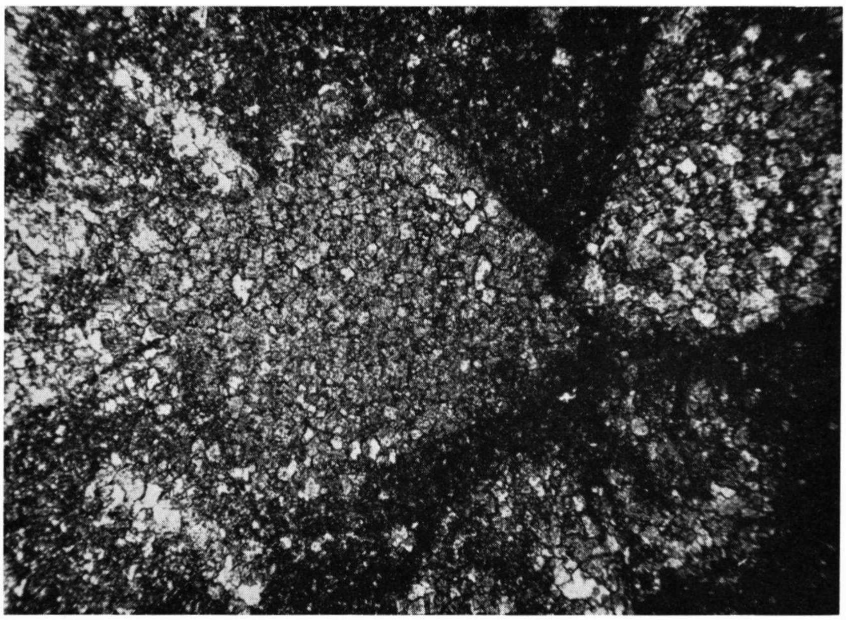


Fig. 18. Rhombohedral forms built by grains coarser than surrounding dolomite. crossed nicols. 80 ×.

ad Type 2. The second type is thought to be the result of grain growth. The process of grain growth has been recognized in metallurgy as a solid-state recrystallization producing a coarser grain size. Bathurst (1958) summarized its characteristics and attributed certain changes in the texture of limestones to the same process. More recently, the same author (Bathurst, 1959) ascribed the origin of pseudobreccias (limestones with small isles of a finer grain) to grain growth also. Although among the dolomites I have observed a pseudobreccia only once, the contacts between the two grain-size ranges in the other hand-specimens suggest that this kind of texture is intermediate between the initial dolomite texture and pseudobreccia texture. I will therefore assume that the mixed textures are grain-growth patterns, the degree to which fine or coarse dominates reflecting the extent of the process.

ad Type 3. Increasing grain growth would certainly lead via the pseudobreccias to the third type. Because of the scarcity of pseudobreccias, there must also be another way by which recrystallization is reached. Some observations may clarify this statement. In the field I noticed patches, a few meters in diameter, which attracted the attention by their gray colour and conspicuously different weathering type (Fig. 19). These patches intersect the beds and have rather irregular outlines.

X-ray photographs revealed that the chemical composition of a given bed is exactly the same outside and inside the patches, but thin sections suggest that the difference is due to a change in grain size, the material inside the patches being coarser. It is remarkable that even calcitic beds show no change in chemical composition but, again, only in grain size. The coarsening is accompanied by a textural change.

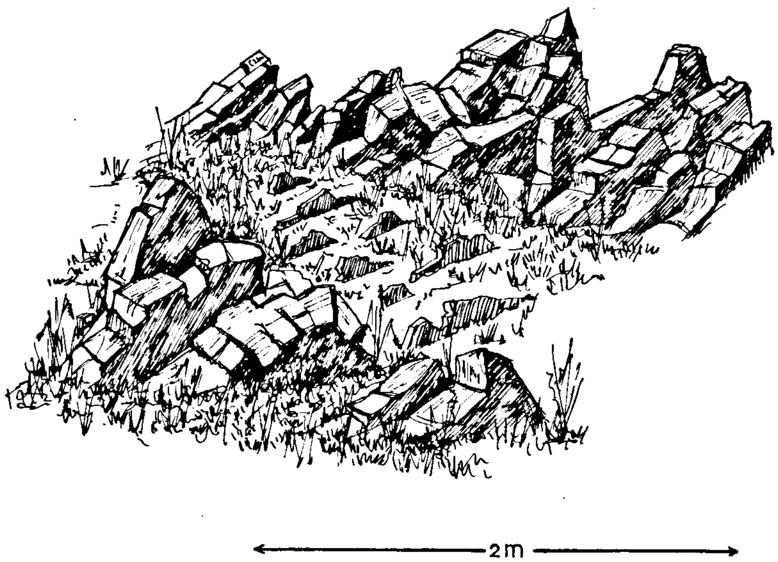


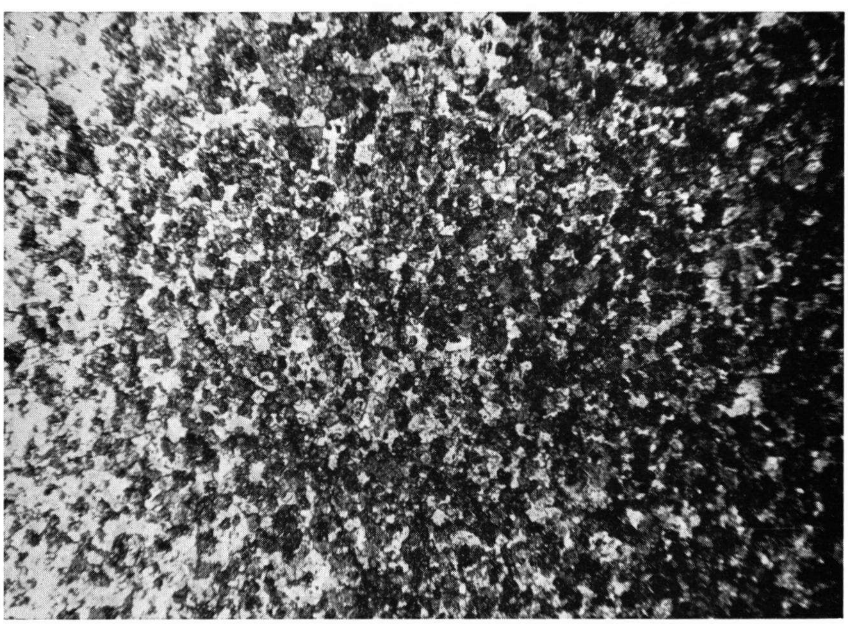
Fig. 19. Patches remarkable for their gray colour and slightly different weathering.

The coarser parts may be assigned to our Type 3, while the finer parts are representative of the other two types. Fig. 20 shows the difference between the first and third textural types in a single bed. The grain size of the beds within the gray patches is not at all uniform, because it appears to be dependent on the initial grain size as we now see it outside the gray spots. The coarsest beds contain crystals showing bent twinning planes, apparently the result of mechanical stress. Consequently, it may be assumed that the third textural type represents a recrystallization caused by mechanical stresses, but it is not clear why the stress acted in such irregular patches.

The question arises as to what constitutes the differences between grain growth and this last kind of recrystallization. The process of grain growth occurs in the diagenetic stage and does not require exceptionally high stresses or temperatures, at least not higher than those normally produced by depth of burial and overburden. The various stages of grain growth observed indicate the slowness of the process. On the other hand, the recrystallization which created Type 3 must have been the reaction to higher stresses, as demonstrated by the strongly bent twinning planes. The local occurrence forms an argument in favour of the short duration of the action of the stresses.

Minor changes. Cavities and cracks produced by solution and perhaps also by a decrease in volume after dolomitization have been filled up by clear calcite. The mineral shows rhombohedral outlines which are accentuated by ferric inclusions aligned along such planes, thus constituting, as they have been called by Cayeux (1935), "crystaux capuchonnés".

A. texture of limestone bed outside the patches. crossed nicols. 28 ×



B. same bed within the patches. crossed nicols. 28 ×.

Fig. 20. Difference in texture inside and outside the patches shown in Fig. 19.

Cross-lamination. Weathered surfaces of the dolomites often show a micro-cross-lamination which must be due to differences in composition. Mostly these differences are too small to be observable, but in a few instances there can be no doubt that the lamination concerns alternating quartz-free and quartz-bearing laminae. On a somewhat larger scale, a kind of gullying has been observed in the La Braña Valley (Fig. 21). These features indicate not only a rough-water environment but also a detrital origin (Harbaugh, 1959). The detrital character implies mechanical transport, but there is no way to determine the length of transport. The depositional environment must have been shallow, as indicated by the cross-lamination.

Calcareous sandstones. Other environmental indicators are some calcareous sandstones which occur between the dolomite beds. The sand-size particles are all limpid quartz grains. Their size reaches a value of 500 μ. Dust-rings of opaque material clearly mark the outlines of the grains and accentuate their sub-rounded nature. The occurrence of the sandstones in combination with the cross-lamination suggests a shallow-neritic environment.

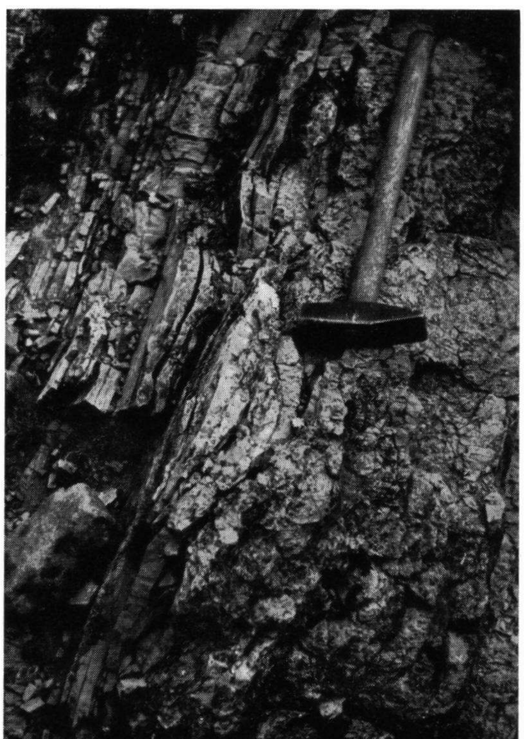


Fig. 21. Gully-like deposits of Lancara dolomite at La Braña.

Dolomitization. This was assumed to have been a diagenetic process here, in agreement with the conclusions drawn from the fossil-bearing dolomites, but actually there is no further support for such a supposition. However, Siegel (1961) determined pH to be the most important factor influencing the precipitation of dolomites, but the pH value must be rather high (9.7 at 25° C) and could not have been reached in the supposed depositional environment. Indeed, Chilingar & Bissell (1963) think such high values of pH unfavourable to the formation of primary dolomites, and state that a low pH value ($\leqslant 8$) is required in an environment with high CO_2 pressure and high salinity. In discussing the oolitic limestones we will find reasons to assume that the waters had a high salinity and, since a pH of about 8 is a common value, conditions would be favourable for precipitation of dolomite according to Chilingar & Bissell. On the other hand, in their note they again draw attention to the opinion of Degens that primary dolomites do not exist at all, dolomites being formed only diagenetically. As long as the question remains open, we may assume the Lancara dolomite to be a diagenetic product.

Oolitic Limestones

A second important group consists of oolitic limestones, which in several respects conform to the oolitic arenites described by Graf & Lamar (1950). Following the terminology of Carozzi (1960), we can distinguish here two kinds of oolites: normal and superficial oolites. The latter have an envelope consisting of only one layer. The oolitic limestones are composed of either normal oolites with a few superficial oolites or of superficial oolites only; gradations between the two extremes were not observed. In section, the oolites themselves are circular, but ellipsoidal forms occur as well.

Size. The oolites equal the Bahamian oolites in size, i.e. about 0.5 mm, and show a high degree of sorting (Table VI).

TABLE VI Size distribution of the oolites as measured in thin section, in percents.

	105—150 μ	150—210 μ	210—300 μ	300—420 μ	420—600 μ	600—850 μ	850—1400 μ
Spec. 121	2	4	3	14	59	17	1
Spec. 496	—	5	14	30	44	6	1

The sorting which follows from these values is very good, for no less than 60 % of the oolites of specimen 121 lie in the 420—600 μ fraction, while for specimen 502 the value is 70 %, comprising only two size grades (300—600 μ).

Core. It is difficult to determine what kind of matter served as the core, because the latter has several appearances. Mosaic calcite, built up of grains of ca. 60 μ or ca. 20 μ is seen, as well as a single crystal formed by grain growth. It is quite possible that the mosaic calcite represents faecal pellets. In some instances shell fragments served as the nucleus (Fig. 22), in others it was quartz grains.

Envelope. The envelope around the core is built up of two kinds of calcite which can be distinguished according to their colour, white or gray. Both the white and the gray calcites are microcrystalline, less than 10 μ in size, but the white appears to be somewhat coarser. Due to alternation of the two in a concentric as well as

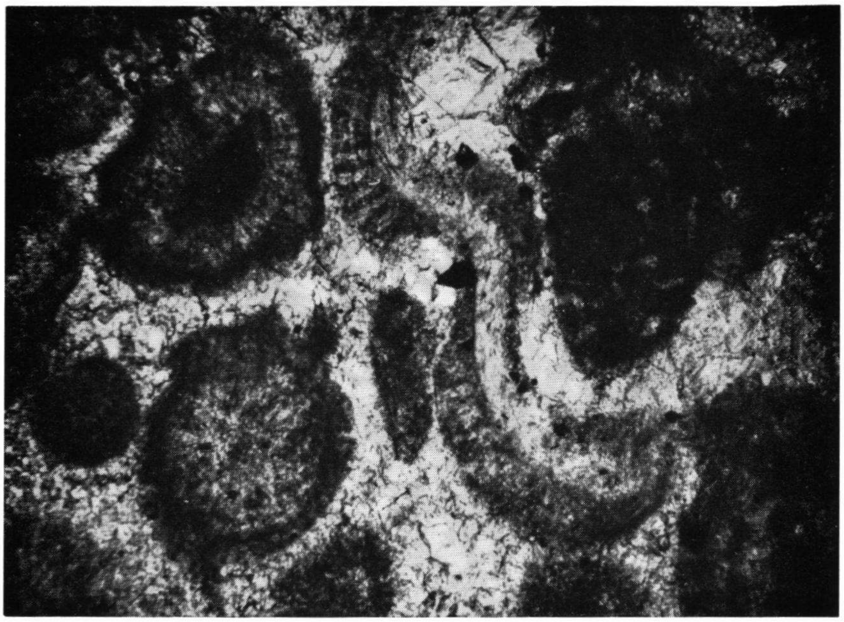


Fig. 22. Shell fragment serving as nucleus for oolite. Oolitic overgrowth partially replaced by cement. 80 ×.

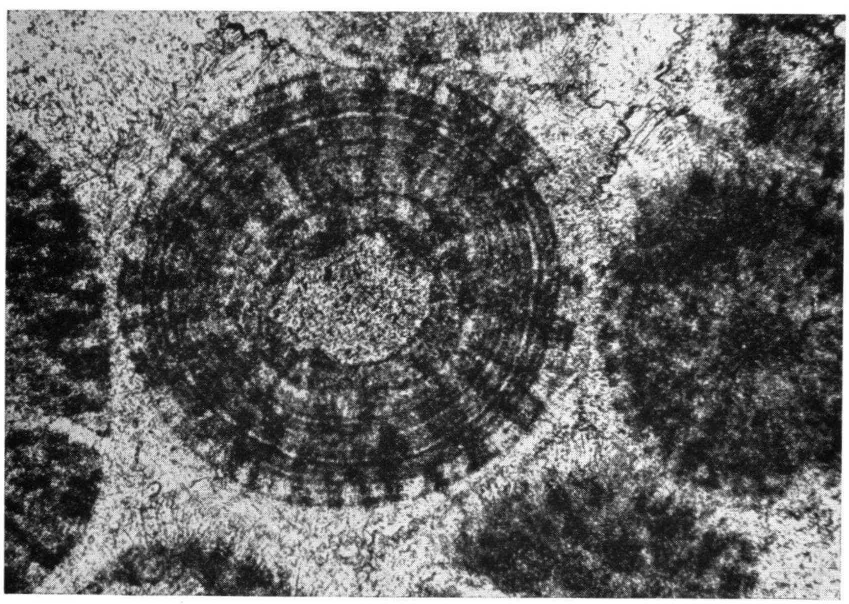


Fig. 23. Radial-concentric pattern of oolite. 200 ×.

radial sense, a radial-concentric pattern is created (Fig. 23). Because of the interrupting character of the concentric rings it is not clear which one is the replacing calcite. That a replacement took place cannot be denied, since Cayeux (1935, p. 225) put forward strong arguments in favour of the supposition of a diagenetic origin of the radial pattern. He observed the extension of the radial sets into the surrounding cement. Because we observe that the material composing both the radii and the rings is exactly the same, it must be assumed that the whole texture dates from the diagenetic stage. Here the dark calcite seems to have been replaced by the white calcite, but there is no strong evidence for this. Our observations clearly support Cayeux's view, but are opposed to the statement of Graf & Lamar, who favour the assumption that the whole structure is depositional but do not discuss the matter. I am inclined to follow Cayeux for two reasons. Firstly, because I myself noticed the continuation of the gray radial sets in the surrounding cement, and secondly, because the white calcite is coarser than the dark calcite, and in general recrystallization gives a coarser mosaic.

Cement. Outside the envelope we can distinguish two phases in the cementing calcite. In a first phase, rims are formed around the oolites; in a second phase, mosaic calcite fills up what is left of the pores. These rims are made of small crystals with crystal planes normal to the oolites edges or they are built by twinned crystals displaying a bending of the twinning planes. Fig. 24 shows two crossing sets of twinning planes, each having a sheaf-like arrangement. In some places the rims wholly fill up the pores between the oolites and there the second-phase calcite is not developed at all. The mosaic pattern of the latter can be converted into a single crystal by grain growth. In some cases the cement has replaced parts of the oolites (Fig. 22). It can be demonstrated that this replacement is the result of solution subsequent to the first cement building phase. The rims first precipitated around the oolites are lacking where the oolites appear to have been attacked, and

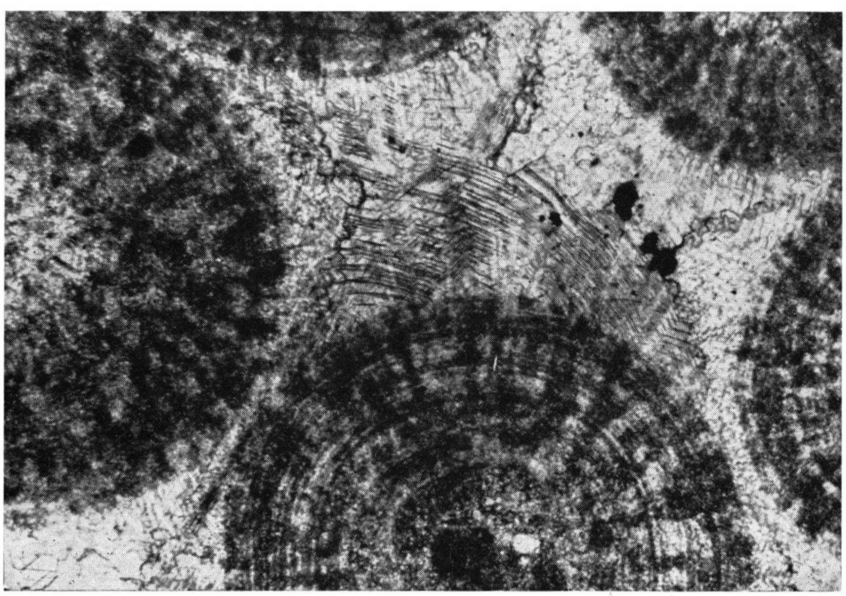


Fig. 24. Bent twinning planes of calcite overgrowths on oolite. 200 ×.

here we see a direct contact between the oolites and the second-phase mosaic calcite of the cement.

A peculiar form of the cement is the gray calcite (Fig. 25). This form shows replacement phenomena at the contacts with the white cement. According to Graf & Lamar, it represents an early cementation phase.

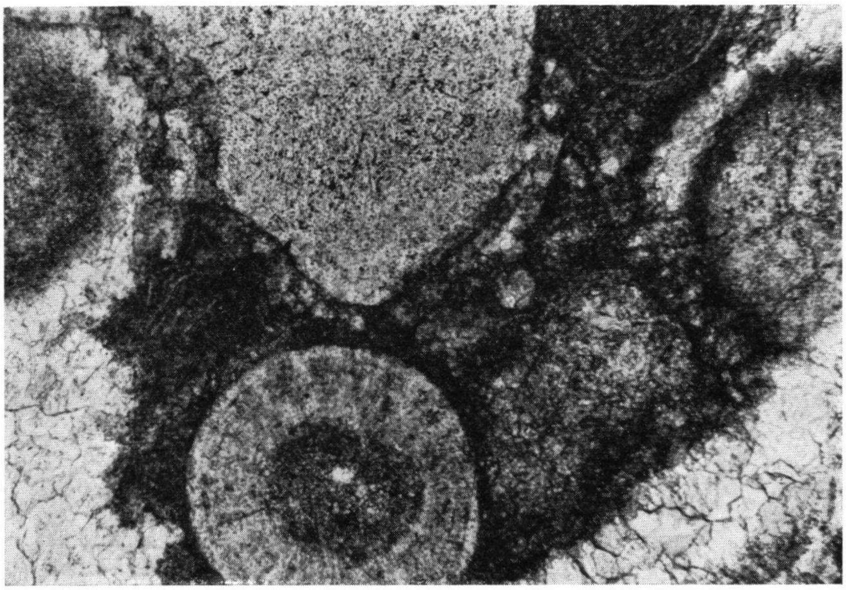


Fig. 25. Oolitic limestones with light and gray cement. 80 ×

Authigenic quartz. Nearly all of the oolitic limestones contain grains of detrital quartz with diameters generally less than 100 μ, but which in a few instances may reach values of as much as 200 μ. Besides the detrital grains, authigenic quartz has precipitated either as secondary enlargements or as well-developed crystals. The latter (Figs. 26, 27), displaying no detrital core, cannot be regarded as resulting from simple overgrowth. Henbest (1945) also observed the formation of quartz crystals in oolitic limestones. The idiomorphic crystals usually contain carbonate inclusions, sometimes aligned in zones according to the crystal faces of the quartz.

There is evidence that the quartz crystals were formed later than the oolites because traces of the concentric rings are sometimes still visible where the quartzes penetrate into the oolites (Fig. 27). It is, however, impossible to prove that they grew prior to the main cementation phase, although this may be assumed in order to simplify comparison with the results of Graf & Lamar.

The presence of authigenic quartz crystals may give information on the chemical state of the sediment, and will be dealt with after a discussion of the mode of formation of the oolitic limestone.

Mode of formation. Summarizing, we get the following series of events:

1. deposition of dark carbonate around the nucleus. *)

* Because of the resemblance of the deposits to those of the Bahamas, the concentric layers probably originally consisted of aragonite here too.

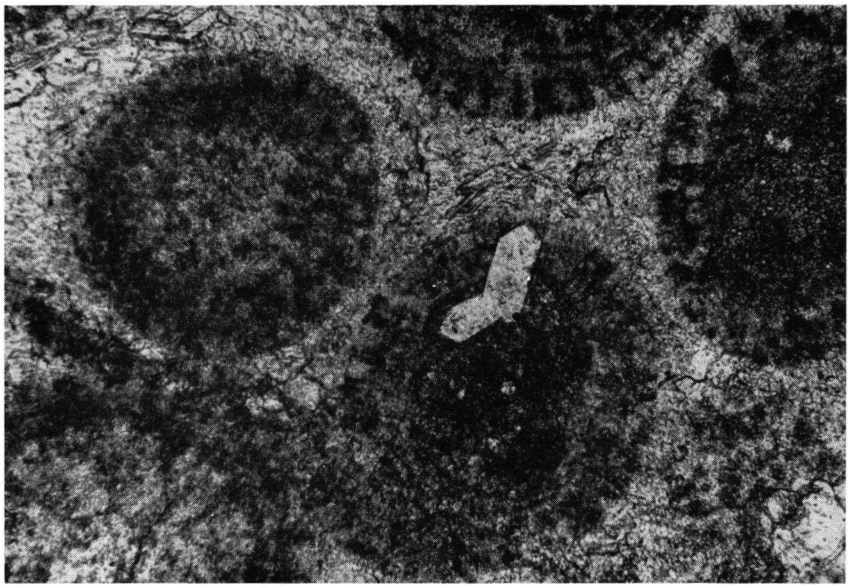


Fig. 26. Authigenic quartz developed in the oolite. 30 ×.

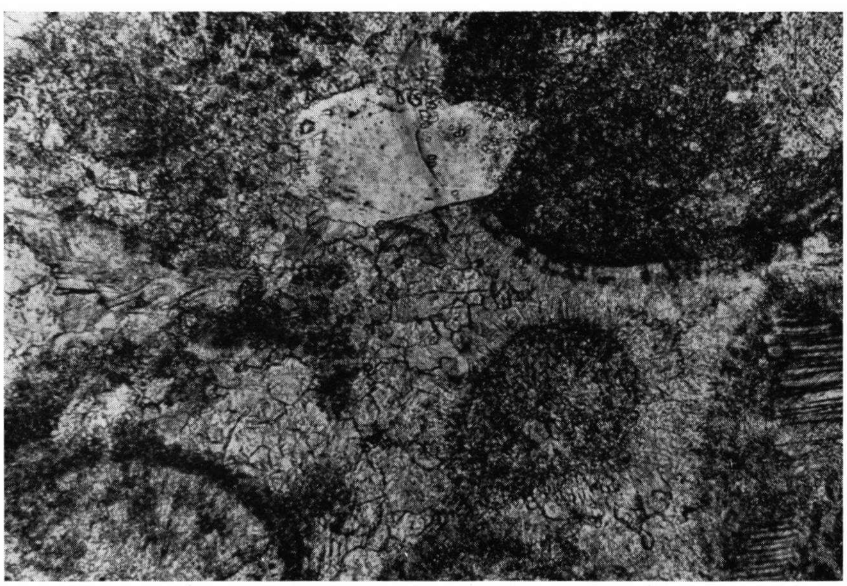


Fig. 27. Authigenic quartz developed in cement and penetrating into oolite. Original boundary of oolite is still visible. 120 ×.

2. local cementation by dark calcite.
3. formation of authigenic quartzes.
4. first phase of the main cementation, building the rims around the oolites.
5. second phase of the main cementation: filling up of the pores and local replacement of older calcite.
6. later diagenetic changes: grain growth; replacement of the gray calcite constituting the oolites by white calcite, thus forming the radial-concentric pattern.
7. veins cutting the limestones undoubtedly represent the last phase of calcite crystallization.

On the whole, there are only slight deviations from the picture given by Graf & Lamar. For instance, I could not observe a solution of the gray cement prior to the main cementation, so it is not listed separately. Secondly, the large twinned crystals in the cement sometimes comprising ghost-structures of oolites with their enveloping cement must be regarded as resulting from grain growth rather than from deformation of the limestones. The grains are similar to those which, according to the description given by Bathurst (1958), were developed by grain growth with syntaxial rims. Interpenetration of oolites is rarely seen, as is splitting-off of outer layers.

Besides the oolites, shell fragments are also found in these oolitic limestones. They are mainly remnants of trilobites and brachiopods, now lying about parallel to the bedding planes. Their presence supports the theory that the oolites formed on or near the shoreline. The observations made of recent oolites by Illing (1954) at the Bahama banks and by Rusnak (1960) along the Texas Gulf Coast, as well as the experiments of Monaghan & Lytle (1956) all strengthen the classical theory of the origin of oolites: accretionary growth in agitated water, supersaturated as to calcium carbonate. These observations, combined with the presence of shell debris and the good sorting, are reasons to assume shallow neritic to littoral conditions during deposition of the Lancara oolitic limestones. The environmental water must have been agitated, as indicated by the arrangement of the shell fragments parallel to the bedding and by the form of the oolites. According to Freeman (1962), quiet-water oolites are highly asymmetrical.

Discussion of the formation of authigenic quartz. The presence of authigenic idiomorphic quartz crystals in limestones has been recorded for several places. For instance, their occurrence is a diagnostic property of the carboniferous Caliza de Montaña in the Cantabrian Mountains of Spain (cf Koopmans, 1962 p. 154). I have also observed them myself in Devonian and Jurassic limestones and dolomites. The cause of the simultaneous occurrence of calcium carbonate and quartz is still under discussion, especially as to the reversible nature of the replacement (Walker, 1962).

Recent investigations into the precipitation of quartz under varying conditions may possibly explain the occurrence of the authigenic quartzes in the oolitic limestones of the Lancara Formation. The crystallization has been assumed to date from the main cementation phase, which probably started early in the diagenesis (early-burial stage of Dapples' scheme, Dapples 1959). This means in the lime-mud. The supposed environment implies that the process must have taken place under normal conditions as to temperature and gaseous content. The most important factor influencing precipitation of quartz appears to be pH (Krauskopf, 1959; Walker, 1962), since small variations in temperature do not essentially affect the solubility of quartz. The influence of the pH, however, becomes important only above the critical value of 9, a figure rarely observed in natural waters. In the range of pH 6 to 9, solubility remains low and almost constant. Because the depositional environment of the Lancara oolites must have greatly resembled that of the Bahama banks (see above), we may use the figures recorded by Cloud (1955 a). The pH, then, appears on the average to reach a value of 8.1 during the daytime in the surface waters; in the lime-mud it ranges from 6.9 to 7.8. Consequently, the variation in pH would have been too small to influence precipitation of the silica.

Greenberg & Price (1957) established that 1 normal NaCl solutions greatly influence the solubility of silica. The more recent experiments of Lovering & Patten (1962) reveal that the mere presence of $Na^{\cdot}$ in neutral solutions supersaturated as to silica, leads to the precipitation of a colloidal silica gel. These results conflict with data given by Krauskopf, who states that the only strange ion of importance is $Al^{\cdots}$. Following Siever (1962), the solutions would not have been supersaturated in silica, but only slightly undersaturated. Therefore, if the silica precipitates, it will be in the form of crystalline quartz (Krauskopf, 1959). Tentatively, I am inclined to assume that $Na^{\cdot}$ also affects slightly undersaturated solutions, resulting in the formation of crystalline quartz. The conclusions reached by Grimm (1962) support this assumption. He states that idiomorhic quartz in sediments indicates a high salinity of the waters from which it precipitated. The environment of the oolitic limestones would then have been rather high in salinity, like that of the Bahama banks.

Breccias

At several localities, breccias consisting of limestone fragments set in a quartz-rich calcareous matrix have been observed. Many of the fragments come from oolitic limestones composed exclusively of superficial oolites. The fragments are angular and have an average size of about 1 cm. The frequent sand grains are sub-rounded to well-rounded, ranging in size up to about 500 μ. According to Pettijohn (1957, p. 405), this concurrence of sands and calcareous clastic fragments is quite normal. Although the total thickness of some successive breccia beds reaches as much as several meters, they must be considered as of only local importance because of the small lateral extension of the beds and their lenticular form. With the exception of the size of the fragments, the breccias have all the characteristics of shoal breccias, according to the description of the latter given by Dunbar & Rodgers (1956), thus fitting very well in the assumed shallow-neritic to littoral environment of the oolitic limestones and dolomites. The sand grains indicate land to have been adjacent, which is in contrast to the Bahama bank deposits.

Glauconite

Of special interest is the mineral glauconite, occurring throughout the Lancara Formation. For some general remarks on this mineral and on glauconitization see page 54.

Description. The glauconite was subjected to x-ray analysis, which confirmed the microscopic determination *). The internal structure appears to be cryptocrystalline

* The picture is the same as that obtained from the glauconites of the Greensands near Swanage (Dorset, Britain).

none of the grains being monocrystalline. This is the more striking because monocrystalline grains are not exceptional in the younger Oville Formation. This contradiction will be dealt with in discussing the latter. The grains, which show dehydration cracks (Edwards, 1945), are lobate, and consequently we must consider them to be autochthonous, although some have a rounded form but still show the cracks. The latter grains I think to be allochthonous. The grains are rather big and visible even to the naked eye, the size being 0.5 to 3.0 mm. Their colour is light green. The glauconite grains are sometimes surrounded by iron rims composed of ferric oxides, as can be determined by means of the ore microscope, and not of ferrous sulfide as recorded by Valeton (1958). (Iron sulfide occurs only in very small amounts). The rims accentuating the outlines facilitate observation of the replacement of glauconite by calcite. However, this kind of replacement took place only occasionally here, whereas elsewhere it is quite common (Pfefferkorn & Urban, 1956; Walker 1960).

Since the conditions governing the formation of glauconites are still under discussion, the presence of this mineral can give little information about the depositional environment except for the physico-chemical properties of the latter at the moment of glauconite formation. The assumption that the glauconites are not synsedimentary but autochthonous, combined with their size and the replacement by carbonates which must have taken place still later, leads to the conclusion that the glauconite is early diagenetic, that is to say it formed while the sediment was still unconsolidated. The chemical circumstances are primarily governed by pH and Eh. The environments of glauconite formation would have been alkaline according to Galliher (1936) and reducing according to Cloud (1955b). The presence of hematite in the Lancara dolomites makes the reducing character rather doubtful.

Source material. The source material of the glauconite remains obscure, but as long as micas of bigger size are not found, it is quite possible that faecal pellets and clay particles were the original material. The latter kind of alteration was also assumed by who studied comparable deposits. Moreover, Houbolt (1957, p. 67) observed Lewis (1962), glauconite originating from black carbonate particles which he held to be faecal pellets. However, this kind of glauconitization takes place only in marly limestones where the necessary silicon and aluminium are available. The presence of corroded quartzes in the Lancara limestones and dolomites would suggest that silica was present in sufficient quantities. The quantity of aluminium seems more doubtful, but apparently the clay present released sufficient amounts to build the glauconite. Another problem arises as to the source of the potassium. Valeton (1958) stresses the fact of the existence of K-poor glauconites. She also notes that Correns (1939) supposed K to be supplied by means of organic matter, which found later confirmation in the observations of Conway (1943, 1945). The latter noticed the alteration of organic matter in seawater resulting in an enrichment of K. Since faecal pellets contain organic matter, and since the Lancara deposits also contain organisms, a source of K is present.

Depositional environment

From the observations reported above it appears that the depositional environment of the Lancara dolomite s.l. is rather well understood. All the various carbonate rocks indicate a shallow neritic to littoral environment. Especially the oolitic limestones resemble the recent deposits of the Bahama banks. Their depositional environment must have been very similar, even though the environment of the Lancara dolomite bordered land, as indicated by the medium to coarse sand grains in the breccias.

LANCARA GRIOTTE

The griotte consists of alternating limestones with a nodular appearance and shale beds with or without calcareous nodules. The nodular appearance of the limestones is produced by small seams of shaly and/or hematitic material. In fact, there is a transition from the nodular limestones to shale beds with nodules and this is the result only of an increase of the shale content. Such a transition is the trend observed in going from the Lancara limestone into the Oville sandstone, making it difficult to decide where to draw the boundary between the two formations. The first sandstone bed has been considered to be the lower limit of the Oville Formation, but this rather arbitrary choice causes differences in the thickness values recorded by the various investigators. In the section W. of Valdoré (Esla Valley), the thickness of the griotte is about 50 metres. The series is characterized by the dominating red colour, although some green beds are intercalated.

Limestone

Mineralogically, the limestone is rather pure because of the strong dominance of the carbonates and the very small amounts of quartz, mica (among which also allogenic glauconite), and hematite. Some chert is also present. Both the quartz and the micas are of silt size, the maximum diameters ranging from 30 to 75 μ, although in one case the size observed was 200 μ. The chert forms irregular patches, about 100 μ in diameter. Because of the irregular outlines and inclusions of calcite, the chert must be authigenic. Calcite is the dominant mineral, but has various appearances (Fig. 28).

Shell fragments. The shell fragments are of great importance, some of them are composed of opal. These fragments are sometimes present in large quantities, giving the limestone an organo-clastic character. It must be said that, besides the fragments, well-conserved specimens of the organisms have also been found. The shells are generally monocrystalline, showing the calcitic cleavage planes, and are discernable because of their purity which gives them a clear white colour. The shell debris is arranged more or less parallel to the bedding planes. The grain size varies from 0.1 to 10 mm, this variation causing the lamination, which is often accompanied by a small change in the colour.

Grain-growth calcite. Another variety of the calcite is grain-growth calcite. This form is seen in fragments having irregular outlines. Like the organic remains, they are monocrystalline, exhibiting the calcite cleavage. Frequently parts of fossils can be traced in the grains, so there can be no doubt that the crystals resulted from grain growth with syntaxial rims (Bathurst, 1958).

Clastic carbonate. Clastic carbonate grains fill up some of the shells as well as some of the pores between the shells. These detrital grains have a size of 20 to 75 μ, in

Fig. 28. Organo-clastic appearance of griotte limestone. Note presence of grain-growth calcite with syntaxial rims. 70 ×.

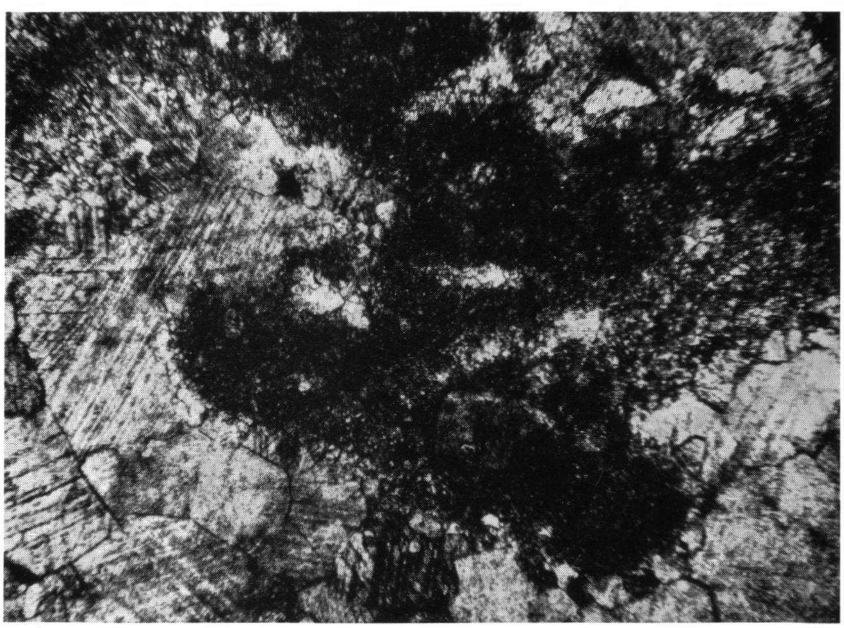


Fig. 29. Patches of fine cement in coarse cement, demonstrating that the latter is due to grain growth. crossed nicols. 80 ×.

general somewhat smaller than the accompanying quartz grains. They are sub-angular to sub-rounded.

Cement. The cement is a fourth type. It is fine-grained or coarse-grained (20 μ and about 300 μ respectively). The latter is produced by grain growth, as can be concluded from the patches of fine cement in the coarse cement (Fig. 29).

Cavity-filling calcite. Another type is the cavity-filling calcite. Fig. 30 shows this calcite filling up a shell. Some investigators might call the calcite directly grown on the shell 'encrusting calcite' (Harbaugh, 1961). The filling up of a cavity, however, will always start at the wall, and consequently the first crystals can easily resemble encrusting calcite, so I think it better to avoid the term in this connection. It is preferable to apply the term 'encrusting calcite' only to surrounding rims made of small fibrous crystals, oriented perpendicular to the wall on which they originated (Harbaugh's second sub-type).

Fig. 30. Cavity-filling calcite. crossed. nicols. 80 ×.

Encrusting calcite. Encrusting calcite also occurs, but only in minor amounts (Fig. 31). The various calcite types of the griotte can be listed as follows:

1. unaltered organic fragments.
2. organic fragments altered by grain growth.
3. clastic calcitic grains.
4. cement $\begin{cases} 20\ \mu. \\ 500\ \mu, \text{ due to grain growth.} \end{cases}$
5. cavity-filling calcite.
6. encrusting calcite.
7. post-lithification calcite (not yet mentioned).

These different types of calcite create a brecciated texture, made the more conspicuous by hematite which occurs not only in laminae but also in small irregular seams. During the process of grain growth, hematite is expelled from the former grains, as indicated by the ferric rims enveloping the grain-growth crystals.

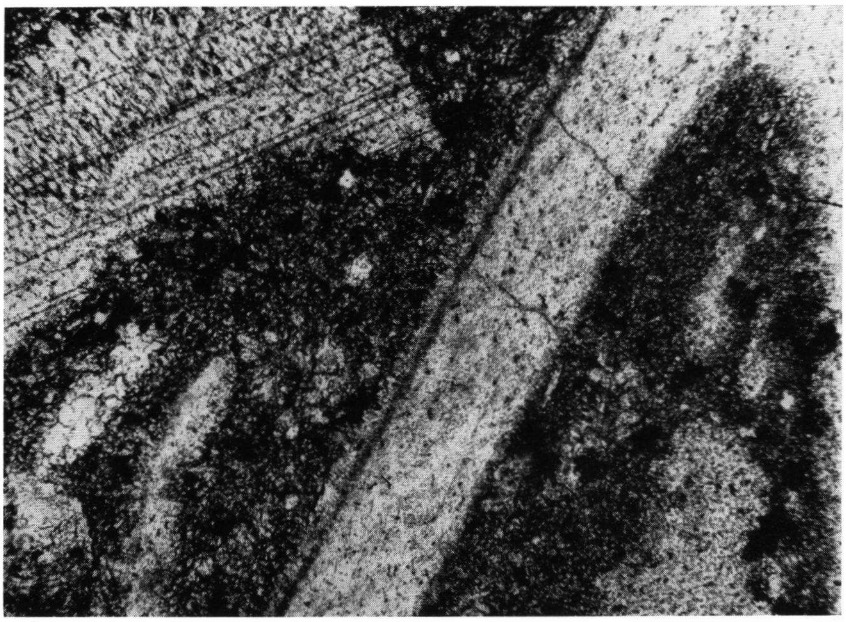


Fig. 31. Calcite encrusted on shell fragment. 200 ×.

Shales

The shales consist mainly of fragments of clay size, but quartz grains of silt size are also distributed through the sediment. Some of the shale beds contain appreciable amounts of calcite, whereas in others lime is almost absent. Hematite, present as a pigment, colours the beds deep red. Some of the beds are green, which cannot, however, be ascribed to the presence of ferrous oxides. Like the Herreria shales, these beds are probably coloured green by illite.

Layering

The stratification of the griotte beds is well developed. The complex is medium- to thin-bedded. The bedding planes are mostly irregular and undulating. Some of the bedding planes show a structure resembling loadcasting, but this can readily be explained by solution along the planes (a kind of stylolites). A minor lamination is sometimes visible due to the presence of shell fragments which are aligned parallel to the bedding planes.

'Nodules'. Some final remarks on the nodules may be made. Their arrangement in the shale beds is of special interest. The longest axes of the nodules lie parallel to the bedding planes. It is remarkable that the lateral distances between the nodules in one bed are about constant, the same being the case as to the vertical distances. Also, the nodules in a given bed are of almost equal sizes. The form of the nodules is ellipsoidal, or irregularly curved. They have, on the other hand, a smooth surface with typically broad but shallow cavities.

Origin of the griotte

The origin of the griotte is still being debated. The most important suppositions are concretionary growth, a detrital origin, and a tectonic origin. Before entering on our discussion, three special characteristics of the griotte may be recapitulated:

a. the gradual transition from the nodular limestones to the nodule-bearing shales.
b. the clastic character of the limestones.
c. the arrangement of the nodules parallel to the bedding planes, the distances between the nodules in one bed being about equal.

These facts as listed enable us to reject several earlier suppositions. The nodules lack a nucleus and do not have a concentric structure, showing, to the contrary, a clearly clastic appearance; hence a concretionary origin is unlikely. Nor are the facts in agreement with a detrital origin. Zwart (1954), Ovtracht & Fournié (1956), Bellière (1957), and Koopmans (1962) have put forward arguments in favour of this origin of Devonian griottes in the Pyrenees, of calcareous nodules in Devonian shales in Belgium, and of the Carboniferous griotte in the Cantabrian Mountains, respectively. Their arguments, however, do not hold for the Lancara griotte. If the limestone nodules were fragments of former beds displaced by the agitation of water, they would show, at least sometimes, one of the following characteristics: higher angularity; absence of such a well-developed arrangement; greater variety of the distances between the nodules in one bed; more frequent occurrence of a high angle between the lamination of the nodules and the stratification of the sequence.

A tectonic origin has also been mentioned. This mode of formation was proposed by Voigt (1962) for nodular limestones in Germany, but it does not explain some of our observations. It is difficult to believe that such a gradual transition could result from tectonic movements. Moreover, one even gets the impression that the nodules have not been rotated at all. As an example there is the observation of Mr. F. H. Cramer (personal communication) who noticed a nodule with a shell fragment protruding into the shale.

The clastic character of the limestones and the indication that the nodules are still lying *in situ* lead to the following conclusion: the nodular limestones and the limestone nodules are remainders of ordinary beds. The only satisfactory explanation of the disrupted nature seems to be the assumption of the removal of part of the limestone by solution processes. The fact that the sequence of the sediments is uninterrupted suggests that the process took place in a submarine environment. Heim (1958) first proposed the action of this kind of solution, and more recently it has been propagated as the mode of formation of some Jurassic nodular limestones in Northern Italy (Hollman, 1962). The latter author gives some pictures of ammonites attacked by solution, and here we see a great resemblance to the surface of the nodules in our shales. In addition, the shape of nodules originated by a solution process according to Hollman, greatly resembles the shape of the Lancara nodules. The cause of the solution, however, remains obscure. Heim suggested the necessity of the action of cold water with a high CO_2 content, but the geochemistry of the carbonates is too complex to enable us to draw any conclusions at present.

Solution could also account for the more or less regular distances between the nodules. Solutional attack takes place most easily along the bedding planes. Vertical solution may possibly be facilitated by fracture cleavage. Voigt (1962), for example, assumes that the fracture cleavage planes develop early during diagenesis. Such planes might be situated at about equal distances from each other.

Depositional environment

The picture of the depositional environment is not so clear here as for the dolomites. The shell fragments and their arrangement point to a shallow neritic, agitated environment. According to Van Houten (1962), the red bed colour indicates an oxidizing environment, which fits well with the supposed agitation of the water. The increasing shale content implies the presence and increasing influence of a borderland. The hematite, giving the red bed colour to the deposits, probably derived from the source area. Normally, it indicates a warm and seasonally humid climate. Some doubt remains, because in the early Cambrian a vegetational cover would not have been present, but Millot, Perriaux & Lucas (1961) assume the same climatic conditions for the Permo-Triassic red beds of the Vosges, also with absence of vegetation. The high percentages of silt and clay favour the supposition of such a climate. This cannot be said with respect to the composition of the clay minerals, but an alteration of the expected kaolinite into the observed illite during the diagenesis would provide a satifactory explanation.

CHAPTER IV

OVILLE SANDSTONE

INTRODUCTION

The Oville Sandstone Formation is not very well exposed because of the high clay content of its sediments. The best exposures can be found along the Esla River, north of the village of Valdoré, and along the footpath from Valdoré to Vozmediano; along the latter the transition from the Lancara Griotte into the Oville Formation is most distinct. Fig. 32 shows the Oville in section, the basal part as it is developed along the footpath, the upper part as it is exhibited north of Valdoré. The total thickness of the formation north of Valdoré measures 180 meters.

The series contains many shales and siltstones between the still dominant sandstone beds. Several sedimentary structures, to be discussed below, are present.

MINERALOGICAL COMPOSITION

The mineralogical composition is given in Table VII. The strong domination of the quartz over the feldspars and rock fragments places the sediment in the group of quartzsandstones, most of them being argillaceous. Another, even diagnostic, property is their carbonate content. With few exceptions, all the beds contain a carbonate cement which, however, varies widely in quantity. In some instances the carbonate is the major constituent, the others forming the matrix, so that the term sandy limestone is more appropriate.

TABLE VII Mineralogical composition of Oville Sandstone in percents by Volume

Number Specimen	Quartz	Sec. Q	Feld-spar	Rock fragm.	Mica	Clay	Carb.	Op.	Glauc.	Acc.
110	67.2	—	—	1.0	3.0	23.4	—	4.8	0.4	0.2
112	36.1	—	—	0.3	3.0	5.8	54.0	0.4	0.2	0.2
114	78.6	—	0.4	2.2	3.4	12.4	—	2.4	—	0.6
612	30.1	0.4	0.8	0.9	10.0	4.2	46.0	7.6	—	—
613	54.7	6.0	1.8	1.5	0.6	3.2	30.2	1.6	—	—
614	38.3	0.4	2.4	1.3	9.4	8.6	36.8	2.8	—	—
615	29.9	—	2.8	0.7	5.4	18.4	18.4	12.4	—	—
481	67.2	2.6	0.8	1.4	1.0	22.0	2.8	2.2	—	—
482	59.4	2.8	3.2	0.8	2.2	15.2	15.6	0.8	—	—
483	57.1	6.4	1.8	1.1	0.2	8.6	23.8	1.0	—	—
489	27.0	—	2.6	0.4	2.4	55.4	1.6	10.6	—	—

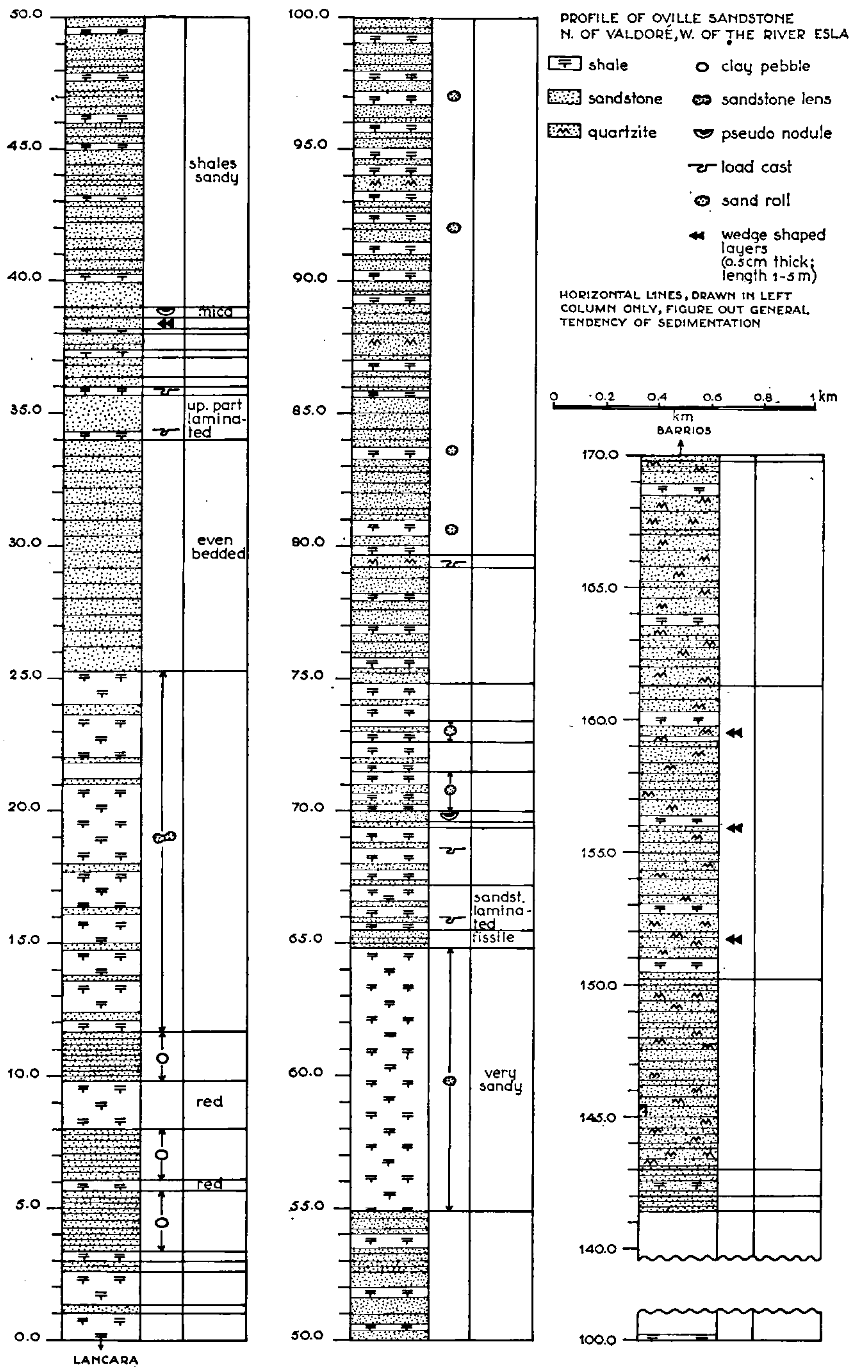

PROFILE OF OVILLE SANDSTONE
N. OF VALDORÉ, W. OF THE RIVER ESLA
shale
sandstone
quartzite
clay pebble
sandstone lens
pseudo nodule
load cast
sand roll
wedge shaped layers (0.5cm thick; length 1-5 m)
HORIZONTAL LINES, DRAWN IN LEFT COLUMN ONLY, FIGURE OUT GENERAL TENDENCY OF SEDIMENTATION
0 0.2 0.4 0.6 0.8 1 km
km
BARRIOS
LANCARA
shales sandy
mica
up. part laminated
even bedded
red
red
sandst. laminated
fissile
very sandy

Fig. 32a.

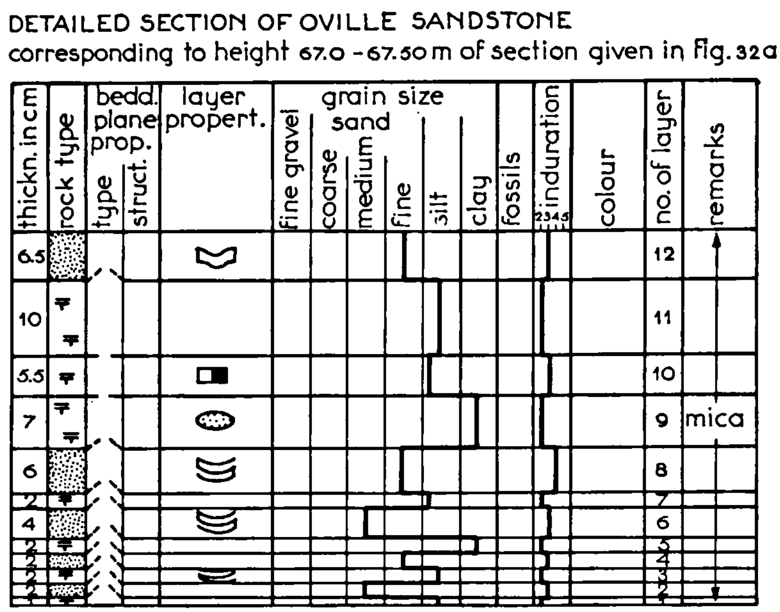


Fig. 32b.

Quartz

The quartz grains appear to be limpid. The irregularity of their outlines reflects a diagenetic solution. Due to this attack the original outlines are no longer visible. Only where an overgrowth can be discerned does the dust-ring reveal the original outline of the grains, which is sub-rounded to sub-angular. It may be noted that the secondary quartz not only lacks calcareous inclusions, but shows features of replacement by calcite. Therefore, this quartz probably originated earlier than the calcite cement. According to Carozzi (1960), it is commonly held that the overgrowths are later than the calcite cement. This, however, is overgeneralized. Gilbert (1949), Siever (1959), and Todd (1963) mention that the calcite as cement postdates the quartz cement. The overgrowths, too, show solution phenomena along their edges. Inclusions are rare; even tiny liquid or gaseous inclusions are found very infrequently. The quartz shows undulating extinction.

Feldspars

The feldspars are exclusively microclines. They too have been strongly attacked by solution, which sometimes progressed so far as to split the grains into two parts. Their limpidity is therefore all the more remarkable: the grains show no signs of kaolinization as do the microclines of the older Herreria Formation. Besides the solution, which often implies a replacement by calcite, I observed an alteration of the microclines into a sericite-like mineral. This appears to take place along the edges and cleavage/twinning planes. From the foregoing it follows that nothing can be said as to the shape of the grains at the moment of their settling. Secondary overgrowths were not observed, except for one grain, faintly showing features suggesting an overgrowth.

Micas

The micas are present in varying amounts. The detrital ones are mainly muscovite, although biotite occurs as well. The latter is pleochroic from green or brown to colourless. The micas show a sheaf-like termination and only exceptionally have rounded edges. The length of the flakes can vary considerably in successive beds (e.g. 350 μ and 75 μ).

Several modes of alteration occur: some of the micas are converted into sericite, others (especially biotite) are subject to glauconitization, while still others are replaced by calcite, which is discussed below

Glauconite. The authigenic mica is glauconite which occurs throughout the whole formation. It often builds cryptocrystalline aggregates, but crystal forms are also present. Determination of the crystal grains was done microscopically, and both kinds of grains were also analysed by means of x-ray diffraction. The aggregate grains are undoubtedly authigenic: they are lobate, displaying desiccation cracks and even desiccation rims; this kind of glauconite fills up the pits in the quartzes. Thus we know that glauconitization took place on an extensive scale.

Several arguments favour the assumption that the crystal forms result from conversion of biotite into glauconite rather than that they too originated from the aggregate (cf Carozzi 1960, p. 53). Firstly, we know that in general biotite is one of the source minerals of glauconite. I myself noticed biotite with glauconitized edges. Secondly, their shape closely resembles the biotite flakes, being elongated, while the aggregate grains have a more rounded appearance. It is noted that Pratt (1962) also concluded that the shape of the glauconite grain depends on the source material.

More convincing, however, is the presence of grains which on the basis of their colour must be considered as being in transition from biotite to glauconite. In such micas we see some blades of only one flake still retaining the characteristics of the biotite, while other blades have the pale green colour typical of the glauconite in these sediments. In this connexion I may also refer to Hodgson (1962), who also observed all transitional stages between muscovite and glauconite in Australian sandstones and concluded that the glauconite originated from the muscovite. It is quite possible that, on the other hand, the aggregate grains represent glauconite which was formed from clay minerals. This would also explain the absence of crystalline forms of glauconite in the Lancara Formation, because there biotite is absent and glauconite is held to originate from faecal pellets.

The glauconites of the Oville Formation have the same colour as those of the Lancara Formation. The presence of light green glauconites is not in agreement with Wermund's statement that the light green colour is characteristic for glauconites occurring in quartzites (Wermund, 1961). In all these sediments the glauconite present has the same colour. This makes a positive correlation between the kind of sediment and the colour of the glauconite hardly credible.

To some extent the physico-chemical properties of the environment are reflected by the colour. Takahashi & Yagi (1929) established a relation between the intensity of the colour and $Fe^{\cdots}$ content, increasing $Fe^{\cdots}$ in the glauconite causing a deepening of the colour. According to Valeton (1958), a low percentage of $Fe^{\cdots}$, as is the case in the sediments mentioned here, is either a result of late origin or of a loss of $Fe^{\cdots}$ during diagenesis. Valeton (p. 111) points out that during diagenesis the ferric content will decrease, so glauconites formed later will contain less $Fe^{\cdots}$, but she also observed glauconites which were dark green in the centre with light green edges, suggesting a loss of iron.

As already stated (see page 44), the glauconite of the Lancara Formation probably originated in an early-diagenetic, i.e. early-burial, stage. The Oville glauconite, however, appears to be late-diagenetic, as concluded from the following train of thought.

In the Oville Formation we observed only a few grains that are allochthonous. Some of the glauconites have desiccation rims. This implies contraction when the matrix was already consolidated. The glauconite sometimes comprises the filling of the pores. These features suggest a diagenetic origin later than the Lancara glauconites.

The iron-poor specimens of the Oville are easily explained by their late moment of origin, as suggested by Valeton, but the same cannot be said for the Lancara glauconites. At the moment of the latter's formation ferric ions must have been present, as can be concluded from the over-all presence of hematite. Therefore, the light green colour must be a result of a loss of $Fe^{\cdots}$. It is possible that the ferric material building rims around the glauconites was derived partly from the expelled iron, but in that case the iron was expelled shortly after the glauconite formation, because still later, replacement by calcite took place.

The glauconites have been partially replaced by calcite, visible due to the dentated edges. Sometimes the grains have sharp points protruding into the surrounding calcite, indicating replacement. The Lancara glauconites show the replacement more clearly, when they have a ferric rim displaying the old grain boundaries.

Rock fragments

As can be read from Table VII, rock fragments are present in only very small amounts. They consist of composite quartz grains.

Binding material

The matrix comprises mainly clay-sized particles. X-ray analyses revealed their composition to be illite with some small amounts of kaolinite. The illite is well crystallized. Dr. P. Hartman has therefore tentatively suggested the illite to be an alteration product of the muscovites, i.e. that the clay originally consisted of muscovite and sericite. Sericite occurs as well, and must be regarded as an alteration product of the micas.

The cement is calcitic. Its content varies considerably, as can be seen from Table VII. It is mostly limpid, but is sometimes turbid due to iron. Siderite is present throughout the whole formation, albeit in minor amounts.

The upper part of the Oville contains some sandstones cemented mainly by quartz. These quartzites build the transition into the younger Barrios Formation.

The opaque matter is mainly limonite, but some hematite is also present in addition to non-determinable matter.

The accessoria in Table VII comprise the heavy minerals and clay pebbles. The most frequent minerals are zircon (sometimes euhedral), blue tourmaline which often is broken, plus staurolites and epidotes.

REPLACEMENT BY CALCITE

All the chief minerals mentioned were attacked during diagenesis and partially replaced by calcite. *Quartz* and *feldspars* have dentated edges suggesting a solution of the mineral; more convincing, however, is the occurrence of calcite inside the dust-rings marking the old grain boundaries. Some grains are split into several parts by the crystal growth of the calcite, as revealed by the shape and the persisting simultaneous extinction of the fragments. The same holds for the *glauconite*.

A difference in the rate of replacement between the authigenic minerals, such as glauconite and secondary quartz, and the others has not been established. Heald (1956) observed that replacement of the secondary quartz is stronger than replacement of the primary grains, due to an attack along the dust-rings as well, but I did not observe such replacement here.

Replacement of muscovite and biotite

The replacement of quartz and feldspar is a common feature and has been described by several authors. The replacement of muscovites and biotites, however, has been recorded only recently by the present author (Oele, 1962). One of the difficulties encountered in observing this replacement is the usually very small size of the micas in sediments and the consequent small scale of the replacement features.

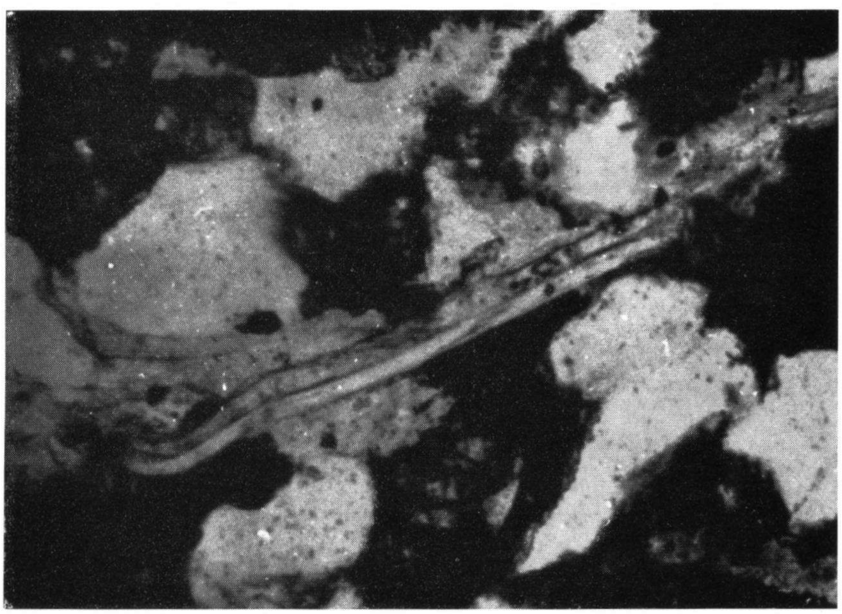


Fig. 33. Calcite bar, pseudomorphic after mica, lying against mica. crossed nicols. 200×.

The first remarkable fact leading to discovery of this type of replacement is the frequent juxtaposition of micas and calcite as it is commonly observed in the Oville sandstone. Often calcitic bars are seen lying against the mica blades, parallel to the basal cleavage of the latter (Fig. 33). In some cases these bars follow precisely the bending of the micas and remain constant in width, so they may in fact be considered as pseudomorphs after mica.

The question arises whether these bars are merely peculiar forms of another frequently observed phenomenon. Mica flakes have been seen several times embedded in the calcitic cement and surrounded by the calcite. The presence of the calcite at this site is probably due to its ability to separate the easily-bent micas from the other minerals by its crystallization force. For instance, Swineford (1947), among others, attributes to the crystal growth of calcite a force strong enough to push quartzes aside.

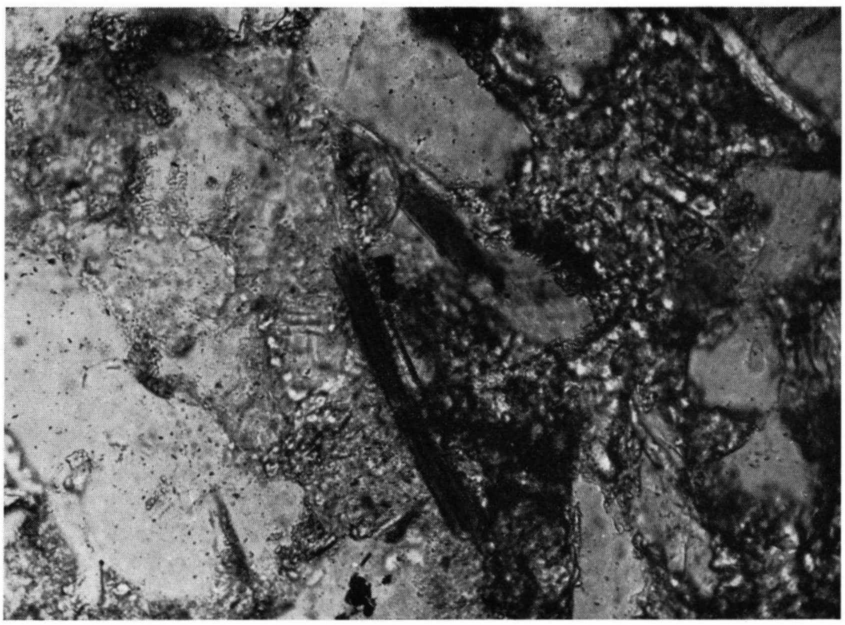


Fig. 34. Biotite, partially replaced. 200 ×.

On the other hand, the calcite can also be the result of replacement, which means replacement of the outer blades of the micas. In some instances there is evidence that the pseudomorphic calcite really originated in this way, as illustrated by Fig. 34. The outer blade of the biotite flake has been only partially replaced, as revealed by the contrasting colour of the biotite.

Another argument in favour of replacement is the presence of dentated micas. Their edges show exactly the same features as the edges of corroded quartzes. Fig. 35 shows such a mica. It may be noted that solution, resulting in dentated edges, can take place not only along but also perpendicular to the basal cleavage. Yet on the whole, replacement appears to take place preferentially along the basal cleavage, as deduced from the dominance of the above-mentioned pseudomorphisms over the micas with dentated edges.

Several times I observed features suggesting replacement which appeared to have been actually produced by other causes. For example, in some instances the micas showed hacksaw edges, but raising and lowering the objective of the microscope revealed the presence of tiny calcitic crystals lying partly on the mica. Another

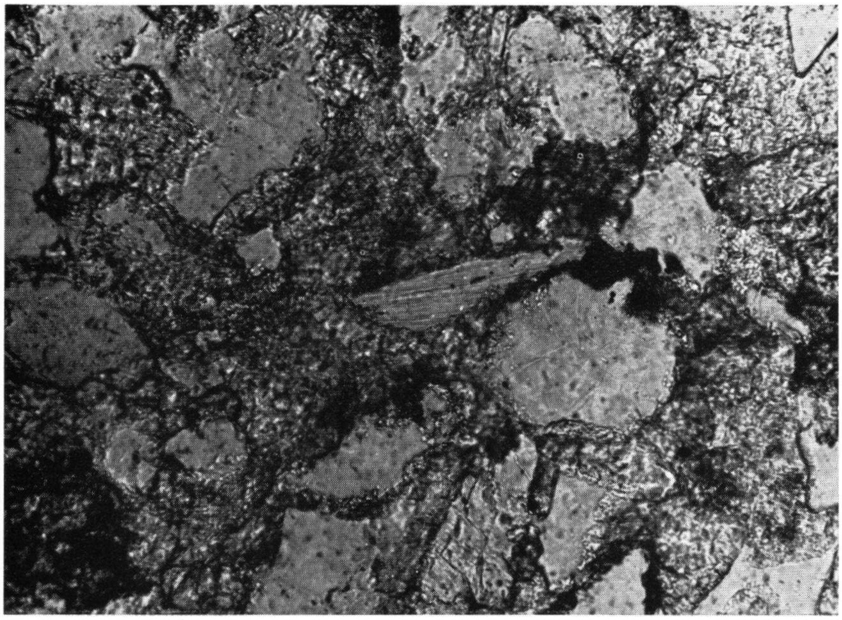


Fig. 35. Dentated mica. 200 ×.

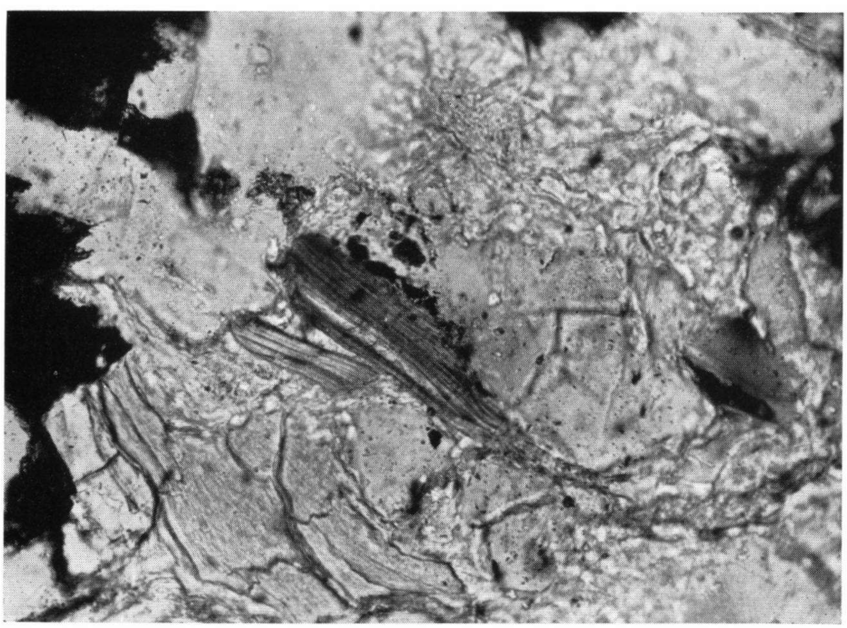


Fig. 36. Mica split open along basal cleavage. Calcite precipitated in the open space. 200 ×.

example is shown in Fig. 36. Here the centre of the biotite seems to be replaced, but the flake has been split open along the basal cleavage and calcite has precipitated in the open space.

Replacement has been observed in very different deposits. In an earlier paper the author recorded it for an Upper-Pliocene caliche of the Spanish Pyrenees. More recently, it has been observed in cretaceous deposits of the Spanish Pyrenees. These sediments are supposed to have been deposited by turbidity currents (Nagtegaal, 1963). Replacement is also known from other Cretaceous sediments deposited in a Wealden facies along the southern border of the Cantabrian Mountains. Consequently, it can be concluded that there is no relation of the replacement to the depositional environment.

It is difficult to fix the time-relations of replacement. Unless we assume that the replacement of the mica took place simultaneously with the replacement of the other minerals, we are unable to fix it at all. For the Oville Formation the time element can be approximated from the observed relations between the various minerals. Since the secondary quartz as well as the glauconites (both being authigenic and thus marking the sequence to some extent) have been replaced, the process must have taken place in a late-diagenetic stage. Unfortumately, it is impossible to obtain evidence as to the time-relations of replacement in the other instances. The resemblance of the caliche deposit in the French Pyrenees with those described by Nicholas (1956) and Alimen & Deicha (1958) leads to the assumption that in this case it is an early-diagenetic process. The authors of both papers report a replacement of quartz by calcite, that, according to them, should have occurred in an early-diagenetic stage. Concerning the replacement in the other deposits, nothing can be said. It seems, therefore, that replacement can take place at any time during diagenesis.

The cause of the solution of the micas is unknown. As long as the conditions governing the more simple replacement reaction of calcite and quartz remain uncertain, we cannot expect to explain the more complex reaction of the mica replacement. Walker (1962) has suggested that the rise in temperature caused by depth of burial was responsible for the reversible replacement of chert and calcite. However, the occurrence of mica replacement in sediments poorly covered by other deposits, like the caliche, excludes such a cause (the caliche has a sedimentary cover measuring no more than 300 meters). In addition, temperature should act throughout the whole sediment, while the replacement probably has a more local character, as suggested by the presence of unattacked micas. This localization corresponds to a statement made earlier by Walker (1960), to the effect that conditions varied within short distances.

THE TEXTURE

Contacts

The detrital grains set in a clayey matrix show several kinds of contacts. The grain contacts occasionally show pressure features, but they are mostly of the common type, i.e. tangential, straight, curved, or zig-zag. A few suturing contacts are found. The contacts between quartzes and micas are mostly straight.

Discussion of the open frame-work. The grains, bound together by calcite, mostly float in the cement. The problem of such an open frame-work is interesting. Cayeux

(1906) and later Waldschmidt (1941) propagated the theory that among the detrital constituents calcitic fragments were once present, solution and subsequent recrystallization of those fragments causing the still more or less uniform appearance of the cement. As additional causes, Carozzi (1960, p. 39) recalls the assumption of a pushing-apart of the other grains by the crystallization force of calcite as also put forward by Waldschmidt and by Swineford (1947) and, secondly, the solution of the quartz.

The supposed presence of calcite fragments is rather doubtful. Unless they were of silt size or clay size, their presence seems rather doubtful because neither the fragments nor ghost structures are present. One would rather expect at least some of the fragments to have been enlarged by grain growth, the more so because this is a common feature in the (Lancara) limestones. Since neither this nor any further evidence is present, the occurrence of sand-size fragments seems very unlikely to me. If the carbonate had a silt size or clay size, the quartz grains would probably have sunk onto each other and not formed an open frame-work.

Grains attacked by solution have actually been observed, as described above. Where the original grain boundaries are still visible, only a relatively small portion of the quartz seems to have been dissolved. This quantity was probably greater in reality, in view of Page & Carozzi's (1962) calculation of the volumetric decrease of quartzes due to solution in dolomitic rocks. Their figures reach values as high as 44.4 %. If these figures are correct, the solution process might be responsible for the texture containig the floating quartzes. Besides these two processes, a third remains. It is generally accepted that the forces created by crystal growth of calcite are rather strong, so they, too, may worth considerating as a possible cause and one whose influence is perhaps often underestimated.

Grain-size distribution

It is difficult to determine the original grain-size distribution. The original outlines of the grains are only occasionally visible, because most of them have been partially solved, while others are secondarily enlarged. Measurements of the grains in the microscope as well as mechanical size-analyses give no more than an idea of the size of the detrital grains. From Fig. 37 it can be seen that the deposits belong to the fine-sand sediments. It is clear that the sorting is very good, although it cannot be expressed by the sorting coefficient of Trask.

Shape of the grains

The shape of the grains is of course also hardly visible. Nevertheless, the original outlines observed are those of sub-rounded grains, but the surface displays pitting. The pits are filled up by the dust-ring material, similarly to those occurring in the Herreria sandstone and their origin is probably the same, i.e. the result of solution. The material building-up the dust-rings is in all likelihood the same as that constituting the dust-rings in the Herreria Formation.

Grain orientation

Orientation of the grains was also observed. Fig. 38 shows some of the results. Although the range of direction of the long axes is rather wide, samples 465, 467,

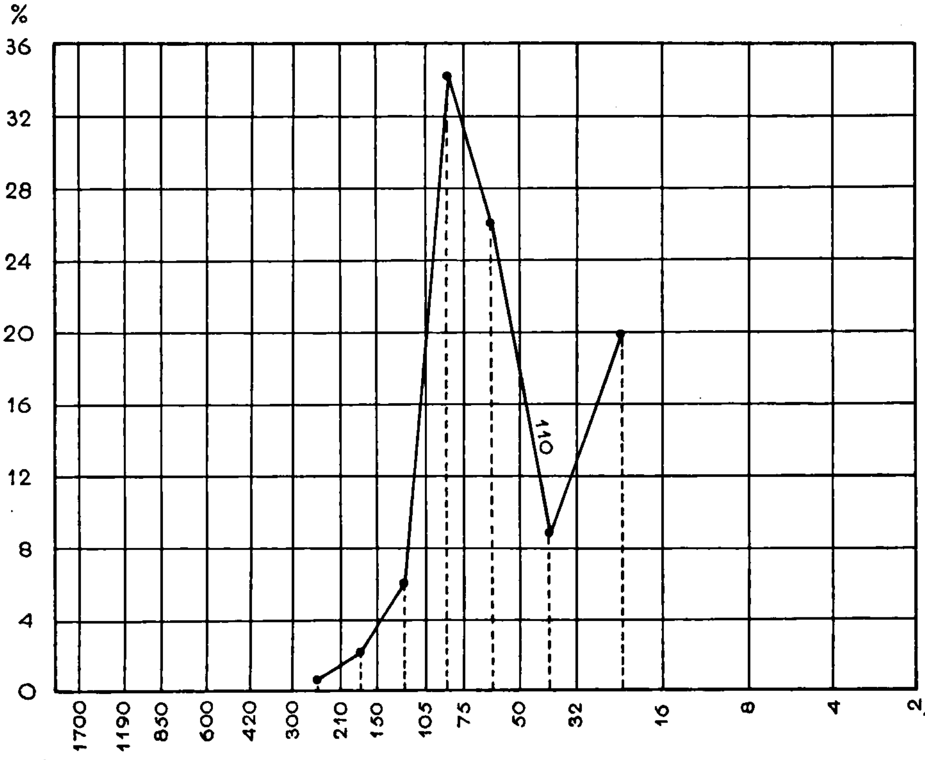


Fig. 37. Grain-size distribution of Oville Sandstone.

and 483 show a dominating N-S direction. This means that the current which deposited the sediment flowed either from N to S or in the opposite direction. Sample 401 displays another direction, which is comprehensible because the sample comes from another locality. While the former samples were taken north of Valdoré, the latter comes from a spot along the footpath from Valdoré to Vozmediano. The number of samples investigated is not sufficient to warrant any conclusions concerning direction, except that orientation is present.

LAYER PROPERTIES

Bedding contacts

The type of contacts between the beds is determined mainly by the rock types involved. If two beds both have a predominating or very high clay content, their contact will be less distinct than if one of them contains only a low percentage of clay. In both the lower and upper parts of the Oville, the contact planes are sharp or distinct, while in the middle part the transitions are mostly gradual or hardly visible. The form of the contact plane is irregular throughout the whole

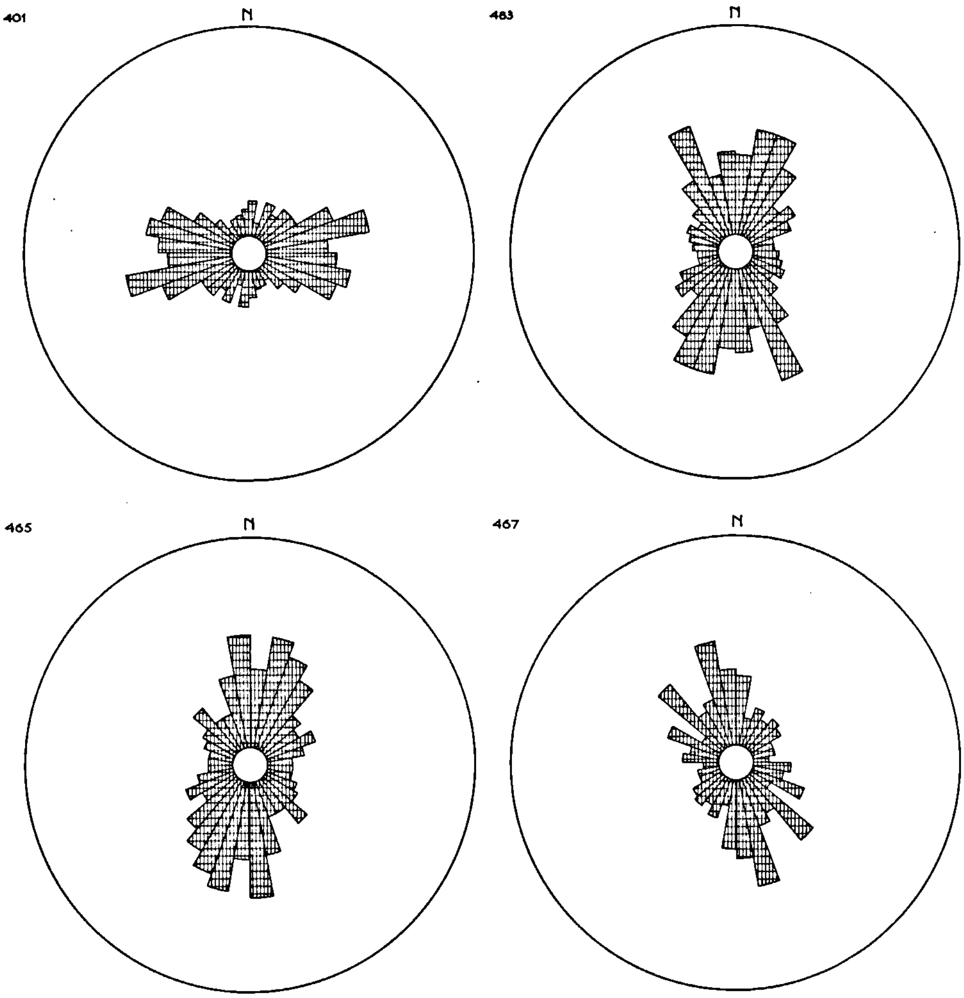


Fig. 38. Orientation of long axes of grains of Oville Sandstone.

series. Flat contact planes occur in the upper part; they are observed especially between micaceous beds. The mica probably causes, or at least favours, those partings.

Thickness

In the middle part of the section the thickness of the beds measures at most half a metre. Here the shale beds occasionally measure several dm, the sandstone beds being not more than a few cm thick. However, these figures are much higher in the lower and upper part, especially for the shale beds. The latter sometimes have thickness values of several metres, but sandstone beds measuring more than a metre are also present.

Stratification

Wedging out of the beds within a distance of several metres was observed in only few instances, but on the whole the complex is parallel-stratified. The internal stratification, or lamination, is often of the parallel type too. It is sometimes wavy and in some instances current ripple laminations were observed (Fig. 39). The small-scale ripples with such lamination are constructed in exactly the same way as the mega-ripple so lucidly described by Niehoff (1958). Here too we see laminae whose topsets cut off older laminae, while their foresets are deposited parallel to those of the older laminae.

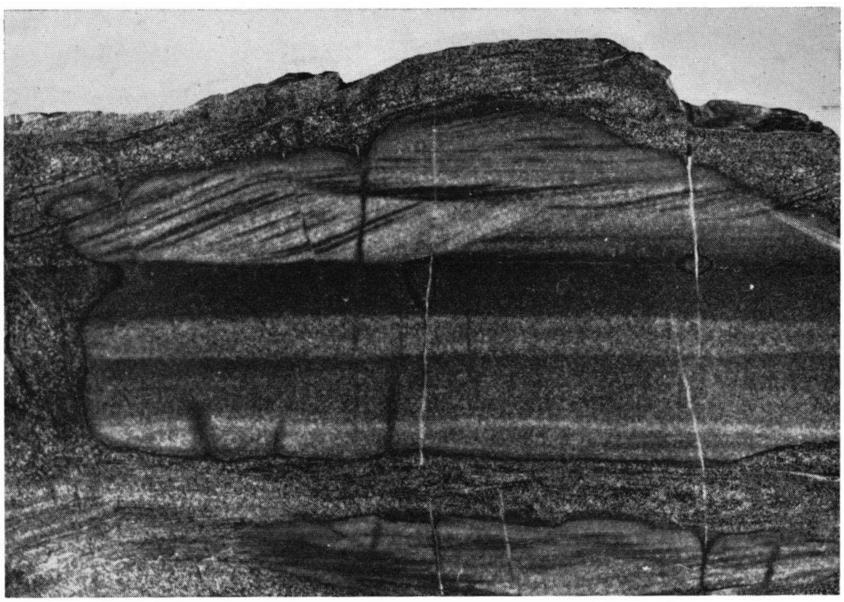


Fig. 39. Parallel lamination and cross lamination in Oville Sandstone. About one-third natural size.

Gully. Only in the exposure north of Valdoré, is such layering visible as can be ascribed to gullying. The depth of the gully measures 80 cm, its width is 200 cm.

Sand rolls. Very locally, typical forms of sandstone are found. Along the Esla River, north of Valdoré, disk-shaped and tubular sandstone fragments are frequently observed in the shales. These fragments lie parallel to the bedding planes with a more or less preferred orientation of their long axes, in this case perpendicular to the strike of the shaly beds. In cross-section, the fragments are elliptical or plano-convex with the convex side downward. The length of the fragments is 10 to 15 cm, the width measuring some cm, occasionally a dm.

Mineralogically, these are calcareous sandstones, their composition being similar to the sandstones described above. An exception is the rather high amount of siderite occurring in the fragments, scattered throughout the rock as small rhombohedra. Granulometrically, the fragments are fine sand.

The relative distances between the rolls are not at all uniform. In a lateral sense they are about 20 cm apart.

Practically all the fragments have internal lamination. The lower part of the example given in Fig. 40 has a wavy lamination corresponding to an oscillation ripple. Although less distinctly, the upper half also contains a ripple-like lamination. A few of the sand rolls contain parallel lamination, but most of them are laminated in the way just described, i.e. they are more or less distinctly ripple-laminated. It is noted that the laminae consist of alternating layers of fine sand to silt and clay. It is concluded from this lamination that the rolls correspond to Reineck's (1960) so-called "Linsen-Struktur", that is, they are eroded ripples. It is even possible that the ripples were of the half-stationary type (Reineck, 1962). This supposition is made because the rolls so often contain an oscillation ripple lamination which is, on the other hand, often asymmetrical.

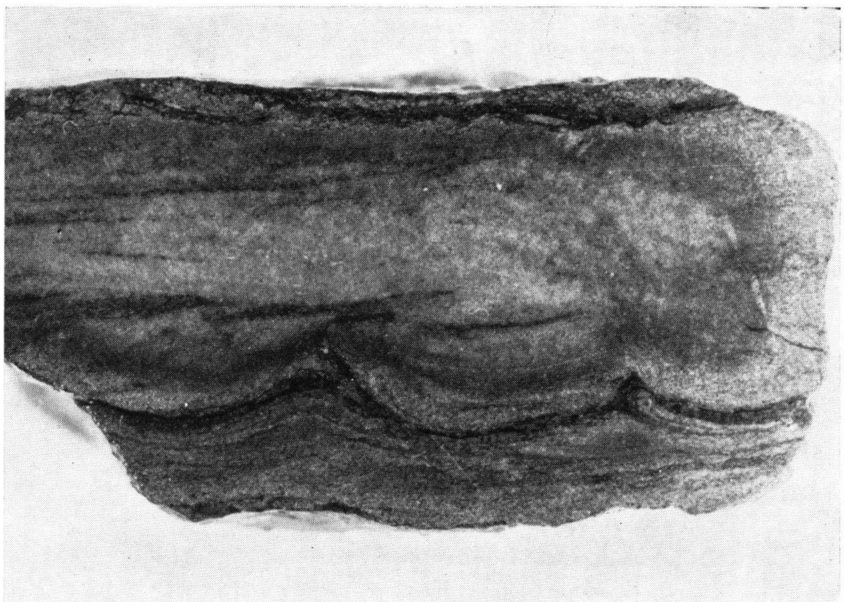


Fig. 40. Sand roll of the Oville Sandstone. The rolls are assumed to be eroded ripples. $1\frac{1}{2}$ × natural size.

Colour

The formation has a rather special colour. Although on a fresh fracture plane the rocks are light grey, most of the sandstones have a light yellowish-brown weathering colour as a result of the presence of limonite. The glauconite so often present gives the rocks a green shade.

BEDDING PLANE STRUCTURES

Tracks and burrows

Traces of worms (?) are frequently observed, usually present on the sole of the beds as tracks. Some burrows have also been found. The diameter of the tracks and burrows is about 2 cm.

Ripple marks

Ripple marks were observed only occasionally. They are small asymmetrically-shaped ripples with straight ridges. Interference patterns also occur.

Cone-in-cone

Among the slump balls of the Oville a structure resembling cone-in-cone as to shape was found. The composition is that of a carbonate-cemented sand containing floating grains. The shape, of course, is conchoidal. In accordance with Pettijohn's description (Pettijohn 1957, p. 209), it shows striations on its surface. Two sets of striations occur.

No internal structure can be detected. It is, therefore, uncertain whether the structure is really a cone-in-cone. Boyd & Ore (1963) described siltstone cones occurring in Permo-Triassic deposits of Wyoming. These are considered to be casts produced by upwelling water, thus resembling sand volcanoes (Gill & Kuenen, 1958) and sandstone-plugged pipes (Allen, 1961). Those cones, however, bear patterns of branching ridges on their surfaces and occur in rather large numbers. Yet the occurrence of such cones would fit well in the supposed depositional environment.

THE OVILLE EXPOSURE N. OF VALDORÉ

Near the bridge over the Esla River, about 1.5 km north of the village of Valdoré, the uppermost part of the Oville is very well exposed. Here the upper 30 metres of the formation exhibit numerous sedimentary structures, the most important of which are the slumped masses. A picture of the section is given in Fig. 41.

Load casts

Throughout the whole section, load casts occur where sandstone overlies shale. They are often asymmetrical, their shapes also suggesting a lateral flowage of the sand during its descending movement. In some instances the vertical movement has gone so far as to separate the load cast from the layer of which it once formed part (Fig. 42). The original layer has generally disappeared. Macar (1948), who did not identify such balls as load casts, used the term pseudo-nodules, a term afterwards commonly applied to this special kind of load casts.

The aligned load casts of Fig. 43 somewhat resemble flute casts. They are, however, too closely spaced and too inconsistent in width to be regarded as erosional furrows, although the occurrence of the latter would be rather normal, even in a depositional environment without turbidity currents (Ksiazkiewicz, 1961; Murphy and Schlanger, 1962).

The load casts and pseudo-nodules are rather small; they measure about 10 cm or less. Generally they lack internal lamination, but it is often possible to observe flow structures in the surrounding clay.

The following observation is of special interest. In several cases in a lateral direction the load casts show a decrease in size, and some metres further they disappear entirely. On the one hand this confirms the assumption that load-

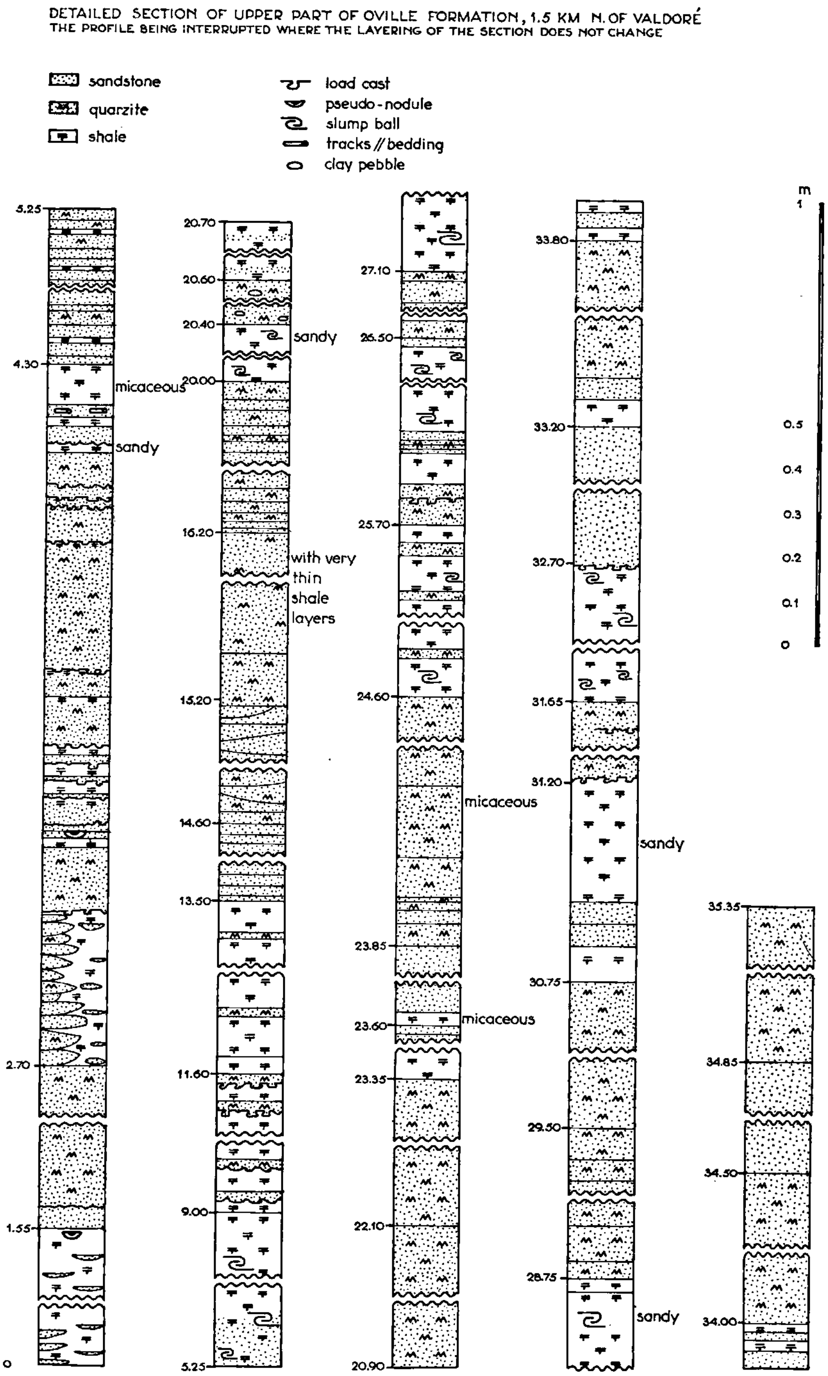

DETAILED SECTION OF UPPER PART OF OVILLE FORMATION, 1.5 KM N. OF VALDORÉ
THE PROFILE BEING INTERRUPTED WHERE THE LAYERING OF THE SECTION DOES NOT CHANGE
sandstone
quarzite
shale
load cast
pseudo-nodule
slump ball
tracks // bedding
clay pebble
micaceous
sandy
with very thin shale layers
m
1
0.5
0.4
0.3
0.2
0.1
0

Fig. 41.

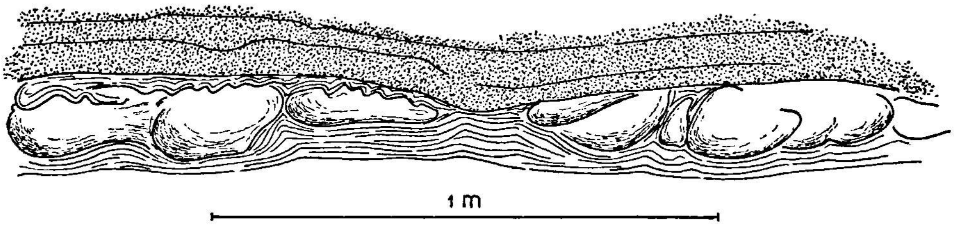


Fig. 42. Pseudo-nodules in the Oville Sandstone.

Fig. 43. Directed load casts of Oville Sandstone. (About half natural size).

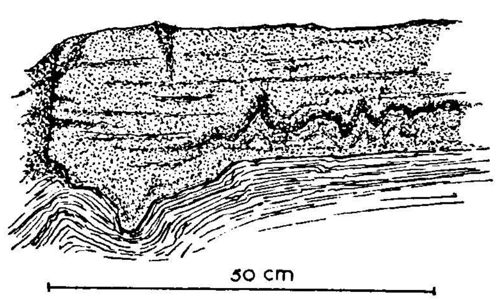


Fig. 44. Load casts of sand in sand, the structure dying out in both lateral directions.

casting is the result of local stresses, but on the other hand the question arises why there is such a regular decrease in size. This can perhaps be explained by assuming that the layers were deposited on a slope, albeit a small one. The layers would then be wedge-shaped and the pressure exerted on their lower surfaces would increase gradually in one direction. Such a regular increase could account for the observations just mentioned. It is quite possible that flowage, parallel to the lower surface, in the sand layers themselves, formed an additional cause.

Loadcasting of sand in sand, as it is so frequently observed in the Belgian Psammites de Condroz, also occurs here. Fig. 44 shows a peculiar example: at first sight the bed seems to be a single depositional unit, but tracing it in the exposure shows that it is divided into two parts. The division can be seen because of the load casting of the upper part into the lower part. The load casts are only locally present because they die out in both lateral directions.

Slump balls

Description. Several beds with slump balls are exposed, but the main bed with slumps lies between 5.26 m to 9.00 m in the section. The bed is composed of a very sandy gray shale (sample 489 in Table VII), in which the slump balls float. Due to their greater induration, the balled-up masses protrude and are easily distinguished from the softer shales.

The slump balls are irregularly distributed throughout the bed, but they show a preferred direction of dip. The length-axes of the balls do not lie parallel to the bedding planes, making a small angle with the latter, but all dip in approximately the same direction, which is roughly N. Moreover, most of the balls have at least one rounded edge. The axes of curvature of these roundings also show orientation perpendicular to the length-axes, thus striking about E-W.

It is remarkable that the slump balls show such variety in their structure. The deformation undergone by the various source rocks evidently led to several quite distinct forms, some of which will be described in detail.

Fig. 45 shows the simplest form of slump balls. This form represents merely a broken part of one sedimentary bed built up by thin layers of alternating sandstone and shale, both being micaceous. The composition is in no way different from that of the other sediments of the Oville Formation. The sandstone layers show parallel lamination. Most of the sandy layers formed load casts in the underlying shale. These load casts do not show features evidencing later disturbance.

Remarkable is the parallelogram-like shape of the ball in longitudinal section caused by an inclination of the vertical planes. This inclination of both planes must have resulted initially from an oblique rupture of the slump ball from the original sedimentary bed. The various internal layers wedge out against the curved plane, which supports the supposition as to the origin of the plane.

Most of the planes bounding the slump balls are very sharp, marking off the blocks very distinctly from the surrounding shales. An exception is formed by one of the two vertical planes (at the right in Fig. 45). Here the material of the slump ball merges into the shale more or less gradually. However, its opposite side is extremely sharp, besides being curved and smooth. This phenomenon may be explained in two different ways. On the one hand, the pressure exerted on the frontal plane of the slump ball during its slide resulted in bending and smoothing of this plane, while the suction at the rear pulled the sand of the slump ball somewhat apart at this side. On the other hand, it may be that the same forces acted in just the opposite way: pressure on the frontal plane forced the ball open, while, suction at the rear evoked flowage of the matrix, which polished the rupture plane.

If the latter explanation is correct, the planes of rupture would be concave in the direction in which the mass is going to move, as observed in landslides, etc. This might be an argument in favour of this view.

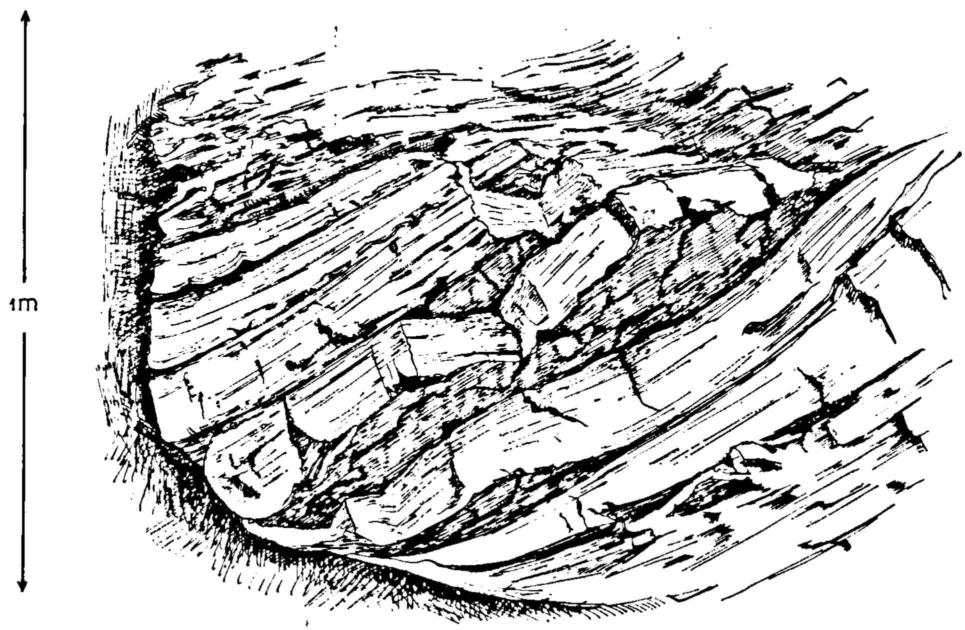


Fig. 45. Slump ball representing undisturbed broken part of sedimentary bed.

Another example of deformation due to the same kind of pressure may be given: Fig. 46 displays a slump ball whose strata form a recumbent fold. Here attention may be drawn to the experiments described by McKee and others (1962 a, b), applied to unconsolidated cross-bedded sands. One of their conclusions is that intraformational folds developed either from drag exerted by an over-riding force or from pushing by a lateral force. According to McKee and others, the folds built in the latter way have crinkled upper limbs. The photographs presented by these authors show clearly that the folds also display the typical feature of lack of space in their core (cf de Sitter, 1956, p. 199), of which the crinkling is but another expression. Although, on the other hand, the folds resulting from an over-riding force are also very sharp, they do not show these features, the sharply-bent laminae having a predominantly chevron-like character.

The folding of the laminae in the slump ball resembles the intraformational recumbent folds developed by pushing. The wedging-out of the laminae against the bottom of the slump ball favours the assumption that, here too, foreset beds were deformed. Earlier, Jones (1962) too explained intraformational isoclinal folds, assuming folding of a cross-stratified series of layers due to sliding with subsequent crumpling and overfolding.

The next figure (Fig. 47) shows another kind of balled-up masses. The several

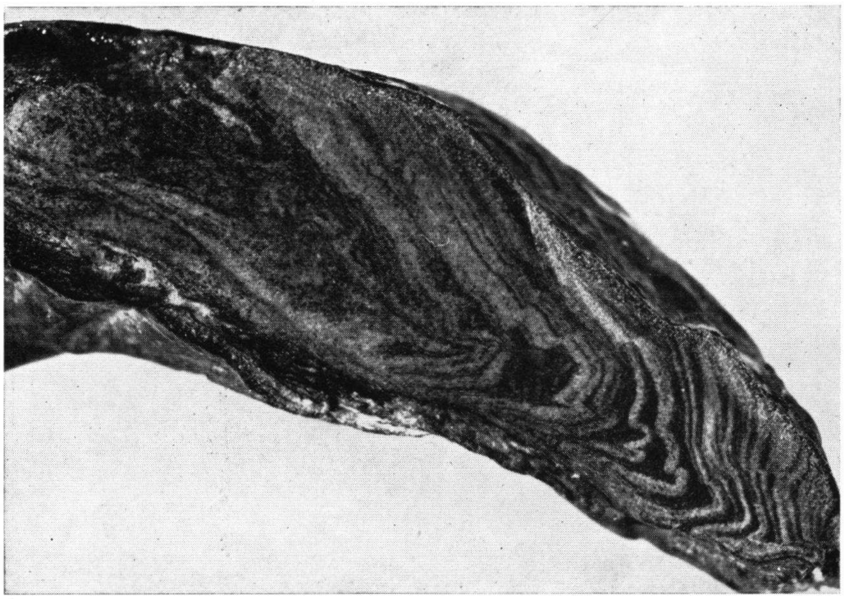


Fig. 46. Slump ball with internal laminae forming recumbent fold caused by the gliding; the folding is assumed to be a deformed cross-bedding. (Natural size).

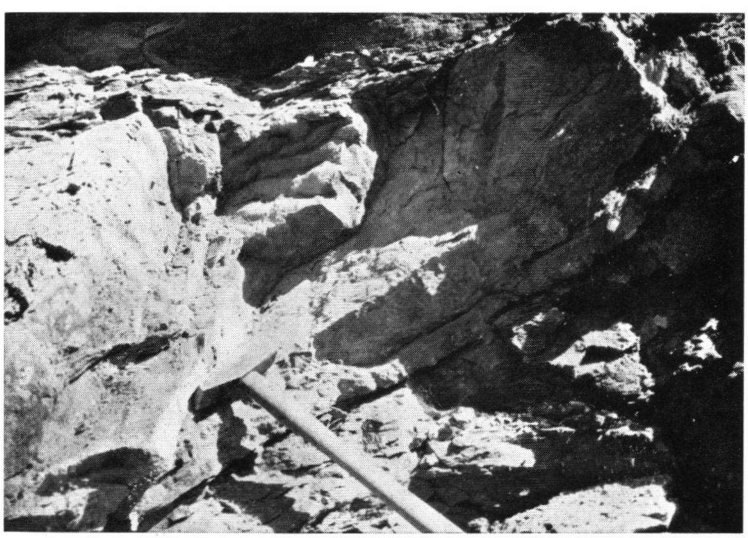


Fig. 47. Saucer-shaped slump ball. Load casts in the centre show upside-down position of the ball.

layers give the ball the shape of a saucer. The material is predominantly sandstone; only a few shale layers are intercalated. The sand layers have loadcasted like those in the ball mentioned above. The load casts observable in the centre of the picture indicate the upside-down position of the slump ball.

The fourth example of slump balls is also interesting for its peculiar shape. The layer in the centre of Fig. 48 wedges out between the two other layers. Since the middle part faintly shows lamination, but one without flow structures, the layer was not pressed out but had initially wedged out between the outer layers.

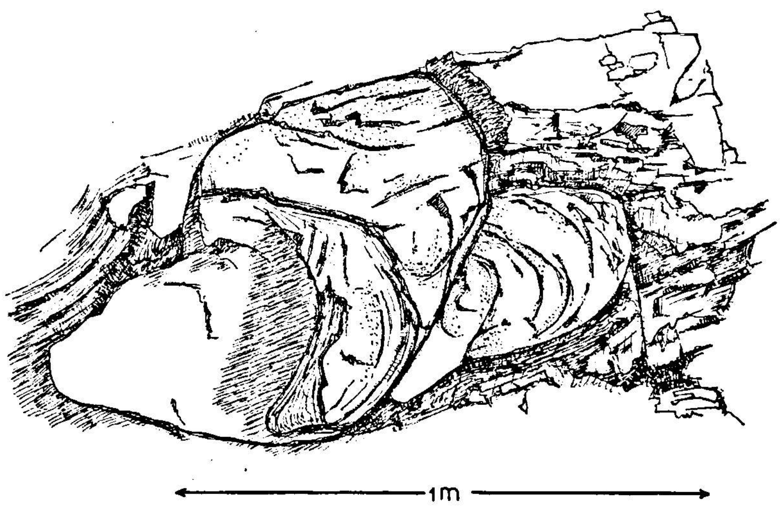


Fig. 48. Slump ball showing deformed cross-bedding.

The last example again shows a slump ball with a contorted bedding (Fig. 49). The upper part of the ball is parallel-laminated, but the laminae of the lower part are intricately folded. In addition, several thrust planes can be observed. The fold axes are horizontal, but others are almost vertical. Since the upper part is

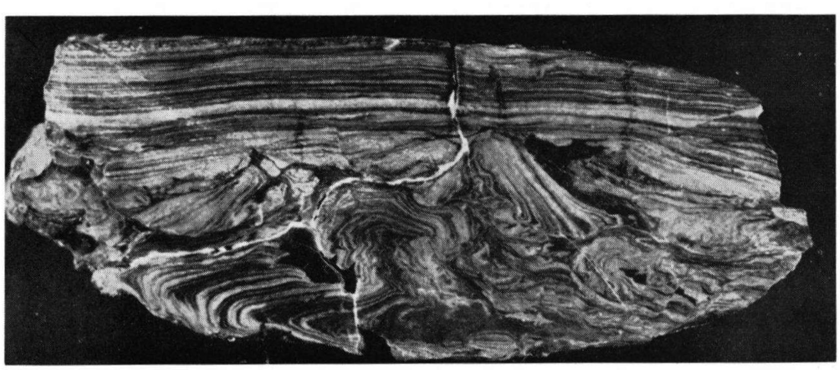


Fig. 49. Slump ball showing flow structures antedating the slumping. The flowage is assumed to be small-scale slumping which is closely related to convolute lamination. (About one-third natural size).

undisturbed, the deformation must have taken place in the original sedimentary layer. This kind of deformation is also due to flowage. It may be termed internal slumping, and will be dealt with below, in the discussion of the state of the sediment at the commencement of the slumping.

Theoretical remarks. In the foregoing, the term slump ball has been used, but the mode of emplacement may be discussed at some length. Since the following must bear relation to the mode of deformation of unconsolidated subaqueous sediments subjected to a shearing force, some of the physical properties of such sediments may be recapitulated.

The way in which fine-grained sediments react when a shearing force is applied, depends primarily on the composition and moisture content of the sediments. Because of the importance of the latter, the limits between the various fields of behaviour are expressed in terms of the moisture content. The moisture content is defined as the quantity of water (in grams), before drying, present on 100 grams of dry solid. Now, depending on the value of the moisture content, a shearing force — assumed it is high enough, of course — produces one of the three principal modes of deformation: rupture, plastic flow, or fluid flow. Plastic flow can be said to occur between two critical values of the moisture content: a. the lower plastic limit and b. upper plastic limit or liquid limit. Below the lower plastic limit the material is so dry, and therefore so brittle, as to permit only rupture (under normal non-confining pressure), while above the upper plastic limit fluid flow takes place *). In the field of plastic flow — and that of false-body thixotropy (see below) — internal folding and crumpling of the sediment is possible.

Suspensions reacting to a shear force by fluid flow can be in very different rheological states. Following the scheme given by Pryce-Jones (1952), these states in order of increasing dispersion are: (plasticity) — false-body thixotropy — thixotropy — newtonian liquid **). The two thixotropic states imply that the suspensions are in the state of a gel, which under the influence of shear (also including tapping, etc.) transforms into either a more fluid gel (false-body) or a sol (true thixotropy) but returns to its original state after the shear ceases. A characteristic difference between the two kinds of thixotropic behaviour is that the breakdown of the gel in the false-body state is far more rapid than the conversion of gel to sol in the thixotropic state.

Transitory stages between the two thixotropic states exist. In the newtonian liquid state, the dispersion has all the characteristics of a liquid whose viscosity is independent of rate of shear or time of shear. It may be remarked that according to Pryce-Jones, a sol too is a newtonian liquid in rheological respect and consequently the sol state of a thixotropic system cannot be made less viscous by applying a higher shear force.

Two more remarks must be made. Firstly, it is seen that the successive rheological states reflect the state of the sediments as the moisture content increases. Boswell (1952, 1961), like Ackerman (1948), assumes that as a result of de-watering, sediments pass from the liquid state at the moment of settling via the thixotropic state to the plastic and even solid state during the subsequent diagenesis.

* Actually, fluid flow starts above the lower thixotropic limit, which has a distinctly higher moisture content than the liquid limit. The zone between the two limits is a transitional zone. Boswell (1952) discusses this difference in some detail.

** Between thixotropy and newtonian liquid is the intermediate state of dilatancy. In this state the dispersion is a newtonian liquid at low rates of shear, but its viscosity increases at higher rates of shear. Since this behaviour has no importance in relation to slumping, the term has been omitted from the scheme.

Boswell states in both papers that geological materials show a false-body, rather than a true, thixotropic behaviour. From this it may be concluded that sediments, after settling, reach the false-body thixotropic state very soon and that the true thixotropic state probably does not last long enough to be of importance in the sedimentary history. If, therefore, in the following a thixotropic state is supposed, it is meant to be false-body thixotropy.

The second remark bears on the reverse change of the states. Besides an increase in moisture content, other causes such as electrolytes may produce the transition from false-body thixotropy to newtonian liquids. It must be realized, however, that such a transition cannot be produced by merely increasing the shearing force as proposed by Dott (1963), because each rheological state is defined by a particular relation between its viscosity on the one hand and rate of shear and time of shear on the other hand (Pryce-Jones, 1952). If an increasing shear produced dispersion in the successive rheological states, this relationship would show a marked trend, which actually is not the case.

The error made by Dott is a fundamental mistake and probably results from the assumption of an incorrect analogy. Consolidated rocks subjected to an increasing shear show a strain that progresses from elastic through elastico-viscous and plastic behaviour to rupture. (Hence, the lower plastic limit is commonly called the elasticity limit).

Now, in the first place, unconsolidated, uncemented, rocks behave in quite a different way, and, in the second place, even when the moisture increase is gradual, rocks with different moisture contents have different physical properties and consequently constitute different rock types.

It is also preferable to avoid introducing the terms elastic and elastico-viscous behaviour in relation to subaqueous movements or, which is the same, to the rheological state. This was done by Dott, who termed simple rockfall or rupture elastic deformation, and gliding along discrete shear planes of rigid and semi-consolidated masses an elastico-viscous behaviour. Since, by definition, distinct rupture planes must develop, it is hardly justified to call these deformations elastic: deformation does not cease after a certain time interval nor does recovery occur.

The origin of the Oville slumps. In many respects the deposits resemble those described by Kuenen (1948) for Pembrokeshire. For instance, here too the slump balls are embedded in a sandy shale, present in beds with distinct flat soles. The shale itself is irregularly cleaved, displaying no signs of internal layering. Kuenen called these beds slump sheets, after their mode of origin. Several arguments favour the assumption that the Oville slump balls were also transported in a flowing mud-stream as assumed by that author.

Firstly, there is the parallelism between the long axes as well as the axes of curvature of the slump balls and the bedding planes. Secondly, the axes of curvature show a strike predominantly E-W. On the other hand, the position of the slump balls, scattered as they are throughout the shale beds, implies that the mud-stream did not have the properties of an ordinary liquid. This is perhaps best explained by the assumption that the clay was in the state of false-body thixotropy.

Another train of thought leads to the same conclusion. It is shown, for instance in the first example given above, that an internal deformation of the balls did not always take place during the gliding. Internal deformation would be expected if the material involved was in a plastic state. So the conclusion to be drawn is that material, lacking such deformation, was at least semi-consolidated, while other slump balls, to the contrary, show features of a plastic deformation as a result of the gliding movement. The difference in degree of consolidation indicates that the sliding involved more beds than just the most superficial ones. The conclusion is therefore that clay beds firm enough to support several sedimentary beds suddenly started flowing, disrupting the overlying beds. Depending on their degree of consolidation, the latter were deformed. The clay, however, must then have been in a state of false-body thixotropy.

The last example given above concerns a slump ball exhibiting contortions (internal slumping) which antedate the formation of the ball. The mode of origin of this contorted stratification is also of interest. The moment of deformation may be discussed first. Since the upper part of the ball shows undisturbed parallel lamination, only the lower part being crumpled, deformation must have taken place before the whole bed was deposited. Consequently, the contortions are a metadepositional structure. Two stages in the sedimentation are then to be distinguished: stage 1 represents deposition of a parallel-laminated series with subsequent deformation; in stage 2 renewed deposition takes place. It is noted that during the second stage the lower part was still mobile, for the upper part loadcasted into the deformed lower part.

With regard to the rheological state of sediment subjected to deformation, it is of importance to know that the contortion occurred in superficial material. During deposition, sedimentary beds are not yet in a plastic state. Yet the deformation suggests such a state. Again, however, a state of false-body thixotropy can explain the contradictory behaviour. The same conclusion has been reached by Nagtegaal (1963) after an extensive study of convolute lamination.

In this connection, some remarks on the relation between internal slumping and convolute lamination are appropriate. All kinds of intermediate stages occur between convolute lamination, as defined by Kuenen (1953), and slumping, as just described. Therefore, I agree with Nagtegaal that the difference is one of rheological state. The slumping is characterized by a more plastic-like flow: it is a false-body thixotropy with a less fluid behaviour under shear than that of the convolute-laminated sediment. Slumps and convolute lamination are to be regarded as the opposite end-members of the state of false-body thixotropy. It is only logical that convolute lamination is more commonly found in connexion with turbidity deposits. Hardly any other kind of sediment would have been deposited at such high rates as the latter. Consequently, their moisture content will be much higher than more regular deposits, and turbidity sediments will therefore approximate the state of true thixotropy more than the other sediments.

According to Nagtegaal, the different ways in which the two movements start constitute another difference. Like many others, he suggests that slumping needs a trigger *), which would not be a requisite in the case of convolute lamination. The turbidity sediments he investigated were deposited on an inclined surface, and therefore underwent a shearing force due to gravity. As the thickness increased the force grew too, until the strength of the sediment, being in state of false-body thixotropy, was surpassed, and the material started flowing. However, experiments have demonstrated that a rather high angle of repose must be exceeded before even a metastable sediment starts flowing. (Boswell, 1961; Moore, 1962; Dott, 1963). As early as 1935, Rettger mentioned a slope of 20°. Although the latter figure is perhaps very high, it is clear a high angle is required. The question is then whether a turbidity current would halt at all and start settling on such a slope.

* Earthquakes are often put forward as having been the trigger, but exceptional tidal waves may have served as well.

On the whole, the assumption of such an angle seems doubtful. Perhaps the turbidity current itself acted as a trigger. When the velocity of the current was low enough to allow sedimentation below a certain wave base, the overlying waters were still agitated. As a result, impulses acted on the water-loaded sediment, causing convolute lamination after the fine-grained, susceptible part had settled.

The parallel-laminated sequence might also be explained by such agitation. Kuenen (1953) explained this lamination on top of convolute lamination by assuming a further decrease of the turbidity current velocity so as to come below the critical velocity of ripple formation. Rettger (1935), however, in his experiments on slump structures was also able to produce a parallel lamination on top of the slumped laminae by stirring the water.

The parallel lamination on top of both slumps and convolute lamination can be explained as follows: the first impulses of the agitational movement, occurring in jerks, cause flowage of the sediment (internal slump or convolute lamination, depending on the moisture content), but, as the intensity of the impulses decreases gradually in the course of time, later, less vigorous, impulses of the water resulted in parallel lamination.

SOURCE AREA AND DEPOSITIONAL ENVIRONMENT

Source area

The mineralogical composition gives little information about the depositional environment or the source area. The strong predomination of quartz among the detrital components designates the rocks as mature. According to Pettijohn (1957, p. 511), maturity "most likely is the product of a warm and humid climate in an area of low relief", but maturity may be reached in another way as well and therefore such a generalization seems doubtful. Pettijohn assumes such source areas because intensive weathering is required to make quartz almost the only surviving mineral. A polycyclic history, however, will also lead to the same mineralogical state. The roundness of the tourmalines indeed favours the latter view. Furthermore, there are no indications of a warm, humid climate.

Hematite is encountered only in slight amounts, as is kaolinite. The sedimentary structures present point to a rapid accumulation and consequently a rapid supply. It is quite possible that the relief was moderate or high instead of low. The conclusion from the foregoing is that Pettijohn's statement does not hold in general, and that the data are insufficient to give us an idea of the kind of source area involved. The question is discussed below (see pag 89).

Depositional environent

The depositional environment deduced from the fauna is marine. The presence of ripple marks, the frequency of load casts, and the observed slumps all agree with a marine environment. Fortunately, this broad environment can be delimited somewhat as follows.

Generally, lamination is held to form at low velocities (e.g. Ksiazkiewicz, 1961). Hamblin (1961), too, assumes that micro-cross-laminations indicate low levels of energy, and he supposes that such structures originate on tidal flats, deltas, etc. The afore-mentioned structures fit well in the supposition of near-shore deposits.

Rapid accumulations such as are required to form load casts and slumps certainly took place near the shoreline, where the rates of supply are the highest. Yet the Oville deposits cannot be correlated with tidal flat deposits as described by van Straaten (1954) and more recently briefly reviewed by McKee (1957). For example, all the sediments of the Oville lack gullies except for one observed in the exposure north of Valdoré. The picture shows more resemblance to a deltaic deposit. All the structures mentioned can be expected in a deltaic environment. Slumping too may start on a delta (Moore, 1961). Therefore, the assumed environment is a rapidly growing delta. Since the Oville deposits have a great lateral extension, it seems very likely that a series of deltas is involved.

CHAPTER V

BARRIOS QUARTZITE

INTRODUCTION

The Barrios quartzites constitute a formation that stands out distinctly in the field because of its high degree of lithification. North of the village of Corniero, the Barrios of the Esla nappe is especially well exposed; a section is shown in Fig. 50.

Quartzites are the dominating rocks in this formation, but more argillaceous rocks are intercalated and can be frequently observed. Near the village of Boñar a conglomeratic series crops out, but appears to be only of local importance.

It is difficult to see exactly where to draw the boundaries of the formation. Both the transition of the Oville Formation into the Barrios and the transition from the Barrios into the overlying Formigoso Formation are gradual, although the transitional zones are not extensive. As lower boundary is taken the limit above which the shales play a minor role, as upper boundary the limit above which the rocks have a distinct argillaceous character. In this sense the thickness measured near Corniero is 305 metres, but elsewhere the values will differ from this figure.

MINERALOGICAL COMPOSITION

Quartz

Table VIII, showing the mineralogical composition of some hand-specimens indicate that quartz is strongly dominant, the rocks must therefore be classed as quartzsandstones.

TABLE VIII Mineralogical composition of quartzites of the Barrios Formation in perc. by Volume

Number spec.	Quartz	Sec. Q	Feldspar	Rock fragm.	Musc.	Biot.	Clay	Op.
333	70.8		5.4	0.8	0.6	1.0	18.6	—
331	74.9	4.4	1.8	1.3	0.4	1.0	15.8	0.4
320	69.3	4.2	2.8	0.7	0.8	2.6	19.6	—
316	73.2	2.2	2.8	0.6	0.4	—	19.8	1.0
313	49.2	0.2	1.3	0.2	14.0	4.9	30.0	—
311	77.6	4.0	2.0	0.8	—	—	24.8	—
308	89.4	5.4	0.2	0.4	0.2	0.2	4.2	—
305	84.2	4.8	1.2	0.6	0.4	—	8.6	0.2
303	83.1	3.6	1.4	0.9	—	—	6.4	—
302	89.1	4.4	1.2	0.7	—	—	4.6	—

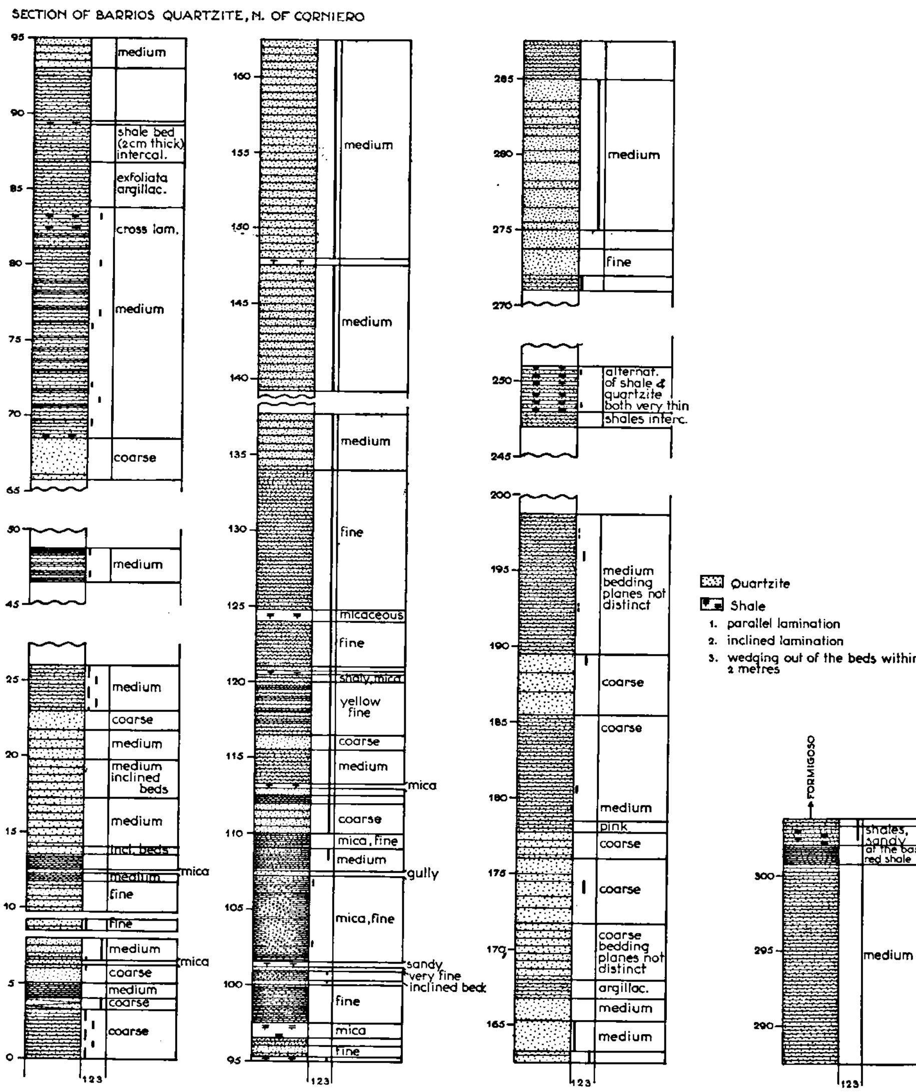

SECTION OF BARRIOS QUARTZITE, N. OF CORNIERO
medium
shale bed (2cm thick) intercal.
exfoliata argillac.
cross lam.
coarse
fine
micaceous
shaly, mica
yellow fine
mica
mica, fine
gully
mica, fine
sandy
very fine
inclined beds
alternat. of shale & quartzite both very thin
shales interc.
medium bedding planes not distinct
pink
coarse bedding planes not distinct
argillac.
inclined beds
incl. beds
Quartzite
Shale
1. parallel lamination
2. inclined lamination
3. wedging out of the beds within 2 metres
FORMIGOSO
shales, sandy at the base
red shale bed
123

Fig. 50a.

DETAILED SECTION OF BARRIOS QUARTZITE
corresponding to height 66.0-69.25 m. of section given in Fig. 50a

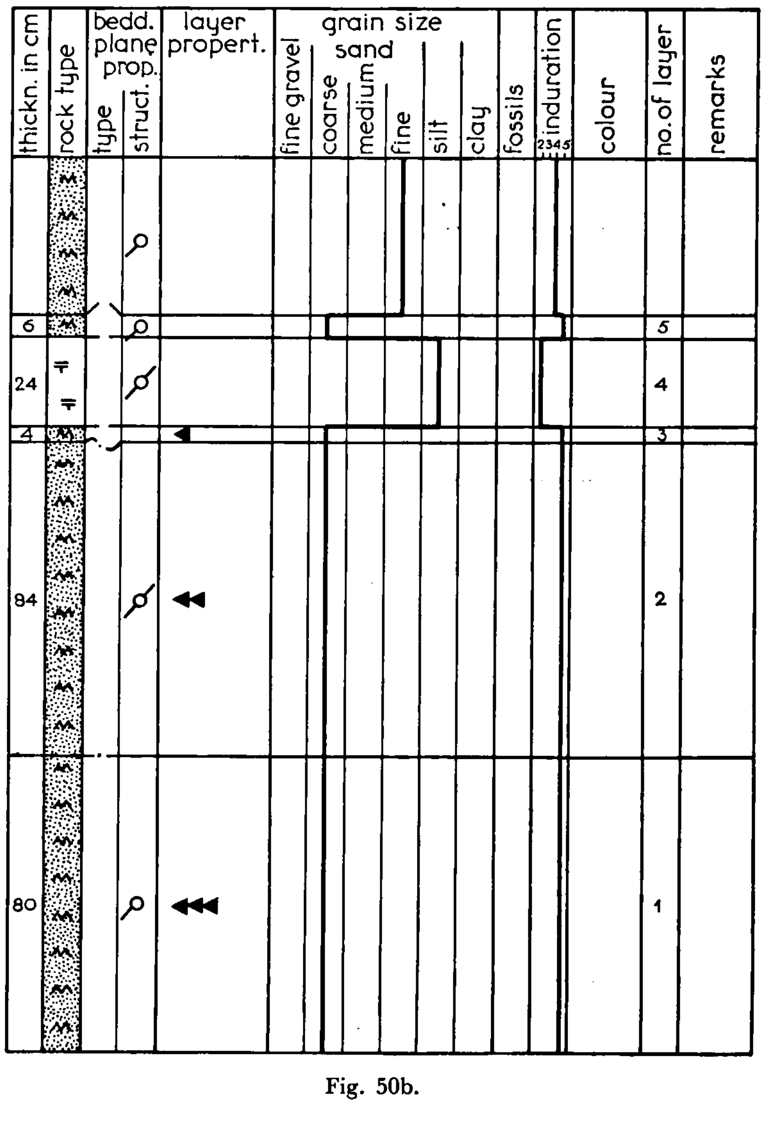


Fig. 50b.

Since dust-rings are often present around the primary grains, the shape and roundness of these grains is known. The quartzes are sub-angular to sub-rounded. Most of the grains are equidimensional, but an appreciable number are elongate. In some beds elongate grains predominate. It is, however, possible that pressure solution somewhat accentuated the elongate character.

The dust-rings just mentioned are composed of opaque matter which is probably manganese, as has already been mentioned. No dust-rings of clay have been observed. Due to the dust-rings, pitting of the grains is visible. Since the pits have been filled up by clay or secondary quartz, the pitting and formation of dust-rings must have taken place in the depositional or early-burial stages.

In contrast with the Herreria sandstones, the grains are only occasionally composite. Inclusions too are scarcely observed, liquid or gaseous inclusions excepted. The extinction is undulating.

Feldspars

Although some plagioclases are present, most of the feldspars are microclines. Unlike those occurring in the Herreria Formation, the microclines are seldom secondarily enlarged, most of the feldspars displaying features of chemical wear along the edges. It could not be determined whether the overgrowths are also composed of microcline.

The roundness of the feldspars as revealed by the dust-rings of the secondarily-enlarged specimens is sub-angular to sub-rounded. Like the shape of the quartzes, most feldspars are equidimensional, but more elongate forms are also present. Rhombohedral outlines are sometimes produced by the secondary overgrowths.

The feldspars are very clear, since alteration took place only on a minor scale. Only a small number of kaolinized microclines have been observed. Kaolinization probably antedated the depositional stage, since the clear and kaolinized forms occur together. On the other hand, decomposition of the grains and formation of clay is visible along the edges and in small fissures. This process proceeds from the edges to the centre of the grains.

Micas

Both muscovite and biotite are present. Generally, the biotite is only weakly pleochroic due to bleaching. The various blades constituting one biotite flake may reflect different stages of bleaching. Because of the optic angle, 2 V measuring 30—35°, and the refractive index n_β which lies between 1.589 and 1.60, the light-coloured mica has been identified as phengite (Tröger, 1956). It is weakly pleochroic, from light gray (n_α) to colourless. It predominates over biotite, which is not surprising since the light-coloured mica is commonly held to be less susceptible to wear than biotite (Pettijohn, 1957, p. 506).

It should be noted that H. Konig (personal communication) was able to demonstrate that the optic angle of muscovite can be smaller than is generally accepted. Koning submerged muscovite (2 V = 45°) in water, and then placed the submerged muscovite on a radiator for several days, after which it was held at room temperature. A week later the optical angle measured only 35°. Koning suggests that submersion over a longer period might have lowered the angle to still smaller values. According to him, the change in the angle is due to the reaction by which $K^{\cdot}$ is replaced by $H_3O^{\cdot}$.

The mica content is generally very low. In some cases, however, the content is very high, e.g. hand-specimen 313 (Fig. 51), especially in shales which must be termed micaceous shales. Again, phengite is in such instances the dominating mica. The appearance of such beds might suggest a diagenetic origin of the mica, but actually no arguments other than the high amounts of mica and the length of the blades can be found. That such a supposition is indeed incorrect follows from the sub-rounded shape of the mica, whereas authigenic mica would have a more euhedral form (cf Clarke & Stevenson, 1960).

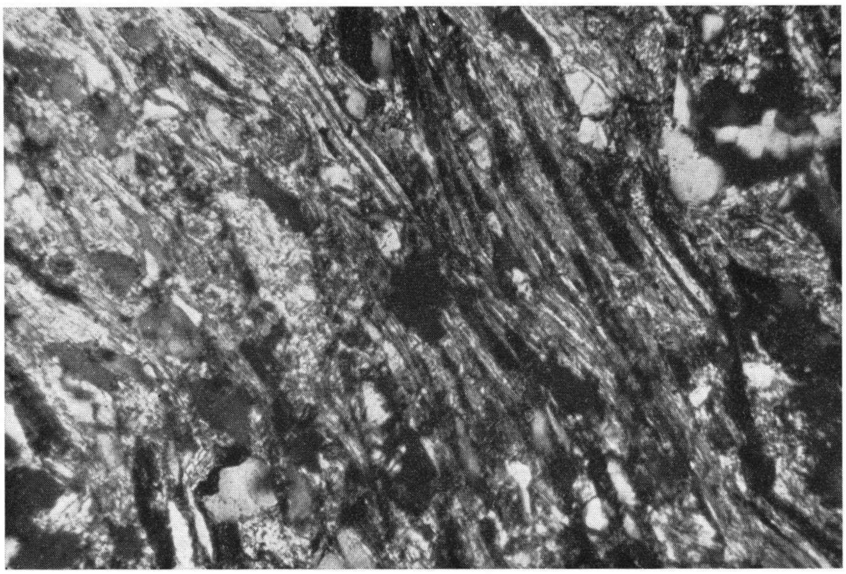


Fig. 51. Micaceous sandstone of Barrios Formation. crossed nicols. 80 ×.

The micas occurring in the sandy sediments are not longer than the specimens forming part of the micaceous beds ($\pm$ 500 μ). The grains are sub-rounded, but they sometimes show a sheaf-like termination.

In addition to bleaching, the micas show other features of alteration. Some of the micas are slightly chloritized, others being decomposed and sericitized. Solution as observed in the Oville Formation does not occur here.

Rock fragments

As in the older formations these are only composite quartzes. It is noted that the percentage is still lower than in the other formations.

Binding material

The *matrix* between the sand-sized grains is clay-sized material consisting of quartz, sericite, and illite.

Cement. Quartz dominates as cementing material. It forms secondary overgrowths, but in some instances the mineral builds minute crystals of clay size, cementing the original grains. In such cases secondary quartz is lacking, the original grains having rounded as well as angular outlines. Here hematite is also finely distributed throughout the beds, which have therefore become red beds.

Minor constituents

Of importance among the minor constituents is the mineral apatite, which is present in the form of primatic sub-rounded grains. The heavy mineral group is represented by tourmaline, blue and brown as in the older formations, zircon, and some rutile, all rounded. Rutile also builds irregular patches, and is seldom observed as detrital grains. The opaque matter, reaching low percent values, consists of magnetite and some pyrite, while hematite is present in a few beds.

It will be noted that rock fragments were not observed except for a few composite quartz grains.

Conglomerate

Mineralogically, the conglomerate of Boñar does not differ from the quartzites. Besides the common constituents they contain composite fragments of quartz, ranging in size up to several cm. Their shape is sub-rounded. The small pebbles do not touch each other and hence the conglomerate has an open frame-work.

The conglomerate resembles the orthoquartzitic conglomerate of Pettijohn (1957, p. 256) in texture and magnitude. He considers such sediments to derive from a transgressive beach over a surface with a low relief. This supposition is discussed below.

TEXTURE

Grain contacts

The sediments being quartzites, the contacts between the quartzes and/or their overgrowths are the most important. As in the Herreria sandstone, all the various kinds of contacts are present. They are tangential, straight, curved, or suturing, but the latter occur in some beds so extensively as to allow use of the term pressolved beds. However, the suturing contact are often contacts between overgrowths, the original grain still being discernable due to the dust-rings. But in these pressolved beds, straight contacts between overgrowths can still be observed. The occurrence of pressolved beds is the more striking because they have not been found in the Herreria Formation, which must have had a still thicker cover of sediments. Since the clay content in both formations is equally low, pressure solution of the Barrios quartzites needs not have been caused by the scarcity of clay, as various authors have argued.

The micas have straight contacts with the surrounding minerals except when they are surrounded by clayey matter where the original outlines may be obscured due to sericitization. In the micaceous shales the mica blades gently curve around the other detrital constituents such as quartz. The micas never abut on such quartzes, which is another argument in favour of the detrital origin of the mica.

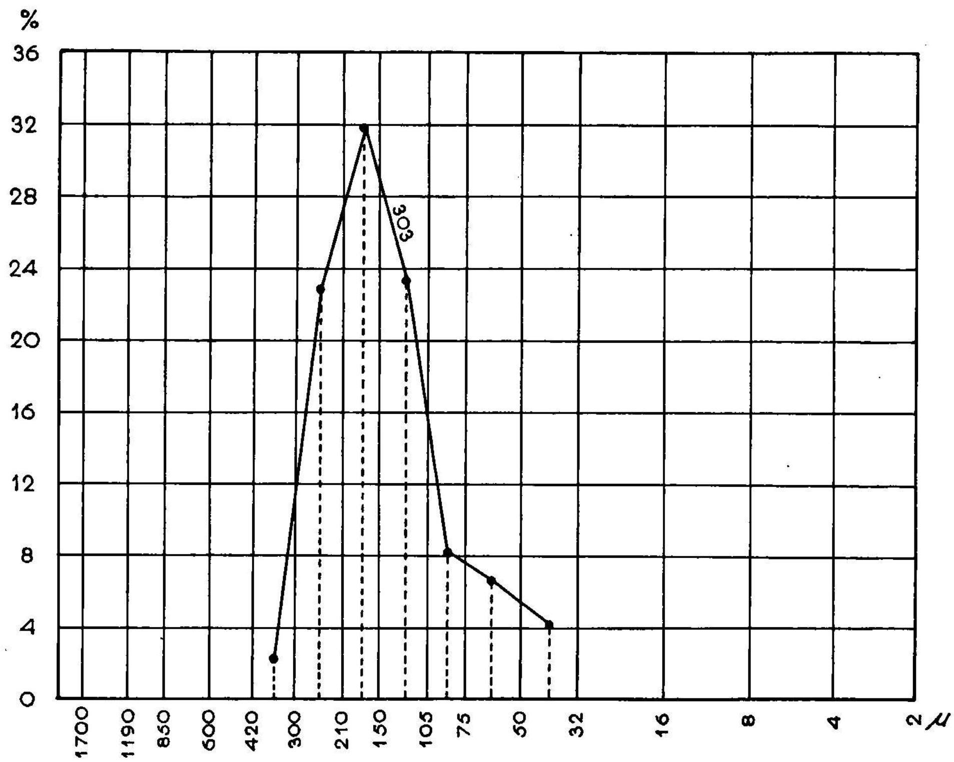


Fig. 52. Grain-size distribution of average quartzite of Barrios Formation.

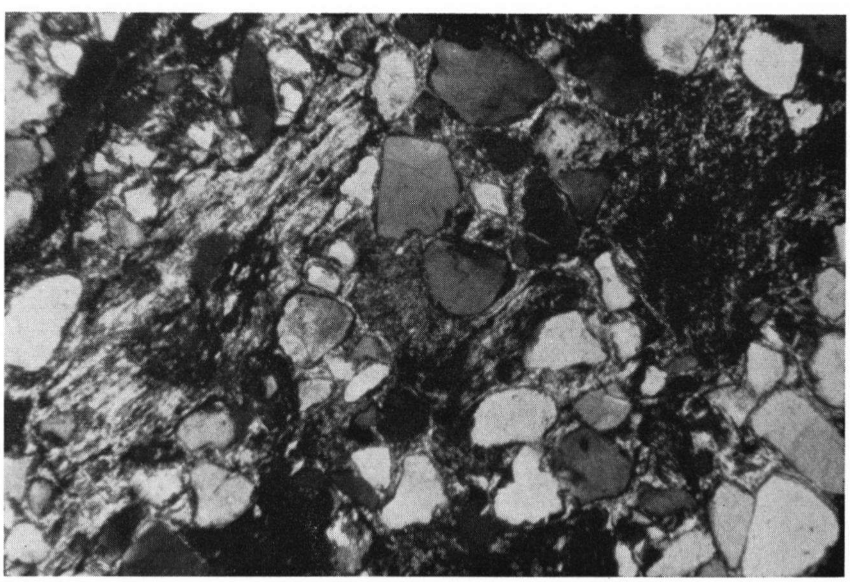


Fig. 53. Badly sorted sandstone with rounded and angular grains. crossed nicols. 32 ×.

Grain-size distribution

Due to the induration of the sediments, the only way to determine the grain-size distribution of the sediments is by measuring the diameter of the original grains with the microscope. The sediments having on the whole about the same grain size, only a few measurements suffice to obtain an idea of the size. The curve in Fig. 52, representing one of the specimens measured, shows that the sorting is rather good.

Some beds deviate from this picture in that they are poorly sorted (Fig. 53). These beds are the quartzites cemented with tiny quartz crystals and the shales, in which the silt-sized or sand-sized grains are floating. These coarser grains of the shales, like those of the quartzites just discussed, can be angular as well as rounded. It is assumed that such beds might represent small subaqueous mudflows.

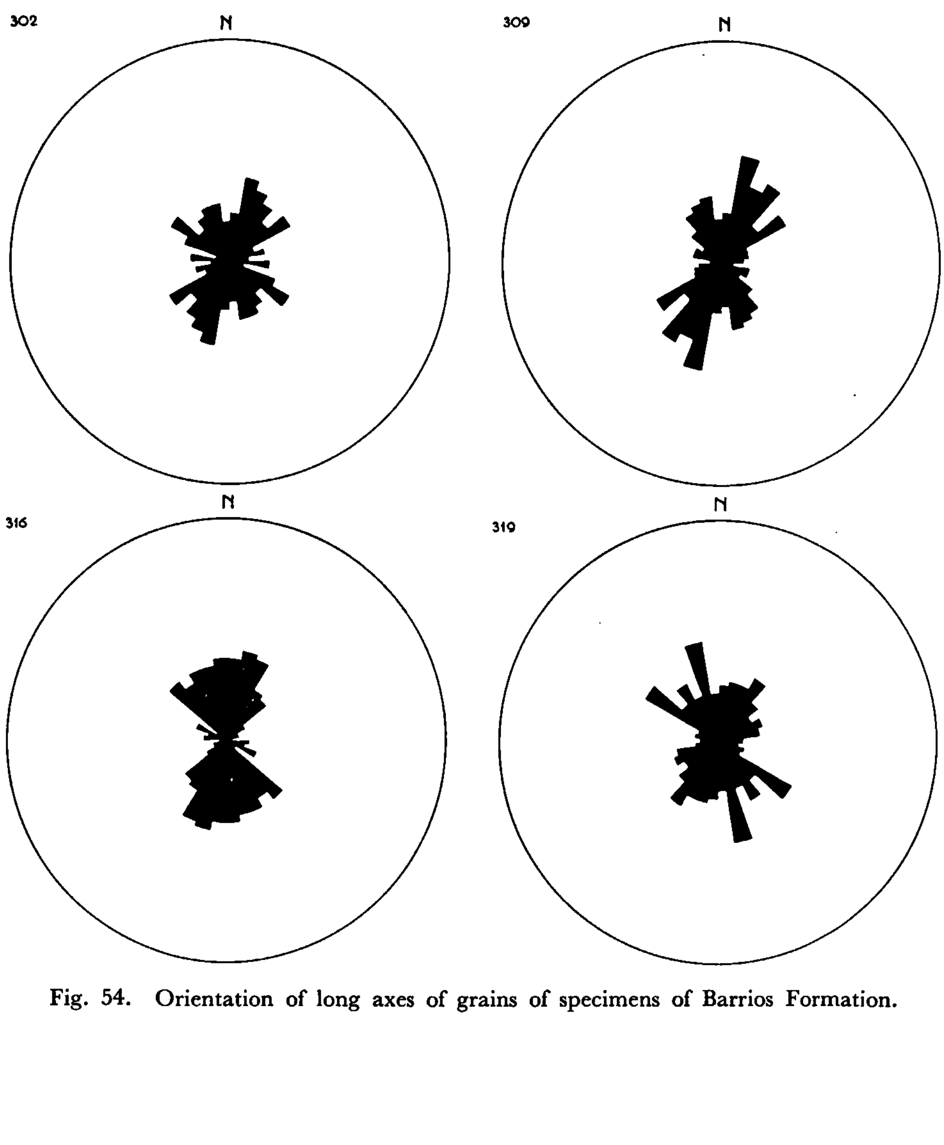


Fig. 54. Orientation of long axes of grains of specimens of Barrios Formation.

Grain orientation

The directions of the long axes of the grains were measured in several specimens; Fig. 54 shows the results. The orientation has a rather wide spread, but there is still a dominating direction, i.e. roughly north-south. Although the number of measurements is too small to permit drawing conclusions, it is noted that the direction given here corresponds to that of the grains of the Oville Formation.

LAYER PROPERTIES

Bedding contacts

The type of contact planes appears to be independent of the rock types involved, at least in the Barrios Formation. All contacts are sharp, some very sharp. The planes have an irregular shape, although undulating forms occur as well. In rare cases they are flat.

Thickness

The formation is a thin-bedded complex, most of the quartzite beds having the average value of about 25 cm. Occasionally some quartzite beds occur whose thickness exceeds the limit of the range defined as thin-bedded, some very thin beds having been observed as well. The shale beds have two dominating thickness values: they measure either about 10 cm or 75 cm and more, values between these figures being exceptional.

Stratification

The wedging-out of the quartzite beds within short distances is a diagnostic property of the formation. In other instances the beds are planar cross-beds forming an angle of up to 20° with the underlying beds (Fig. 55). The internal stratification may be a lamination parallel to the bedding planes or inclined (up to 25°), cross lamination with dips in varying directions rarely occurring. An appreciable number of layers, however, lack any internal stratification.

In the section W. of Cremenes, some shale beds contain quartzite lenses whose diameter may range up to 20 cm. It is difficult to detect the original lamination because small seams of limonite cut the lenses irregularly. It is clear, however, that they contain a wavy lamination or cross lamination. In this respect they resemble the lenticular bodies found in the Oville Formation. The Barrios lenses may have originated in the same way, i.e. the lenses are remnants of eroded beds.

Gullies cutting the stratification occur in both the sections already dealt with, but their number is much higher in a section near Nocedo in the Curueño Valley. Here the dimensions are rather small: the depth measures no more than half a metre, the width is at most 1.50 metres.

Fig. 55. Cross-bedding of some beds of Barrios Formation.

Fig. 56. Spheroidal weathering of Barrios quartzite.

Colour

Most of the quartzites show a white colour on fresh rupture planes. In the upper part of the formation some iron-rich quartzites have been observed with the typically red bed colour. The shales are mostly gray, although a few red shale beds are also present.

Weathering structure

Although the rocks prove to be resistant to weathering, one feature of this type is interesting here: spheroidal weathering. This kind of weathering is noted in the section N. of Corniero; Fig. 56 shows a photograph of the structure.

BEDDING PLANE STRUCTURES

Ripple marks

Some bedding planes with asymmetrical ripples have been found. The crests form straight lines but they flatten out laterally. The distance between the crests is about 7 to 10 cm, their width 0.5 to 1 cm. Lamination is not always present.

Load casts

Quartzite beds overlying shale may have loadcasted into the latter. This structure, however, has no great importance in the Barrios Formation, and has not led to the production of pseudo-nodules.

Clay pebbles

Frequently, small pebbles of clay lie horizontally aligned in the quartzites. They are well rounded, but their dimensions are small. Usually, the diameter measures several millimetres, but some clay galls have a length of 3.5 cm.

Burrows and tracks

Occasionally, burrows and tracks are present. A track, resembling that ascribed to *Cruziana*, was found only in one bed in the section N. of Corniero, but there several specimens are present.

SOURCE AREA AND DEPOSITIONAL ENVIRONMENT

Source area

Here again, the mineralogical composition gives little information on the kind of source area that delivered the material of the Barrios Formation. The predominance of the alkali feldspars over the plagioclases and the presence of the mica phengite may indicate that the source rocks were low-temperature igneous rocks, but a wide range of rock types may deserve consideration. It is also possible that the source rocks too were also sediments, which would find confirmation in the high degree of maturity and, perhaps, the presence of rounded blue tourmaline. Transport over a long distance could also account for the latter observations, however.

Depositional environment

The observations suggest that the Barrios sands were deposited nearshore. The presence of conglomerates, even though they occur locally, the rapid wedging-out of the beds, the cross-bedding, and the occurrence of clay pebbles, all these features taken together seem to point to shallow-water deposits.

Despite the various papers dealing with depositional environments of recent sediments (Thompson, 1937; McKee, 1957; van Straaten, 1959), it is still difficult to determine the depositional environment of the Barrios Formation. Because of the large quantities of material involved, as well as the grain size of the deposits, sedimentation probably took place near the river(s) which supplied the material. On the other hand, since we know the formation to be marine, the very sandy nature of the beds would suggest an environment providing a winnowing-out of the finer fractions such as occurs on beaches. The Barrios deposits are perhaps comparable to the foreshore beach of the Nile delta as recorded by Soliman (1964) in a paper presented at the sedimentological congress at Amsterdam. In such case, the more lenticular character of the Barrios deposits might be explained by assuming a higher degree of roughness of the waters. In this connexion it is noted that Dzulinski & Zak (1960) considered a quartzitic formation in Poland, resembling the Barrios quartzites in many respects, to be storm deposits. If our supposition is correct, the whole formation would represent a mixture of upper foreshore and lower foreshore deposits of a large delta, occasional mud flows having extended from the rivers into these deposits.

CONCLUDING REMARKS

The four formations dealt with in this paper grade into each other without showing any indications that sedimentation was interrupted. On the contrary, the transitional zones between the various formations strengthen the view that the conditions changed gradually. According to Comte (1959), an important hiatus occurs between the Barrios Formation and the overlying Formigoso Formation. He observed an unconformity in the Bernesga Valley, but in the Esla region, unfortunately, no sharp boundary between the two formations can be drawn due to the sandy character of the Formigoso Formation at this site. It is in any case certain that earlier in the Paleozoic disturbances of this kind did not take place.

The three detrital formations among the four, i.e. Herreria Sandstone, Oville Sandstone, and Barrios Quartzite, have one important property in common: they all consist of mature sediments. Mineralogically, they all are quartzsandstones in the classification as used here. Rock fragments and feldspars, and also detrital micas (clay not included), are present only in small amounts. Notwithstanding the fact that the mineralogical differences are small as a consequence of their maturity, these differences are so distinct that they indicate a change of the rock type exposed in the source area in the course of the sedimentary history. (As will be discussed below, the existence of only one source area is assumed). Some of the mineralogical and textural differences may be recapitulated.

Most of the quartzes of the Herreria Formation contain liquid or gaseous inclusions which, according to Comte, point to source rocks analogous to the gneisses cropping out in Galicia. Many other kinds of inclusions are also present. The younger detrital deposits, however, contain quartzes almost completely lacking inclusions. The composite quartz grains, the only rock fragments present, constitute another difference: they occur in smaller amounts in the two younger formations than in the Herreria Sandstone; the Barrios Formation in particular contains but very few of these grains. The presence of kaolinized feldspars in the Herreria Formation is a diagnostic property and in thin section facilitates distinction between rocks of this formation and of both the younger ones.

Mineralogically, the Oville and Barrios Formation differ mutually in that the former contains the mineral glauconite, and the latter the mica phengite. The phengite constitutes the only difference between the formations as far as detrital minerals are concerned.

Texturally, the three detrital formations differ greatly. The Herreria Sandstone consists of quartzsandstones mainly cemented with quartz, the matrix being subordinate. In the Oville Sandstone, clay is important in binding the grains together, and in addition there is the calcite cement which is also rather diagnostic. It is remarkable that the percentages of secondary quartz in the Barrios Formation are rather low compared with the Herreria Sandstone.

Despite these data, it remains difficult to determine the source rocks of the sediments. For instance, the inclusions of the quartzes do not provide such a distinctive clue as Comte thought. However, the source rocks were probably non-sedimentary. It may be argued that the high degree of maturity was due to a polycyclic origin, i.e. the direct source rocks were also sediments. Such an assumption would also explain the roundness of the blue tourmaline ubiquitously present in these sediments, but leaves open the question of the difference between the Herreria quartzes and those supplied in later times. A more mixed character of the quartzes with and without inclusions in the Herreria Formation would be expected, since it is known that tilting took place before deposition of the Herreria Sandstone as revealed by the unconformity between the Precambrian and the younger Herreria, thus exposing various kinds of sediments. On this basis it is assumed that the source rocks were not sedimentary rocks. The blue tourmaline itself is derived from pegmatites, according to Krynine (1946), which supports our supposition; its roundness on the other hand is difficult to explain, because tourmaline is one of the most resistant minerals with respect to mechanical weathering. The only satisfying assumption seems to be long-distance transport.

If the rocks in the source area were non-sedimentary, the maturity of the Paleozoic deposits, too, must be the result of long-distance transport. It is possible that climatic conditions favoured weathering of non-resistant minerals, thus activating the production of mature sediments.

No further evidence is present concerning the rocks of the source area. Some slight indications may favour the assumption that they included acid igneous rocks, at least during suppletion of the Herreria material. Comte found rhyolitic pebbles in conglomerates of the Herreria Formation, and blue tourmaline is indicative of pegmatites. Unfortunately, no other distinctive features have been found.

The situation of the source area also remains obscure, but fortunately here we can rely on the study of the Cambrian in Spain by Lotze & Sdzuy (1961). These authors find reasons to assume that the source area of the Cantabro-asturian sediments was situated in the North. It will be noted that our data on the orientation of long axes of the grains are in agreement with this hypothesis.

It is remarkable that the depositional environment of these various deposits did not undergo important changes. It has been assumed that all sediments were deposited near the shore in a shallow neritic environment. The persistence of such an environment implies that the rate of supply and of sedimentation kept pace with the subsiding movement of the sedimentary basin. Some tentative remarks on the relation between this subsidence, rate of supply and of sedimentation, and movements in the source area will be made.

During deposition of the Herreria sandstone, the relief being in a state of relative youth, sediment-loaded rivers debauched into the sea, where the material settled; it did not, however, built deltas but rather formed continuous sand layers. Currents spread the material over the sea bottom. Occasionally the fluvial influences extended further and built fluviatile beds, now intercalated between the marine layers.

When deposition of the Lancara limestones started, the source area no longer delivered detrital material. Uplifting of this area had probably ceased; subsidence of the basin continued, but more slowly, which follows from the fact that the environments remain at about the same depth, although the sediments involved settle slowly.

The griotte indicates the start of new tectonic activities which continue during deposition of the Oville sandstone. At first, lime deposition is regularly interrupted by sedimentation of argillaceous and later more sandy material. Since this coarser material settled at higher rates than the lime, and since the environment remains about the same, a more rapid subsidence of the sedimentary basin took place. In addition, the source area was perhaps lifted up again. If our supposition is correct we see once again a combination of tectonic movements and a warm, humid climate, as indicated by the red colour of the griotte beds. Such a concurrence is has also been observed elsewhere.

The tendency towards an increasing rate of sedimentation continued until it reached a maximum during deposition of the Barrios Formation. It is quite possible that the lower clay content here also reflects a climatic change. According to Comte, after sedimentation of the Barrios a long period of non-deposition started before renewed deposition, which built the Formigoso Formation, took place.

SAMENVATTING

De structuren en petrografie van de sedimenten der oudste vier formaties van het Paleozoicum in het noordelijk deel van de provincie León (Spanje) vormen het onderwerp van dit proefschrift. Drie van de vier formaties bestaan uit detritisch materiaal, terwijl de vierde is opgebouwd uit kalken en dolomieten. Mineralogisch gezien zijn de detritische sedimenten rijp. Bij gevolg zijn de verschillen tussen de formaties onderling slechts klein, maar desondanks kenmerkend. De gesteenten zijn afgeleid van niet-sedimentaire gesteenten.

De *Herreria Zandsteen Formatie* is de oudste formatie. Hier worden slechts de bovenste 200 meter beschreven. Dit gedeelte bevat middelkorrelige kwartsieten met inschakelingen van grove kwartsieten, conglomeratische lagen en schalies.

De detritische kwartsen bevatten veel vloeibare en gasvormige insluitsels en zijn vaak samengesteld. Microklien, de meest voorkomende veldspaat, is veelal gekaolinizeerd. Beide mineralen hebben secundaire aangroeiingen. De bron der secondaire kwarts wordt ter discussie gesteld: de silica is gedeeltelijk aangevoerd; vervolgens heeft neerslag uit de formatiewaters plaatsgevonden. De saliniteit der formatiewaters neemt toe tijdens de diagenese, waardoor de oplosbaarheid der silica juist afneemt.

Over het algemeen vertonen de lagen een parallelle laminering, maar kriskras laminering komt eveneens voor. In twee delen van het pakket is een snel uitwiggen van de lagen waargenomen.

Het afzettingsgebied was vermoedelijk ondiep, in de nabijheid van de kust met fluviatiele invloeden.

De *Lancara Kalk Formatie* kan worden onderverdeeld in Dolomiet s.l. en de Griotte. De *Lancara Dolomiet s.l.* omvat dolomieten, kalken, oölitische kalken en breccies. Het diagenetische proces "grain growth" vervormde de oorspronkelijke textuur van deze sedimenten. De dolomitisatie vond na de afzetting plaats. Rekristallisatie ten gevolge van mechanische krachten is ook waargenomen.

De *Lancara Griotte* bestaat uit knobbelige kalken en schalielagen met kalkknobbels. De kalken zijn in oorsprong detritisch. Het ontstaan der griotte wordt nagegaan: de griotte is het gevolg van oplossingsprocessen.

De afzetting van de Lancara Dolomiet s.l. wordt vergelijkbaar geacht met de recente neerslag op de Bahama Banks. Het milieu van de griotte is minder duidelijk, maar moet ondiep neritisch zijn geweest. De rode kleur zou op een warm, humide klimaat wijzen.

De *Oville Zandsteen Formatie* wordt gekenmerkt door het hoge kleigehalte, een betrekkelijk hoog kalkgehalte en het voorkomen van het authigene mineraal glaukoniet. Een vervanging van de mica's door carbonaten is waargenomen, hetgeen een onbekend proces is. De conclusie omtrent de ontstaanswijze van de glaukoniet is als volgt: cryptokristallijne aggregaten zijn oorspronkelijk klei geweest, terwijl de kristallijne vormen ongevormde mica's zijn.

Belangwekkend zijn de slump structuren. Aangezien zij het gevolg zijn van een thixotroop gedrag van de sedimenten, wordt in het kort ingegaan op enige grondregels der rheologie. Bovendien wordt er een verband gelegd tussen intern slumpen en vorming van convolute laminaties in die zin, dat beide een "falsebody" thixotrope toestand van het sediment ten tijde van de beweging weergeven. Een dergelijke toestand kan verwacht worden binnen een bepaald bereik van het

vochtgehalte: intern slumpen geschiedt bij de laagste waarden, vorming van convolute laminaties bij de hoogste waarden van dit bereik. Convolute laminaties worden frequenter waargenomen in afzettingen uit troebelingsstromen als gevolg van het feit, dat deze afzettingen sneller zullen sedimenteren dan andere sedimenten, zodat het vochtgehalte van eerstgenoemde sedimenten ook hoger zal zijn.

In dit dunbankige complex overheerst een parallelle laminering maar een kriskras laminering, zij het op kleine schaal, komt ook voor. Andere sedimentaire structuren zijn load casts, pseudo-nodulen en Linsen structuren.

Het afzettingsmilieu zal deltaisch zijn geweest, dat wil zeggen de formatie omvat een keten van delta's.

De *Barrios Kwartsiet Formatie* bestaat uit kwartsieten met een enkele schalielaag en zeer lokaal een conglomeraat. De kwartsen zijn helder en bevatten geen insluitsels. Samengestelde korrels zijn zeldzaam. De veldspaten zijn niet gekaolinizeerd, doch slechts gesericitizeerd. Het voorkomen van het mineraal phengiet is karakteristiek.

De meeste van deze lagen zijn wigvormig, hetgeen de formatie een eigen aspect in het veld geeft. Evenals de Oville afzettingen zijn ook de Barrios zanden afgezet op een delta.

RESUMEN

Las estructuras y la petrografía de los sedimentos de las cuatro formaciones más viejas del Paleozoico en la parte norte de la Provincia de León están descritas en esta tesis del doctorado. Tres de las cuatro formaciones están construidas de material detrítico, la otra contiene calizas y dolomitos. Respecto de la mineralogía los sedimentos detríticos están maduros. Por consiguiente las diferencias mineralógicas entre las formaciones son pequeñas, pero con todo características. Las rocas que han proveido el material no fueron sedimentos mismos.

La *Formación de Arenisca de Herrería* es la formación más vieja del Paleozoico. Solamente la parte superior está descrita aquí. Este parte contiene cuarzitos de grano mediano, en las que están intercaladas cuarzitas de grano grosero, conglomerados y esquistos arcillosos.

Los cuarzos contienen muchas inclusiones líquidas y gaseiformes. Hay muchos cuarzos, que son compuestos. El microclino, el feldespato más frecuente, se encuentra kaolinizado muchas veces. Los dos minerales han sido aumentados durante el diagénesis. La procedencia de cuarzo secundario es discutida: en parte el excedente de silice resultó del transporte de los ríos; después el cuarzo se precipitó por el agua de formación. Generalmente el contenido de sal de las aguas de formación aumenta durante el diagénesis, lo que reduce la solubilidad de silice.

En su mayoría las capas tienen una laminación paralela, aunque laminación inclinada ocurre también. En dos partes de la serie las capas son cuneiformes.

El ambiente de deposición fué probablemente poco profundo, cerca de la costa, con influencias de ríos de vez en cuando.

La *Formación de Caliza de Láncara* puede ser subdivida en el Dolomito s.l. y la Griota. El dolomito s.l. contiene dolomitos, calizas, calizas ooliticas y brechas. El procedimiento diagenético llamado "grain growth" transformó la textura original de esos sedimentos. La alteración de caliza en dolomito ocurrió después la deposición. Fuerzas mecánicas también causaron recristalización.

La Griota contiene calizas nudosas y esquistos con linchazónes de caliza. Las calizas son detríticas en origen. El origen de la Griota resultó de procedimientos de solución.

Se puede comparar la deposición del Dolomito s.l. con la precipitación calcarea que acontece ahora en las Bahamas. El ambiente de la Griota está menos claro, pero debe haber sido poco profundo y nerítico. El color rojo demonstrará un clima cálido y húmedo.

La *Formación de Arenisca de Oville* está caracterizada por su contenido de arcilla, su contenido bastante grande de cal y la presencia del mineral autigénica glauconita. Aconteció substitución de micas por carbonatos, pero las condiciónes precisas están desconocidas todavía. El origen de glauconita se piensa así: los aggregados criptocristalinos han sido originalmente arcilla; las formas cristalinas son micas transformadas.

Son interesante las estructuras de "slumping". Han sido mencionados, porque resultaron de un estado tixotropico de los sedimentos, algunos principios de la reología. Hay una relación entre "internal slumping" y formación de "convolute lamination": los dos reflejan un estado de "false-body" tixotropía del sedimento durante el movimiento. Este estado se puede esperar entre ciertos grados de humedad. "Internal slumping" acontecerá en grados bajos, formación de "convolute lamination" en grados más elevados. "Convolute lamination" está observada más frecuentemente en deposiciónes de "turbidity currents", porque estos sedimentos son depositados más rapidamente que los otros sedimentos. Por consiguiente esos sedimentos seran también muy húmedos.

Las capas de esa formación son tenues y tienen una laminación paralela. Hay tambien laminaciones inclinadas. Otras estructuras sedimentarias están "load casts", "pseudo nodules" y "Linsen-Strukturen".

El ambiente de deposición fué deltaico, q.e. la formación constituye una cadena de deltas.

La *Formación de Cuarzitas de Barrios* contiene cuarzitas y algunas capas de esquistos y en algunos sitios un conglomerado. Los cuarzos están claros, faltan inclusiones. Los granos compuestos son raros. Los feldespatos no están kaolinizados, sino sericitizados. La presencia del mineral "phengite" es característica.

En su mayoría las capas son cuneiformes, lo que da a la formación un aspecto distinto. Como las deposiciones de la Formación de Oville la Formación de Barrios representa también un ambiente deltaico.

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