LEIDSE GEOLOGISCHE MEDEDELINGEN, Vol. 45, pp. 73—154, published separately 22-4-1970

SEDIMENTOLOGY AND PALEOGEOGRAPHY OF AN UPPER CRETACEOUS TURBIDITE BASIN IN THE SOUTH-CENTRAL PYRENEES, SPAIN

BY

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ABSTRACT

The present study deals with the primary lithology, sedimentary structures, depositional history and paleogeography of an Upper Cretaceous turbidite basin in the south-central Pyrenees, and presents a brief review of the lithology and depositional environment of surrounding contemporaneous deposits.

During Coniacian to lower Maastrichtian, over a longitudinal distance of about 80 km along the present strike of the Pyrenees, thick series of calcareous turbidites, pebbly mudflows, slumps, sedimentary limestone breccias and marls accumulated, constituting what is termed the Vallcarga Formation.

Sedimentary structures occurring in this formation indicate a relatively proximal site of deposition.

A close study of the primary lithology of the Vallcarga sediments shows that they are largely composed of calcareous material derived from a shallow shelf, such as fossils, limestone rock fragments, intraclasts, pellets and micrite, with an additional admixture of terrigenous quartz and muscovite grains.

A brief study was carried out in areas surrounding the Vallcarga Formation in which sediments of the same age are cropping out. Terrigenous material was supplied in areas to the NW and NE of the Upper Cretaceous basin, derived from granodioritic massifs exposed there, whereas in the southern part of the basin limestones were deposited, largely composed of reefal detritus.

It is concluded that the turbidite basin was a small trough, and was brought about by increased local subsidence. In the westernmost part, this subsidence led to the formation of a fault-controlled submarine canyon, filled up with sedimentary breccias,possibly derived from the collapse and submarine erosion of an uprising diapire-like anticlinal structure.

Field observations as well as measurements of current directions show that the basin was composed of three parts, separated partially or entirely by submarine swells. Each of these parts was filled up by currents flowing in a different direction, thus resembling the present basins off the coast of southern California.

SUMARIO

En este estudio se trata la litología primaria, las estructuras sedimentarias, la historia de deposición y la paleogeografía de una cuenca de turbiditas del Cretáceo superior en los Pirineos surcentrales, dando al mismo tiempo una versión resumida de la litología y el medio ambiente de deposición de sedimentos contemporáneos que rodean esta cuenca.

Durante el Coniaciense y hasta el Maastrichtiense inferior, sobre una distancia aproximada de 80 km a lo largo del rumbo actual de los Pirineos, fue depositada una potente serie de turbiditas calcáreas, *mudflows*, *slumps*, breccias calcáreas sedimentarias y margas. Estos sedimentos constituyen la denominada Formación de Vallcarga. Esta formación puede ser dividida en una parte inferior (Mascarell Member) que consiste principalmente de depósitos resedimentarios y margas autóctonas y una parte superior (Salas Member) formada por margas grises azuláceas en las que las turbiditas se ven rara vez.

Las estructuras sedimentarias ocurrentes en la Formación de Vallcarga, tales como *flute-* y *groovecasts*, canales, *loadcasts*, granulo-clasificación, laminación paralela, laminación oblicua a pequeña escala, convoluciones y *slumps* son descritas en relación al medio ambiente en el que fueron formadas. Se saca como conclusión que indican un sitio de deposición relativamente proximal ya que se encuentran frecuentemente canales con relleno grueso, breccias calcáreas de gran espesor que pueden ser graduadas o no graduadas y microbreccias. Esta proximalidad también es indicada por la presencia de capas gruesas de laminación oblicua a gran escala en algunas partes de la serie sedimentaria. Hay depósitos que teniendo la apariencia de turbiditas proximales y de turbiditas distales ocurren juntos en la misma sucesión, indicando así que la magnitud de una corriente de turbidez es el factor principal que determina la aparente proximalidad de sus depósitos. Apenas se han encontrado argumentos que sugieren un transporte por otras corrientes de fondo. Por consiguiente, la hipótesis de las corrientes de turbidez explica satisfactoriamente las propiedades de la mayoría de las capas resedimentarias. Los *slumps* están presentes en todas las partes de la formación; sus ejes son normalmente paralelos a la dirección de corriente. Se han observado todos los estados de transición desde simples roturas hasta mezclas caóticas de diversas turbiditas y margas.

Un estudio minucioso de la litología primaria de los sedimentos de la Formación de Vallcarga revela que estos estan principalmente formados por material calcáreo procedente de un *shelf* poco profundo tales como fósiles, fragmentos de roca calcáreos, *intraclasts*, *pellets* y micrita, conteniendo adicionalmente también cuarzo y muscovita de origen terrigénico. Los componentes calcáreos y algunos fragmentos de roca de edad Triásica y Albiense, igualmente como granos de glauconita, proceden de una erosión que se produjo dentro de la cuenca misma. El mineral arcilloso mas importante que compone las margas es ilita, que puede provenir de rocas más antiguas o por autigénesis.

Los procesos diagenéticos que condujeron a la litificación de los depósitos incluyen la recristalización de micrita, la cementación por calcita y algunos otros fenómenos como son reemplazamiento y autigénesis de cuarzo y disolución por efecto de presión.

Se hizo también un breve estudio en las regiones que rodean a la Formación de Vallcarga, en las que estan expuestas sedimentos de la misma edad. Su contenido fosilífero indica que fueron depositados en un mar poco profundo, bien aerado y caluroso. En regiones al NW y NE de la cuenca del Cretáceo superior fue suministrado material terrigénico, proveniente de macizos granodioríticos expuestos aquí, mientras que en la parte meridional de la cuenca fueron depositadas calizas principalmente constituídas por material detrítico procedente probablemente de arrecifes.

El area de máxima sedimentación se desplazó hacia el oeste durante un lapso de tiempo comprendido entre el Cretáceo inferior y el Eoceno. Se concluye por tanto que la cuenca de turbiditas fue una canaleta estrecha originada por un hundimiento local aumentado. La formación de esta cuenca empezó en la sección central de los Pirineos meridionales, extendiéndose después hacia el oeste y el este. En la parte más occidental este hundimiento, que posiblemente puede ser relacionado con una migración de evaporitas del Keuper, condujo a la formación de un cañon submarino originado entre fallas y rellenado por breccias sedimentarias posiblemente derivadas del colapso y la erosión submarina de una estructura anticlinal diapírica. Las observaciones en el campo, tanto como las mediciones de las direcciones de corriente, indican que la cuenca estaba compuesta por tres partes, separadas parcialmente o totalmente por umbrales submarinos. Cada una de estas partes fue rellenada por corrientes que se dirigían en una dirrección diferente, mostrando así una semejanza con las cuencas que actualmente estan presentes delante de la costa de California.

El último estado de relleno de la cuenca de turbiditas fue marcado por una potente serie de margas y la desaparición de turbiditas, probablemente debido tanto a la disminución del hundimiento como a la ausencia de material que pudo ser emplazado en corrientes de turbidez.

Estos depósitos son seguidos por calcarenitas arenosas de grano grueso que constituyen la Formación de Arén y que fueron depositadas en un medio ambiente marino mas bien poco profundo.

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CHAPTER I

INTRODUCTION

PURPOSE OF THE PRESENT STUDY

The present study concerns itself with a local turbidite basin of Upper Cretaceous age, situated within an area of calcareous shelf sedimentation. Because of the good exposures of the basin deposits, consisting of turbidites, marls, slumps, mudflows and very coarse to fine-grained sedimentary breccias, it may serve as an example of the development of, and the mechanisms active in, such a basin.

During Upper Cretaceous times the largest part of the south-central Pyrenees was the scene of a predominantly calcareous sedimentation, which took place on a slowly subsiding marginal basin, more or less parallel to the present strike of the Hercynian axial zone of the Pyrenees (de Sitter, 1964). The basin was bordered in the west by the so-called Aragón swell (Misch, 1934) and in the east by the Ampurdán swell (Ashauer, 1934), which consequently were areas of little or no deposition. Part of this basin, however, was subjected to a high local rate of subsidence, leading to the deposition of a thick series of mainly calcareous turbidites and marls called the Vallcarga Formation (Mey *et al.*, 1968). This formation was laid down in a relatively narrow trough extending over a longitudinal distance of about 80 km in a NW-SE striking direction. To a basin of this type we shall apply the term turbidite basin given by Potter and Pettijohn (1963, p. 241).

The purpose of the present study has been to investigate the origin and sedimentation history of this turbidite basin by means of the primary lithology, paleocurrents and sedimentary structures as well as a general review of the surrounding formations, in order to reconstruct the paleogeography of the basin and the possible source areas of the turbidites.

The south-central Pyrenees are extraordinarily well-suited for such a study because of the many outcrops in the folded Upper Cretaceous formations all over the area (Enclosure I) and the rather limited region in which marine sedimentation took place, so that an overall picture of the distribution pattern could be obtained.

STRATIGRAPHY

Broadly speaking, the Vallcarga Formation consists of a lower member composed of an alternation of turbidites and marls with frequent intercalations of slump sheets, pebbly mudflows and breccias, termed Mascarell Member, and an upper part composed of bluish-grey marls called the Salas Member. In some areas, however, other members may be discerned locally. In the Esera region the lowermost 400 m of the Mascarell Member show a succession of coarse thick-bedded limestone breccias which were formerly recorded as a separate formation (Mey *et al.*, 1968), but which in the present study are considered as part of the Vallcarga Formation and named the Campo Breccia Member, since its depositional history is intimately related to that of the Mascarell Member. In the Ribagorzana and Pobla de Segur areas, the middle part of the formation (between the Mascarell and Salas Member) is composed of a thick sequence of mudflows, slumps and olistostrome levels which sequence is called the Pumanous member (Wiersma, 1965). Almost everywhere the lower boundary of the formation rests conformably on top of Upper Cretaceous formations, except in the Esera area, where an angular unconformity may be observed locally. The upper boundary of the Vallcarga Formation is drawn at the contact between the blue marls of the Salas Member and the thick-bedded coarse-grained calcarenites of the Arén Formation (Mey *et al.*, 1968).

The age of the Vallcarga Formation is fairly well known by the stratigraphic work of Souquet (1967). The lower boundary of the formation is oldest in the Ribagorzana Valley, where the common occurrence of *Globotruncana coronata* Bolli, *Praeglobotruncana imbricata* (Mornod) and *Praeglobotruncana* sp. and the absence of *Globotruncana concavata* indicate a middle Turonian to early Coniacian age (B. Kuhry, written comm.). In a westerly and easterly direction, deposition of the formation began later. According to Souquet (1967), the lower boundary in the Esera Valley is of lower to upper Santonian age, whereas the base of the formation in the Pobla de Segur area was deposited during the upper Santonian to lower Campanian.

No exact age determinations are available to locate the Santonian-Campanian boundary, which may tentatively be taken somewhere beneath the base of the Salas Member (Souquet, 1967), neither is this the case with the Campanian-Maastrichtian boundary. It is only known that the base of the Arén Formation was laid down during the upper Maastrichtian.

Consequently, the bulk of the Vallcarga Formation was deposited continuously during the upper Santonian to lower Maastrichtian. Its lateral transition into contemporaneous limestones, formed in different depositional environments, is summarized in Chapter V; the variations in total thickness of the Vallcarga Formation are shown in the enclosed sections (Enclosures II, III, IV, V, VI).

PREVIOUS AUTHORS

The rock units with which this study concerns itself have been summarily described by various authors in the course of their regional geological surveys of the folded Upper Carboniferous to Lower Tertiary strata of the south-central Pyrenees. The studies by Dalloni (1910, 1930), Misch (1934), Selzer (1934) and Rosell Sanuy (1965) largely cover the area in which the Upper Cretaceous is exposed. Parts of the area are covered by geological maps prepared by members of the Department of Structural Geology of the University of Leiden (Wennekers, 1968; Mey, 1968; Hartevelt, in prep.). A practical lithostratigraphic subdivision of the post-Hercynian succession was proposed by Mey *et al.* (1968). The extensive stratigraphic work by Souquet (1967) has largely contributed to the knowledge of the Upper Cretaceous rocks, and was also of great assistance in the interpretation of facies distributions. Internal reports of the subsection of Sedimentology of the University of Leiden on parts of the Vallcarga Formation were prepared by Wiersma (1965) and Masselink (1965) for the Pobla de Segur area and by Ditzel (1966) for the Ribagorzana region. Nagtegaal (1963) reported on the occurrence of convolute lamination and metadepositional ruptures in Vallcarga turbidites exposed north of Pobla de Segur, while a brief preliminary description of the Campo Breccia was given by Van Hoorn (1969).

The present paper forms part of a project undertaken by students of the subsection of Sedimentology of the University of Leiden with the object of elucidating the sedimentary processes of Upper Cretaceous and Lower Tertiary formations in the southern Pyrenees.

EXPLANATION OF TERMS AND SYMBOLS USED

Only a few remarks have to be made concerning the terms used in this paper, since they are widely found in geological literature. The description of bedding thickness was made according to McKee and Weir (1953), while the terms relating to roundness and size distribution (Wentworth scale) of clastic rocks are those listed by Pettijohn (1957). The terminology used in the description of the rock types occurring in the Vallcarga Formation is presented in Chapter III. The classification of limestones as proposed by Folk (1959) was followed for the calcareous rock sequences surrounding the turbidite basin (Chapter IV), whereas the symbols used to express their fossil content and sedimentary features are shown in Fig. I-1. Sandstones were classified according to Gilbert (1958).

ACKNOWLEDGEMENTS

In the first place I wish to express my gratitude to Prof. A. J. Pannekoek, whose numerous critical suggestions

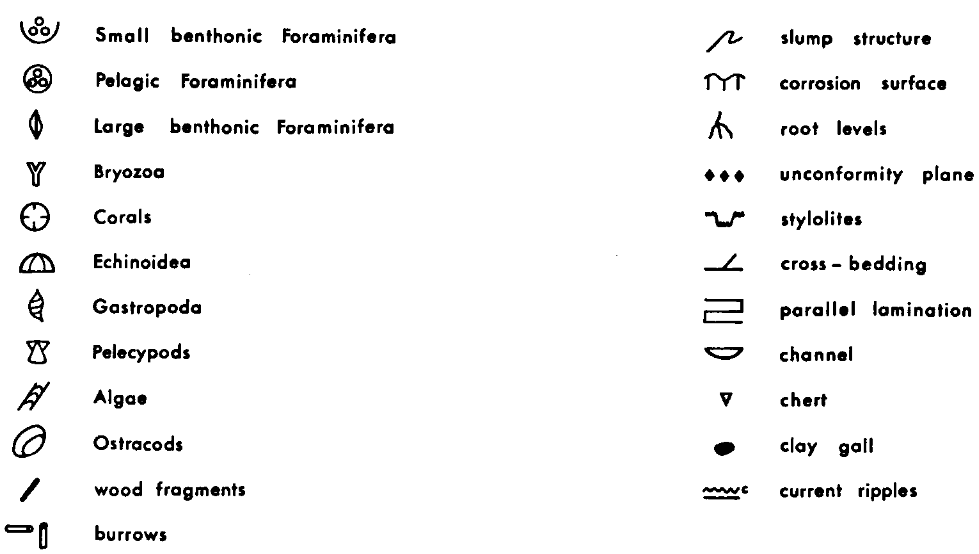


Fig. I-1. List of symbols used.

greatly contributed to the completion of the present study.

Dr. J. D. de Jong, whose enthusiasm for sedimentology was infectious, rendered valuable assistance in the field and provided me with many useful suggestions for this manuscript.

In my activities in the south-central Pyrenees I was introduced and greatly inspired by Dr. P. J. C. Nagtegaal, who also aroused my interest in the microscopical study of sedimentary rocks.

My thanks are also extended to Prof. A. Brouwer for his critical reading of this thesis.

I am also greatly indebted to Miss E. van der Wilk, Dr. J. W. A. Bodenhausen, Mr. J. J. A. Hartevelt, Mr. C. M. Kooter and Mr. A. J. van Loon who were of regular assistance during the preparation of this paper, as well as to Mr. P. A. M. Gaemers and Mr. B. Kuhry for their valuable paleontological advice.

The execution of the X-ray analyses by Mr. A. Verhoorn and their interpretation by Mr. R. O. Felius merits grateful acknowledgement. I wish to extend my thanks to Mr. B. G. Henning and Mr. M. L. Brittijn for their skilful draughtmanship, to Mr. W. C. Laurijssen and Mr. W. A. M. Devilée for the photographs, and to Mr. J. Verhoeven and Mr. J. Schipper, who prepared the thin sections.

Special gratitude is due to Mrs. E. M. Prins-v. d. Drift who rendered me the invaluable service of typing the document, and to Mr. A. V. Zimmermann who corrected the English text.

CHAPTER II

SEDIMENTARY STRUCTURES: OCCURRENCE AND GENESIS

In the present study sedimentary structures will be described in relation to the environment in which they were formed. For practical reasons the classification proposed by Kuenen (1957), based on the moment of origin, was not followed. Instead a more or less descriptive division into external structures (flute- and groove-casts, channels, load casts, trails and burrows), internal structures (turbidite intervals) and deformational structures (slumps) was used. Detailed descriptions of these types of structures may be found elsewhere (see Potter and Pettijohn, 1963; Dzulynski and Walton, 1966), whereas our study concerns itself mainly with the particular features found in the Vallcarga Formation, relating them to their genetic significance.

EXTERNAL STRUCTURES

The term external structure is used here to designate all structures present on bedding surfaces (Dzulynski and Walton, 1966), inorganic or organic in origin, which may be present at the sole of a layer as well as at the upper surface. In the case of organic activity evidence can be found throughout the layer.

Flute casts and groove casts

Most workers make a distinction between "scour marks" (mainly flute casts; also other structures such as rill marks, small channels, etc.) attributed to the effect of sediment-laden turbulent eddies within a suspension flow (Dzulynski and Sanders, 1962) and

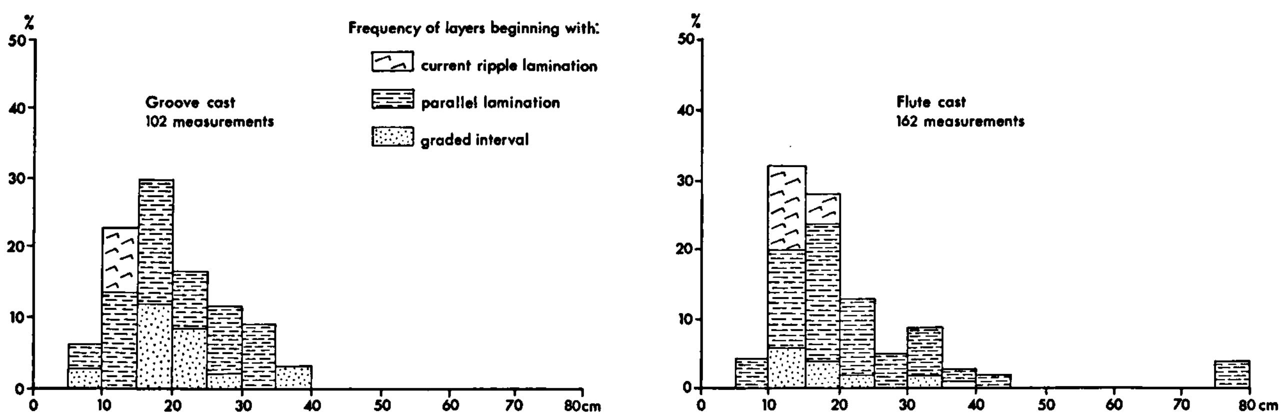


Fig. II-1. Relation between type of lowermost interval and layer thickness in flute- or groove-cast bearing turbidite layers.

Area W of Turbón ridge (Campo, Llert)

cm	abcd	bcd	cd	a	ab	ac	acd	abc	b	bc	c	a → e	Total
0 – 5	–	2	–	–	20	–	–	2	140	4	3	56	227
5 – 10	1	3	–	31	51	2	–	9	175	14	4	–	290
10 – 20	2	6	1	9	33	1	–	3	65	16	1	–	137
20 – 40	–	6	–	5	15	3	–	6	20	10	–	–	65
40 – 80	–	–	–	4	11	–	–	6	–	2	–	–	23
80 – 160	–	–	–	–	3	–	–	1	–	2	–	–	6
160 – 320	–	–	–	–	–	–	–	1	–	–	–	–	1
> 320	–	–	–	–	–	–	–	–	–	–	–	–	–
	3	17	1	49	133	6	–	28	400	48	8	56	749

Area E of Turbón ridge (Vilas de Turbón, Isábena Valley)

cm	abcd	bcd	cd	a	ab	ac	acd	abc	b	bc	c	a → e	Total
0 – 5	–	–	4	6	8	–	–	–	91	2	15	45	171
5 – 10	–	16	30	50	48	3	1	3	252	44	76	–	523
10 – 20	3	20	9	23	45	2	–	6	122	54	40	–	324
20 – 40	6	19	–	30	42	2	–	6	49	27	12	–	193
40 – 80	3	2	1	15	12	–	–	3	8	4	–	–	48
80 – 160	1	–	–	4	4	–	–	1	3	–	–	–	13
160 – 320	–	–	–	3	–	–	–	–	–	–	–	–	3
> 320	–	–	–	–	–	–	–	–	–	–	–	–	–
	13	57	44	131	159	7	1	19	525	131	143	45	1275

Table II-1. Relation between layer thickness and type of sequence. Sequence terms according to Bouma (1962). a→e refers to structureless layers less than 5 cm in thickness included in interval c (Walker, 1967). b-intervals may include upper horizontal lamination d not recognized in the field.

"tool marks" (groove casts, chevron marks, prod marks, bounce marks, etc.) attributed to the effect of traction of larger particles within a current with traction carpet. The presence of both flute casts and groove casts at the sole of a layer obviously indicates a current with mixed loads: finer-grained material in suspension and larger particles moving by traction (Dzulynski and Sanders, 1962).

A relation between the thickness of the layers and the occurrence of flutes and grooves is apparent in the Vallcarga Formation (Fig. II-1), most of these being present at the base of layers with thicknesses varying between 10 and 20 cm, and which begin with a lowermost parallel laminated interval. This thickness value is somewhat larger than the mean thickness frequency of the Vallcarga turbidites (Table II-1). Flutes are eroded by the head of the current, whereas filling up and deposition on top of the flutes takes place behind the head, at a lower stage of flow regime, leading to the formation of parallel laminated beds (Middleton, 1967).

No careful measurements were carried out to obtain a correlation between grain sizes and the presence of flute casts. Broadly speaking, it may be stated that about 80% of layers with flute casts can be classified as medium-grained deposits with grain-sizes between $\frac{1}{2}$ and $\frac{1}{4}$ mm, less than 20% being coarse-grained. In all cases where flute casts were recorded, the layers contain 20% or more of terrigenous detritus (mainly quartz and chert fragments).

The relative abundance of flute casts varies greatly from area to area. In the Noguera Ribagorzana and Isábena sections their occurrence is quite common, whereas in the region west of the Turbón ridge they are scarcely found. According to Allen (1968) currents making flutes are turbulent and are controlled by high Reynold's numbers. Hence, variations in these values must have occurred in order to obtain a notable absence of flutes. A possible explanation has been offered by Dzulynski and Sanders (1962) who stated that, when in high-density currents the optimum concentration of tools which may be carried in suspension is exceeded, a traction carpet may be formed. In this traction carpet moving grains follow a linear path on the muddy bottom, which is therefore not eroded; i.e. there is no turbulent scouring effect. In agreement with this, Sanders (1965) stated that the presence of small objects moving at the base of the current prevents the formation of eroding eddies on the underlying mud, which explains the absence of flutes. An alternative interpretation based on a low current velocity does not seem to agree in the area W of the Turbón ridge, most sediments having been deposited proximal to the source area under conditions of an upper flow regime which evidently favours a greater scouring action (see Chapter V). Otherwise, the absence of groove casts in the same area may be explained by the uniformity in grain size of the passing material in which no suitable tools were present, leading to a fully protective traction carpet.

In some cases it was found that the material filling up the flute mould is coarser than the material composing the lowermost part of the turbidite layer. Evidently the sediment-laden current following the eroding currenthead did not flow sufficiently fast to prevent the coarser grains present in the lowermost bed load from being carried on, and being consequently captured in the hollow, protecting it from further erosion. To Sanders (1965), the presence of coarser-grained flute casts is proof of the existence of an inertia-flow layer together with a turbulent suspension. The term inertia-flow refers to transportation of sediment in which grains move passively within the flow above the muddy bottom, following linear paths. Hence, all coarse grains transported by this flow are able to outdistance some of the finer grains.

Channels

According to the definition of Dzulynski and Sanders (1962), channels are longitudinal scour marks, showing a varying pattern in size and shape. They are frequently recorded in the Vallcarga Formation. Four types of channels have been encountered, differing notably in size and type of infilling material:

a. At the base of parallel-laminated layers channels occur, their curved and relatively smooth bottom surfaces being accentuated by loading, and showing bended laminated internal structures, in which parallel alignments of clay galls are common (Fig. II-2). Mean width and depth values observed are 60 cm and 30 cm respectively. Width dimensions are generally twice as large as the corresponding depth values. In some cases the lower surface of a channel presented flutes formed prior to its infilling.

b. Wide channels with gently dipping flanks (Fig. II-3), filled with coarse-grained material (up to 2 mm in diameter) showing an internal lamination. The lower surface of the channel is highly irregular and erosive, and sometimes displays tool marks. Width and depth can be as great as 10 m and 50 cm respectively.

c. Deeply eroded channels with steep flanks and relatively smooth lower surfaces, filled with microbreccia material consisting mainly of small (mean diameter between 2 mm and 2 cm) rounded to subangular rock fragments and abundant bioclasts (Fig. II-4). Clay galls and internal bedding features are common (Fig. II-5), as is loading at the base of the channel. The maximum width observed was up to 6 m, the depth being not more than 2 m. On the top the channel fill in all cases changed abruptly into a laminated calcarenite layer extending, with no particular change in thickness, over great distances beyond the maximum channel extension.

d. Very coarse-filled highly erosive channels, with dimensions varying from 5 m in width and 50 cm in depth (Fig. II-6) to several metres in both directions in

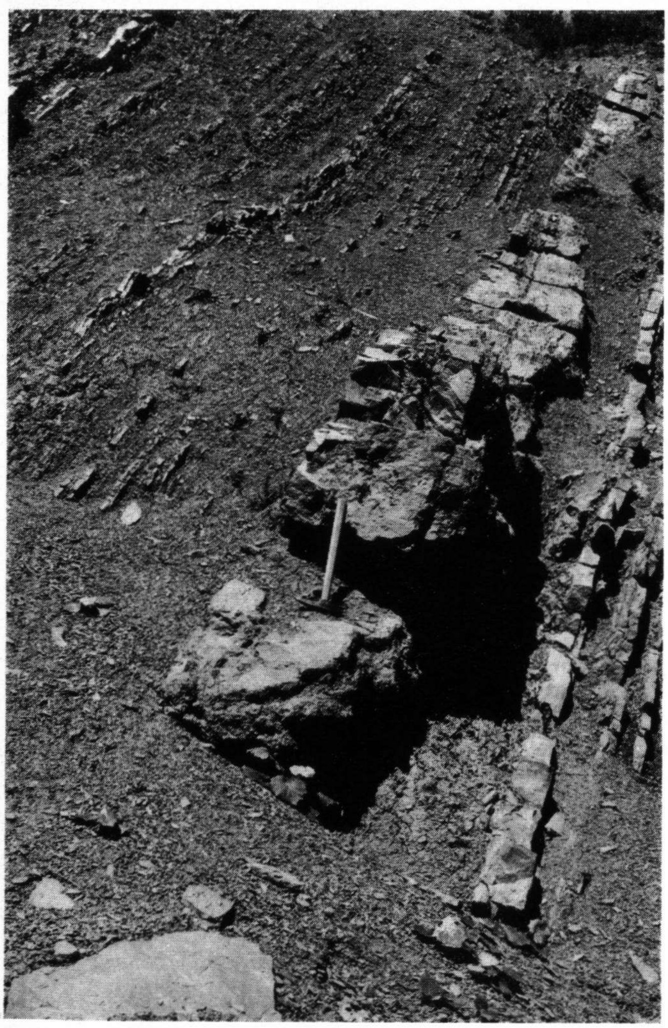
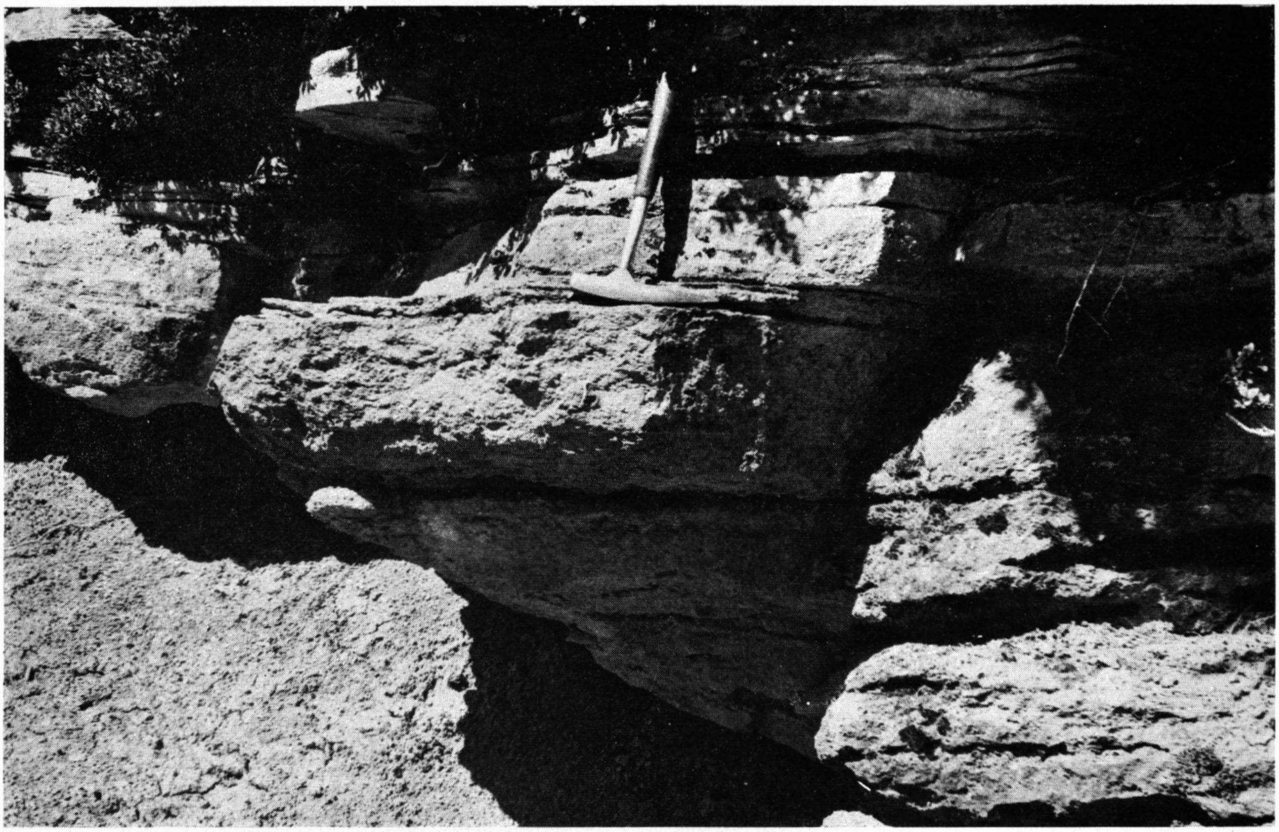


Fig. II-2. Small channel-like scour at the base of a coarse parallel-laminated sandy calcarenite (Photograph by P. J. C. Nagtegaal).

Fig. II-3. Flat channel filled up with coarse sand to microconglomerate within the turbidite sequence near Obarra, Isábena Valley (Photograph by P. J. C. Nagtegaal).

Fig. II-4. Strongly erosive channel filled with calcareous rock fragments, followed at the top by coarse laminations. Exposure N of Vilás del Turbón.

Fig. II-5. Clay galls aligned parallel to bedding in a channel, filled with coarse material, with a parallel-laminated top. Surrounding material is marl. Photograph taken N of Campo, Esera Valley.



one particular case (Fig. II-7). The infilling material consists of angular breccia boulders, slabs of fragmented layers, and a lesser amount of calcareous muds.

From the descriptions above it would seem obvious that, though not in all cases, channels can be formed as a result of extreme localization of flow lines in a turbidity current (Dzulynski and Sanders, 1962). Other mechanisms must be held responsible for the enormous dimensions of some breccia-infilled channels.

The erosional and depositional history within channels has been reviewed by Walker (1966) in the light of the turbidity current hypothesis. For theoretical reasons he concludes that turbidity currents capable of eroding channels must flow sufficiently fast, have no traction carpet, and be underloaded. These conditions can be fulfilled in the most proximal environment, at the base of the slope, within a submarine fan. In some cases channels may even be expected after the current has flowed over a short distance on to the basin floor (Walker, 1966). It may therefore be concluded that the presence of channels in a turbidite sequence is indicative of proximality. Upfilling occurs by deposition of coarse material of the denser base of the turbidity current, whereas the top of the current which is less dense, spreads out over the channel levees forming a finer grained layer. These observations are in accordance with recent examples reported by Menard (1955) of the north-eastern Pacific basin.

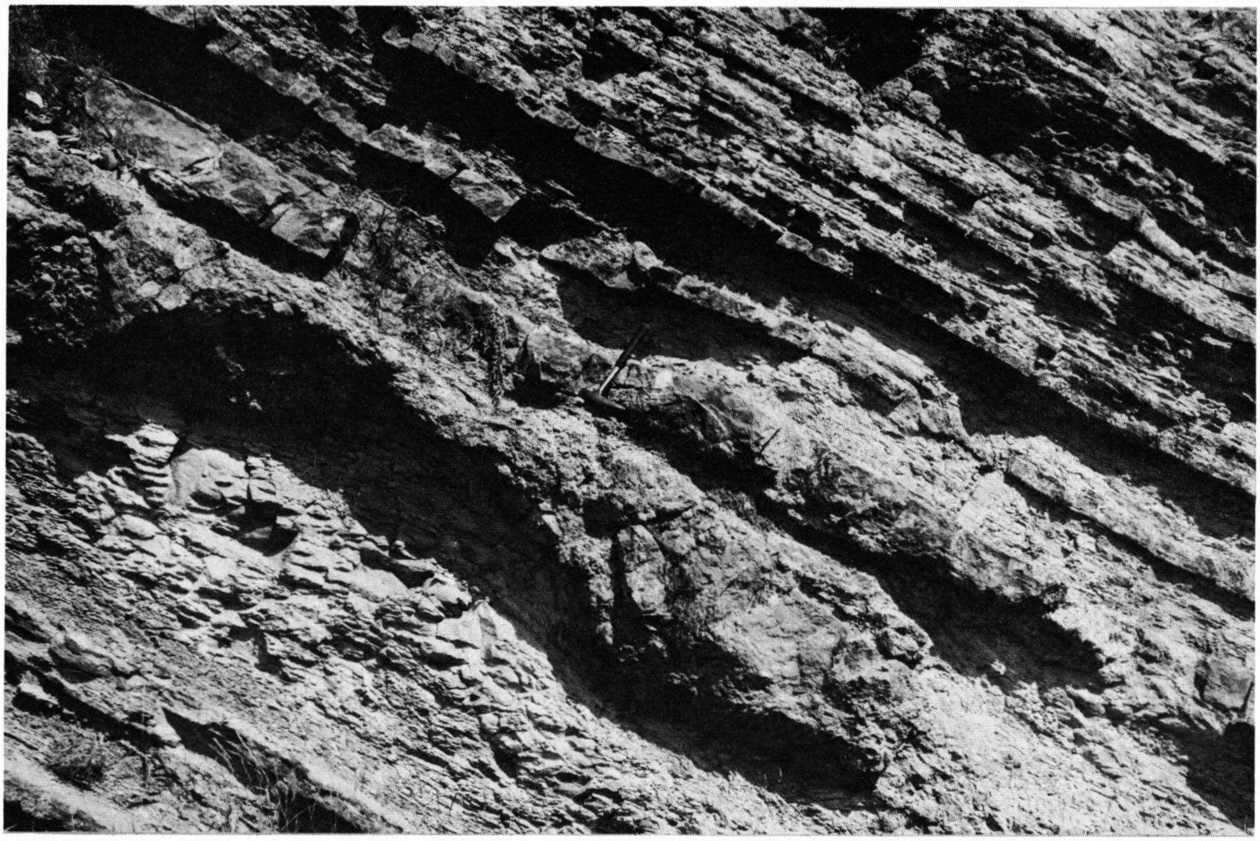
In relation to the channels found in the Vallcarga Formation the explanation advocated by Walker (1966) and confirmed by observations in present basins by Menard (1955, 1964); Gorsline and Emery (1959); Shepard and Einsele (1962); Hand and Emery (1959); and Shepard and Dill (1966) seems to agree quite well for the second (b) and third (c) type of channels described previously. The first type (a) of channel may be considered as a major scouring as compared with flutes, in which coarse material was captured which otherwise would have passed by, whereas the fourth type (d) requires some different explanation. The breccia boulders and pebbles upfilling these channels probably were not transported in a true suspension but, mainly due to their dimension, as a kind of subaqueous rockfall (Dott, 1963), submarine slide (Kuenen, 1956) or large tractional carpet (Sanders, 1965). These traction carpets are likely to be preceded by turbulent suspensions forming the current-head flowing faster than the rolling and saltating, or even sliding, breccia boulders, which undergo a drag effect exercised by the frictional basal boundary in conjunction with the substratum (Bagnold, 1962).

Consequently, the channels are formed by the erosive action of turbulent suspensions which preceded the main bulk of breccia material, and flowed along determinate flow lines. Their velocity was sufficiently high to erode and carry the eroded material further on (demonstrated by the remarkable absence of clay galls in the channel fills). This evidently indicates an underloaded turbidity current, capable of scouring and of picking up the scoured material, which is likely to exist in a proximal environment. As noted by Johnson (1962), fast-flowing turbidity currents gradually increase in volume downslope by the entrapping of sea water, consequently decreasing their sediment concentration and becoming underloaded. The slower-moving carpet of breccia fragments tended to be confined to the channels previously eroded, filling them up. This conclusion is in accordance with Middleton's experiments (1967) which showed that the head of a current is a region of erosion and grooving of the underlying strata and is therefore non-depositional. Deposition takes place behind the head.

The breccia-filled channel in Fig. II-7 cannot be explained by the previous mechanisms. Its dimensions are similar to those described by Walker (1966) in Upper Carboniferous formations in England, but differs in the type of upfilling material. Walker's channels have been cut into turbidites and were gradually filled by material deposited partly by turbidity currents, sand creep or tractional currents. In the present case the channel was also cut into a turbidite sequence but was at once filled by a breccia-laden current (denoted by the absence of any layering). Slumping from the walls into the channels has also been observed (Fig. II-8). In the case of this particular channel only a sudden catastrophic event can be imagined, as prior to the formation of the channel only turbidite deposition is observed and no major erosion features. In conjunction with the approximately E-W orientated channel axis great olistoliths are encountered, deposited on the northern levee, within intensively deformed turbidites and marls. Measured slump axes of these beds show orientations roughly

Fig. II-6. Breccia-filled channel between parallel-laminated turbidites. Diameter of limestone boulders up to 10 cm. Exposure N of Campo, Esera Valley.

Fig. II-7. Large-scale limestone-breccia-filled channel N of Campo near the electric power station, with an almost vertical northern flank. Depth at least 15 m. Photograph looking in an easterly direction.




LST. BR.

similar to that of the main channel axis, but give no indication of the direction of movement. Along the channel walls traces of vertical shear were observed (Fig. II-8), which tentatively suggest downward sinking along faults in part of the basin floor leading to the formation of a small graben. Fault strikes were roughly parallel to the basin axis. Immediately afterwards, as a consequence of the same unstable conditions, coarse material varying from angular polymict limestone pebbles and boulders to immense olistoliths were brought in and deposited in or nearby the newly formed channel. Almost no fragmented components of the Vallcarga Formation were encountered, so that it could not be determined whether the source was intrabasinal. Renz *et al.* (1955) and Marchetti (1957) described gravity sliding over a distance of about 35 km of olistoliths of enormous dimensions, eroded from active fault scarps. However, such a long transport distance does not seem to fit in this case, as is demonstrated by the angularity of the breccia components. Similar breccias deposited in subsiding fault-controlled parts of a marine basin have been described by Blount and Moore (1969) in the Upper Cretaceous of north-western Guatemala, and in recent environments such as the Gulf of Paria (v. Andel and Postma, 1954) and near the coast of Japan (Yamasaki, 1926).

Loadcasts

This type of post-depositional sole marking is often found at the base of microbreccias (Fig. II-9) and coarse-grained calcirudite and calcarenite layers. Their occurrence is dealt with in relation to the external structures, since they are closely related to flute casts and channels. The most common form is that of an irregular downward bulge at the sole of a layer, which is supposed to be the result of vertical and lateral adjustment of the basal material to unequal loading of relatively coarser material in finer-grained hydroplastic sediment (Kuenen and Prentice, 1957). Evidently, the extreme variations in the settling velocity of the sediments of the Vallcarga Formation favour this process. The presence of loadcasts seems to be related to the grain size of the rapidly deposited material. No appreciable loading was observed in fine-grained calcarenites and calcisiltites, whereas breccias only show a slight deformation of the underlying sediment.

Deformation of flute casts and channel bases caused by loading is commonly encountered, in exceptional cases leading to an exaggeration of their dimensions. Pressing upwards of the underlying marls into flame structures generally occurs at the sole of microbreccias. Pseudonodules have rarely been observed.

Trails and burrows

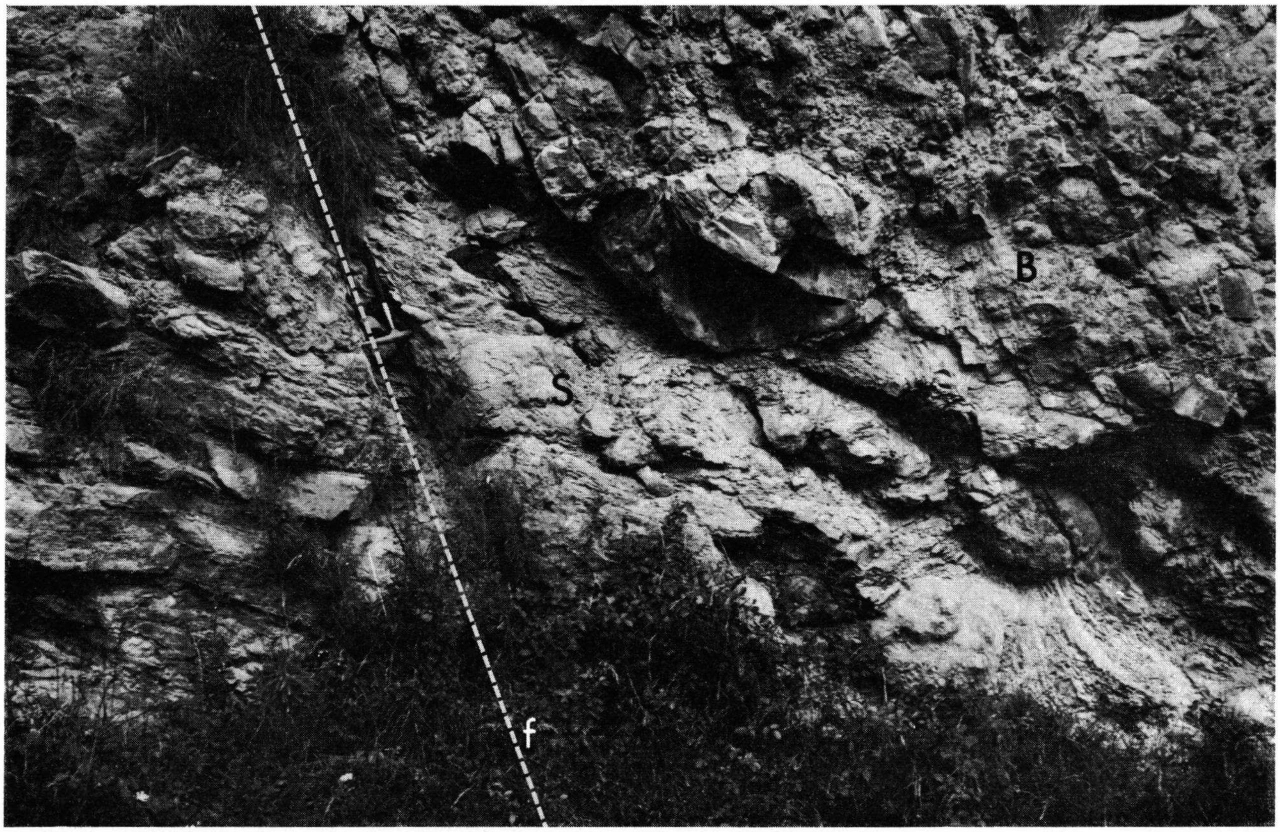
Extensive studies relating to the occurrence of trails and burrows in flysch sequences have been carried out by Seilacher (1959, 1962, 1964, 1967). In the present formation trace fossils are observed either at the sole of turbidites, at the top of these layers or intersecting them partially or totally. In the latter case burrowing has evidently taken place after the deposition of the turbidite bed, partly destroying the internal parallel lamination of the sediment. This intensive reworking is extremely frequent in thin-layered structureless calcisiltites. Presedimentary feeding trails are most common at the sole of the layers, occurring even at the base of scour marks, indicating a lapse between erosion and deposition. Possibly some organisms were able to resist erosion and early sedimentation (Seilacher, 1959). Detailed descriptions of trace fossils of the Vallcarga Formation in the Pobla de Segur area have been given by Farrés (1963). The bathymetry of some trace fossils has been reviewed by Seilacher (1967). The characteristic Nereites community which he describes for depth ranges in recent oceans up to 4.000 m is present throughout the entire formation.

INTERNAL STRUCTURES

According to Bouma (1962), the ideal turbidite sequence is composed of five intervals. Three of these, the graded interval (a), the parallel laminated interval (b) and the current-ripple laminated or convoluted interval (c) are easily distinguished in the turbidites of the Vallcarga Formation. The upper parallel lamination (interval d) is extremely difficult to detect when not preceded by a c-interval, whereas the e-interval may theoretically be expected but was not distinguished in the field. More than 2000 observations concerning internal structures were recorded in two different areas (Table II-1).

Fig. II-8. Detail of Fig. II-7. To the left undisturbed turbidite/marl sequence. Hammer lies on the fault-plane. To the right, slumped turbidites followed in an upward direction by breccia boulders up to 60 cm in diameter.

Fig. II-9. Loadcast at the base of a calcareous microbreccia disturbing underlying thin turbidite and marl layers. Exposure in the Barranco de Llert.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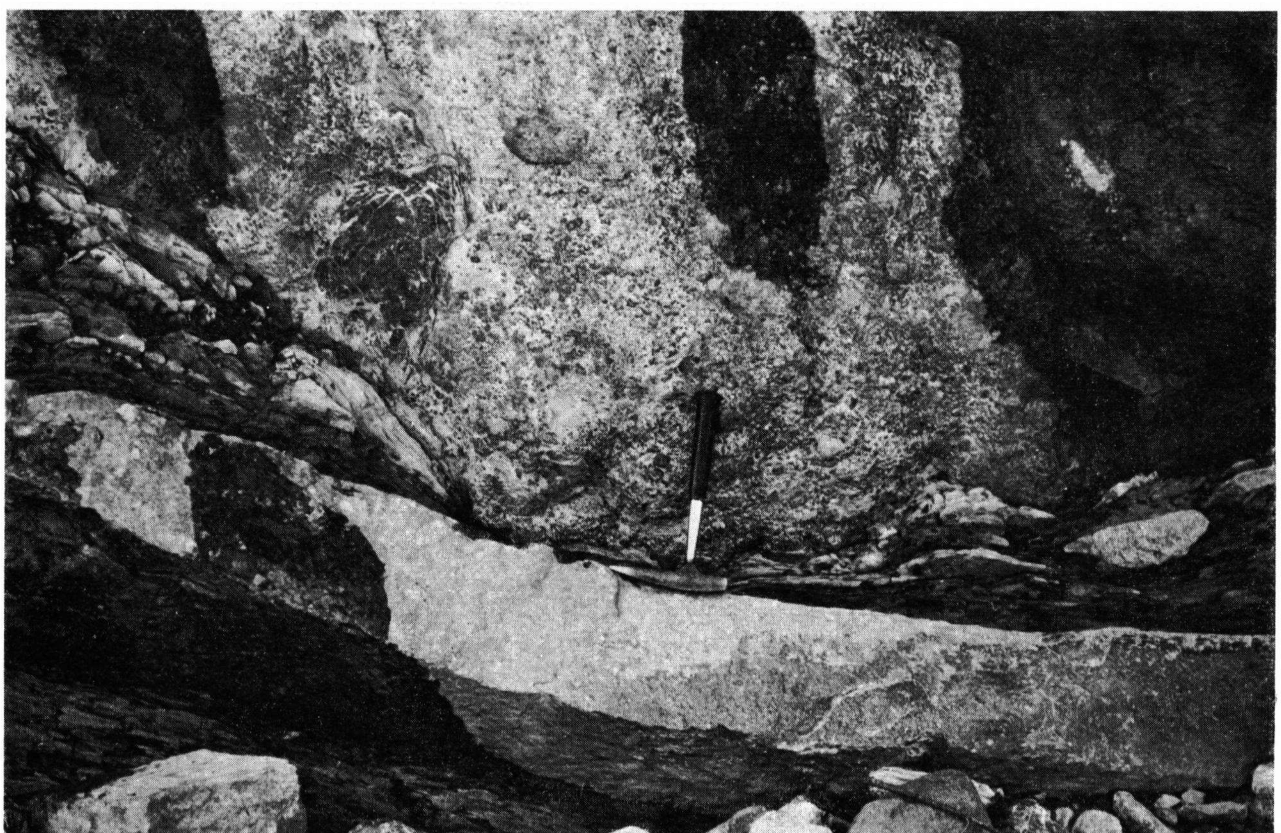


Graded interval

All layers of the Vallcarga Formation show conspicuous grading which may or may not be restricted to the lowermost interval of the turbidite sequence. Two main types have been distinguished in this interval:

a. Increasing amount of matrix in an upward direction, sometimes accompanied by a slight decrease in the maximum grainsize (Fig. II-10).
b. Non-graded interval with an abrupt transition into a parallel laminated upper part (Fig. II-11).

The first type occurs chiefly in sandy calcarenites and biocalcarenites in which the largest grains are constituted by:

1. fragments of corals, Bryozoa, Algae, Echinoidea, oysters, rudists, well-preserved Foraminifera (Orbitoidea, *Prealveolina, Lacazina*)
2. micrite intraclasts
3. quartz and chert fragments

They decrease in number towards the top of the a-interval, being replaced by smaller organic components (mainly benthonic Foraminifera, pelagic Foraminifera, sponge spicules and ostracods (Fig. II-12).

The second type occurs predominantly in coarse calcarenites, microbreccias or even in limestone breccias. No grading, whether in the quantity of matrix, or in the grain size, has been encountered. In between the rock fragments intraclasts, shell fragments, large Foraminifera, quartz granules, etc., are quite common. At the top there is a sudden transition into the parallel-laminated interval in which hardly any coarse components of the types occurring in the underlying deposit are found.

The first type of graded bedding presents no serious difficulties to interpretation and seems to fit well within the normal concept of deposition of material from a decelerating turbidity current (Kuenen, 1953) at an advanced stage of flow (Walker, 1965), whereas the non-graded second type may be attributed to deposition of large grains mixed with finer material from an overloaded turbidity current (Kuenen, 1967), which deposition mainly occurs at early stages of flow when there is little separation of material according to size in the current (Walker, 1965). However, the fact should be stressed that the stage of flow mentioned may not reflect the distance from the source area. According to the experiments of Middleton (1967), the type of grading depends on the concentration of the flow. Both types of grading previously described seem to belong to Middleton's coarse-tail grading produced by a high-concentration flow. Only thin-layered structureless beds grading on the top into a pelitic interval, which have also been observed in our area, may belong to the distribution grading caused by low concentration flows. High concentration of material in a turbidity current therefore prevents the separation of grain sizes in a vertical direction but leads to lateral grading which has sometimes been observed in the Vallcarga Formation. It may consequently be concluded that the occurrence of both types of grading in one vertical section, with an analogous composition reflecting a similar source area, tends to illustrate not merely a difference in flow stage depending on the flow path, but more or less the combined influence of:

1. the initial velocity
2. a difference in concentration of material in the turbidity current
3. the grain size of the particles transported

Experimental results concerning the relation between the velocity of a current and the concentration of a suspended load have been listed by Kuenen (1967, p. 210). All arguments should be examined from case to case; some ungraded layers do not show any erosional features at the base and may express deposition by a sudden "freezing" of an overloaded current with an equal velocity gradient throughout the entire transported mass, whereas other layers with strongly erosive bases were deposited by currents which became overloaded due to their ability to pick up the freshly eroded material. In the latter case it is obvious that no large traction carpet existed at the base of the fast flowing current head, since this would have prevented erosion of the muddy bottom by the turbulent eddies of the current. Middleton (1967) even states that the head of a current is a well-defined region of non-deposition and bottom erosion, whereas deposition takes place immediately after the head has passed by.

Fig. II-10. Graded bedding in a limestone turbidite. Coarse components, mainly bioclasts, some intraclasts and rock fragments (original size 5 × enlarged).

Fig. II-11. Non-graded a-interval composed of rounded limestone pebbles and fossil fragments. Fine-grained material in between. Abruptly overlain by parallellaminated calcarenite (original size 3 × enlarged).

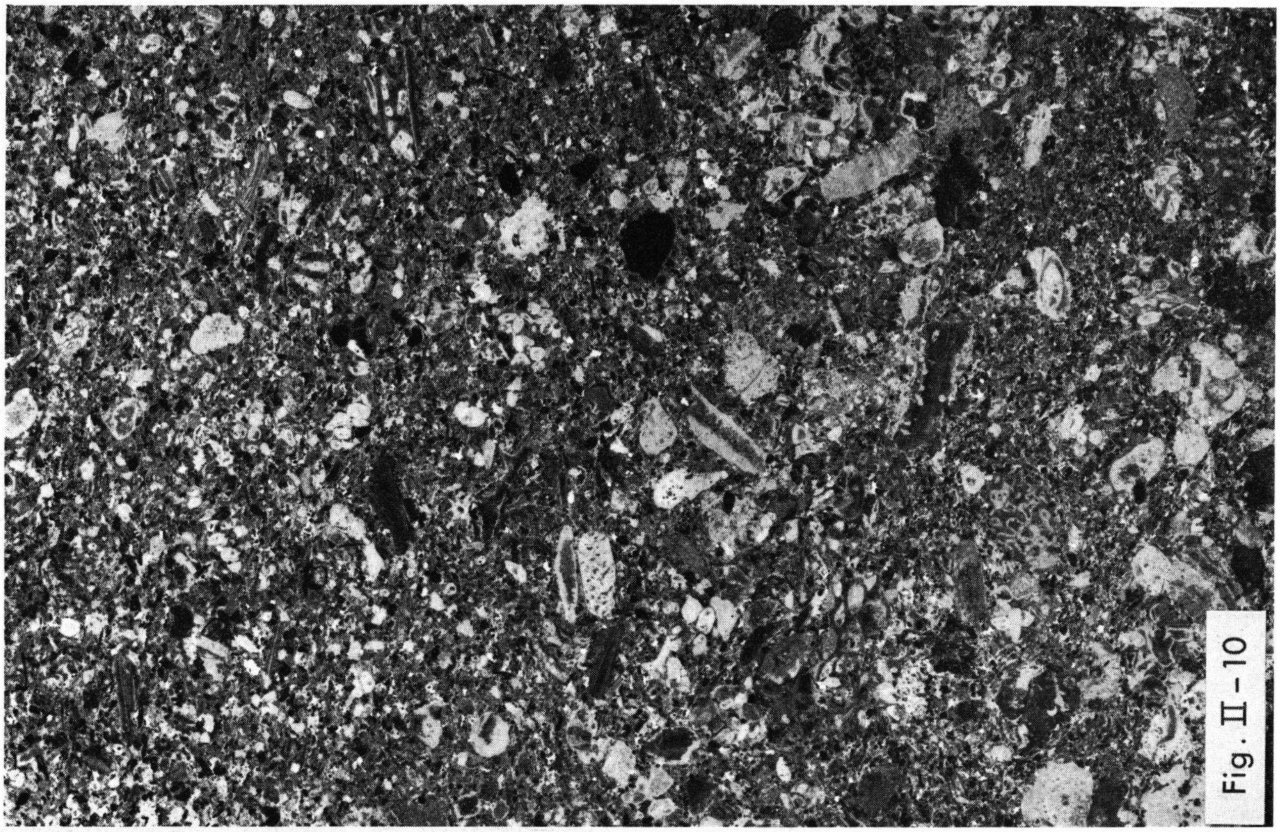


Fig. II-10

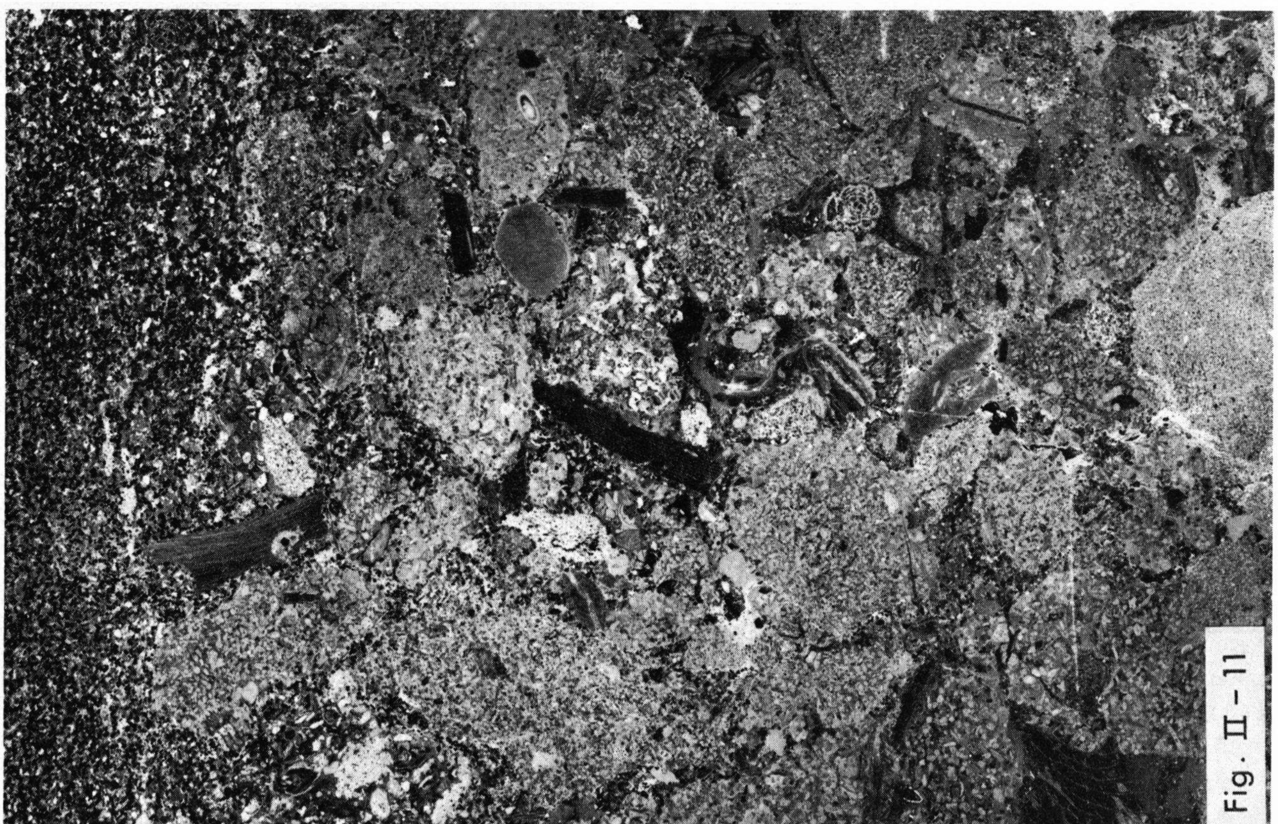


Fig. II-11

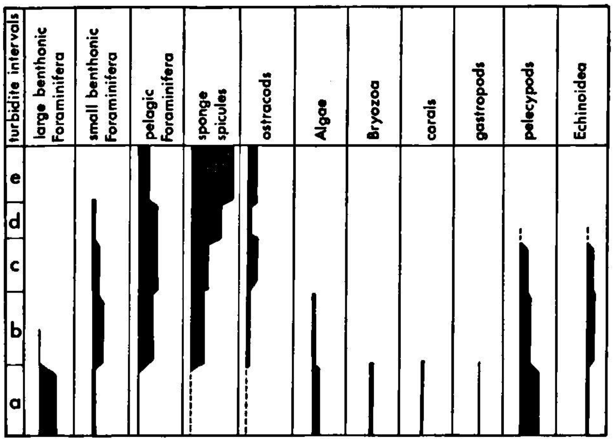


Fig. II-12. Frequency of fossil groups in a turbidite layer. Average of 5 sampled beds. Method according to Weidmann (1966).

Other graded units observed are limestone breccias chiefly present in the Campo area (Van Hoorn, 1969), occurring at the base of the Vallcarga Formation (Fig. II-13). Transport of the breccia components has evidently taken place as a watery slide or subaqueous rockfall (Dott, 1963). Depending on the transport distance some segregation took place, in which the greatest boulders moved on first, followed by a saltating and rolling carpet of finer material, which built up the graded top of the limestone breccias. Almost all layers have a sandy parallel laminated interval following the graded top, indicating that at least part of the finest material was transported in a turbulent suspension moving much slower than the breccia-transporting current. The great bulk of fine material was presumably not deposited but flowed further on, being deposited in relatively distant areas.

Similarly graded breccias were encountered by Gwinner (1961) who also suggested that a better grading reflects a larger transport distance which causes the breccia components to attain a greater freedom of movement within the current. A similar conclusion was postulated by Remane (1960).

Special type of resedimented deposits lacking turbidite features

According to Sanders (1965), the bulk of the transported material was deposited from large traction carpets. This is actually demonstrated by the presence of coarse, thick, ungraded layers with a well-developed large-scale cross-stratification. Calcareous microbreccias in the Llert area, sandy calcarenites in the Isábena valley and Pobla de Segur region (Mutti and Rosell Sanuy, 1968) display this feature. The thickness of these layers may be up to 1 m, the inclination angle of the dipping-cross-laminae may vary between 5° and 20°, the entire bed being cross-laminated or only the upper part following an ungraded structureless lower part (Fig. II-14). The direction of dip of the cross-lamination is similar to the orientation of groove casts present at the sole, although flutecasts do not occur. Sometimes small current ripples are present at the top of the sandy calcarenites.

Analogous cross-bedded layers in other turbidite sequences have been described by Dzulynski *et al.* (1959); Stanley (1963); Ojakangas (1968) and Hubert (1966), who interpreted them as having been deposited under conditions of a lower flow regime. Dzulynski and Walton (1965) give a general description suggesting that they may be explained as small bars originated from swift currents, or, in the case of their being large-scaled as in the present case, as a dune phase during tractional transport (see also Simons *et al.*, 1965).

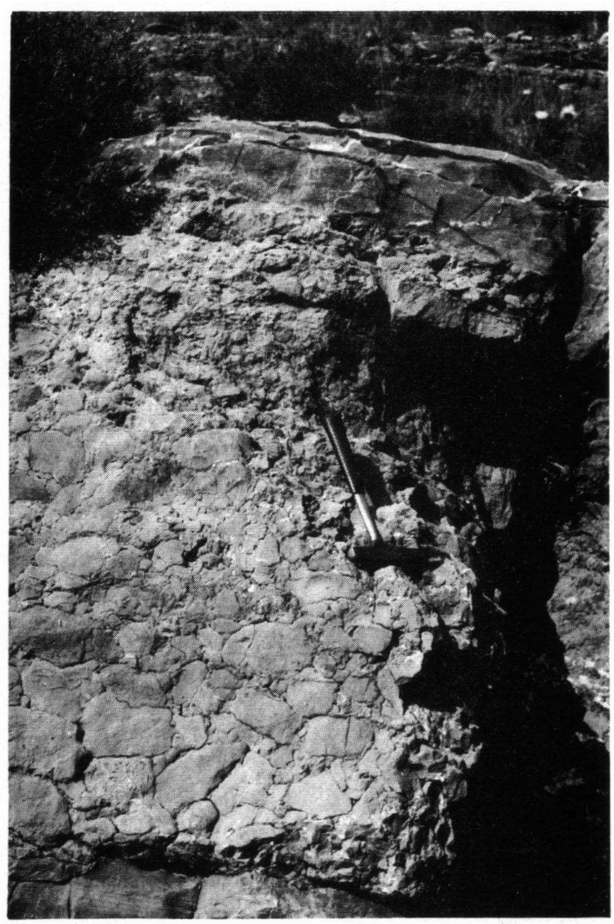


Fig. II-13. Graded limestone breccia transitional at the top into a laminated interval. Erosive base on a preceding turbidite. Stylolitic contacts between polymict limestone boulders. Exposure along the road, N of Campo.

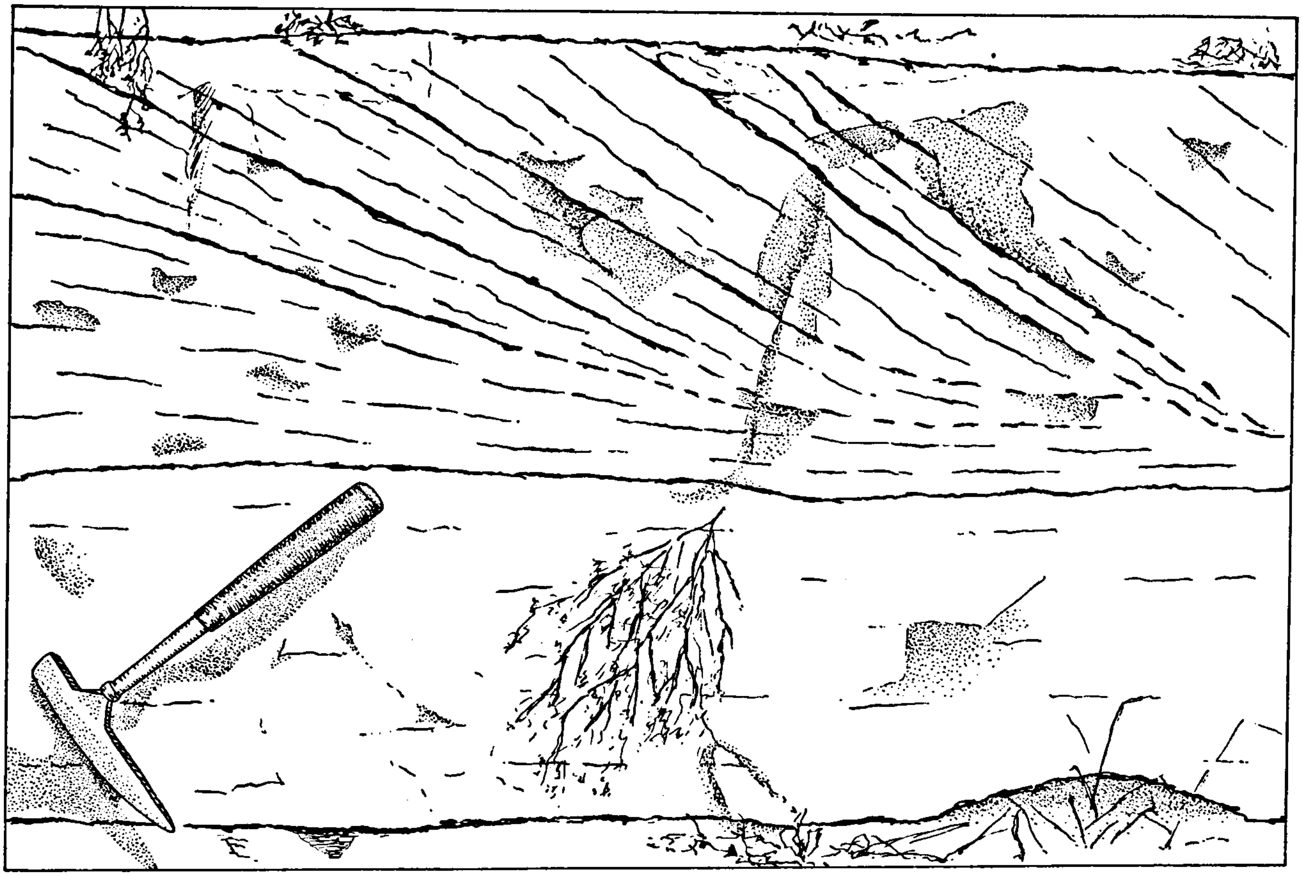


Fig. II-14. Large-scale cross-stratification at the top of a sandy calcarenite within the turbidite sequence exposed near Las Herrerías, Isábena Valley. Drawing after photograph.

According to Allen (1969) the various types of bedform are related to the depth of flow. In the case of dunes the greatest flow depth at which they may form is about 80 m. If this picture proves to be correct it may be concluded that the depth of the Vallcarga basin at the location of deposition of the large cross-bedded layers did not exceed this depth value.

Alternating with these layers in the Vallcarga sequence are coarse-grained, mainly parallel laminated, sandy calcarenites or calcareous quartzarenites (Fig. II-15) which are encountered in the Víu area (van Hoorn, 1969) (section 1) and in the Isábena Valley (section 12). Variations in flow regime must be held responsible for the presence either of large-scale cross-lamination or parallel lamination. Even structureless ungraded thick beds are encountered, which show an erosional base, frequently containing clay galls and rock fragments. The top boundary is extremely well-defined, not showing any transition into finer deposits. Flutes are quite rare, whereas grooves, crescent moulds and other tool marks are common, mainly at the somewhat channelling base. Current ripples of varying size are present at the top, with the same orientation as the tool marks. The thickness of these layers varies between 25 m in one case (near Torre la Ribera) to 5 cm, but is mainly between 30 cm and 1 m. The grainsize is up to coarse sand, although granules, pebbles and cobbles also occur. Clayey and silty material does not exceed 2% of the total rock volume. The properties just mentioned closely resemble those of fluxoturbidites (Dzulynski *et al.*, 1959), (Unrug, 1963); grain-flow or mass- flow deposits (Stauffer, 1967, 1968); proximal turbidites (Walker, 1967) or sandflows (Dzulynski and Walton, 1965).

The absence of grading, scour marks and the presence of tool marks, scour-and-fills, large-scale cross-bedding, the clean-washed texture, the sharply defined tops as well as bases, the great thickness, are all features indicating traction transport. The original velocity of the current appears to determine whether material is transported in a turbidity current or by tractional processes. This may be demonstrated by the alternation of layers displaying all the features of normal turbidites and layers with a large-scale cross-bedding, both of

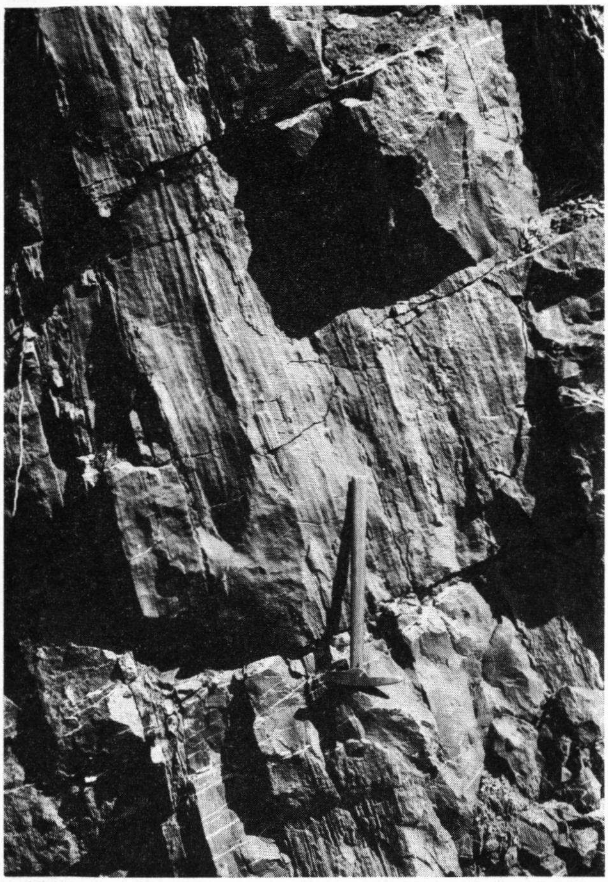


Fig. II-15. Coarse, parallel-laminated thick layers near Víu, W of Esera Valley (Photograph by P. J. C. Nagtegaal).

these types of deposits being of the same petrographic composition. These differences may also be explained as the result of different velocities at which auto-suspension[1] ceases (Bagnold, 1962).

Parallel-laminated interval

The parallel-laminated interval occurs in most turbidites of the Vallcarga Formation (Table II-1); at the top of limestone breccias, pebbly mudflow deposits, microbreccias, graded or non-graded sandy calcarenites, or constituting the base interval of many layers. The individual laminae are sometimes inversely graded, whereas in the entire interval the maximum grain size decreases towards the uppermost lamination.

Lamination is due to the concentration of different grain sizes. The coarser laminae are chiefly composed of grains of quartz or chert, fossil fragments and intraclasts, mixed with micrite, whereas micrite, clay, silty material and pelagic fossils are to be found in the fine-grained laminae. The mean grain size of the coarse laminae is quite uniform, varying between 0.16 mm to 0.10 mm (for quartz grains). Well-preserved microfossils and intraclasts may have a slightly greater diameter. The coarse parallel-laminated beds previously dealt with in connection with the graded interval show grain sizes of up to 0.8 mm throughout their entire thickness.

The lower parallel-laminated interval (b-interval of Bouma) is considered to have been deposited by a plane-bed phase of transport under conditions of an upper flow regime (Walker, 1965, 1967; Harms and Fahnestock, 1965). The origin of the lamination has been subjected to various different investigations. Lombard (1963) and Sanders (1965) advocate pulsations of the current velocity as the primary cause; Wood and Smith (1959) and Unrug (1959) offer an explanation based on the discontinuous density of the tail of the current which would cause sand from the denser part and clay from the part of lesser density to settle down; whereas Kingma (1958) and Mangin (1962) assume separate flows to be responsible. Walker (1967) suggests two mechanisms responsible for the formation of laminae in turbidites: either reworking by the current of previously deposited sediment, or primary deposition from the current. The former mechanism is argued by Hubert (1964), Hsu (1964) and Scott (1966) as the main cause of lamination.

These various opinions have been reviewed by Kuenen (1966) who attempted to explain the turbidite lamination by experiments in a circular flume. He postulates a "kind seeks kind" principle by which lamination is brought about by the tendency of material to concentrate according to its weight, density and shape. Laminae will therefore be formed by the concentration of particles possessing similar properties. This theory resembles the one proposed by Moss (1963) who also claims that each sedimentary lamina is characterized by one single population. The difference in composition to that of the next lamina is due to the rejective action of the former lamina, which refuses to accept particles not possessing suitable properties.

According to our own observations, concentration of particles in one lamina according to shape cannot be advocated, since angular to subrounded quartz grains,

[1] According to the auto-suspension theory, grain collisions cause a dispersive stress under conditions of a high shear stress during tractional transport in a high flow regime. When this stress decreases below a certain critical value no more dispersive stress acts upon the grains, leading to a sudden deposition of the transported material, the grain concentration being too high to permit of gradual deposition. Remaining material could be absorbed up into a turbidity current moving further on into the basin (Walker, 1967).

well-rounded Foraminifera and very angular, elongated shell fragments are found together. Weight and density may be a major factor but it is quite difficult to establish any relationships. For instance, two fossil fragments of different shape and size may have been of the same weight at the moment of deposition, since one may already be filled up with calcite cement while the other is still an empty shell.

The thickness of the coarse laminae is about three times that of the finer-grained laminae in the lower parallel-laminated interval, whereas this relation becomes inversed in the upper laminated interval. Here, fine laminae can be as much as four times the thickness of the coarser ones.

Most peculiar is the inverse grading from a fine lamina to a coarse one towards the top of the lower parallel-laminated interval sometimes observed, and the presence of coarse laminae in between and at the top of this much finer-grained interval (Fig. II-16). Grain sizes in these laminae are the same as those encountered in the preceding graded interval. They probably indicate an immature current in which not all coarse material had been able to move into the front but instead flowed with random distribution throughout the length of the current behind the greatest bulk of material, being deposited at a later stage of flow. This has been particularly well observed in layers with a lowermost non-graded interval showing evidence of deposition from a traction carpet. In these currents deposition occurs of the material available at that moment. Small scour structures, up to 1 cm in width and 5 mm in depth, filled up with coarser material indicate that, at least occasionally, variations in current velocity as advocated by some authors, occur, which may conceivably be local vortices. The presence of rounded intraclasts consisting of coarse-grained material of the graded interval in between the parallel-laminated interval may also be attributed to a higher local velocity followed by erosion of freshly deposited sediment. The development of lamination can better be observed in fine-grained (up to fine sand) limestone turbidites than in coarser ones. In the former case rather sharp boundaries between succeeding laminae are common.

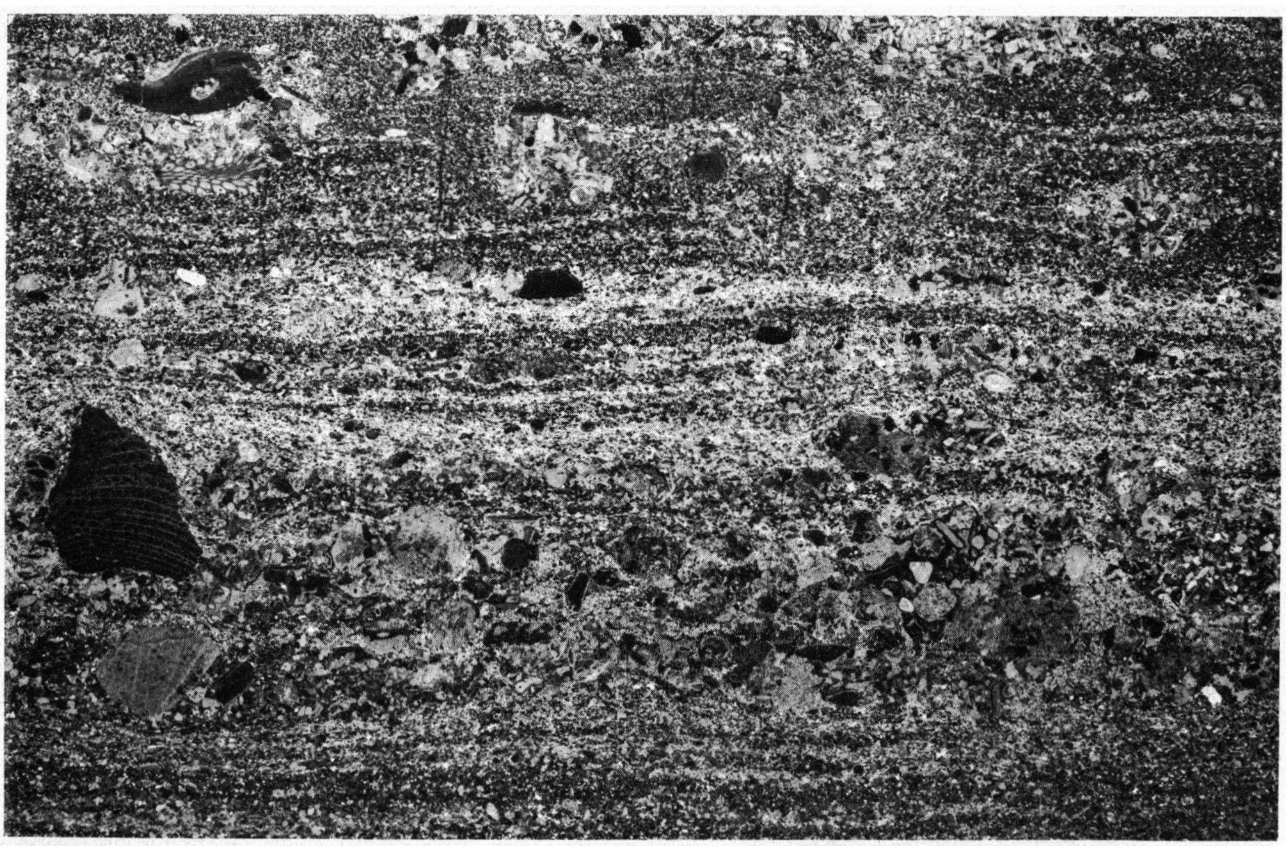


Fig. II-16. Coarse lamina in between a predominantly fine-grained parallel lamination, composed mainly of fossil fragments. Darker laminae are rich in clayey and silty material, in contrast to the lighter laminae composed of small fossils and quartz grains cemented by sparry calcite. 8 ×.

Our observations do not raise serious objections to Kuenen's "kind seeks kind" theory; on the other hand no additional support can be provided to improve his interpretation. Possibly an alternative explanation can be offered, accounting mainly for the rare development of inverse grading from one lamina to a following one. According to an interpretation given by Bagnold (1954), grain to grain collisions will occur in flows containing a high concentration of particles. In this case the smaller grains tend to concentrate near the deposition boundary which is the zone of greatest shear strain, in contrast to the larger grains which are to be found further upwards, where the shear strain is low. Such a mechanism may also account for the development of sharper defined laminae. The thicker fine-grained laminae are, at any rate, not easily explained by a "kind seeks kind" principle. The only suggestion made in this direction by Kuenen (1966), assuming that lutum could be transported and deposited as large flakes or small pellets, weighted by coarse silt (in the present case quartz grains, small pelagic Foraminifera and sponge spicules), requires further experimental demonstration.

c-interval

The third unit in a complete turbidite sequence has been termed c-interval (Bouma, 1962) and is characterized by the presence of current ripple lamination which may have been deformed into convolutions. In the present study the descriptions of ripple laminae, convoluted ripple laminae and convolutions formed from original horizontal lamination are presented together, since they all constitute the c-interval.

Current ripple lamination. – 30% of the layers studied in detail containing a c-interval show more or less distinct current ripple lamination (Fig. II-17). The origin and geometry of this small-scale cross-lamination in turbidites is now well understood as a result of the extensive studies by Walker (1965, 1967, 1969), Kuenen (1967) and Middleton (1967); detailed descriptions and further references may be found there. Our own observations indicate that the material composing the ripple laminae consists mainly of fine sand to coarse silt, which were deposited under conditions of a lower flow regime (Walker, 1965). Allen (1969) states that sedimentary structure and grain size may be correlated, which applies to most turbidites observed in the Vallcarga Formation. However, the fact should again be emphasized that, if deposition of a waning turbidity current is accepted, sedimentary structures will be formed under conditions of a flow regime prevailing at that moment, with a grain size available at that same moment (Fig. II-18). Consequently, vertical fluctuations in grain size may occur in a turbidite layer, depending upon the degree of maturity of the current.

Convolute lamination. – The presence of convolute lamination following Bouma's a- or b-interval is much more common than current ripple lamination (about 70% of c-intervals) in the Vallcarga turbidites. Their occurrence in the same formation in the Pobla de Segur area has previously been described by Nagtegaal (1963). His interpretation, based mainly on gravity-induced lateral movements during and shortly after deposition under influence of a false-bodied thixotropy, accounts for 65% of convoluted layers, especially if the distorted laminae are composed of silty material.

A similar deformation of horizontal lamination into convolutions has been recorded previously (Dzulynski and Slaczka, 1958; Nederlof, 1959; Holland, 1959; Sanders, 1960 and Dzulynski and Smith, 1963) and has been interpreted as the result of a shearing action of the current over the bottom surface, leading to irregularly distributed pressure and suction giving rise to convolute lamination.

Deformation of current ripples into convolutions formed by swiftly deposited sediment, as advocated by Kuenen (1953a), is apparent in only 35% of the convolute laminations observed. Layers showing this feature are somewhat coarser (mainly medium-grained to fine sand).

Pelitic interval

Walker (1965) defined the pelitic interval as the structureless mudstone following the top of a turbidite layer. Part of the mud was deposited by the turbidity current at a motionless final stage, part of it represents the normal sedimentation in the basin. The presence and distribution of the dominantly planktonic fauna

Fig. II-17. bc-turbidite showing climbing current ripples. In between the parallel lamination small current ripples are present. Top of the layer towards the left (Photograph in the Isábena Valley by P. J. C. Nagtegaal).

Fig. II-18. Turbidite layer grades from medium-grained sandy laminae to fine sand, turning again into medium grain sizes in the current-ripple interval, grading afterwards into coarse silt and fine sand. Concentration of coarse material in parallel laminae is higher than in current ripples. (natural size of sample).

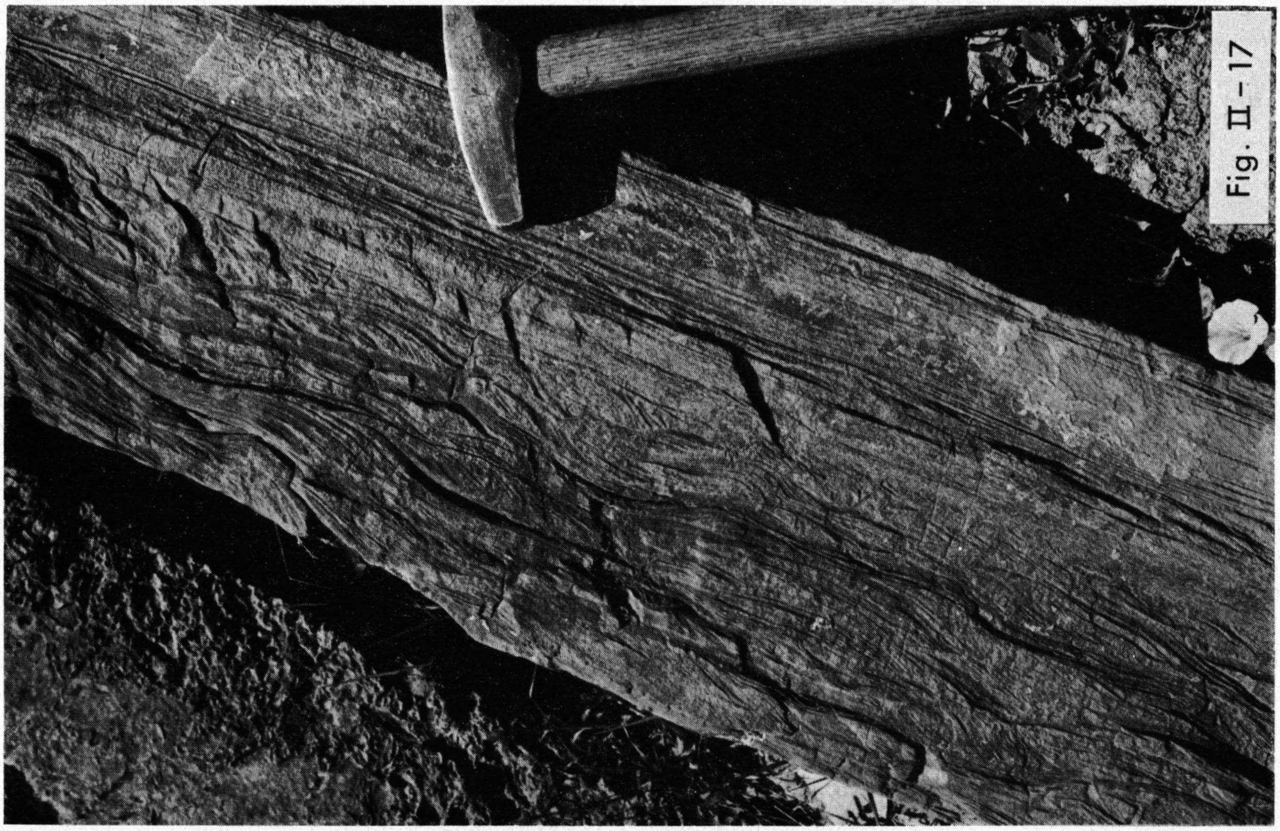


Fig. II - 17

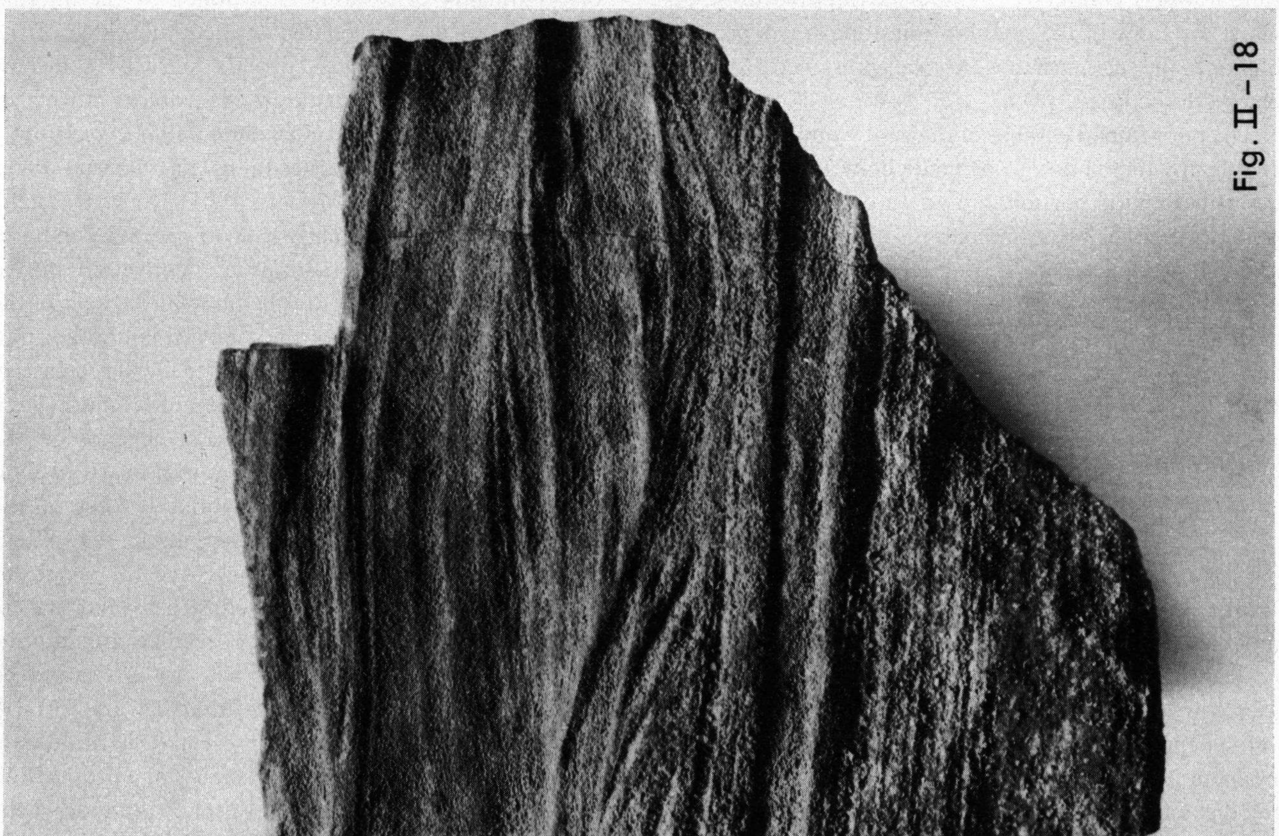


Fig. II - 18

sample	Ca Co3 content	> 75 μ	75-50 μ	50-37 μ	37-2 μ	< 2 μ
III	34.9	22.4 *	1.5	9.5	60.3	4.6
II	31.4	14.8	1.1	3.1	75.2	4.4
I	30.2	13.6	4.7	4.8	69.5	5.1

* Probably due to the higher concentration of micaceous material

Table II-2. $CaCO_3$ content and grain size of three pelitic samples between two turbidite layers. Sample I collected at the top of a turbidite, sample II 12 cm higher up and sample III beneath the next turbidite, 25 cm above the previous one.

found in these beds has been the object of extensive studies (Kuenen, 1964, 1968; Nesteroff, 1963; Weidmann, 1966; Meischner, 1964, 1967; and Brouwer, 1965, 1967). Our own observations tend to confirm Ksiazkiewicz's conclusion (1961) that these deposits are almost devoid of rich benthonic life. Only a small quantity (less than 1% of each sample) of Globotruncanidae, ostracods and sponge needles have been found. Consequently conditions in the Vallcarga basin were not very conducive to life. An exception has to be made for some of the nodular limestone layers present higher in the sequence in which well-preserved Echinoidea were recorded.

The poor faunal content of the pelitic interval did not enable any distinction to be made between a turbidite-deposited pelitic portion and an autochthonous pelitic portion. Neither grain size analysis nor $CaCO_3$ percentages showed any marked differences (Table II-2). No relationship was observed either between the thickness of a pelitic interval and that of the preceding turbidite. The most likely conclusion seems to be that most of the mud had already been deposited within the lower intervals (a, b and c), or had been swept away by next turbidity currents (Walker, 1965).

Proximality and distality of turbidites

Walker (1967) assigned a numerical value (the ABC index) to the proportion of beds beginning with a graded interval, a lower parallel-laminated interval or a cross-bedded interval. He states that this numerical value depends on the flow regime of the current at the moment of deposition, which is more or less related to the distance from the source of the current. Turbidites deposited near the source, called proximal, tend to be thick and begin with graded interval a, reflecting an upper flow regime, whereas in distal areas layers are thinner and begin with the cross-laminated interval c formed under conditions of a lower flow regime. The ABC index may therefore be used in conjunction with a basin analysis. For instance, a gradually decreasing parameter value in a longitudinal basin suggests one single source area, whereas a fluctuating ABC index indicates the influence of laterally located source areas. However, one has to be cautious regarding the degree of proximality or distality determined by the internal structures (which are related to a certain value of the flow regime). Turbidity currents start as slides or slumps triggered by earthquakes, tropical storms or tsunamis (Hayes, 1967; Coleman, 1968), or simply overloading of quickly accumulated material on a gently falling slope (Moore, 1961), and gradually increase in velocity, mainly due to a gravity-induced force. These slides or slumps are converted into a turbidity current by a hydraulic jump at the base of the initial slope when most of the material is brought into suspension by a turbulence generated by a decrease in mechanical flow energy (van Andel and Komar, 1969). The velocity of the turbidity current depends on the energy which the suspended grains impart to the fluid in the case in which the gravity-induced force on the sediment exceeds the fluid force supporting the suspended load (Bagnold, 1962). Hence, due to its greater weight, a large high-density current will flow faster than a small lower-density one, and proceed further into the basin until the critical settling velocity is reached by which the succeeding turbidite intervals are formed. An analogous conclusion was presented by Middleton (1967), who demonstrated the existence of a proportional relationship between the physical magnitude of the current, the settling velocity of the sediment carried in suspension by the current, and the velocity of the current head. The arrangement of the internal turbidite intervals formed by distinct flow regimes in a waning current is consequently determined by the magnitude of the current and may be relatively independent of the distance to the source area. The presence of turbidite tails within coarse-graded proximal-like layers may therefore be logically explained, as well as tractional deposits which may be interpreted as thick slides with a density not permitting of the transition into a turbidity current by a hydraulic jump.

Concluding remarks on the internal structures

The sequence of internal structures observed fits quite well into Bouma's conception of a deposition by a diminishing turbidity current. Some slight variations as shown in Fig. II-17 do not alter the general lines but merely indicate that small fluctuations in the flow regime have to be taken into account. The figure showing the relations between layer thickness and type of the lowermost interval (Fig. II-19) demonstrates that most turbidites are not thicker than about 20 cm. Layers beginning with an a-interval may cover a greater range in thickness than layers starting with a current ripple lamination or convolution. Recently certain authors (Hsu, 1964; Hubert, 1964, 1967; Scott, 1966) have advocated normal bottom currents as a major agent in the emplacement of flysch-type sequences, assuming that deposition occurs grain by grain from traction currents with decreasing or pulsating velocities. In the opinion of the present author, many of the coarse-grained microbreccias may in fact have been deposited by a sudden freezing of such a traction carpet which may have been brought about by overloading of a normal turbidity current below a critical flow velocity or by the absence of the hydraulic jump, as previously indicated. This mechanism, however, does not exclude the presence of true turbidity currents depositing material according to the classical point of view. One could, therefore, state that each turbidite should be considered separately, as differences in the mode of deposition may occur throughout the entire sequence, depending upon the initial velocity and the quantity of material transported. Whether reworking by bottom currents of parallel-laminated intervals into climbing current ripples had taken place could not be determined, since no major discrepancy in flow directions (determined from sole markings) was observed. Reworking of the top of a graded unit into a deposit with parallel lamination or current ripple lamination (Walker, 1965) does not seem plausible, as there usually is a considerable jump in grain size between these intervals. The good preservation of fragile shell fragments of large dimensions in many graded layers also indicates transportation by a turbulent suspension rather than by a tractive bottom current.

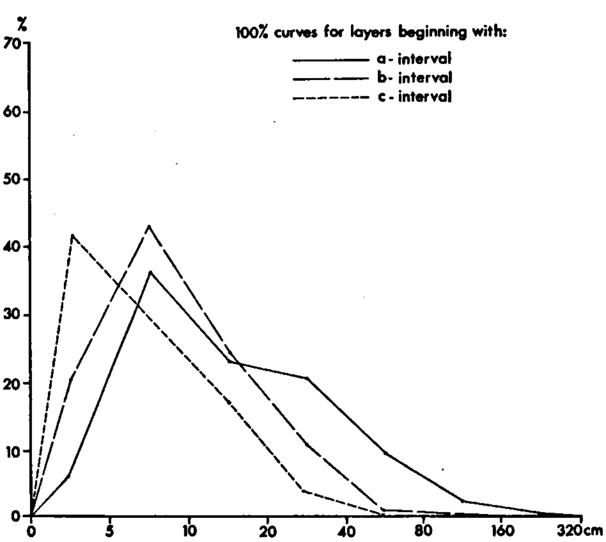


Fig. II-19. Frequency distribution of layer thickness for three groups of turbidites, each with a different lowest interval.

The only alternative which may be posed to the turbidity current hypothesis is the idea of the hydraulic jump proposed by van Andel and Komar (1969), and previously discussed. One could imagine that the density of a slide will determine whether the conversion into a turbidity current at the base of a slope will take place or not. In the latter case the slide will continue to exist, moving slightly further into the basin prior to depositing its material, which afterwards could be drawn into suspension or traction by a bottom current. However, this picture is highly speculative and has not yet been proved experimentally. Extremely high current velocities would be required for setting the coarse-grained fraction in motion, but such velocities appear to be quite rare (Kuenen, 1967). Possibly only fine-grained turbidites can be explained by such a mechanism.

Normal turbidity currents therefore seem the most logical way to explain most flysch-type beds showing a particular behaviour at the moment of deposition which may vary from case to case. The argument advanced by the advocates of normal bottom currents according to which turbidity currents have recently been recorded nowhere except some indirect evidence such as the Grand Banks turbidity current (Heezen and Ewing,

Fig. II-22. Metadepositional rupture of a slightly consolidated graded turbidite layer. Photograph taken along the road, N of Campo.

Fig. II-23. Slump sheet composed of fragmented turbidites and crumpled marls. Exposure near Campo.

Fig. II-20. Small-scale deformation of a silty turbidite layer between undisturbed marls.

Fig. II-21. Intensively folded marls and thin silty turbidites due to slumping. Exposure near Senz, Campo area.

1952) is not valid if we consider the roughly calculated time interval between succeeding turbidites. For instance, Sujkowski (1957), calculated a mean interval of 4000 years in the Polish Carpathians, and figures varying between 800 and 10,000 years have been given for Cretaceous sediments of the Sacramento Valley (Ojakangas, 1968). Sedimentation rate values based on ^{14}C determination in the western Mediterranean also show a mean value of one turbidity current each 4000 years (Eriksson, 1967).

DEFORMATIONAL STRUCTURES

Deformation of primary sedimentary structures or sediments may be due to vertical motion caused by loading, or to lateral displacements which may have been brought about by gravity-induced movements (Potter and Pettijohn, 1963). Some sedimentary disturbances, such as loadcasts and convolute lamination, have already been described in relation to sole markings and internal structures, whereas structures generated by deformation in areas at a certain distance from the site of deposition (pebbly mudflow deposits and olistostromes) are to be dealt with in the following chapter. No particular attention will be paid to present structures which have been thoroughly described in the geological literature, such as pull-aparts and sand volcanoes (Kuenen and Natland, 1951; Gill and Kuenen, 1957).

Slump structures

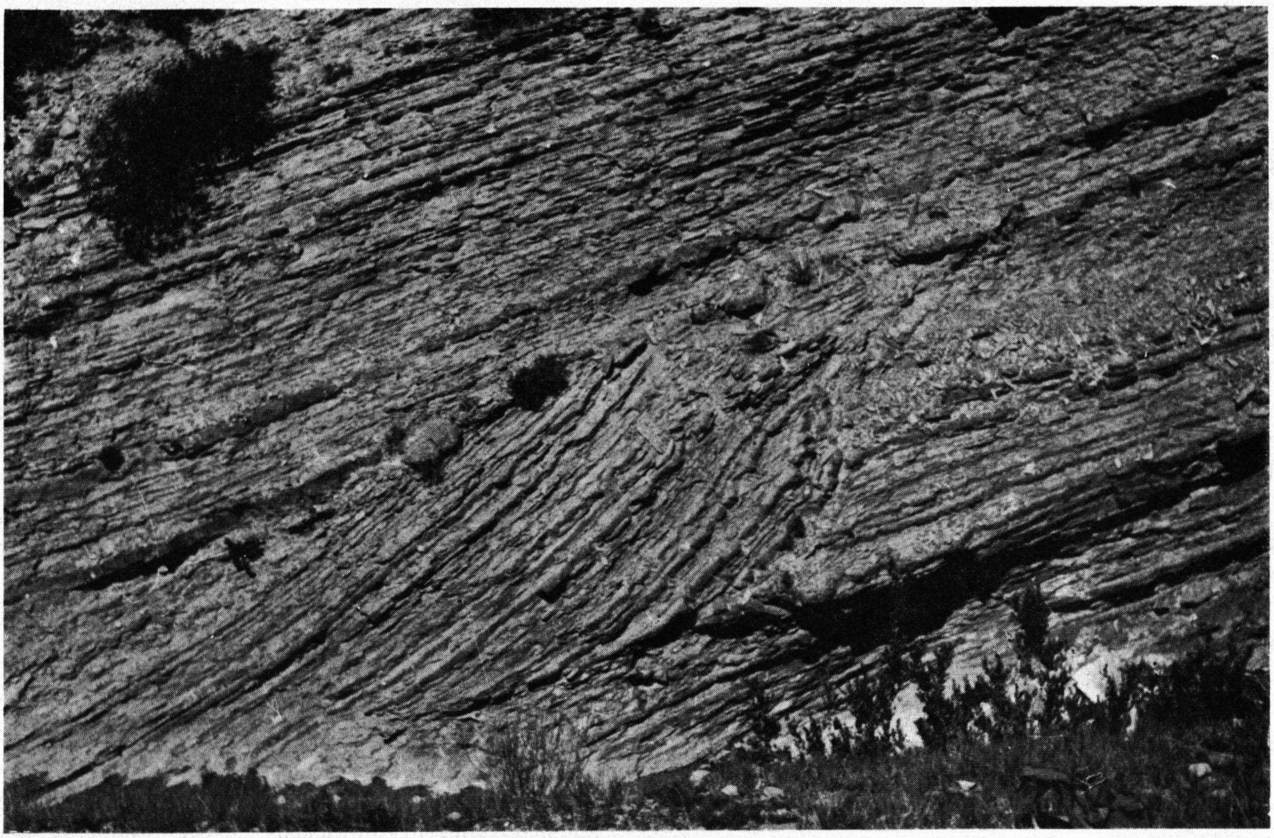
Slump structures are normally described as the result of lateral displacements generated by gravitational forces (Potter and Pettijohn, 1963). Their occurrence and genesis has been widely discussed elsewhere (see Potter and Pettijohn, 1963, p. 155). Examples in the Vallcarga Formation of the Pobla de Segur area have been recorded by Nagtegaal (1963).

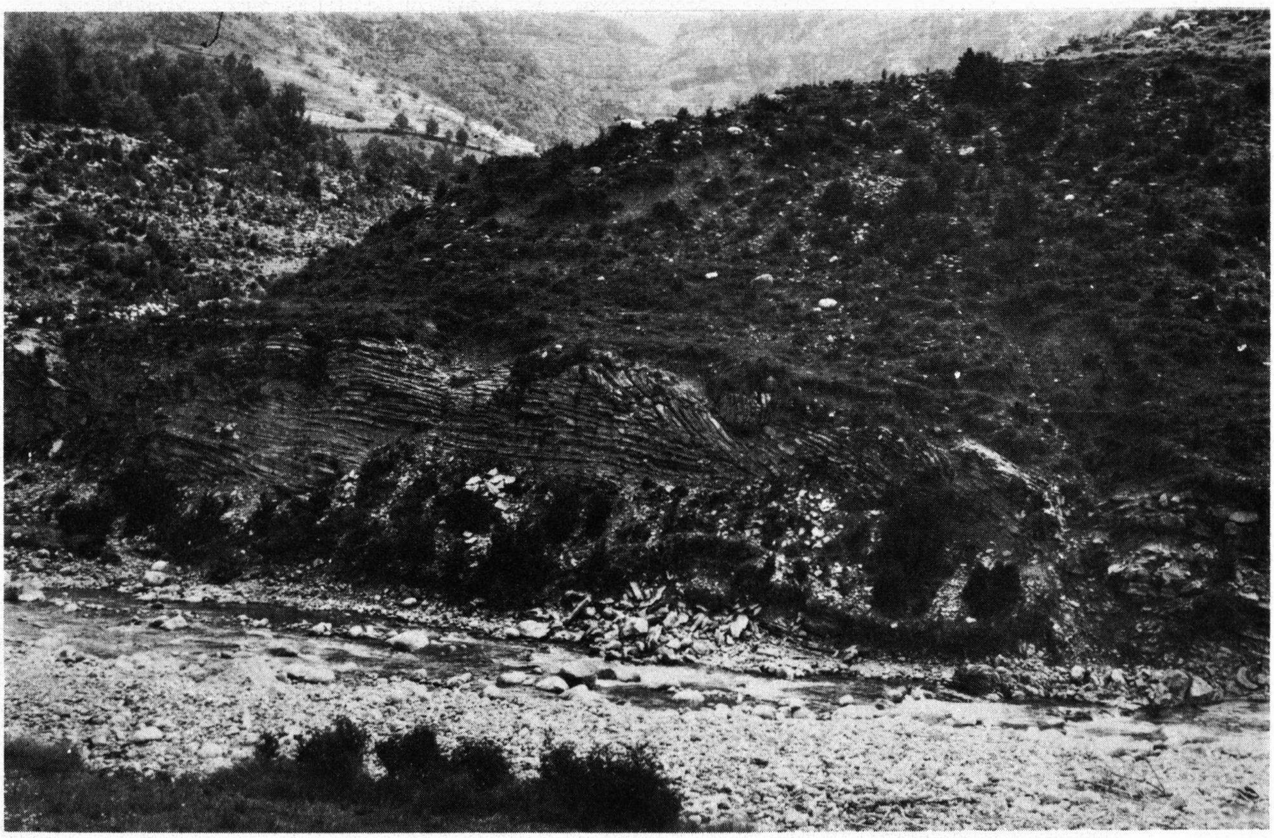
Slumps may involve one layer or an entire sequence of layers. In the former case all variations from deformation of a non-consolidated sediment immediately after deposition (Fig. II-20), and intensive folding of plastic material (Fig. II-21), to displacements of a slightly lithified sediment along metadepositional faults (Nagtegaal, 1963) may be observed (Fig. (I-22). This mechanism leads to a chaotic mixture of turbidites and pelitic marls, in which the more competent layers are broken into slabby fragments (Fig. II-23), or to the rotation of packets of strata, etc. unconformably lying in between the normal sequence (Fig. II-24). Evidently sediments were deposited under stable conditions, becoming partly consolidated before sliding due to tectonic tilt turned them into their present position. Similar intrabasinal features have been described by Laird (1968) in Silurian rocks of western Ireland, and by Berger *et al.* (1968) in the Lower Carboniferous of the eastern Iberian Chain. The total thickness of turbidite and marl sequences involved in one single slump mass may reach great dimensions (20–30 m), as in the Llert region (section 3), the Isábena Valley (section 12) or in the Ribagorzana area (section 13). Measurements of slump axes indicate an orientation of displacement more or less perpendicular to the main current direction (p. 141), which is in accordance with the results of earlier studies (Dzulynski and Slaczka, 1959; Murphy and Schlanger, 1962). The presence of crumpled turbidite/marl units in the slump sheets indicates that the slumps were triggered by local tilting of the trough floor, possibly due to small longitudinal faults or flexures (Kuenen, 1967). Tectonic instability of varying intensity obviously existed in the present basin. According to Gregory (1969) gravitational displacements may occur on slopes of less than 3° and are inevitable on slopes of 5°.

Slump folds have also been observed in some of the slid down nodular limestone slabs found in the Isábena and Ribagorzana Valleys (Fig. II-25) which may be described as olistostromes (Beneo, 1956; Marchetti, 1957). Sedimentary folding evidently occurred using intercalated marls as lubricants.

Fig. II-24. Rotated packet of strata in between a turbidite sequence, NE of Campo.

Fig. II-25. Sedimentary folds in an olistostrome exposed in the Isábena valley.





CHAPTER III

PETROGRAPHY AND DIAGENESIS OF THE VALLCARGA FORMATION

The petrographical study of sedimentary rocks concerns itself with two groups of aspects: those related to the original sediment and those related to diagenesis (Nagtegaal, 1969). These two categories will be dealt with in the same order in the present chapter, following a brief description of the lithologies of various sediment types.

A general review of the Vallcarga Formation shows two alternating lithological types, brought about by different modes of deposition: a slowly settling clay, rich in carbonate, which may be considered as the autochthonous sedimentation in the Vallcarga basin, and incidental incursions of gravity-induced turbidity currents, pebbly mudflows, breccias and olistostromes, all composed of clastic material originating from subaerial or submarine erosion from adjacent parts of the basin or its borders. A close study of the clastic components and their vertical and lateral distribution throughout the entire formation will consequently reveal a great deal concerning the lithological nature of the supply areas, and will also be useful in reconstructing the paleogeography. More than 700 samples were collected, chiefly from the basal portions of beds in which the coarsest components and the greatest variations in composition occur. Consequently, most data on the source areas may be obtained from these samples. Upper parts of beds (in the case of turbidites: Bouma's b, c and d intervals) are better sorted due to the flow conditions in which they were deposited. Only part of the total quantity of material transported in the current is therefore represented here, giving no additional clues on the clastic components in the supply area.

PRESENT LITHOLOGIES

The Vallcarga Formation consists of an alternation of resedimented deposits and marls. Individual components of these sediments will be described presently, as well as the diagenetic changes leading to the present lithology. It is therefore expedient to briefly describe these lithologies in order to afford an idea of the mean composition of various types of deposits building up the formation. Evidently, composition and texture reveal the type and the capacity of the transporting mechanisms, as well as the lithological components of the source area.

Calcarenites

Calcarenites are rather poorly sorted deposits forming the graded interval or the lower parallel-laminated interval of most turbidites. They are mainly composed of terrigenous detritus (quartz, chert, muscovite, feldspar), micrite intraclasts, fossil fragments and 30–60% matrix (Table III-1). The material building up the matrix consists of silty quartz, fossils, clay and microcrystalline calcite. In some levels of the Isábena area the terrigenous admixture may exceed 50%, so that the rocks should be classified as calcareous quartzwackes.

Calcisiltites

Calcisiltites possess more or less the same composition as calcarenites but vary in mean grain size, whereas the matrix content is much greater (more than 60%)

TABLE 1

ORIGINAL SEDIMENT			DIAGENESIS		
Texture	Detrital grains	%	Authigenesis	%	Texture
Matrix (micrite, clay, quartzsilt, fossils) 59.5 %	Quartz	7.9	Calcite	27.1	Cement 2.2 %
	Intraclasts	7.6			
Max. grainsize quartz 0.214 mm	Fossils	7.6			Several types distinguished: drusy, granular, fibrous or syntaxial rim
	Muscovite	2.1			
Max. grainsize rock fragments 0.9 mm	Wood fragments	1.3			
	Rock fragments	1.1			Recrystallization of micrite 17.5 %
Quartz: angular	Chert	0.6			
Rock fragments: subangular	Feldspar	0.3			Replacement of quartz 9.3 %
	Glauconite	0.3			
Matrix content increases toward the upper part of the layer	Tourmaline	0.1			

Table III-1. Vallcarga Formation: calcarenite. Mean values derived from samples 4.66L; 25.65E; 6.65E; 28.65E; 36.65E.

TABLE 2

ORIGINAL SEDIMENT			DIAGENESIS		
Texture	Detrital grains	%	Authigenesis	%	Texture
Matrix (micrite, clay and quartz) 64.1 %	Fossils	9.5	Calcite	12.0	Recrystallized micrite 5.5 %
Quartz : angular	Quartz	4.2			Replacing quartz 6.5 %
Max. grainsize quartz : 0.045 mm	Muscovite	3.0			
	Wood fragments	2.2			
	Pellets	2.0			
	Intraclasts	1.0			
	Glauconite	1.0			
	Chert	0.7			
	Feldspar	0.3			

Table III-2. Vallcarga Formation: calcisiltite. Mean values derived from samples 87.67V; E56; 31.65E; 29.66L.

(Table III-2). They form the uppermost intervals of turbidite layers (c-interval; d-interval when distinguished) or thin-layered structureless beds which may be regarded as turbidite tails.

Microbreccias

The term microbreccia refers here to coarse layers composed of more than 50% of rock fragments, whose mean size varies between 2 mm and 2 cm, and an additional content of large fossil fragments, intraclasts, quartz and chert. Microbreccias form the lowermost non-graded interval of turbidites, usually with an erosive, channelling base, always followed by a laminated calcarenite top interval.

Their most common occurrence is in the sequences of the Esera and Llert areas where they immediately follow the Camp Breccia Member, closely resembling the limestone breccias here in their composition. Texture and lateral extension of microbreccia layers vary according to the distance from the supply area. Microbreccias in the Esera area generally form limited channels at the base of parallel-laminated calcarenites, showing a relatively high content of matrix (Table III-3), whereas microbreccias deposited further from the supply area (near Llert) show a larger layer extension and are mainly clean-washed (Table III-4). The degree of roundness of rock fragments also increases in this direction.

TABLE 3

ORIGINAL SEDIMENT			DIAGENESIS		
Texture	Detrital grains	%	Authigenesis	%	Texture
Matrix (micrite and quartz): 17.0 %	Limestone rock fragments	52.5	Calcite	2.5	Cement 1.0 %
Quartz : angular	Fossils	17.5			Recrystallized micrite : 1.5 %
Rock fragments : angular	Intraclasts	6.5			
Max. grainsize quartz : 0.5 mm.	Quartz	2.5			
Max. grainsize rock fragments : 12 mm	Ophite fragments	1.0			
	Chert	0.5			

Table III-3. Vallcarga Formation: microbreccia of the Esera section. Mean values derived from samples E60; 4.65E; 1065E; E55.

TABLE 4

ORIGINAL SEDIMENT			DIAGENESIS		
Texture	Detrital grains	%	Authigenesis	%	Texture
Matrix (micrite): 2.0 %	Limestone rock fragments	50.5	Calcite	18.0	Cement
Quartz : angular	Fossils	17.0			
Rock fragments : subrounded to subangular	Intraclasts	4.5			
Max. grainsize quartz : 0.4 mm.	Ophite fragments	3.5			
Max. grainsize rock fragments : 9 mm.	Greenshale fragments	2.5			
	Quartz	2.0			

Table III-4. Vallcarga Formation: microbreccia of the Llert section. Mean values derived from samples 40.66L; 61.66L; 324L; 11.66L.

TABLE 5

ORIGINAL SEDIMENT			DIAGENESIS		
Texture	Detrital grains	%	Authigenesis	%	Texture
Matrix (micrite): 0.6 %	Fossils	46.0	Calcite	14.5	Cement
	Polycrystalline quartz	12.8			
Quartz : angular	Quartz	9.1			
	Intraclasts	7.0			
Polycryst. quartz : rounded	Non-calcareous rock fragments	6.2			
Rock fragments : angular	Chert	1.8			
	Limestone rock fragments	1.4			
Max. grainsize polycryst. quartz: 2 mm	Muscovite	0.4			
Max. grainsize rock fragments: 5.8 mm	Feldspar	0.2			

Table III-5. Vallcarga Formation: calcirudite. Mean values derived from samples 547I; 550I; 555I; 537I; 533aI.

Calcirudites

Calcirudites are thick (up to 4 m), well-bedded, parallel-laminated, large-scale cross-bedded or structureless coarse layers, composed of large fossil fragments (mainly deriving from a reefal fauna), rounded quartz and chert granules and rock fragments (less than 25%), cemented by sparry calcite and a small quantity of matrix (Table III-5). The mean grain size of the clastic components exceeds 2 mm. Calcirudite layers occur at certain levels in the Viú and Isábena areas (section 1, 10).

Breccias

Breccia layers forming the bulk of the lowermost part of the formation in the Esera area (Campo Breccia Member) consist mainly of limestone rock fragments (mean proportion of 95%). The original quantity of matrix is difficult to estimate since it has been partially or totally dissolved by pressure solution, but it probably did not exceed 1% of the total rock volume. Most breccias show grading at the top of the layer, and are followed by laminated calcarenite. Dimensions of the rock fragments may vary considerably (p. 108, and sections 1, 2, 3).

Pebbly mudflow deposits, olistostromes

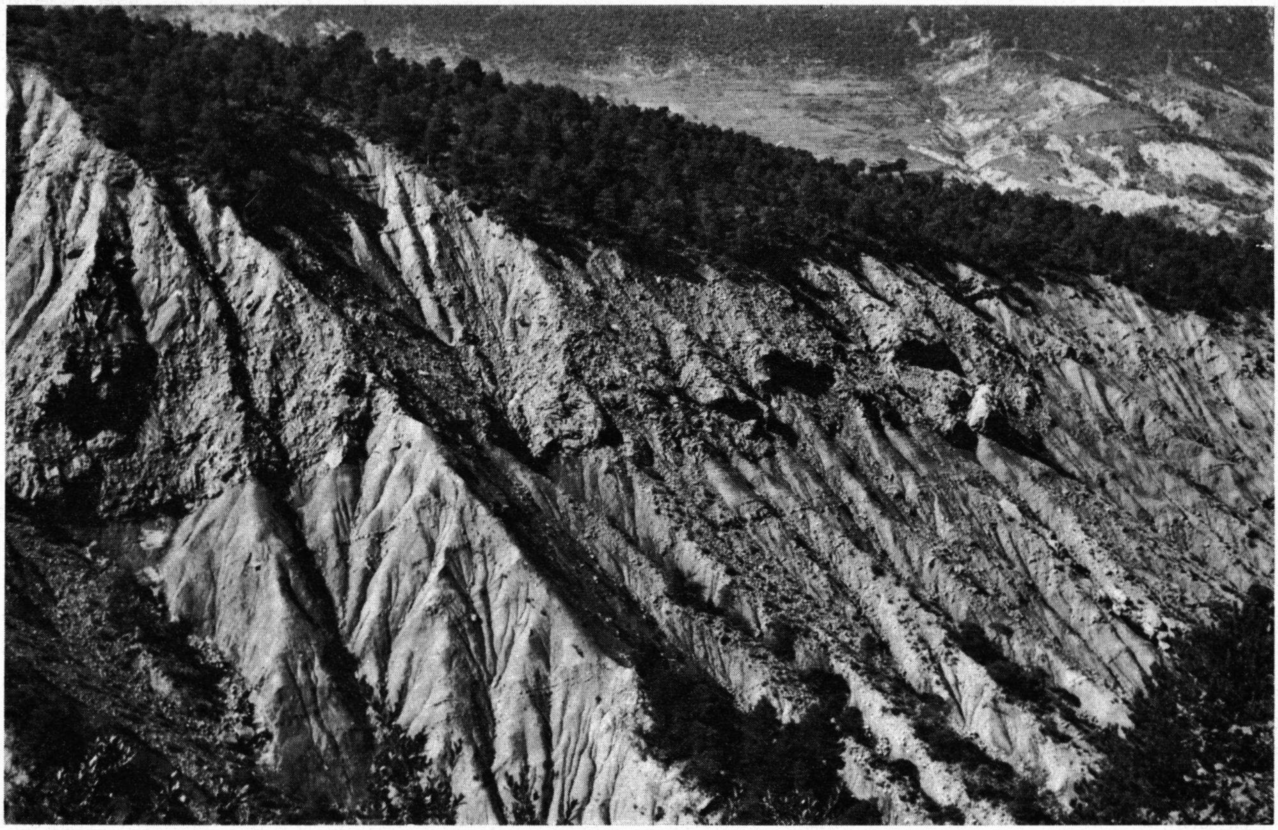
Pebbly mudflow deposits are composed of rounded limestone pebbles and cobbles lying in a muddy matrix (Fig. III-1) which has a composition analogous to that of the autochthonous marls. The total amount of rounded limestone rock fragments varies between 5% and 70%. These deposits are described in literature as pebbly mudstones (Crowell, 1957). In the present study the term pebbly mudflow deposit is introduced to illustrate its transport mechanism as well as the content of lithified rock fragments.

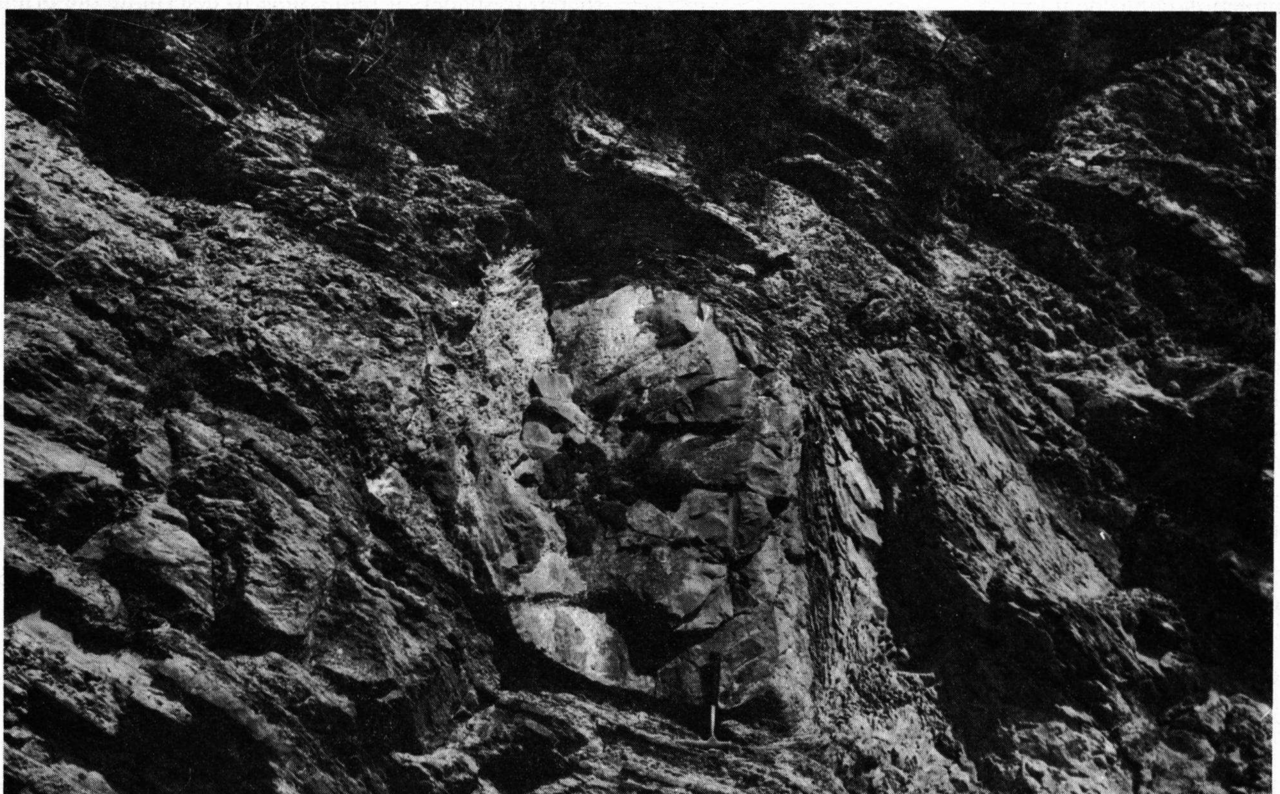
The limestone pebbles and cobbles occurring in mudflows in the lowermost breccia member of the formation closely resemble the Santonian components found in the breccia layers, and therefore suggest a similar supply area, but a different mode of transport. The breccias were presumably deposited by a kind of avalanching following a very rapid removal of angular components of the parent mass, probably due to sudden tectonic movements, whereas the pebbles and cobbles of the mudflows are rounded indicating that they underwent an abrasion before sliding down into the deeper basin within a plastic muddy mass. Components of mudflows in the breccia-lacking parts of the sequence are micrites with some pelagic fossils, similar to the nodular limestones found elsewhere in the Upper Cretaceous basin (Seira section, p. 116) or in the same Vallcarga Formation. Transformation of nodular limestones into pebbly mudflows has been recorded elsewhere (Laubscher, 1961). Even crumpled turbidite cobbles have been found here, suggesting disruption of these layers after deposition and incorporation of the torn-up sediment into a mudflow (Crowell, 1957).

Pebbly mudflows have been explained (Dott, 1963) as a plastic flow of a cohesive mass moving as a unit, and with a transporting power capable of carrying the limestone blocks. Their presence in the formation indicates that the liquid limit was not exceeded during movement, since in that case a turbidity current would have formed due to an increasing admixture of water, leading to the rupture of the cohesive forces within the muddy mass. The transporting power of mudflows is

Fig. III-1. Part of the sequence exposed at the western flank of the Turbón ridge. Alternation of thick pebbly mudflow deposits and marls. Typical badland erosion.

Fig. III-2. Limestone olistolith within thin turbidites and marls. Exposure N of Campo, along the road to Seira.





enormous, as evidenced by the occurrence in some layers of gigantic blocks. Wiersma (1965) described blocks of up to 50 m in diameter in the Pobla de Segur area whereas even larger blocks were found in the Esera region (p. 108). These components, however, were probably not carried up in a mudflow but merely slid down due to their own weight. In the Pobla de Segur area they occur beneath the uppermost marl member, forming a unit of deposits up to 250 m in thickness. These blocks, which slid down, have been called olistoliths whereas the entire deposit enclosing them is termed olistostrome. Extensive literature exists on the subject of the olistostromes, their origin, transport and deposition, which will not be fully quoted here (see Beneo, 1956; Marchetti, 1957; Flores, 1959; Jacobacci, 1965; Görler and Reutter, 1968). The olistoliths composing the olistostromes consist of massive limestone or nodular limestone/marls retaining their original stratification. The fossil content of these limestones indicates an Upper Cretaceous age. Plastic deformation is usually absent, suggesting that the olistoliths were already lithified during transport (Fig. III-2). The olistoliths of various sizes as well as the limestone pebbles in mudflow deposits are distributed at random in the muddy matrix, showing no features such as imbrication or tool marks from which a flow direction could be inferred.

Marl

Bluish-grey marls constitute the autochthonous sedimentation in the Vallcarga basin. However, part of it may have been laid down during the ultimate depositional stage of a turbidity current, or by slumping processes. Compositional analyses from 37 samples show that about 40% of the marls is composed of calcareous matter, mainly microcrystalline calcite with some additional sponge spicules and pelagic Foraminifera. It was not possible to determine whether the micrite was detrital or had originated as a primary precipitate. The non-calcareous fraction of the marls may range up to 60%. According to grain size analyses (37 samples), the average clay content is 18.6%, the average silt content 22.9% and that of fine sand 14.2%. Mineralogically it consists of illite, quartz and muscovite.

COMPOSITION AND PROVENANCE OF THE ORIGINAL SEDIMENT

Terrigenous, non-carbonate minerals

In the Vallcarga Formation this group is chiefly represented by the most resistent minerals: quartz, muscovite and chert, attaining grain sizes varying between the silt and the granule fraction. They rarely constitute more than 25% of the total rock volume. Some minor accessory minerals such as zircon and tourmaline as well as biotite and feldspar occur too, but usually form less than 3% of the rock composition.

Four types of quartz grains may be discerned: non-undulatory angular grains; angular grains with an undulatory, strain extinction; rounded polycrystalline quartz grains; and large fragmented idiomorphic quartz grains. The former two types occur throughout the entire formation in a more or less equal proportion, whereas polycrystalline quartz grains are restricted to certain distinct levels in the Viú area (section 1) and the Isábena region (section 12), mixed with the former two types. Grain size analyses show that the dimensions of non-undulatory quartz grains extend up to the fine sand range (<0.25 mm), whereas undulatory quartz grains may be somewhat coarser, extending up to the coarse sand range (<1.0 mm). Polycrystalline quartz occurrence may be classified between the very coarse sand fraction and the granule fraction (1.0 – 4.0 mm). Detailed studies by Blatt and Christie (1963) and Blatt (1967) have shown that the presence of undulatory extinction in clastic quartz grains indicates a wide range of igneous and metamorphic source rocks, and therefore renders it impossible to determine the origin of sediments.

Three types of polycrystalline quartz grains are common in the Vallcarga Formation:

a. a great number of fine-grained quartz crystals composing one single polycrystalline quartz grain
b. a number of coarse quartz crystals forming one polycrystalline quartz
c. polycrystalline quartz grains composed of quartz crystals with a bimodal size distribution

According to Blatt (1967), the first and third types derive from gneisses and schists, whereas the second type occurs in igneous rocks. A metamorphic source rock for the finely divided polycrystalline grains may be proved by the inclusions of staurolite. However, nothing can be stated concerning the direct provenance of quartz grains, since recycling may have played an important role. The difference in the degree of roundness between individual quartz and polycrystalline quartz (Fig. III-3) shows that the latter type of grains has probably been recycled, extensively reworked in the basin of deposition or transported over large distances (Kuenen, 1959; Hatch and Rastall, 1965),

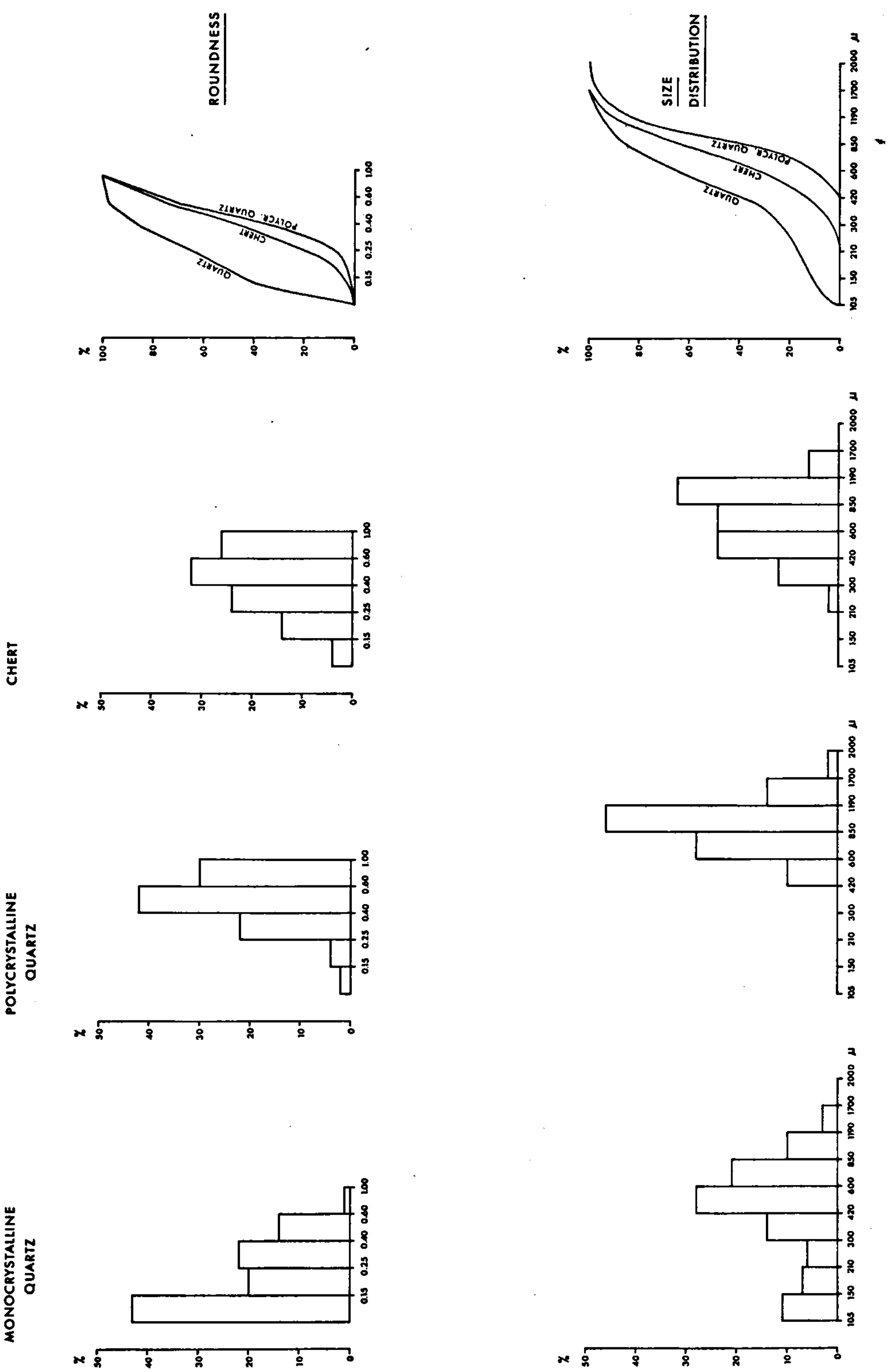


Fig. III-3. Roundness and size distribution of monocrystalline quartz, polycrystalline quartz and chert fragments in non-graded turbidites. Mean values of 100 readings in thin sections: 527I, 528I, 547I.

whereas the angular monocrystalline quartz grains may have been brought about by a first cycle from a crystalline source area. Size distribution analyses (Fig. III-3) also suggest different original sources, since the monocrystalline quartz grains occur in a wide range of grain sizes (median diameter 0.5 mm), whereas the size distribution of polycrystalline quartz grains shows a higher degree of sorting (median diameter 0,9 mm); this is also suggested by the vertical distribution of the grains in the formation. As was previously indicated, monocrystalline quartz was supplied from the beginning of deposition of the Vallcarga Formation, whereas polycrystalline quartz appears halfway the formation in the Viú area (285 m above the base of the section) and the Isábena Valley (at an altitude of 990 m in the section), and is almost exclusively associated with debris of reefal Campanian fauna: Bryozoa, Algae, rudists, corals; large benthonic Foraminifera (Orbitoidea) and algal clasts (Fig. III-4). Since the base of the section in the Isábena Valley is much older (lower Santonian, Souquet, 1967), the supply of polycrystalline quartz presumably started simultaneously in both regions. A small quantity of polycrystalline quartz associated with the same fossil groups has also been observed in coarse-grained turbidites at 1000 m above the base of the Ribagorzana section, corresponding in age to the exposures previously mentioned. A supply area for these quartz components in the Isábena and Ribagorzana sections is probably to be found in a southerly direction, where the base of the transgressive Upper Cretaceous, deposited near the basin border, is composed of quartz-conglomerates containing a large quantity of rounded polycrystalline quartz (p. 130). Polycrystalline quartz components laid down in the Viú area were transported by turbidity currents flowing towards the NE. In a south-westerly direction no exposures are known in which a large quantity of quartz was deposited there in Campanian times (p. 121). The quartz possibly originated from the north-westerly Salinas area (p. 121) and was transported to the southern part of the basin, from which it was carried further by turbidity currents.

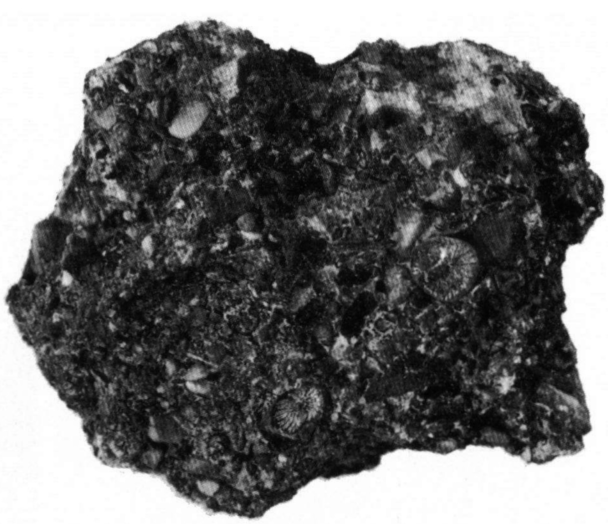


Fig. III-4. Coarse-grained base of a turbidite layer located at 990 m above the base in the Isábena section. Corals, Bryozoa, algal clasts and shell fragments form the bed, together with rounded polycrystalline quartz granules. Natural size.

Large idiomorphic quartz grains are sometimes present in turbidites of the Llert and Isábena regions, apparently originating from the Upper Triassic outcropping in the southern part of the basin (p. 124). They are also associated with debris of a reefal fauna of Campanian age, and occur at the same level as was previously indicated for the polycrystalline quartz (in the Llert area, at 350 m above the base of section 3).

Chert grains occur throughout most turbidite layers of the formation, mainly as medium to very coarse sand grains, attaining a degree of roundness between that of the individual and of the polycrystalline quartz grains (Fig. III-3). No provenance criteria could be established. Angular feldspar grains never exceed 1% of the rock volume. The predominant feldspar (about 80% of samples measured) is andesine or oligoclase (25–40% An), and additionally some microcline. This agrees with the main feldspar constituents of granodioritic rocks of the axial zone of the Pyrenees (Mey, 1968).

Muscovite and biotite rarely exceed 2% of the total rock volume and occur in almost every turbidite of the Vallcarga Formation, whereas heavy minerals such as tourmaline and zircon, both with a well-rounded shape, are found very sporadically.

Clay minerals

X-ray analyses of 11 samples show that clayey material present in the matrix of turbidites (estimated at 5%) or constituting about 18% of the marls is composed of illite and traces of quartz, muscovite and chlorite. It is well-known (see e.g. Fairbridge, 1967) that illite appears to be the most stable of the clay minerals in a marine environment, here being, however, no evidence of the formation of illite *in situ* in a marine environment (Griffin *et al.*, 1968). In general, clay mineral distributions depend on climatic zoning and not on the hydrochemistry of the basin (see e.g. Rateev, 1964). Weaver (1967) states that montmorillonite, rather than illite, was formed in Mesozoic oceans. According to Müller (1967), it is only after deep burial that potassium ions combine with clay minerals to form illite, which occurs at a depth of burial of about 5000 m. It is quite improbable that such a thick sediment sequence has actually overlain the Vallcarga Formation. At present, illite is mainly found in the oceans at mid-latitudes (Biscaye, 1965). It is well known that illite is absent in tropical soils formed by a primary weathering cycle. Its presence in tropical seas is possibly the result of the breakdown of montmorillonite and kaolinite (see e.g. Fairbridge, 1967).

These controversial remarks show that the overwhelming presence of illite in the Vallcarga Formation is a puzzling problem. Conceivably the illite is partly detrital, originating from older rocks such as the Devonian in the Pyrenees in which it occurs widely (Roberti, pers. comm.), or was partly brought about by the transformation of clay minerals under high pH and Eh conditions of the depositional environment. This agrees with the findings of Millot (1953), that ancient limestones formed under such physico-chemical depositional conditions are generally associated with illite.

Traces of chlorite found by X-ray analyses may either have been formed as a product of the recent weathering of illite or as a primary mineral deriving from metamorphic or sedimentary rocks (Biscaye, 1965). Both possibilities may fit here, though a transformation from other clay minerals does not appear very probable, as chlorite is considered to be a typical mid-latitude clay mineral (Griffin *et al.*, 1968), therefore not likely to occur in the warm, humid conditions prevailing during the deposition of the present formation.

Another clay mineral found exclusively in the graded basal part or in the lower parallel-laminated interval of turbidites of the formation is glauconite. It may even occur in the form of small pellets in the matrix of limestone breccias and microbreccias, but is not present in the marls, suggesting an allochthonous origin in the Vallcarga sediments. The type of glauconite most commonly found is that of spheroidal or even ovoidal grains with a random microcrystalline internal structure (Triplehorn, 1966). Sometimes squeezed grains were encountered, brought about by compaction in between detrital particles, suggesting that the glauconite was still plastically deformable at the moment of deposition. Two types occur:

a. glauconite grains distributed throughout the matrix

b. glauconite filling up or replacing biogenic material, such as Foraminifera, sponge spicules and fragments of pelecypods and echinoderms.

The first type of glauconite encountered in practically all turbidites of the formation at a rate of 5 grains in each sample, occurs relatively frequently (more than 30 grains counted in each sample) in intervals lying between 345 m and 990 m above the base of the Isábena section and between 680 m to 980 m of the Ribagorzana section (section 13). The grain size of glauconite was noted to be slightly smaller than or equal to the grain size of the quartz grains (Fine to coarse sand). Glauconite replacing fossils is sporadically found throughout the formation, not reaching distinctly higher concentrations at certain levels.

As was stated an autochthonous formation of glauconite in the Vallcarga Formation does not seem very probable, since its occurrence is restricted to the lower part of turbidite layers, while clay and ferromagnesian minerals from which it may have been formed (Hatch and Rastall, 1965) occur throughout the entire bed, moreover, its absence in the marls also suggests that no submarine alteration of clay minerals took place. A primary precipitation of glauconite appears highly improbable, since in that case it should be expected in the slowly settling marls and not in sediments reflecting conditions of a high deposition rate such as turbidites or limestone breccias. It may therefore be concluded that the first type of glauconite is of allochthonous origin, and was carried into the Vallcarga basin by turbidity currents. Its relative abundance suggests that at least part of the basin supplying material for the turbidity currents was characterized by a low rate of sedimentation (Müller, 1967). Glauconite is normally formed during a preburial stage of diagenesis in areas of slow deposition, commonly in shelf areas (Fairbridge, 1967) with little or no detrital sedimentation (Porrenga, 1967). It may also occur at the change in gradient between a shallow shelf and a slope such as on the Orinoco shelf (van Andel and Postma, 1954) or in the Niger delta (Porrenga, 1967). In the present basin, such a location for the formation of glauconite may tentatively be sought somewhere between the shallow shelf deposits exposed in the southern part of the Pyrenees (Chapter IV) in which no glauconite was found, and the turbidite basin in the north.

Glauconite replacing or filling up biogenic material also seems to be of allochthonous origin at the place of deposition, but was probably formed within the basin.

All stages in the formation of glauconite may be observed here, from pigments of glauconite developed in empty shells to glauconite with an internal texture suggesting replacement of sparry calcite cement.

Rock fragments

Rock fragments occur most commonly in the lower part of the Vallcarga sequence, especially in the Esera area where they form a 385 m thick member of breccias; however, they decrease in size and frequency higher up in the formation where they are found as components in the lowermost turbidite interval (forming microbreccias), in pebbly mudflow deposits or in olistostromes.

Limestone rock fragments. – The term limestone rock fragments has been used here to designate all fragments deriving from the erosion of lithified carbonate sediments and which may vary, according to their size, from small limeclasts to olistoliths. The limestone rock fragments deposited in the Vallcarga basin originate from sediments of which it may only be concluded that they are not older than Cenomanian and cannot be younger than Santonian, except sporadically occurring cream-coloured dolomitic limestone fragments closely resembling those found at the base of the Upper Triassic in the southern Pyrenees. With a few exceptions no exact age could be determined for the Upper Cretaceous rock fragments since they all lack specific time-markers, mainly showing an abundant fauna of benthonic organisms characteristic of the Cenomanian to the Santonian. Petrologically they may all be classified as reddish, greyish, black or cream-coloured bioclastic limestones. The rocks from which they originated were probably all deposited in a shallow environment as is suggested by the faunal content (Bryozoa, corals, rudists, Algae, Foraminifera such as Miliolidae, Orbitolinidae, etc.). Consequently, no distribution chart could be made of the vertical and lateral occurrence of rock fragments of a specific age in the Vallcarga Formation. The only rock fragments whose source may be established are pelagic micrites deriving from the immediately underlying Aguas Salenz Formation in the Esera Valley, which occur in most breccia layers of the Campo Breccia Member, and fossiliferous micrites with an abundant fauna of fissurines and Globotruncanidae belonging to the Turonian part of the Baciero Formation.

According to Souquet (1967), most limestone rock fragments may be recognized by comparing them with Cenomanian-Santonian formations outcropping in the Cotiella area (about 7 km to the NW of Campo). However, our own comparative investigations suggest that many other rock types occur which are quite different from those formations. A lithological comparison made with continuous Upper Cretaceous sequences exposed to the N and the E of the Esera Valley, as well as with outcrops in the southern part of the central Pyrenees (Chapter IV) showed that none of these specific areas could have provided coarse material for the Vallcarga basin. Consequently, a source area of the limestone rock fragments has to be sought in a westerly direction (see Chapter V).

A close study of the shape and dimensions of these rock fragments shows a variation between enormous olistoliths with a maximum observable diameter in one particular case of 1000 m (2.5 km NW of Campo) to more or less subangular or even subrounded granules in microbreccias. The largest rock fragments within the limestone breccias occur especially in the lowermost 100 m of the Esera section (section 2) where diameters of up to 15 m are frequently encountered. Large limestone slabs preserving their original upper and lower sedimentary boundaries also occur there. They were evidently carried over merely a small distance. A lateral decrease in size of the rock fragments is observable from the Esera section (section 2) to the Llert area (section 3), which together with current marks indicates the direction of transport. Since all rock fragments evidently originated from the same source area, the difference in shape and roundness between the limestone breccias constituting the lowermost part of the sequence and the microbreccias chiefly occurring higher up in the formation must be related to the type and the duration of transporting agencies acting upon them, and the intensity of the mechanical processes leading to the fragmentation of the lithified deposits from which they were derived. The sudden appearance of limestone breccias and subsequently of turbidites in the Upper Cretaceous of the southern Pyrenees is directly attributable to a short period of rapid sudsidence leading to the formation of a deeper basin. Erosion of older carbonate sediments along the probable fault-controlled slopes was the result of these relatively rapid tectonic movements (p. 139). The intensity of submarine erosion and resedimentation was evidently highest at the first moment of subsidence, slowing down later when subsidence was balanced by the rate of sedimentation.

Non-calcareous rock fragments. – Non-calcareous rock fragments do not occur as frequently as the limestone rock fragments but are persistently found in most turbidites and breccias of the lower part of the Vallcarga Formation (Fig. III-5). They may constitute as much as about 3% of the rock volume and show dimensions varying between 0.5 mm and 2 m, with generally a very angular shape. They derive from formations which may be more or less easily traced.

Ophite. The term ophite has been used in French literature to designate greenish basic rocks consisting mainly of plagioclase and clino-pyroxene with an ophitic texture, commonly present as large bodies in the Upper Triassic of the southern Pyrenees (Mey, 1968). Their emplacement may have occurred either simultaneously with the deposition of the Upper Triassic or later (Mey, 1968), but their occurrence as eroded angular fragments in the lower part of the Vallcarga Formation (Campo Breccia Member), or beneath the base of the Campanian limestones lying transgressively upon the Upper Triassic in the Sierras Zone near Graus (p. 124), indicates that they are at least pre-alpine. Ophite fragments occur as very angular pebbles and cobbles in limestone breccias of the Esera area, whereas in turbidites they reach sizes varying between 1 mm and 5 mm. The distribution chart (Fig. III-5), shows that ophites are more or less restricted to the Esera and Llert areas, whereas in the Isábena region they are scarcely found. Further to the east no ophites have been encountered within the formation.

Red sandstones and mudstones. Red sandstones and mudstones closely resembling those described in the Lower Triassic (Bunter Formation, Nagtegaal, 1969) are restricted to the Viú and Esera regions, where they occur in breccias of the Campo Breccia Member.

Their shape may vary from angular sandstone or mudstone fragments to blocks having retained their original sand/mud stratification (Fig. III-6). This is clearly observed in breccias in the Viú section (section 1), whereas in an easterly direction their frequency and

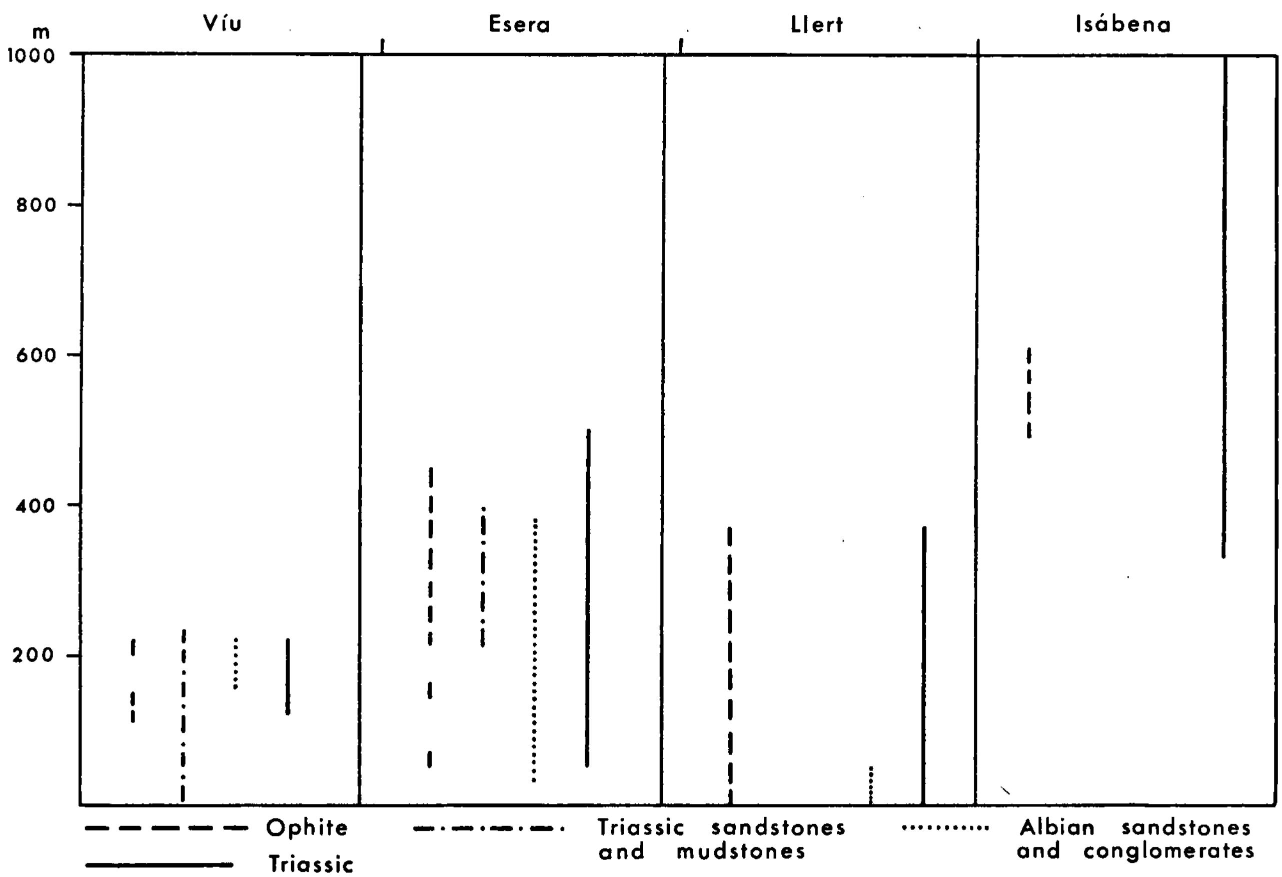


Fig. III-5. Distribution range of non-calcareous rock fragments in the lowermost part of the Vallcarga Formation.

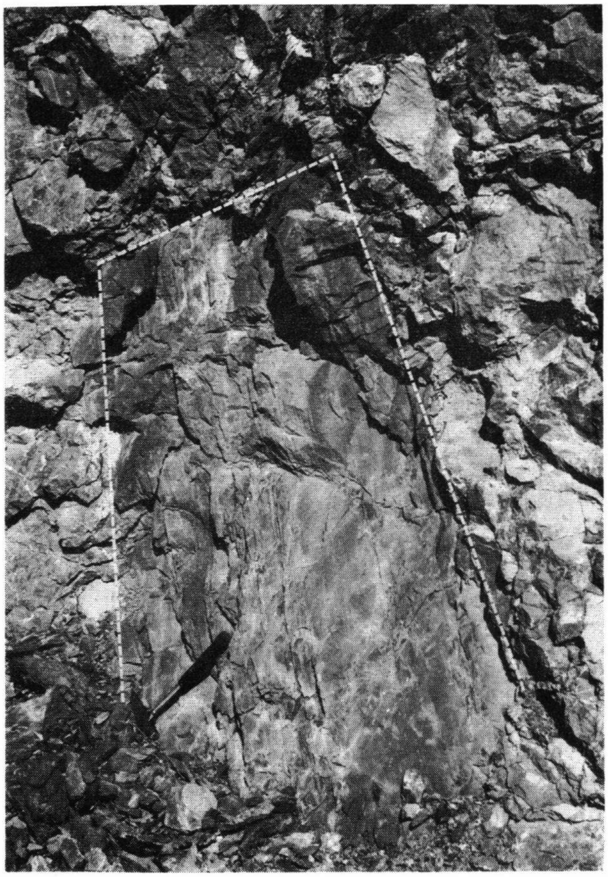


Fig. III-6. Large block of Triassic sandstone within a breccia layer at the lowermost part of the Viú section. Original stratification visible due to thin mudstone laminae.

dimensions decrease. These features indicate a very nearby source area.

Green mudstone. Small slabs of green mudstone occur persistently in the Campo Breccia Member and in many turbidites in the Esera and Isábena regions (Fig. III-5). Their sizes vary between 0.5 mm and 5 cm, generally showing a better degree of roundness than the rock fragments previously described. Their commonest shape is that of elongated rounded granules or pebbles. X-ray analyses show that the green mudstones are composed of quartz, sericite and chlorite, with some additional gypsum. They closely resemble the green mudstones which occur in the upper part of the Triassic.

Phyllite fragments. Small elongated slabs (up to 2 mm) of calcareous phyllite (Fig. III-7) have been encountered in sandy turbidites of the Isábena region between 1160 m and 1650 m above the base of the section (section 12). Their presence evidently indicates that they were brought about by a more or less rapid mechanical erosion and transportation from the source area, the warm and humid conditions which prevailed during Upper Cretaceous times (Lowenstam, 1964; Schwarzbach, 1961; present study, p. 131) making it very unlikely that the highly unstable phyllite grains could have been exposed to weathering on a land mass. Calcareous phyllites are reported in the Devonian of the Pyrenees and as fragments in sandstones of Upper Carboniferous and Permian age (Nagtegaal, 1969). It is not clear whether the phyllite grains of the Vallcarga Formation derive from one of these sources, since at present no outcrops of Devonian, Upper Carboniferous or Permian age are found at the southern margin of the Upper Cretaceous exposures from which the turbidites of the Isábena area were supplied (see Paleocurrents, p. 141).

Quartz conglomerate and quartzwacke fragments. Angular fragments of quartz sandstone and conglomerate are sporadically encountered in the Campo Breccia Member (Fig. III-5). The light-coloured quartz conglomerate fragments are composed of polycrystalline quartz pebbles, idiomorphic quartz grains enclosing anhydrite, gypsum clasts, green mudstone grains and fragmented manganese oxyde concretions. Large, broken Orbitolinidae suggest an Albian age (Souquet, 1967). Angular greenish or light-coloured quartzwacke fragments may also enclose large Foraminifera suggesting an equal age. They sometimes lack fossils but show persistent coal seams, thus resembling Albian rocks recorded elsewhere (Wennekers, 1968). Consequently, it appears that in the source area of the Campo Breccia Member Albian sediments occurred,

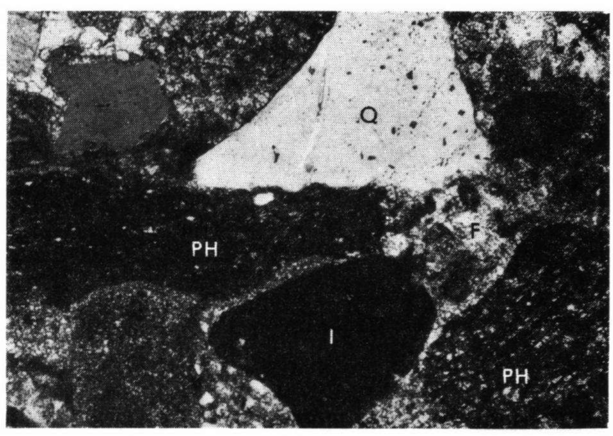


Fig. III-7. Calcareous phyllite fragments (PH) in a coarse-grained turbidite in which angular quartz grains (Q), fossils (F) and intraclasts (I) also occur. 25 ×.

probably of the San Martín Formation (Mey, 1968), directly overlying the Upper Triassic as is suggested by the fragmented idiomorphic quartz grains and gypsum clasts, which are typical of the Keuper, in the quartz conglomerate fragments. A similar sequence is only known to the W of the Campo area (Misch, 1934), whereas in a northerly and easterly direction a more or less continuous Mesozoic development is found (Souquet, 1967).

Fossils

The faunal content within the calcareous turbidites and in the matrix of limestone breccias of the Vallcarga Formation clearly suggests that it is of allochthonous origin, having been carried by turbidity currents from some distant source area into a deeper basin. This is well illustrated by the overwhelming presence of reefal detritus consisting of fragments of rudists, Bryozoa, Algae and well-preserved large benthonic Foraminifera (Orbitoidea, Orbitolinidae, Prealveolinidae), small benthonic Foraminifera (Miliolidae, Textularidae, Valvulinidae, Rotaliidae, etc.), Echinoidea fragments and ostracods within the largest part of the formation, which fossils are absent in the intercalated autochthonous marls in which only a poor fauna of pelagic Foraminifera occurs.

These fossils are sorted according to grain size in the turbidite layers (Fig. II-12). An additional admixture of pelagic fossils (mainly Globotruncanidae) and sponge spicules, which admixture was probably picked up by turbidity currents on their way into the basin, also occurs.

A close study of the occurrence in turbidites of the Vallcarga Formation of reefal fossils such as rudists, Algae, Bryozoa and Orbitoidea shows that they constitute about 50% or more of the total rock volume (Fig. III-8) of layers halfway the sequence (Isábena section, 990–1775 m; Viú section: 295 –520 m), associated with the first large-scale occurrence of rounded polycrystalline quartz. This corresponds in age with upper Santonian-lower Campanian deposits in the southern and western part of the Upper Cretaceous basin (Chapter IV) in which reefal organisms similar to those found in the Vallcarga turbidites were abundant, probably supplying material for many of the turbidity currents. Similar turbidites composed of calcareous fossil detritus have been described by

Fig. III-8. Calcareous turbidite composed entirely of bioclastic material (Foraminifera, Echinoidea, Bryozoa, rudists). 8 ×.

Meischner (1964) as allodapic limestones in the Carboniferous of Germany, and by Szulczewski (1968) in the Upper Devonian of Poland. Turbidites of the lowermost 80 m of the Ribagorzana section and the basal turbidites in the Isábena area do not contain any of the typical shallow water organisms previously mentioned, but only an abundance of pelagic Foraminifera, ostracods and sponge spicules, in a matrix of clay, micrite and quartz silt. They probably indicate the deposition of small turbidity currents triggered in the deeper part of the basin as is suggested by their composition, analogous to that of the intercalated marls. This origin is also suggested for the thin turbidites which occur sporadically in the uppermost marl member of the Vallcarga Formation, with a predominantly pelagic faunal content.

Intraclasts, clay galls and pellets

The term intraclasts was introduced by Folk (1959) to designate weakly consolidated or non-consolidated limestone fragments, sometimes plastically deformed, derived from a penecontemporaneous source which is within or related to the basin in which the clasts occur. They may be differentiated from clay galls by their better rounded shape, their size (not more than 0.7 mm, contrary to clay galls which may measure up to some cm) and composition. Intraclasts are merely composed of lime mud, sometimes armoured with quartz grains, whereas clay galls are composed of a mixture of micrite and clay. These differences may suggest that the intraclasts originated from the same source area as the turbidity currents, in contrast to clay galls which were

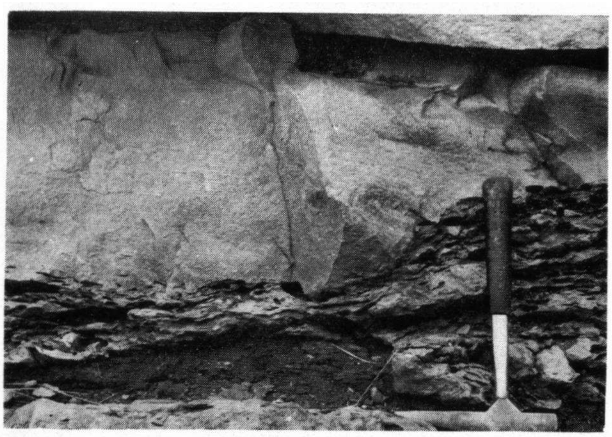


Fig. III-9. Basal part of a turbidite composed of a mixture of sandy material and clay galls (recognizable as hollows due to weathering). Top of the layer slightly laminated, but showing no grading. Head of the hammer lies on the top of a preceding turbidite.

eroded by the turbidity current in the deeper part of the basin, relatively near to the site of deposition, their composition being analogous to that of the marls. Intraclasts occur in all layers, decreasing in size in an upward direction in graded beds, and constituting about 5–10% of the total rock volume, whereas clay galls are present at the base of some coarse-grained thick turbidites (Fig. III-9), indicating a higher erosive activity of some turbidity currents.

Pellets are rounded or spherical micrite grains without internal structures, restricted to a uniform grain size of between 0.04 and 0.08 mm (Folk, 1959). Most authors conclude that they are fecal pellets; in the present case they must be of allochthonous origin, having been transported into the Vallcarga basin by turbidity currents. They mainly occur in fine-grained turbidites lacking any typical internal structures or in the uppermost laminae of graded layers.

Wood fragments, bauxitic components, collophane grains

Wood fragments are dark, opaque grains which are easily identified macroscopically in many siltstones constituting the upper part of turbidite layers. Their occurrence has frequently been noted in the Vallcarga Formation.

Brown, opaque grains closely resembling the ferrigenous grains found in lateritic soils of the southern margin of the Upper Cretaceous basin (p. 130) have been found in coarse-grained sandy turbidites of the Isábena section (990–1030 m). Their frequency in these turbidites is quite low, usually one or two grains in each sample.

Collophane grains are not restricted to certain deposits but occur in most turbidite layers of the formation, varying in size between coarse silt and granules. Their colour is black-brown, and they may enclose small fossils. This calcium phosphate mineral is formed by percolation of phosphoric acids through freshly formed limestones (Fairbridge, 1967). As they are restricted to turbidites it may be concluded that they are of allochthonous origin.

Carbonate mud

Fine-grained carbonate mud is the main component of the matrix of turbidites and an important component of the marls (up to 35%). The estimated grain size is less than 0.004 mm, though this is obscured by recrystallization. The origin of carbonate mud has long been a debated question (Chilingar *et al.*, 1967). This mud may be of biological, biochemical or physicochemical origin, being the result of normal chemical precipitation, the conversion of aragonite needles, the disintegration and abrasion of fossil debris and algal growth, or of a bacterial precipitation.

DIAGENESIS

The diagenetic processes leading to the present-day state of the sediment may be classified in three stages: an early diagenetic stage related to the depositional environment; an advanced diagenesis leading to lithification; an epigenesis comprising all diagenetic changes in the lithified sediment (Nagtegaal, 1969). In the present study only diagenetic features related to the first and second stages of diagenesis will be dealt with.

Diagenetic processes leading to the formation of SiO_2 components and feldspar

Authigenesis or replacement by silica or feldspar has only rarely been observed. Only a brief description of their features will therefore be given here.

Small idiomorphic quartz grains (largest diameter up to 0.25 mm) are sporadically encountered in the microcrystalline calcite matrix or cement of calcareous turbidites of the formation. As these grains have partly been replaced by calcite cement, their origin is most probably early diagenetic.

Relatively speaking, the most common occurrence of silica is as a replacement product in silty turbidites or even in limestone breccias. Primary deposition of aligned detrital chert such as described by Swarbrick (1967, 1968) was not observed. Banded chert occurs, replacing the carbonate matrix but leaving the fossil fragments more or less unchanged. This phenomenon was particularly well observed in one distinct level of the Viú section (160–175 m) and in some layers of the Esera area, whereas in other parts of the formation it is rarely encountered. A relatively large quantity of chalcedony and opal replacing rudist fragments and sponge spicules in these turbidites possibly acted as a silica source. Silicification of these fossils did not originate *in situ* but took place before these fossils were placed in their present depositional environment by turbidity currents. Otherwise other fossil fragments would also show silica replacement. The absence of silicified sponge spicules in the marls also indicates an allochthonous formation of chalcedony and opal.

Silica solution of siliceous skeletal debris followed by precipitation as chert or chalcedony has been reported by several authors (reviewed by Chilingar *et al.*, 1967). According to Newell *et al.* (1953), Walker (1962) and Krauskopf (1967), solution of silica is due to a particular physicochemical environment (high pH) during diagenesis. SiO_2 will therefore migrate to zones with a lower pH and precipitate there, whereas a reverse process is undergone by $CaCO_3$. The occurrence of banded chert restricted to some of the Vallcarga turbidites consequently proves the necessity of considering certain physicochemical variations during diagenesis, related to a relatively higher concentration of silicified sponge spicules in these layers. The present author believes that the occurrence of banded chert may also be indicative of interruptions in or slowing down of the sedimentation rate, leading to more or less permanence of a horizon of the bottom sediment in the CO_2-zone, in which horizon conditions of a lower pH prevail (Braun, 1964). Under these circumstances sufficient time was available to permit silica to be precipitated. At least thin levels of small reddish Fe-Mn-oxyde crusts in marls following cherty turbidites were found which may suggest a period of temporary non-deposition. Such features have also been described by Jurgan (1968).

However this be, it may be concluded that the silica precipitation occurred during an early diagenetic stage as a product of reactions occurring in the interstitial saturated fluids (Dapples, 1967). According to Folk and Weaver (1952), an initial close spacing of the centres of crystallization favours the formation of microcrystalline silica. An extreme example of silica replacement, which can only be explained by a rarely occurring, extremely saturated interstitial fluid, is shown in Fig. III-10. Chalcedony filling up the cavities between the clasts was conceivably formed as a primary precipitation, or closely following the sparry calcite cementation.

The formation of authigenic albite was noted in one particular instance, in a biomicritic turbidite layer (sample 479I). It may be related to a low pH during early diagenesis (Füchtbauer, 1957). According to the same author, feldspar is formed under reducing conditions in layers in which Na_2O is derived from the pore water. As its occurrence in the formation is so much restricted, it must be concluded that conditions favouring its formation were most rare.

Lithification of the Vallcarga sediments

Processes leading to the lithification of the deposits studied mainly comprise the diagenesis of calcareous

Fig. III-10. Chalcedony (CH) filling up pore spaces between clastic grains. Intraclasts (I) and fossil fragments (F) replaced by chert· 25 ×, crossed nicols.

mud to micrite and the precipitation of sparry calcite in pore spaces. (see e.g. Chilingar *et al.*, 1967). The $CaCO_3$ filling up the interparticle pore space was precipitated as ferroan calcite. Staining by the methods proposed by Dickson (1966) shows that most fossil fragments were already cemented by non-ferroan calcite at that stage, as well as the initial cement lining the pores. This is indicated by its sometimes fibrous appearance, growing away from fossils or intraclasts towards the pore centre, which suggests an early diagenetic lithification (Gevirtz and Friedmann, 1966). The introduction of iron during an advanced diagenetic stage is also indicated by the presence of syngenetic pyrite in the sparry calcite.

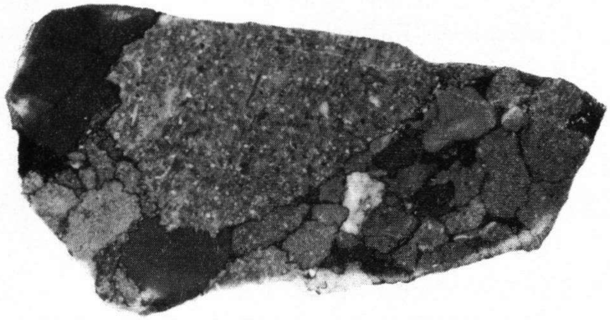


Fig. III-11. Stylolitic contacts between polymict limestone rock fragments in a breccia layer. Natural size.

Subsequent recrystallization of microcrystalline calcite took place, leading to the formation of microsparite (pseudospar). This process has been called neomorphism (Folk, 1965). A gradation was usually noted from microsparite lining the particles to large sparite crystals in the centres of the pores.

A typical feature present in some levels of the uppermost marl member of the Vallcarga Formation is that of slightly nodular primarily deposited limestones. These limestones may be due to a local decrease in the clay supply and consequently a relative increase in the carbonate mud deposition. An alternative explanation based on changes in water depth may be rejected since these changes correlate poorly with the $CaCO_3$ percentage of sediments (Smith *et al.*, 1968). The origin of the somewhat nodular layer boundaries is dealt with elsewhere (p. 118). A diminishing rate of sedimentation at the moment of deposition of the limestones is indicated by the finely distributed ferric oxyde crusts in the carbonate mud. A similar explanation may be favoured for certain levels of siderite concretions in the marls, which may be explained as hiatus-concretions (Voigt, 1968).

Replacement features

The replacement occurring commonly during diagenesis is that of detrital quartz grains by calcite. All stages of replacement may be encountered, from rim replacement to complete replacement forming ghost structures. A process similar to that of chert replacement may be inferred, based on a variability of physicochemical conditions (Walker, 1962). The only puzzling problem is to find out whether the dissolved silica was precipitated elsewhere, since silica precipitates are found only very locally in the formation. It was possibly carried into solution by sea-water, suggesting in this case that replacement took place during an early diagenetic stage. No evidence being available to support this view, the moment of replacement remains an open question.

Pressure solution, stylolites

Pressure solution phenomena are extremely frequent in coarse deposits such as limestone breccias (Fig. III-11; Fig. II-13), leading to stylolitic contacts between the limestone rock fragments, whereas in finer-grained deposits such as calcarenites, calcisilitites or marls these phenomena are rarely encountered.

According to the classical theory (Pettijohn, 1957; Trurnit, 1968) solution takes place due to the overburden pressure, depending upon certain factors as have been summarized by Park and Schott (1968, p. 188). In the case of breccias the depth of burial, the shape of the rock fragments and their manner of packing may have been important conditions for the formation of stylolites. Stylolitic contacts are observable surrounding the entire surface of limestone fragments. Partial solution of calcite occurred, bringing about stylolitic contacts in which a thin film of solution residue, composed of clay and quartz, and coated with iron oxyde, was formed.

Authigenic silica has sometimes been observed in the stylolitic seams indicating that the pore solution was rather undersaturated with respect to calcite so that silica, when present in solution, was precipitated and calcite dissolved (Park and Schott, 1968). Since stylolites are very common in the lowermost breccia member of the Vallcarga Formation, it is possible that at least some pressure was exerted due to the rapid deposition of succeeding breccia layers. However, this is not in agreement with the normal concept of a late-diagenetic origin of stylolites. Stylolites also occur in breccias present halfway the formation (Fig. II-13), in which the large quantity of marl composing the sequence there suggests a lower rate of sedimentation. This may indicate that the most important factor controlling the development of stylolitic contacts in limestone breccias was their manner of packing, since the pore fluid could have circulated more easily in these deposits than in finer-grained sediments.

CHAPTER IV

LITHOLOGY AND DEPOSITIONAL ENVIRONMENT OF UPPER CRETACEOUS FORMATIONS SURROUNDING THE VALLCARGA FORMATION

Data for a paleogeographical reconstruction of the Upper Cretaceous basin have been collected in areas surrounding the Vallcarga Formation outcrops (enclosure I), in which areas rock units of the same age were reported (Misch, 1934; Selzer, 1934; Souquet, 1967). Their significance for the overall basin picture is summarized in Chapter V, whereas the present chapter concerns itself mainly with the lithology and depositional environment.

UPPER CRETACEOUS SEQUENCE NEAR SEIRA
(Localities A and B)

About 8 km NNE of the Campo area, a thick limestone unit followed by grey mudstones is exposed along the Esera Valley (Fig. IV-1). This sequence has been dated by Misch (1934) and Souquet (1967) as being stratigraphically equivalent to the transition of the Aguas Salenz Formation into the Vallcarga Formation found 6 km southwards.

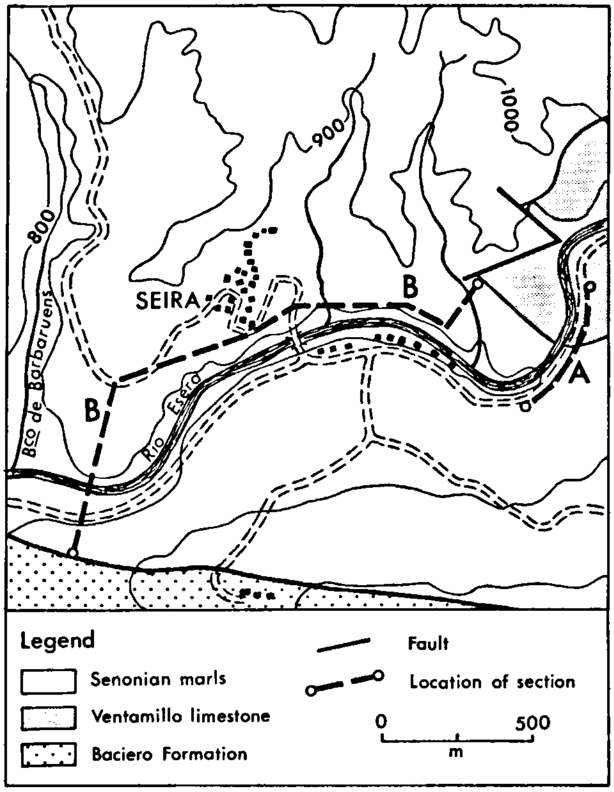


Fig. IV-1. Geological sketch-map of the Seira area (after Wennekers, 1968).

The general occurrence of *Lacazina elongata* M. Ch., *Monolepidorbis sanctae pelagiae* Astre and *Nummofallotia cretacea* Schlumb. indicates a Santonian age for the lower limestone series (Souquet, 1967). Souquet (1962, 1964) introduced the term Ventamillo Limestone to designate this unit, which consists mainly of grey bioclastic limestones with several consistent silex horizons (Wennekers, 1968). The upper 30 m of this unit show a gradual transition from biomicritic limestones, containing an abundant fauna of echinoderms, ostracods, pelecypod fragments and planktonic Foraminifera, to clean-washed fossiliferous intrasparites in which echinoderms, Algae, Bryozoa and small benthonic Foraminifera (mainly Miliolidae) are fairly common (Fig. IV-2).

The uppermost 6 m of the limestone sequence are composed of sandy biosparites forming a succession of widely spaced flat channel fillings, conformably overlain by the so-called "Senonian marls" (Misch, 1934). In a westerly direction (Fig. IV-2, section B) this contact passes into what may be considered to be a small non-angular unconformity, indicated by a strongly corroded surface on the upper layer of the Ventamillo Limestone, closely resembling the corroded surfaces described by Heim (1959), Hollmann (1962) and Jurgan (1968) (Fig. IV-3). According to Heim (1959), these discontinuities are brought about by the submarine solution of bare limestone deposits by weakly pulsating currents rich in CO_2. Thin films of glauconite are found, filling up the depressions of the irregular surface. Their presence is in accordance with a diminished rate of sedimentation (Müller, 1967), which is apparent from a comparison with section A. Section B shows a hiatus of 50 m of sediment over a lateral distance of only 250 m.

At locality A the lowermost 50 m of the Senonian marls consist of homogeneous calcareous sandy mudstones and massively bedded quartzarenites or quartzwackes. A gradual transition from mudstones into quartzwackes has been observed, whereas the quartzarenites generally show rather sharp upper and lower boundaries. Homogeneity of mudstones in which about 20% to 30% of fine sand occurs, is apparently due to the intense reworking and mottling by organisms. The quartzarenites are well sorted, with a grain size distribution falling within the medium sand range. Subangular to subrounded quartz grains are the main components, with some additional muscovite, biotite, feldspar and chert, all cemented by sparry calcite. There is no increase in grain size in an upward direction throughout the sandstone layer. Small scour-and-fill structures and cross-stratification are to be seen at the top. Deposition evidently took place under conditions of higher turbulence than in the case of the quartzwackes, whose tranquil deposition is indicated by the clayey matrix content and intense reworking by organisms. This sequence is followed by a thick series of thin-layered sandy calcareous mudstones grading upwards into calcareous shales.

The lowermost part of this unit passes laterally (section B) into an alternation of nodular limestones and calcareous mudstones which rest immediately upon the subsolution subsurface at locality B. The mean $CaCO_3$ content is 65% in the nodular limestones and 35% in the calcareous mudstones. Thick pebbly mudflow deposits occur locally between sections A and B. The rounded boulders and pebbles in these are similar to the nodular limestones, indicating postdepositional disturbances. The nodular limestones may show an irregular nodular upper surface or a very flat one. Lateral transitions between both forms have been observed. A common feature is the lower gradual transition from calcareous mudstone into micritic limestone, and a rather sharp upper boundary with pronounced nodularity. Burrowing is frequent in these limestones. In one case a faint cross-stratification was

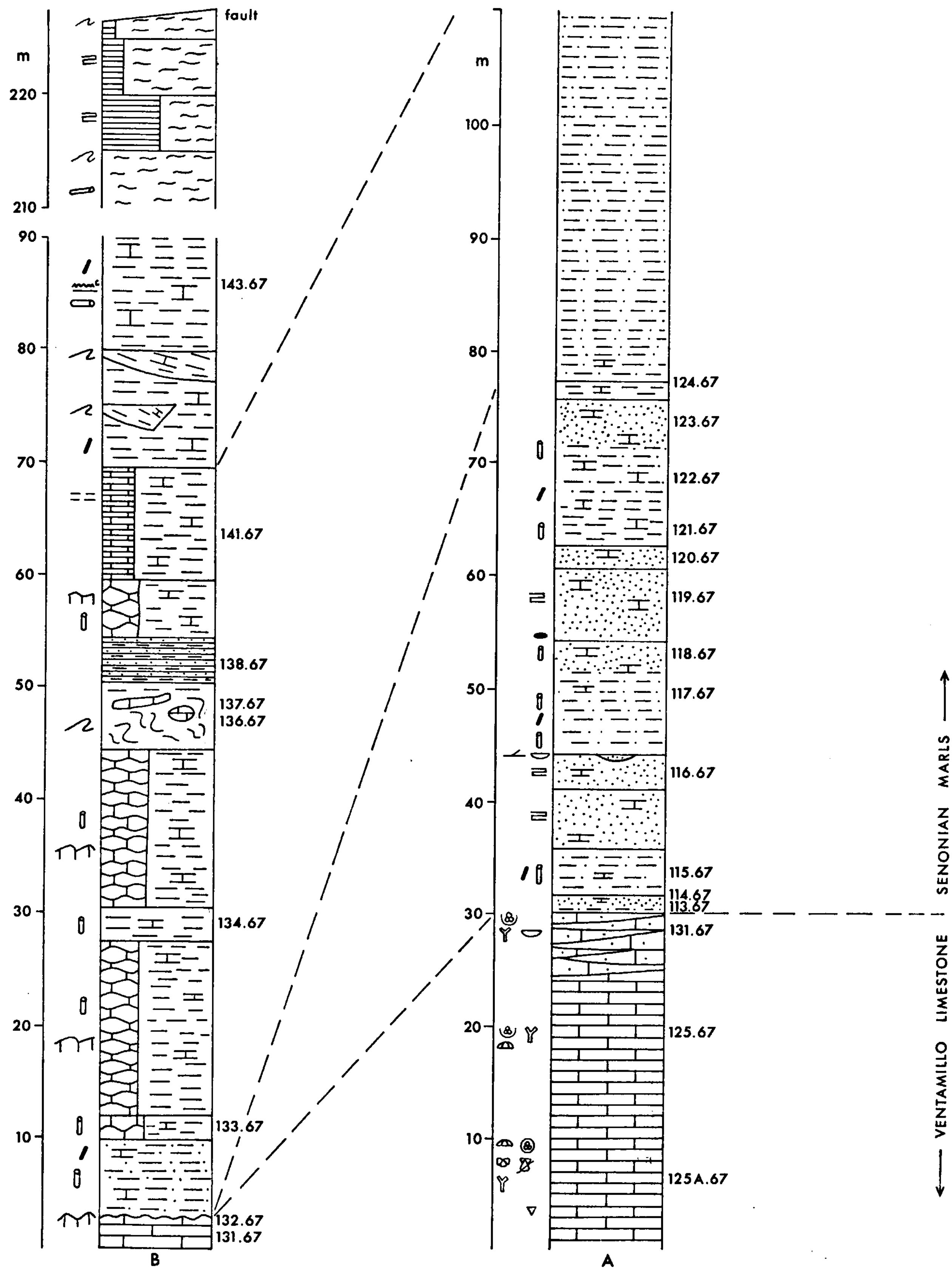


Fig. IV-2. Columnar sections of the Ventamillo Limestone-Senonian marls transition in the Seira area.

observed within a nodular sandy limestone. Pressure solution strings parallel to the bedding are common in the calcareous mudstones, but do not occur in the nodular limestones. Fibrous sparite overgrowths on muscovite are very common here, pushing aside the micrite by their crystallization pressure, which indicates that the sparry calcite must have crystallized when the sediment was not or only slightly lithified in order to permit of such enlargement (Folk, 1962).

The primary or secondary origin of nodular limestones in a limestone/shale rhythm has been the subject of various papers. In the present sequence burrowing, and in one case cross-stratification, proves to be a primary origin of the limestone/shale rhythm; such primary rhythms have been explained alternatively as the result of periodic changes in water depth (Hallam, 1964), periodic variations in $CaCO_3$ precipitation (Seibold, 1952), climatic and tectonic influences (Ziegler, 1958), or cyclic changes in the environment (Schwarzacher, 1964). The present author believes that variations in terrigenous clay supply (e.g. by a periodical shifting or accretion of a delta) against a background of constant $CaCO_3$ precipitation appear to be plausible in this sequence.

Early diagenetic processes account for the nodular character of the limestones. According to Weeks (1957), an increase in the pH of the pore water by the decomposition of organic material in stagnant marine basins leads to the precipitation of $CaCo_3$. Limestone nodules are consequently formed as an early diagenetic feature around fossil fragments. Braun (1964) advocated the existence of CO_2-zone in between an uppermost oxydizing zone and a lowermost reduction zone in bottom sediments in non-stagnant seas. Upward migration of the CO_2-zone (by continuous sedimentation) through already deposited lime-mud leads to the solution of $CaCO_3$, causing the formation of a nodular upper boundary of the limestones. Absence of nodularity probably testifies to a high local sedimentation rate, so that the sediments were not long enough in the CO_2-zone to be dissolved.

The limestone/shale unit at locality B is followed by a thick sequence (70–215 m) of laminated calcareous mudstones and marls, containing an abundant fauna of pelagic Foraminifera. Lamination is expressed in the alternation of silty and clayey material. Small symmetrical ripple marks, plant fragments and trace fossils on the bedding plane are common. Slumped and rotated packets of mudstone strata, possibly triggered by a sudden shock, occur in many places (Fig. IV-4). A south-eastern direction of movement has been established. The slumped horizons may be traced laterally over only a very short distance (3 m), indicating a local origin.

The top of the exposed section, composed of an alternation of thin-layered calcareous quartzwackes, quartzarenites and blue marls, is cut off by a fault, SW of Seira. A faint parallel lamination in the calcareous quartzwackes may be encountered. Deformation of the layers by slumping has also been observed. The total thickness of the Senonian marls cannot be estimated in this area, but may be determined about 5 km in a north-westerly direction, near Barbaruens, where about 1000 m are exposed. The sequence there is composed of calcareous mudstones followed at the top by calcareous sandstones corresponding to the Arén Formation of upper Maastrichtian age (Misch, 1934). A little further in this direction the Senonian marls pass laterally into dolomitic sandy limestones containing well-preserved macrofossils (*Micraster*, *Ananchytes* and *Rhynchonella*) and decrease in thickness to 750 m near Plan.

Depositional environment

The Ventamillo Limestones may be considered as having been deposited in a regressive shallow environment, in which a transition took place from a slow and tranquil sedimentation (biomicrites, with an abundant microfauna) to higher energy sandy limestones (fossiliferous intrasparites), which at the top of the sequence show a marked current activity by the presence of wide flat channels, probably formed near a shore. Together with the lithological indications, the increasing quantity of Miliolidae towards the top of the sequence suggests progressive shallowing of the warm marine environment (Cushman, 1948).

Locally the overlying Senonian marls show a remarkable difference in sedimentation rate between A and B, as was previously described. The lowermost 50 m exposed along the Esera Valley seem to be characteristic of a depositional environment in which oscillations in regressive sedimentation took place. Several units occur, grading from sediments deposited in a relatively calm environment (sandy mudstones) to deposition in

Fig. IV-3. Upper surface of the Ventamillo Limestones at locality B, corroded due to submarine solution. Hammer lies on bedding plane.

Fig. IV-4. Rotated packet of mudstone strata in between the Seira section.

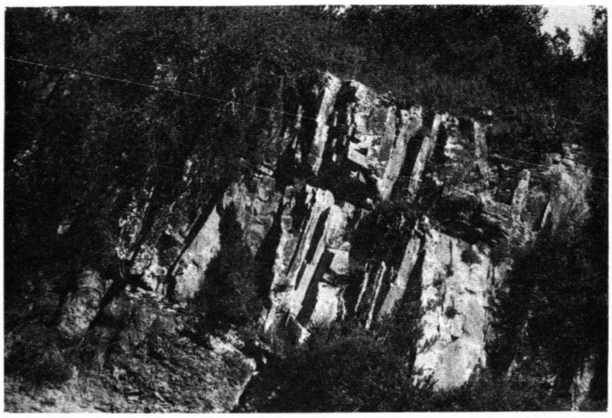


Fig. IV-5. Subaqueous sandstone bar following sandy mudstones with a sharp contact (lower part of the photograph).

a high energy environment (quartz-sandstones), which may be interpreted as subaqueous bars (Visher, 1967) (Fig. IV-5). However, the features typical of such bars, as listed by Shelton (1967), viz. a mottled structure in the lower part, an upwards increase in grain size and gradational lower parts, were not observed. Their modest thickness (about 5 m) indicates that they were not formed as large-scale coastal sand barriers.

A gradual deepening of the marine environment is expressed by the transition into calcareous mudstones and shales, with pelagic fauna. Laterally these sediments pass into nodular limestones and marls following a plane of non-deposition (Fig. IV-2). Rotational slumps within mudstones have been interpreted by Laird (1968) as movements occurring at the change in gradient between shallow shelf and slope deposits. This is in accordance with the occurrence of a thick turbidite sequence at a distance of 6 km in a southerly direction (Campo area), indicating an even deeper depositional environment.

The presence of thin-layered calcareous sandstones alternating with marls at the top of the exposed section may be indicative of turbidity current activity within a deepened environment. However, no sedimentary features were observed to support this idea.

UPPER CRETACEOUS N OF SALINAS, ALONG THE RIVER CINCA (Location C)

About 25 km NW of Campo, 1 km N of Salinas, along the road to Ainsa, an almost complete section through the Upper Cretaceous is exposed, lying upon dolomitic limestones and cellular dolomites of Middle Triassic age (Fig. IV-6) (van Lith, 1966). According to Misch (1934), this contact is an unconformity,

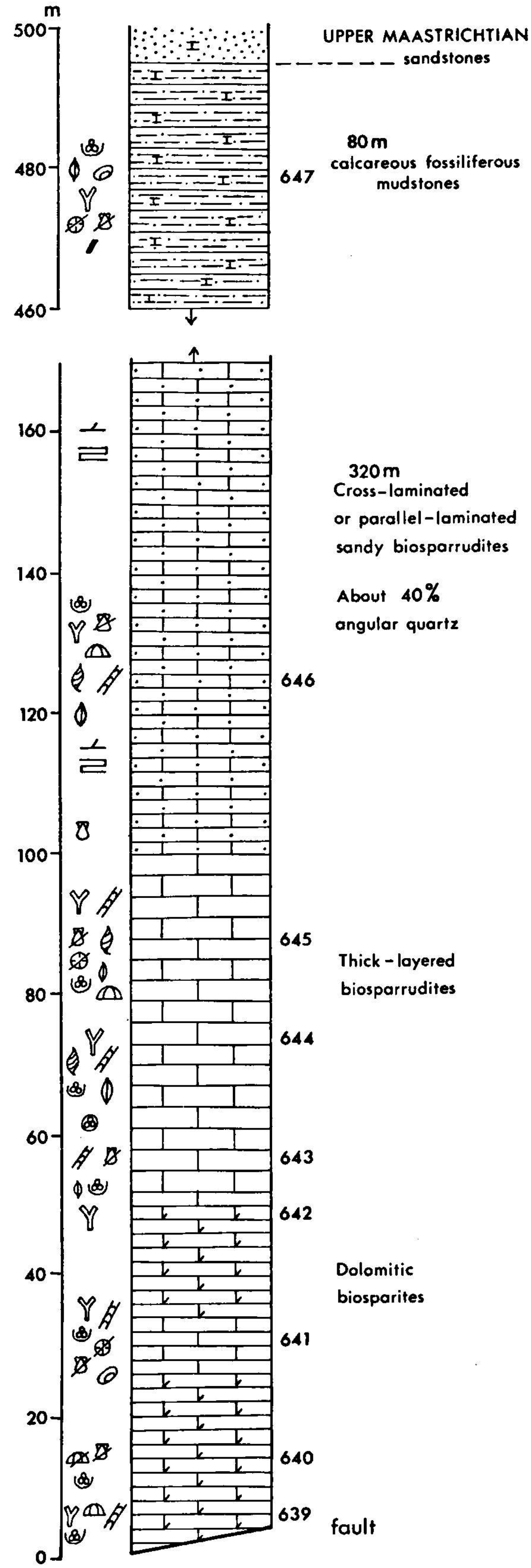


Fig. IV-6. Columnar section of the Upper Cretaceous N of Salinas (along the Rio Cinca).

whereas van Lith (1966) suggests the presence of a fault cutting off part of the base of the Upper Cretaceous sequence. The missing 30 m are to be found in an area W of Bielsa, in which an unconformity between the Upper Cretaceous and Middle Triassic is evident.

The sequence begins here with a conglomerate up to 40 cm in thickness, composed mainly of dolomite pebbles derived from the underlying Middle Triassic, and some additional coarse quartz sand (van Lith, 1966). This basal layer is followed by 60 m of dolomitic biosparites, somewhat reddish in colour, containing an abundant fauna of Miliolidae, Bryozoa, Algae and rudists and which are partly exposed along the Cinca. A Santonian age has been established (Mish, 1934). The reddish colour of the limestones is due to the presence of ferric oxyde coating and filling up of the fossil components. An early diagenetic feature is the common presence in the sparry calcite cement of dolomite rhombi.

These biosparites are followed by a 50 m thick series of biosparrudites, composed of large fossil fragments with a characteristic predominance of Bryozoa, Algae and Orbitoidea. Additional well-rounded polycrystalline quartz granules are present in the lowermost part of the biosparrudite member.

The largest part of the Upper Cretaceous in the Cinca Valley is composed of well-bedded sandy biosparrudites with large-scale cross-lamination or parallel-lamination. Individual laminae are composed entirely of coarse terrigenous material (angular quartz, chert and feldspar) or of bioclasts with a small amount of quartz. Concentration of coarser grains is observed at the base of the foresets. Layer thickness may be up to 2 m. The entire member has been dated as Campanian (Misch, 1934).

The top of the sequence is transitional into calcareous mudstones in which well-preserved fossils occur, chiefly Bryozoa and Orbitoidea. These mudstones pass rather abruptly into sandstones of the upper Maastrichtian.

Depositional environment

The lowermost 100 m of the sequence were deposited in a very shallow warm sea, as indicated by the abundance of Bryozoa, Algae, rudists, corals and large Foraminifera. The absence of calcareous mud or clay as matrix obviously reflects a high energy environment, well aerated as is suggested by the red colour of the fossils. The succeeding thick series of mainly cross-laminated sandy biosparrudites bears evidence of a relatively large terrigenous supply, probably from Paleozoic granites which were at least partly uncovered during this time. The angularity of the terrigenous grains and the persistent SSE direction of the cross-lamination support the location of this nearby source area. Upper Cretaceous limestones directly overlying granites in this region have been reported by van Lith (1966). Intense weathering of the flat granite area may account for the presence of only the most stable minerals such as quartz, chert, muscovite and some additional heavy minerals (tourmaline, zircon). The features of the cross-stratified layers closely resemble those of the Lower Greensand of SE England as described by Allen and Narayan (1964), who interpreted them as being deposited by migrating large-scale ripples in a shallow tidal sea. In the present exposures, the abundance of shallow-water organisms (Miliolidae, rudists, Bryozoa, Algae, Orbitoidea) in the laminae seems to support this idea. Temporary upper flow regime conditions may account for the formation of parallel-laminated beds in these coarse-grained deposits.

The uppermost calcareous mudstones of the sequence were evidently deposited in a non-agitated, shallow sea. Well preserved Orbitoidea, Bryozoa and Miliolidae are in agreement with the latter interpretation. The large quantity of terrigenous silty grains and small wood fragments indicate the proximity of a land mass.

UPPER CRETACEOUS S OF MEDIANO (Location D)

Between Puy de Cinca and the Monte Perdido, 20 km SW of Campo, a NW striking uparching in Eocene strata is found, which has been named the Peña Sestrales upfold (Selzer, 1934). S of the Pantano de Mediano, diapire-like Upper Triassic breaks through the overlying strata. Locally a continuous sequence of Upper Cretaceous rocks unconformably following the gypsiferous marls of the Upper Triassic may be traced here (Fig. IV-7). According to Souquet (1967), the base of this sequence may be dated as Santonian, thus agreeing with Selzer (1934).

The lowermost 80 m of the Upper Cretaceous sediments are composed of badly layered, dolomitic biomicrudites with an abundant fauna of fragmented rudists, Bryozoa, Algae, corals and small well preserved benthonic Foraminifera (Miliolidae). These pass upwards into thick-layered fine-grained limestones composed mainly of micrite intraclasts and Miliolidae (80–210 m). Fragments of rudists, Algae and Echinoidea only represent a small amount of the total fossil

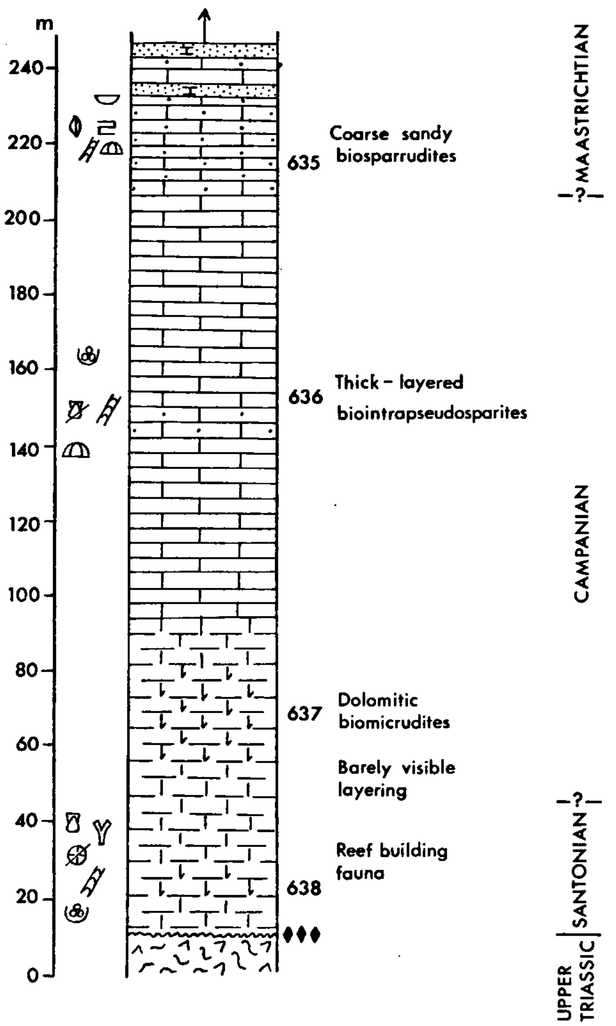


Fig. IV-7. Upper Cretaceous sequence S of Mediano.

content. Some terrigenous admixture, mainly rounded polycrystalline quartz, angular feldspar and muscovite, occur locally.

The uppermost unit (from 210 m) is composed of thick beds of sandy biosparrudites and fossiliferous quartzwackes, representing the transition from Campanian to lower Maastrichtian (Souquet, 1967), and attaining a thickness of about 200 m. The terrigenous components are well-rounded polycrystalline quartz granules and quartz pebbles, with some additional angular quartz, feldspar, chert and muscovite. Large well-preserved Orbitoidea and shell fragments are the common fossils. Practically all layers show a persistent parallel lamination. Scour-and-fill structures have also been observed. Channel axes are orientated in a NW-SE direction.

Depositional environment

The faunal content of the lowermost biomicrudites probably indicates the proximity of a reefal facies as a supply area of fragmented fossils. Deposition evidently occurred in a warm shallow sea with no major current activity, as is demonstrated by the micritic matrix. The features of these rocks probably indicate a transitional facies-zone between a reef-wall and a fore-reef shoal (Henson, 1950).

The same calm conditions prevailed during the sedimentation of the biointrapseudosparites constituting the bulk of the Campanian sequence. The small quantity of reefal debris and the overwhelming occurrence of Miliolidae suggests a very shallow warm environment closely resembling the environment described by Henson (1950) as a back-reef shoal.

The uppermost parallel-laminated sandy biosparrudites have been deposited under conditions of an upper flow regime. The abundance of Orbitoidea together with well-rounded terrigenous dedritus suggest a shallow open-littoral environment. The NW-SE orientation of the channels may be indicative of currents of varying velocity flowing parallel to the ancient shore line.

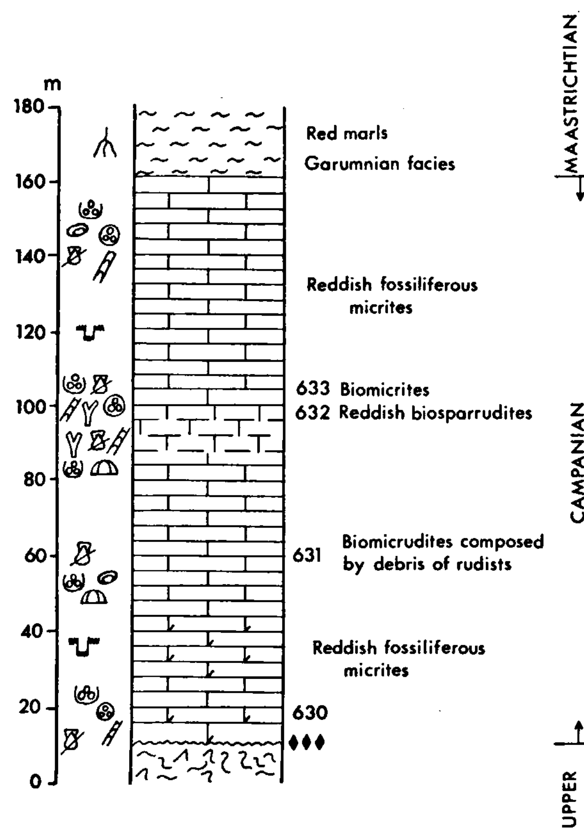


Fig. IV-8. Upper Cretaceous sequence N of Salinas de Hoz.

UPPER CRETACEOUS N OF SALINAS DE HOZ
(Location E)

A similar development of the Upper Cretaceous as is to be found S of Mediano is exposed in the diapiric structure between Salinas de Hoz and Naval, about 30 km SW of Campo (Fig. IV-8). About 150 m of reddish bioclastic limestones unconformably overlying the Upper Triassic are found here, which have been dated by Souquet (1967) as being of Campanian age. At the top these limestones abruptly pass into red marls with characteristic root levels suggesting a continental environment. Since no fossils are encountered it cannot be determined whether these sediments could represent the Maastrichtian in a Garumnian facies (Souquet, 1967).

The sequence begins with reddish fossiliferous micrites, sometimes slightly dolomitic, containing mainly Miliolidae and Rotaliidae. Reef builders like rudists, Algae and Bryozoa occur sporadically. The micrites abruptly pass into biomicrudites composed almost exclusively of rudist fragments with some additional benthonic Foraminifera (Fig. IV-9). No Miliolidae were encountered. They are followed by badly stratified biosparrudites composed of fragmented reef builders like rudists, Bryozoa and Algae. Benthonic Foraminifera have occasionally been recorded.

The uppermost 60 m of the Campanian sequence show a development similar to that of the fine-grained limestones forming the base, consisting of reddish micrites containing Miliolidae and Rotaliidae.

Depositional environment

The classification of facies-zones within reef-complexes as proposed by Henson (1950) has been followed in the present study. The fossiliferous micrites, present at the base and top of the Campanian limestones, indicate quiet deposition in a sheltered environment, probably a back-reef facies. This is in accordance with Henson (1950a), who states that a large quantity of Miliolidae represents deposition in shallow lagoons between fringing or barrier reefs and the shore.

The middle part of the sequence, composed of biomicrudites and biosparrudites, has probably been deposited as reef-talus somewhere between a reef-wall and a fore-reef shoal. The reddish colour of the lime-

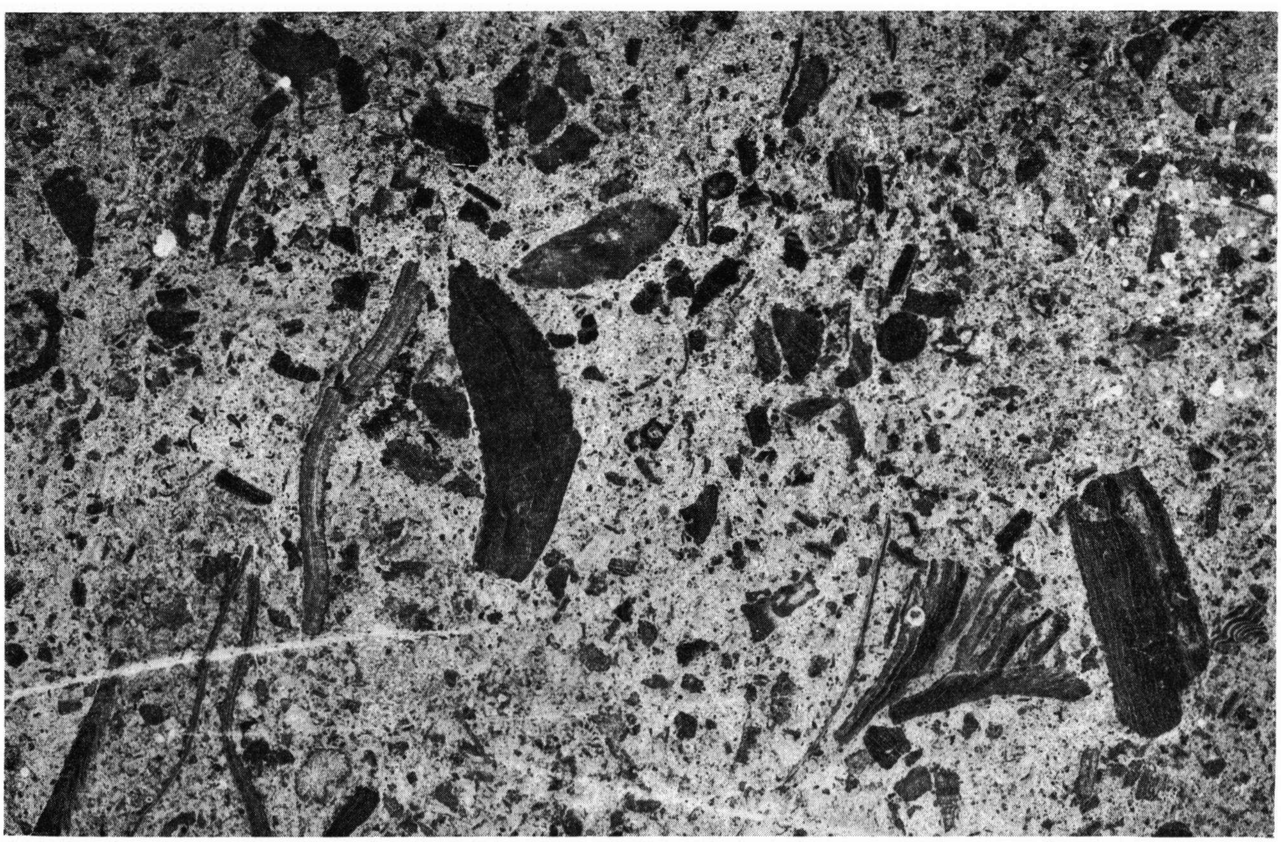


Fig. IV-9. Typical bioclastic limestone composed of fragmented rudists and micrite. Common lithofacies along the border of the Upper Cretaceous basin in the south-central Pyrenees. 8×.

stones suggests a well aerated, warm shallow environment which is evidently in accordance with the faunal content.

UPPER CRETACEOUS S OF AGUINALIU (Location F)

A uniform lithological sequence of the Upper Cretaceous is found in the southern Sierras zone of Aragón, between the Esera river in the W and the Noguera Ribagorzana river in the E (Selzer, 1934). Outcrops are present due to the more or less diapiric behaviour of the immediately underlying Upper Triassic. One representative section has been studied near Aguinaliú, S of Graus, about 35 km S of Campo, where an uninterrupted sequence is found (Fig. IV-10) which, just as the Salinas de Hoz section, has been dated as Campanian (Souquet, 1967).

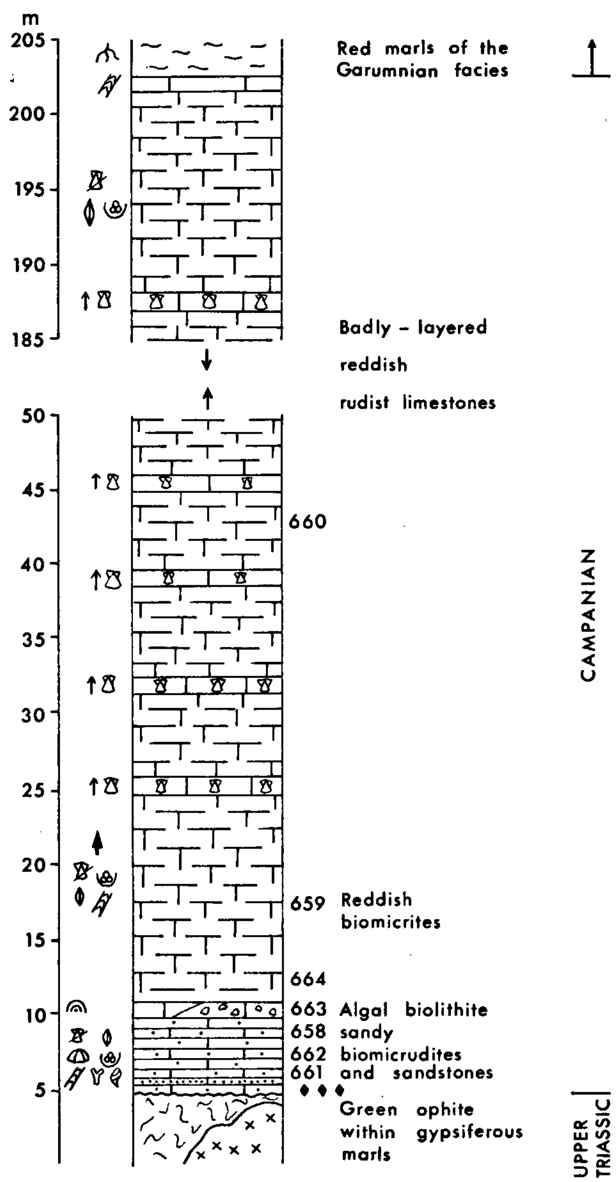
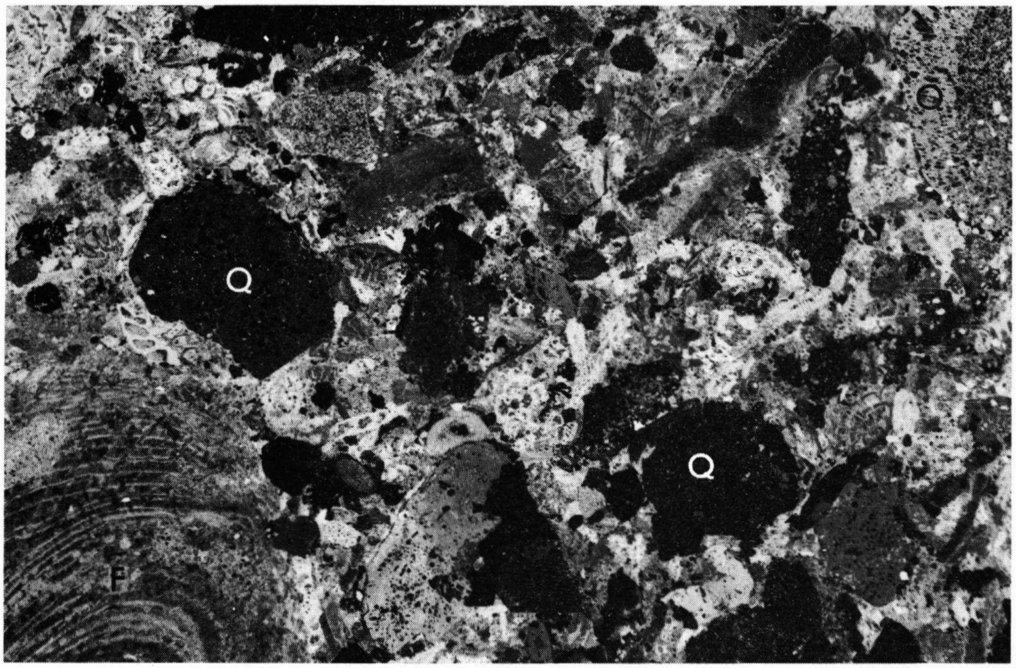


Fig. IV-10. Upper Cretaceous sequence S of Aguinaliú.

The lowermost 5 m of the Upper Cretaceous, lying unconformably upon intensively deformed gypsiferous marls in which ophite bodies occur, consists of irregularly layered sandy biomicrudites and biomicrites. Locally more than 50% quartz components were found, constituting a calcareous sandstone. The quartz grains are idiomorphic crystals, sometimes broken or slightly rounded (Fig. IV-11), with a poikilitic texture of enclosed anhydrite, occasionally replaced by hematite. Rim replacement by microcrystalline calcite is frequently observed. The idiomorphic quartz grains derived from Upper Triassic saline deposits as a result of weathering and erosion, and were carried away during the Upper Cretaceous transgression, being deposited shortly afterwards together with bioclastic material. Such erosion is also suggested by gypsum clasts evidently of Upper Triassic origin. Although the presence of idiomorphic quartz grains normally suggests an authigenic origin in a hypersaline environment (Grimm, 1962), it is unlikely that saline conditions prevailed in the Upper Cretaceous sedimentary environment since the large quantity of oxydized fragmented large Foraminifera (Orbitolinidae, *Prealveolina*), small well-preserved benthonic Foraminifera (Miliolidae), Gastropoda, Algae and Echinoidea fragments in these sandy layers indicate deposition in a well aerated, non-restricted sea. Another explanation, assuming that the idiomorphic quartz grains were formed late-diagenetically by saline solutions rising from the underlying Upper Triassic, may be rejected since a somewhat rounded or broken shape was noted, indicating some degree of transport. Grimm (1962) states that in the case of late-diagenetic origin due to rising solutions, an irregular distribution of authigenic quartz in the overlying rock will be found irrespectively of the stratigraphy, whereas the quartz grains present are restricted to the lowermost beds. One would also expect enclosures of material available in the host rock, in this case microcrystalline calcite, and not merely anhydrite. A final argument against a diagenetic origin is the absence of any growth features such as penetration or pushing aside of carbonate matrix and bioclasts.

These basal layers are followed by a biolithite, 1 m in thickness and composed of limeclasts, fragments of rudists and Algae, benthonic Foraminifera, all bounded by a micrite matrix (Fig. IV-12). This coarse deposit laterally wedges out, being replaced by fine-grained limestones with abundant Miliolidae. The limeclasts



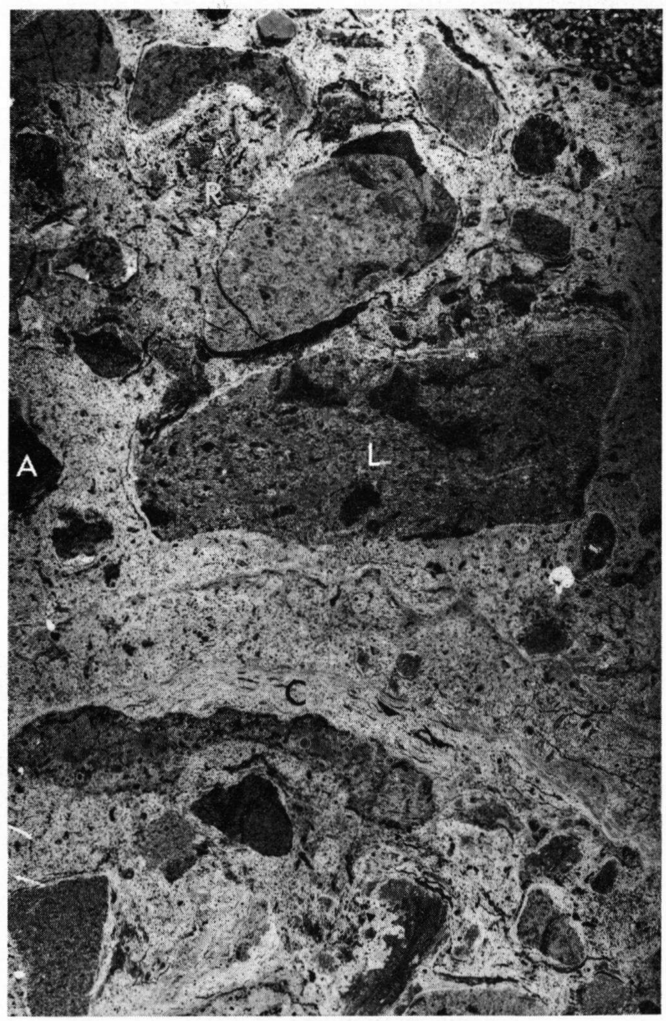


Fig. IV-11. Texture of the basal layer of the Upper Cretaceous transgression near Aguinaliú. Broken idiomorphic quartz grains (Q) and fragmented fossils (F) in a micrite matrix. 8×.

Fig. IV-12. Limeclasts (L), Algae (A) and rudists (R) forming a biolithite layer probably deposited as a talus-slope sediment. Irregular layering due to algal crusts (C). 8×.

derive from an intra-environmental erosion, so that they have to be genetically classified as intraclasts (Wolf and Connolly, 1965). This conclusion is based on the petrology, all clasts having originated from some authochthonous reef-complex source (algal limeclasts, micrite clasts containing rudist fragments circumcrusted by algal calcilutite, bioclastic pebbles with an abundant algal and foraminiferal fauna), and on stratigraphic information, no older limestones having been found in this area. The biolithite layer shows a faintly visible stratification due to algal crusts, which also developed around all clasts, sometimes mixed with a hematite coating.

These are followed by 10 m of badly layered reddish biomicrites with fragments of a reef-building fauna (Algae, Bryozoa, rudists, corals) and benthonic Foraminifera (Orbitolinidae, Miliolidae) gradually passing, in an upward direction, into a thick sequence of about 185 m of reddish rudist-limestones. Rudist fragments as well as rudists in a growth position have been found. In the latter case they are restricted to thin though extensive horizons. Well-preserved Orbitolinidae, Orbitoidea, Miliolidae and small pelagic Foraminifera are present within the micritic matrix. Just as in the section near Salinas de Hoz, no sediments of Maastrichtian age have been determined (Souquet, 1967). The upper layer of the Campanian limestones is composed of brownish fine-grained limestone containing *Chara* and *Microcodium* (Souquet, 1967), followed by red marls of the Garumnian facies.

Depositional environment

The lowermost sandy deposits of the Upper Cretaceous sequence in this area were evidently laid down in a very shallow sea which slowly transgressed the Upper Triassic land mass. The faunal content is probably composed of authochthonous elements (Miliolidae) and washed-in fossil fragments suggesting an open littoral environment. Vigorous current activity did not exist, which is indicated by the presence of fine-grained matrix and absence of current-generated structures.

The biolithite layer resembles a talus-slope deposit, a kind of reef-breccia, its components broken off by breaker erosion and deposited below breaker level, and its interstices filled with calcareous mud. An alternative explanation may be that the coarse detritus of authochthonous organisms was swept by currents into small banks, thus forming a badly sorted fossil breccia. In between the breccias, calcareous mud was deposited with small benthonic foraminifera.

The bulk of the sequence is composed of rudist shoal-reefs associated with foraminiferal limestones containing abundant rudist detritus (the 'type subrécifal" of Souquet). Rudist limestones are well known from various Cretaceous deposits of the Tethys sea (Bergquist and Cobban, 1957), and are commonly associated with large Foraminifera (Orbitolinidae, Prealveolinidae, etc.). They have been deposited in warm, clear, shallow waters of normal salinity. According to Matthews (1951), the normal depth at which rudists occur varies between 17 and 34 fathoms with a mean paleotemperature of the sea-water during Campanian times of about 19–22° (Damestoy, 1967; Lowenstam, 1961).

The top of the exposed sequence was evidently deposited in a lacustrine environment as is indicated by the presence of *Chara*, followed by continental deposits containing root levels.

SANTONIAN AND CAMPANIAN DEPOSITS IN THE MONTSECH AREA (Location G)

South of the Eocene basin of Tremp, an E-W striking steep mountain chain is found, showing a continuous Upper Cretaceous sequence lying unconformably upon Jurassic sediments (Misch, 1934; Souquet, 1967). Since the present study concerns itself with deposits isochronous with the Vallcarga Formation, only the uppermost part of this sequence was examined, corresponding in age to the Santonian and Campanian (Fig. IV-13). A characteristic section was surveyed along the Noguera Pallaresa river, where the greatest thickness is found (955 m).

Three major members may be discerned, a lower sandy one, a nodular limestone/marl unit, and uppermost a thick bioclastic limestone series.The lower sandy limestone member passes into bioclastic limestones in a westerly direction, i.e. toward the Ribagorzana area, whereas the other two members show no lithological differences in the Pallaresa and Ribagorzana valleys (Souquet, 1967).

The base of the sandy member, corresponding to the Coniacian/Santonian boundary (Souquet, 1967; Hottinger, 1966), is developed locally as slightly nodular biomicrudites and marls (10 m), in which an abundant fauna of benthonic Foraminifera (Miliolidae, Rotaliidae, Valvulinidae), ostracods and rudists occurs, and in an upward direction passes into a 20 m thick alternation of reddish sandy biointrasparites and marls containing benthonic Foraminifera (Miliolidae) and some additional fragments of reef building organisms

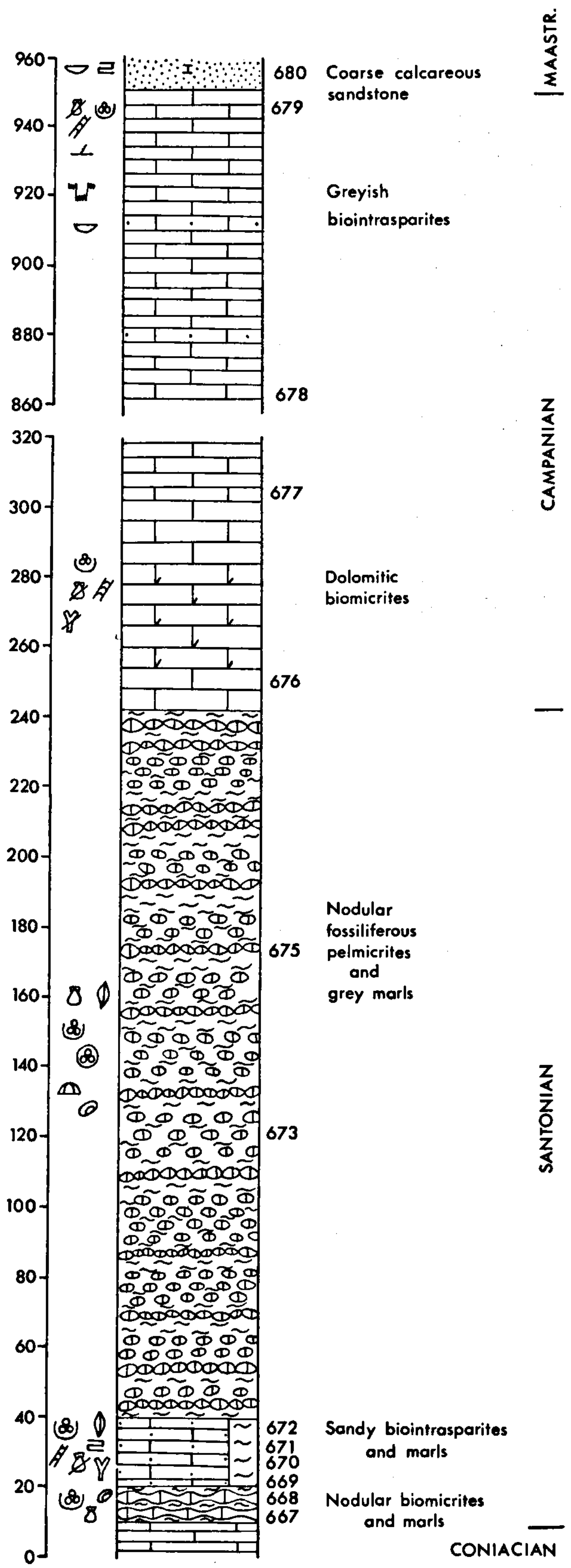


Fig. IV-13. Columnar section of the Santonian and Campanian deposits in the Montsech area.

(Bryozoa, Algae, rudists). A faintly visible parallel lamination occurs in most sandy layers.

This is followed by 200 m of fossiliferous pelmicrites and grey marls which have probably been laid down as a primary alternation of thick-layered fossiliferous lime mud and thin clayey deposits. During an early diagenetic stage the pelmicrites acquired their present nodular shape (Fig. IV-14). The increasing nodularity towards the top of each limestone layer probably indicates a higher clay content in this direction in the original sediment. Nodules were formed with organic components as a core, especially well-preserved rudists, and large benthonic Foraminifera, whereas small Foraminifera coated with ferric oxyde are encountered in the clayey strings between the nodules. A decrease in large fossils is apparent in the uppermost part of the limestones, which is evidently determined there by a higher rate of clay sedimentation. More isolated nodules were therefore formed.

The largest part of the section is composed of massively bedded white, reddish or greyish bioclastic limestones attaining a thickness of 625 m (Fig. IV-15), which have been dated as Campanian (Souquet, 1967). The lowermost 400 m of the sequence show a monotonous succession of fine-grained, sometimes dolomitic, biomicrites with well-preserved benthonic Foraminifera (Miliolidae, Rotaliidae, Textularidae, Orbitoidea) and fragments of Algae, Bryozoa and rudists. Sporadic coarse sandy layers have been observed. The uppermost 225 m of clean-washed biointrasparites contain about 3% of fine-grained terrigenous material (angular quartz, feldspar and muscovite), their faunal content being similar to that of the lower biomicrite unit, with a large quantity of rudist fragments. Large-scale cross-bedding (about 1–2 m high) and small channels filled with coarse sand are indicative of currents flowing in a persistently NW direction.

Depositional environment

Sedimentation in the Montsech area during Santonian times must have occurred in a shallow warm sea on a stable flat shelf.

The lower part of the sequence in the Pallaresa area reflects current activity bringing in terrigenous material from a nearby land mass, mixed up with a shallow water fauna, in contrast to the western part of the Montsech (Ribagorzana area) in which bioclastic limestones were deposited.

The upper part of the Santonian sediments indicates quiet sedimentation of lime mud and marl on a slowly

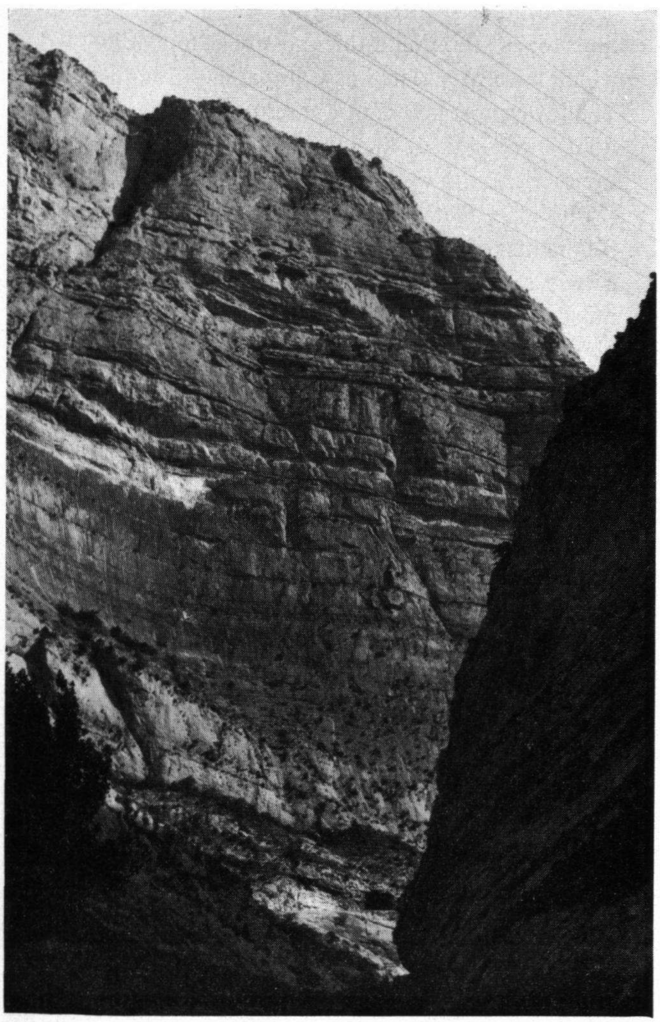


Fig. IV-14. Typical aspect of the Santonian nodular limestones/marls member in the Montsech area. Primary alternation of thick-bedded limestones and marls still visible. Degree of nodularity increases towards the top of a limestone layer.

Fig. IV-15. Exposure of Campanian bioclastic limestones along the Noguera Pallaresa Valley. Photograph taken in a northern direction.

subsiding shallow shelf floor. This picture also agrees for the lower part of the Campanian sequence, except that the clay supply came to an end. This latter part is built up by benthonic Foraminifera and debris of reef-building organisms. According to Henson's classification (1950), the depositional environment of these limestones may be regarded as a fore-reef transitional zone or an open reef-shoal, characterized by the mixture of littoral organisms, reef-building fauna and large Foraminifera. Cross-bedding occurrent in many layers, up to 2 m in thickness, of the uppermost Campanian deposits is indicative of large transporting currents in a shallow sea, probably producing the same type of migrating ripples as postulated by Allen and Narayan (1964) and which has previously been described for sediments in the Cinca area (Section C). Transport over large distances does not seem very likely, since the fossil content is the same as that encountered in intercalated bioclastic limestones lacking current features.

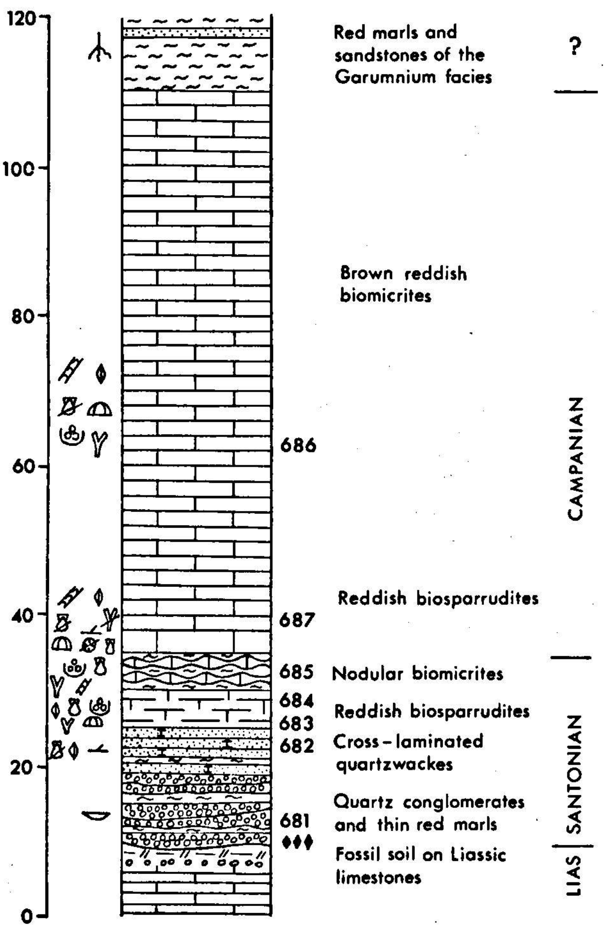


Fig. IV-16. Upper Cretaceous sequence near Os de Balaguer.

Fig. IV-17. Base of the Upper Cretaceous transgression near Os de Balaguer. Unconformity plane marked by top of the hammer. Handle lies on a soil profile on top of Liassic limestones.

UPPER CRETACEOUS SEQUENCE IN THE SYNCLINE OF OS DE BALAGUER (Location H)

Upper Cretaceous sediments may be traced southwards into a W-E striking chain of hills known in geological literature as the Sierras Zone (Misch, 1934). One characteristic section located here, E of the small village of Os de Balaguer, has been surveyed in order to examine the most southerly occurrence of the Cretaceous rocks with which this study concerns itself. Other exposures present in the Sierras Zone show a lithological development similar to that near Os de Balaguer.

The total thickness of Santonian and Campanian deposits in this area is about 100 m (Fig. IV-16). The sequence begins with 15 m of thick-layered channelling calcareous quartz conglomerates, grading upwards into finer-grained calcareous quartzwacke layers. Sandy red or purple marls, in which no fossils were encountered, occur in between. The base of this unit, which has been dated by Souquet (1967) as Santonian, on account of the benthonic Foraminifera enclosed in the sparry calcite cement, rests unconformably upon a fossil soil profile developed on Liassic limestones (Fig. IV-17). Souquet (1967) reported laterite soils at the top of the Lias in other areas (Sierra de Boada, Tragó de Noguera), which in the present section, however, were probably washed away by the advancing sea. This is demonstrated by the occurrence of bauxitic and ferruginous clasts in the lowermost quartz conglomerate. The grain size of these deposits may be up to 5 cm. The large polycrystalline quartz granules and pebbles are slightly rounded, whereas the individual non-undulatory quartz grains are more or less angular. The uppermost well-layered finer-grained quartzwackes (mean size of 0.6 mm) contain up to 10% of fossil fragments (rudists and large benthonic Foraminifera). A well-developed cross-bedding may be observed. The current orientation inferred from these structures shows a very consistent flow direction towards the SW (240°), which agrees with the trend of the channel axes of the quartz conglomerates.

These coarse sandy deposits are followed by 5 m of poorly layered reddish biosparrudites composed of large fossil fragments (Bryozoa, rudists, Echinoidea, large Miliolidae, Orbitolinidae, and rounded algal clasts) and additional angular quartz grains (up to 5%). On top of this unit, about 5 m of nodular biomicrites and marls occur in which small Miliolidae are extremely common. Well-preserved rudists, Bryozoa and Algae were also encountered.

The bulk of the sequence is formed by a 75 m thick succession of reddish-brown well-layered biomicrites and biosparrudites dated as Campanian (Souquet, 1967). At the base of this unit rudist colonies in a growth position are to be found. The biomicrites contain an abundant fauna of benthonic Foraminifera (Miliolidae, Orbitoidea), large algal fragments, rudists, and Bryozoa, whereas the cross-bedded biosparrudites are composed of large debris of reef-building organisms (Bryozoa, Algae, rudists, corals, Orbitoidea) (Fig. IV-18).

The top of the sequence passes abruptly into red continental marls in the Garumnian facies, which were probably partly deposited during Maastrichtian times (Souquet, 1967).

Fig. IV-18. Typical reef-flank deposits composed of fossil debris of reef-building organisms. 8 ×

Depositional environment

The development of laterites and bauxites as soil profiles prior to the deposition of the Upper Cretaceous sequence suggests an emerged landmass exposed continuously to the weathering cycle. Lateritization occurs in areas of low or moderate topographic relief and with a minimum of erosion, requiring a persistently warm climate with temperatures above 25 °C, and heavy rainfall almost continuously exceeding evaporation (Hatch and Rastall, 1965). This flat area was flooded by the Upper Cretaceous transgression, during which a coarse basal unit of conglomerates and sands was deposited. The land mass did not supply the sandy material since only fine-grained Liassic limestones and, in a southerly direction, Upper Triassic gypsiferous marls formed the surface. Consequently, the source area must be sought in a north-easterly direction, as is suggested by the cross-bedding orientation and the channel axes. Material was probably transported by currents flowing along the ancient coastline (see Paleogeography, Chapter V). The absence of fossils in red marls in between the sandy layers might indicate that they were deposited in a continental environment.

The bioclastic limestone series which constitutes the largest part of the sequence shows a fossil content genetically associated with a transgressive reef-complex undergoing subsidence (Henson, 1950). The biosparrudites and nodular biomicrites following the littoral clastics may have been deposited in a back-reef facies on to which rudist reefs and fore-reef limestones gradually extended. The coarse biosparrudites probably reflect reef-flank sediments as is indicated by the coarse, poorly sorted, broken skeletons. (Nelson *et al.*, 1962).

A comparison of this section with the Montsech area shows that sediments decrease in thickness in a southerly direction, as well as possessing a more littoral character. The Montsech sediments were deposited somewhat further into the basin on a flat shelf, which, prior to Santonian times, already was a subsiding marine area, whereas in the southern Sierras Zone the sea transgressed an ancient land mass. According to Souquet (1967), no marine sedimentation took place 5 km south of Os de Balaguer during the Upper Cretaceous (Fig. IV-19). Only continental deposits are found here, providing no evidence as to whether sediments of Upper Cretaceous age are absent or were developed in a Garumnian facies.

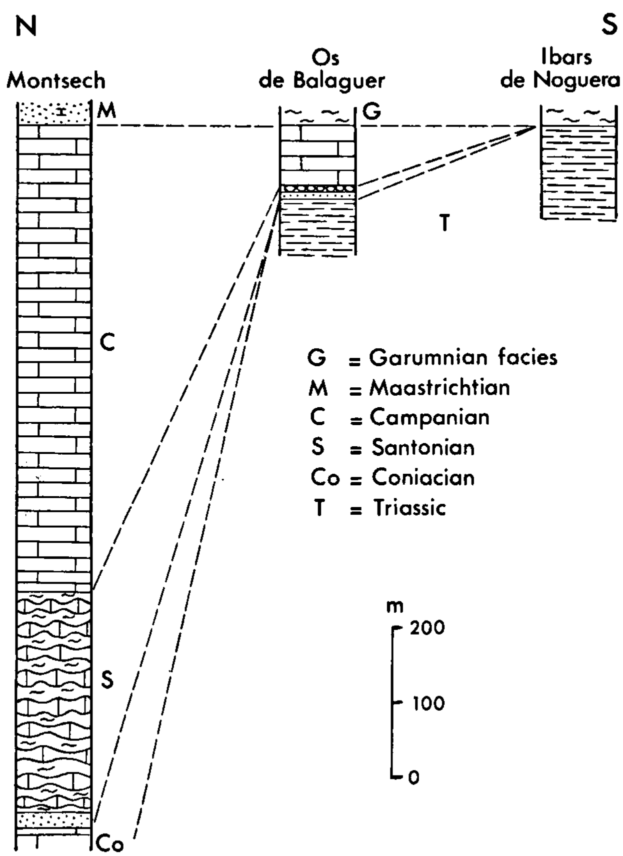


Fig. IV-19. Diagrammatic stratigraphic cross-section in the southern part of the Upper Cretaceous basin.

STRATIGRAPHIC RELATIONSHIPS OF SANTONIAN AND CAMPANIAN SEDIMENTS S AND W OF COLL DE NARGO (Location I)

In the Sierra de Aubens, along the Segre river, Santonian deposits have been reported unconformably overlying Jurassic dolomites (Souquet, 1967). A lithological sequence similar to the one in the Montsech area is developed here. The lowermost sandy member, which in this area shows frequent intercalations of quartz conglomerates, and a basal unit of bioclastic limestones, increases in thickness from 20 m to 70 m, whereas the uppermost member of nodular biomicrites and marls decreases in thickness from 200 m to 100 m, the total thickness of Santonian sediments therefore being less than in the Montsech area.

In the present area, Campanian rocks differ laterally in thickness and lithology, being differentiated in a southerly (Sierra de Aubens) succession of reddish bioclastic limestones, 350 m thick and similar to the Montsech sequence, and a northerly facies type (near Valldarques) of about 900 m of grey nodular sandy limestones and marls. A well-exposed transition between the two lithofacies types is to be seen along the Rialp river. The nodular limestones may be classified as sandy biomicrites or biomicrudites, depending upon the size of the fossils enclosed. An abundant macrofauna has been ascertained (Dalloni, 1930; Souquet, 1967) containing *Terebratula, Rhynchonella, Pecten, Exogyra,* as well as a microfauna composed of benthonic Foraminifera (Textularidae, Orbitoidea, Valvulinidae) and debris of Echinoidea, gastropods and Bryozoa. Terrigenous admixture, up to 15%, is composed of angular quartz, muscovite, biotite and feldspar. Authigenic glauconite filling up fossils or randomly distributed in the micritic matrix is quite common.

The transition from bioclastic limestones in the south in which rudists and Miliolidae are common, to sandy nodular limestones in the north lacking these fossils probably indicates deposition somewhat deeper in this latter area, further from the southern coastline on a slowly subsiding shallow basin. Subsidence was almost balanced by the rate of sedimentation.

The uppermost part of the exposed sequence shows an alternation of sandy bioclastic limestones and marls, passing abruptly into thick parallel-laminated or cross-bedded coarse calcareous sandstones of the Arén Formation. These deposits have been dated as Maastrichtian (Souquet, 1967).

UPPER CRETACEOUS SEQUENCE NEAR ADRAHENT (S OF SEO DE URGEL) (Location J)

A brief survey was made in this area in order to investigate the lithology of the most easterly exposures of the Upper Cretaceous in the south-central Pyrenees. A characteristic section is found near the small village of Adrahent, composed of a lower conglomerate unit which has been called the Adrahent Formation, followed by a limestone sequence called the Bona Formation (Mey *et al.*, 1968).

The Adrahent Formation, tentatively dated as Santonian by Guérin-Desjardins and Latreille (1961), here reaches a total thickness of 180 m, and is composed of white quartz conglomerates, quartzwackes and black shales. Plant remains chiefly occur in the finer-grained sandstones in the form of dark laminations. Small-scale cross-stratification has been found indicating a south-westerly current orientation (210°). However, since only 3 layers with cross-stratification were found, this figure should be considered with some reserve. The mineralogical composition of the quartz conglomerates perhaps preferably termed quartz-breccia because of the angularity of its components, closely resembles that described in the Os de Balaguer section (large polycrystalline quartz pebbles and granules and smaller individual angular quartz grains and feldspar). In the latter area, however, large quartz pebbles show a higher degree of roundness. The absence of fossils in the Adrahent Formation suggests non-marine deposition of the coarse terrigenous material, which probably was derived from the Andorra granodiorite (Hartevelt, pers. comm.).

The Bona Formation, attaining a thickness of 250 m near Adrahent, is mainly composed of biomicrites with a characteristic fossil content of rudists, benthonic Foraminifera (Miliolidae, Textularidae, Valvulinidae, Orbitoidea), algal fragments, Bryozoa, Echinoidea and corals, suggesting a shallow marine depositional environment. They have been dated as Campanian-Maastrichtian (Mey *et al.*, 1968). The uppermost part of the section shows a sharp transition into red shales of the Garumnian facies.

This upper Cretaceous sequence slowly decreases in thickness towards the E. The most easterly occurrence was reported by Ashauer (1934) in the Fresser river area, near the small village of Montgrony (at 50 km E of Adrahent), where only 20 m of quartz conglomerates are present in between Triassic and Garumnian deposits. Thinning off in this direction is connected with the occurrence of the Ampurdán swell during Upper Cretaceous times (Ashauer, 1934).

CHAPTER V

SEDIMENTATION AND PALEOGEOGRAPHY

INTRODUCTION

Prior to the deposition of the Vallcarga Formation, thick units of Mesozoic sediments were laid down in a southern Pyrenean marginal basin, located more or less parallel to the strike of the axial zone of the Pyrenees (de Sitter, 1964). During a considerable part of the Lower Cretaceous (Valanginian-lower Aptian), the area of greatest subsidence was located in the east, in the present Segre area, where a thick series of fossiliferous micritic limestones called the Prada Formation (Mey *et al.*, 1968) was deposited (Fig. V-1), diminishing in thickness in a westerly direction, i.e. towards the N-S striking Aragón upheaval (Misch, 1934). The area of greatest subsidence subsequently migrated towards the west, leading to a maximum thickness of deposition during the Upper Cretaceous in the Ribagorzana area, amounting to almost 4800 m. Sedimentation also continued in the Segre region but does not show the same thickness as in the Ribagorzana area. This migration also affected, to a certain degree, the western part of the south-central Pyrenees: the Upper Cretaceous sea flooded the Aragón upheaval, on which upheaval shallow water limestones were deposited (Salinas de Hoz and Mediano sections, Chapter IV). Later still, during the Paleocene and Eocene, the area of greatest subsidence moved even further to the west, reaching a thickness of at least 4100 m near the Cinca Valley (Fig. V-1). This development during the Paleogene can also be observed as a facies transition from shallow marine and continental deposits in the east to a thick series of turbidites and marls in the west. Simultaneously with the E-W migration of the marginal basin during the Cretaceous and Paleogene, it shifted slowly in a southerly direction (de Sitter, 1964), probably in relation to an emergence of the axial zone of the Pyrenees.

The origin of the Vallcarga Formation may consequently be related to this shifting of the marginal basin during Upper Cretaceous times, during a period in which the equilibrium between subsidence and rate of sedimentation was suddenly disturbed in parts of the basin.

Stratigraphical studies by Souquet (1967) as well as our own observations indicate that the Vallcarga Formation almost everywhere rests conformably on older Upper Cretaceous rocks, except for the local intra-Santonian unconformable contact in the Esera area. This continuous sequence is also present in contemporaneous sediments of the Montsech area, from where an uninterrupted Upper Cretaceous sequence is reported (Souquet, 1967; Misch, 1934), whereas in southern areas Upper Cretaceous rocks lie unconformably upon tilted Jurassic limestones and dolomites and intensively deformed gypsiferous marls of Upper Triassic age (Fig. V-2). Consequently, prior to the deposition of the bulk of the Upper Cretaceous, deformation of these rock units occurred (Fig. V-3), which may possibly be related to a diapiric movement of plastic material of the Upper Triassic (Keuper), which continued until it reached the surface, prior to the deposition of the Upper Cretaceous. This rise of Keuper material in the southern and western part of the south-central Pyrenean foreland may possibly be considered as a response to the accumulation of thick sedimentary sequences during the Lower Cretaceous and the lower part of the Upper Cretaceous in central parts of the basin. This accumulation exerted the pressure required for starting mobilization and migration of the plastic Keuper material (evaporites, possibly including rock salt) towards the flanks of the trough, where low anticlinal diapire-like structures were formed, sometimes with a complexly faulted core. An identical state of affairs has been described in the western Pyrenees (Brinkmann and Lögters, 1967) where diapirism of Keuper deposits has been observed, starting in the Lower Cretaceous and attaining its maximum during the Upper Cretaceous when most diapirs reached the surface in the southern and eastern part of the basin. Here, too, no diapirs are encountered in areas with a thick sedimentary cover, but on the flanks of the trough.

The present author thinks that the initial trigger effect leading to the origin of a deeper turbidite basin was possibly exerted by the migration of evaporites in a south-westerly direction as a consequence of the continuous sedimentation in the subsiding Upper Cretaceous basin. This sudden deepening of the basin was afterwards followed by a normal subsidence until equilibrium was reached in the upper Maastrichtian leading to a regressive sedimentation (Arén Formation).

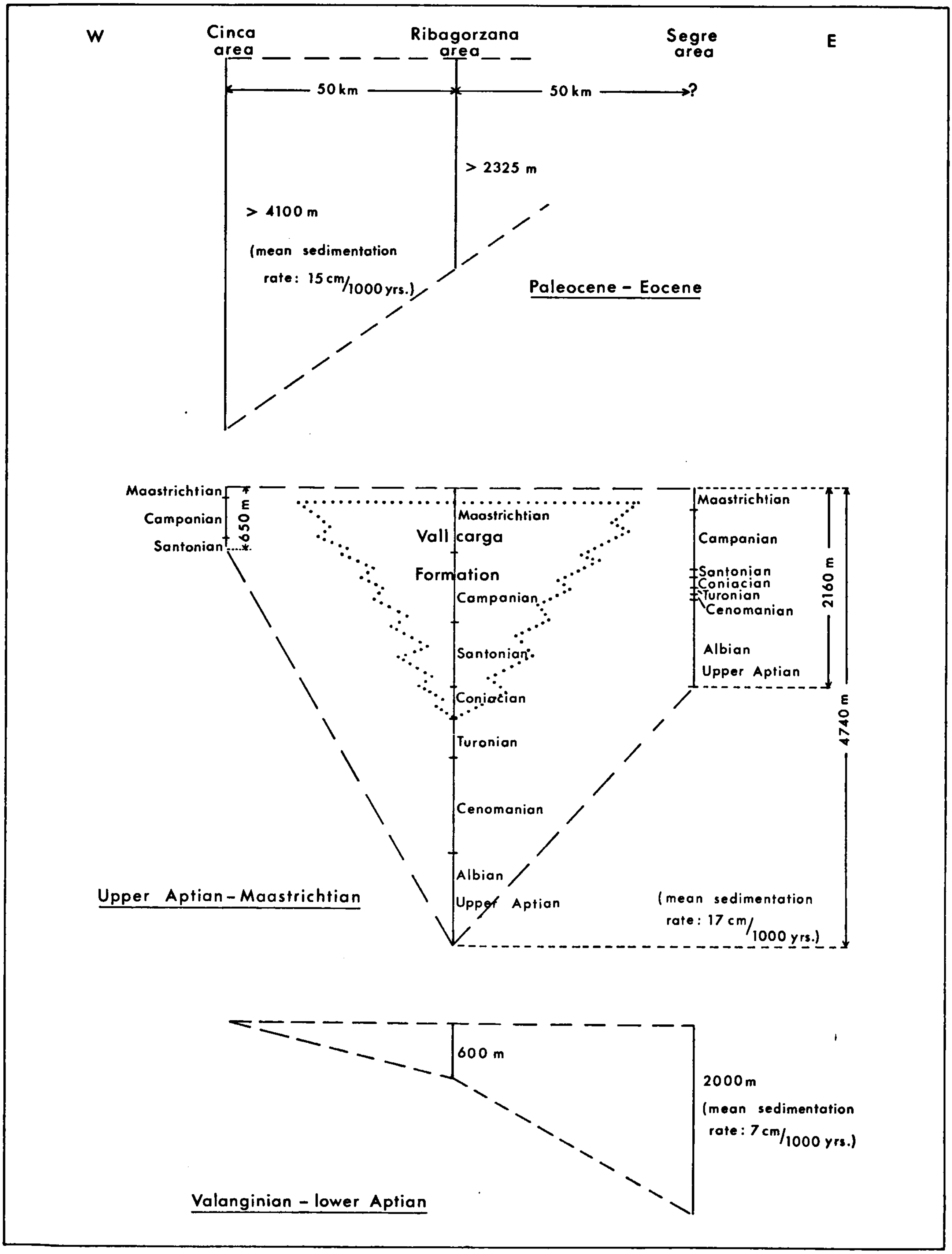


Fig. V-1. Schematic diagram of sediment thicknesses of the Cretaceous and Paleogene in the south-central Pyrenees.

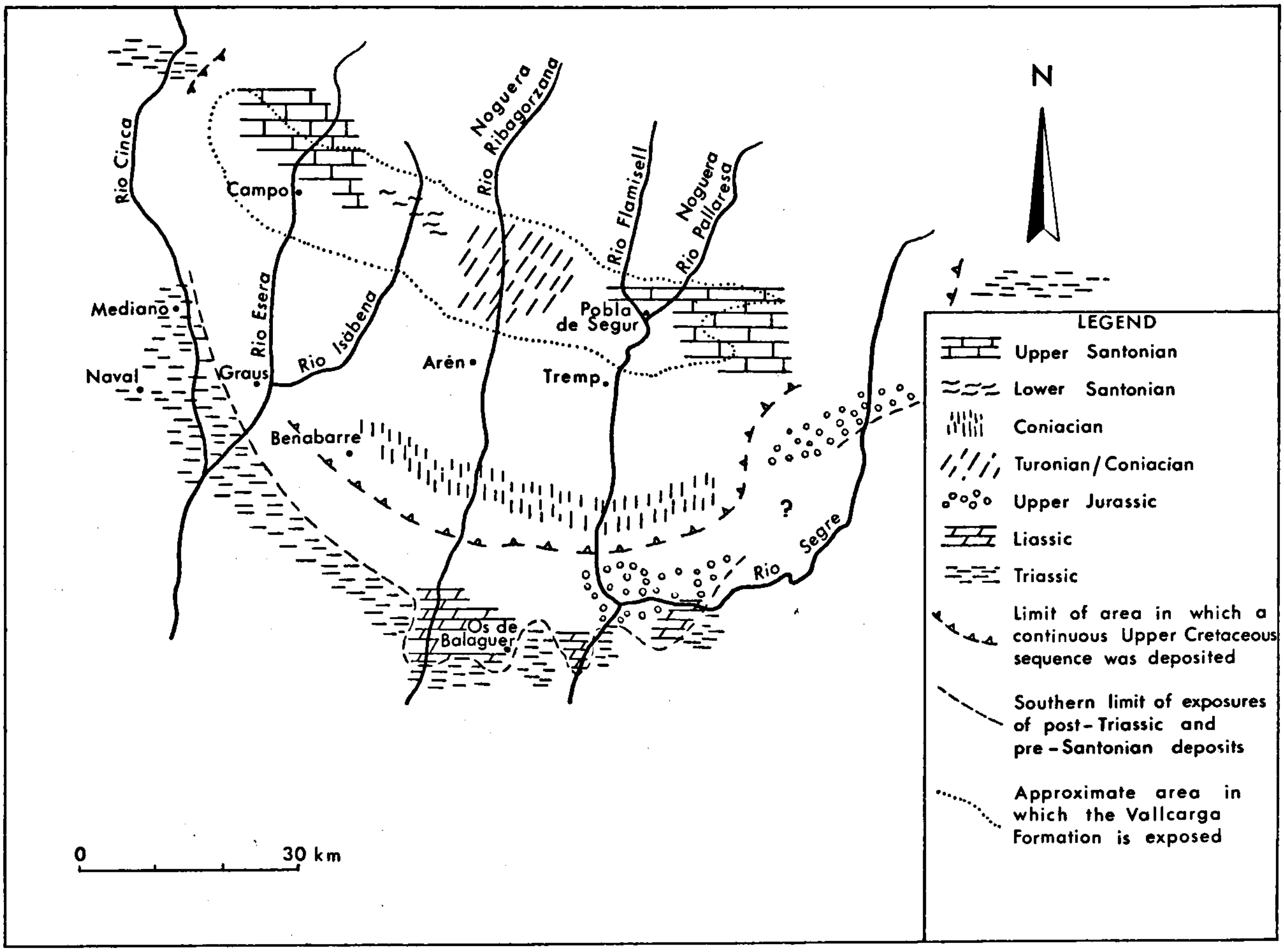


Fig. V-2. Substratum of the Vallcarga Formation and contemporaneous Upper Cretaceous formations in the south-central Pyrenees (after data by Souquet, 1967).

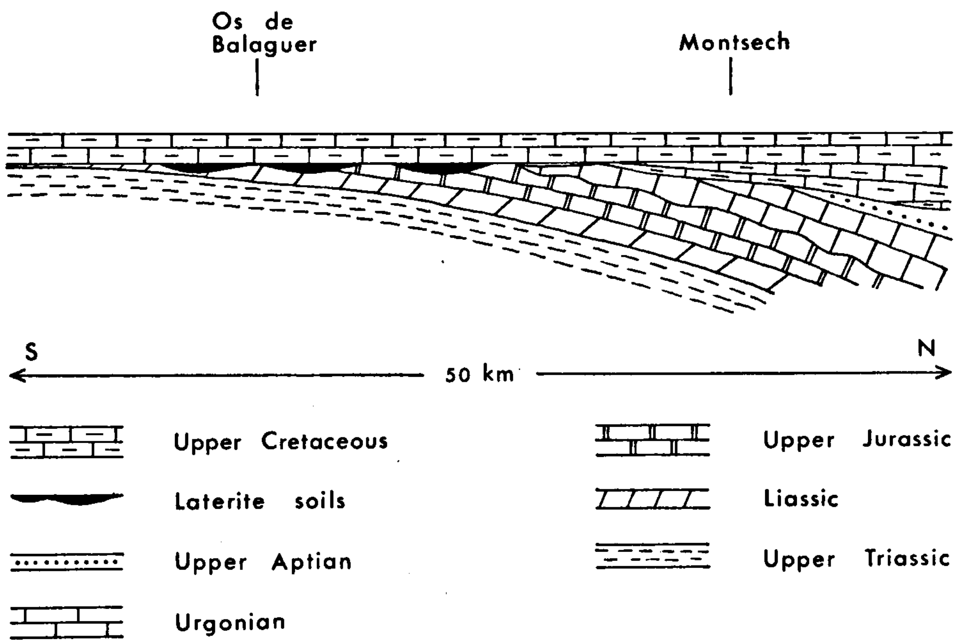


Fig. V-3. Schematic diagram showing the relationships of Mesozoic strata in the Sierras Zone of the south-central Pyrenees (after Souquet, 1967).

DEVELOPMENT OF THE VALLCARGA BASIN

Deposition of turbidites and marls first started in the Ribagorzana area, at a time somewhere between the middle Turonian and early Coniacian, following, or laterally transitional into, a thick micritic limestone series (Santa Fe Limestone Formation, Mey *et al.*, 1968) which in the upper part is strongly slumped and brecciated (Ditzel, 1966). The slump axes measured show an E-W orientation suggesting that a submarine slope was developed striking in that same direction, and dipping either north or south (Ditzel, 1966).

The lowermost 80 m of the Vallcarga Formation (section 13) are here composed of thin-layered fine-grained turbidites and marls. The turbidites are rather poorly graded, showing a fossil content of pelagic Foraminifera similar to those encountered in the Santa Fe Limestone Formation, and a small quantity of additional fine-grained components such as glauconite, quartz, chert fragments, intraclasts and pellets. They were probably deposited by small turbidity currents rather near to the source area, as is demonstrated in exposures on the right bank of the Ribagorzana Valley where a transition was observed from the micritic limestones of the Santa Fe Formation in the east to turbidites and marls in the west. Sporadic flute casts (4 measurements) indicate a current direction from east to west.

During the subsequent period (Coniacian-lower Santonian) limestones were deposited east and west of the Ribagorzana area, whereas in this area turbidite sedimentation continued (Fig. V-4). The limestones lack any turbidite features and reflect tranquil and slow deposition in a protected environment. Some of the limestone formations bear features of a shallow water deposition such as the Congost Formation (upper Coniacian-lower Santonian) exposed in the Pobla de Segur area, built up by small bioherms and reef talus (Mey *et al.*, 1968), whereas other limestone sequences are indicative of a slow sedimentation of lime mud such as Coniacian and lower Santonian deposits in the Isábena region, which towards the west (Esera Valley) are replaced by reddish bioclastic limestones with an abundant shallow water fauna. The area of greatest subsidence slowly shifted in a westerly direction leading to the appearance of turbidites in the Isábena region during the upper Santonian (Fig. V-4). The base of the Vallcarga Formation there is developed as a succession of slumped nodular limestones, coarse breccias and fine-grained turbidites (section 12). During subsequent periods (upper Santonian-Lower Campanian) the Vallcarga basin extended itself over a large area, over a longitudinal distance of about 80 km, more or less parallel to the strike of the axial zone of the Pyrenees.

The birth of this turbidite basin must in all these places have been considered a rather sudden event, marked by the appearance of coarse resedimented deposits (turbidites, breccias) or slumps as a first stage of filling up. This may be observed particularly well in the Esera area, west of the Turbón ridge (Fig. V-5).

Prior to the Vallcarga Formation a thick series (up to 1000 m) of pelagic limestones was deposited there, called the Aguas Salenz Formation (Misch, 1934), gradually decreasing in thickness towards the Turbón area, where they are replaced by a much thinner unit of bioclastic deposits containing an abundant reefal fauna and small bioherms (Souquet, 1967). Consequently, the N-S striking Turbón ridge acted as a shoal in the Santonian sea, bordered in a westerly direction by a

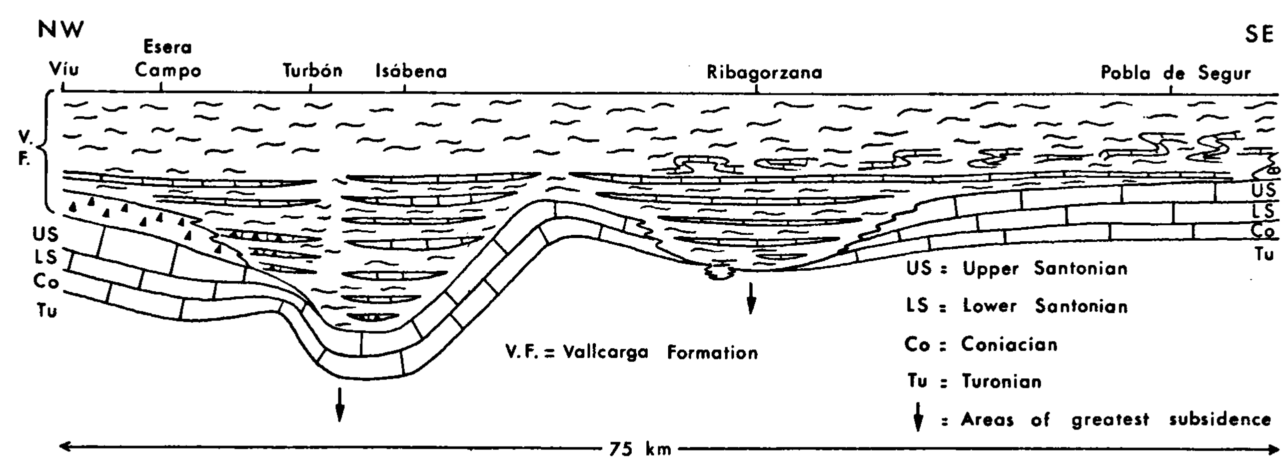


Fig. V-4. Schematic NW-SE cross-section through the Vallcarga Formation. Not to scale.

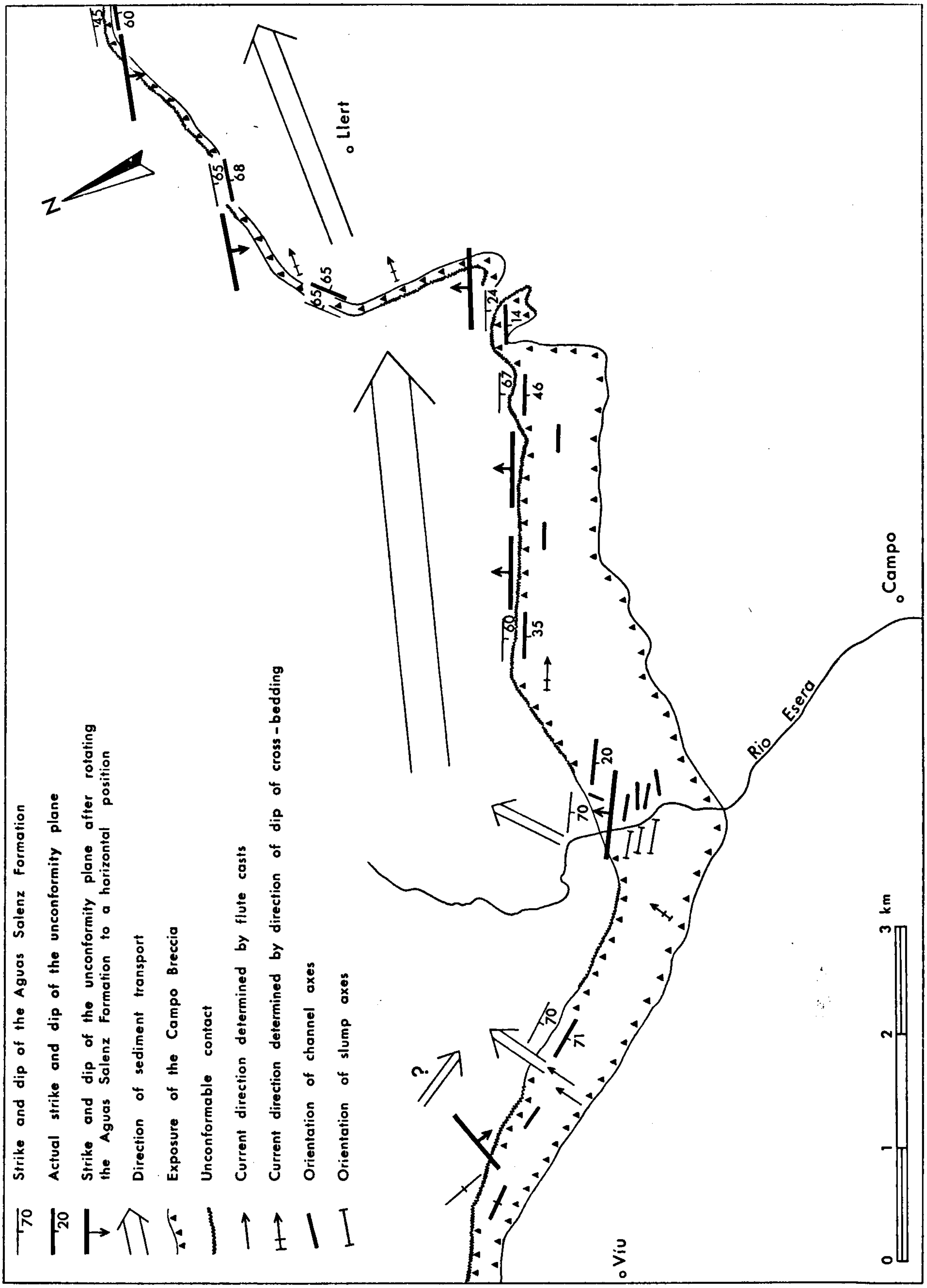


Fig. V-5. Geometry and paleocurrents of the breccia-filled submarine canyon north of Campo.

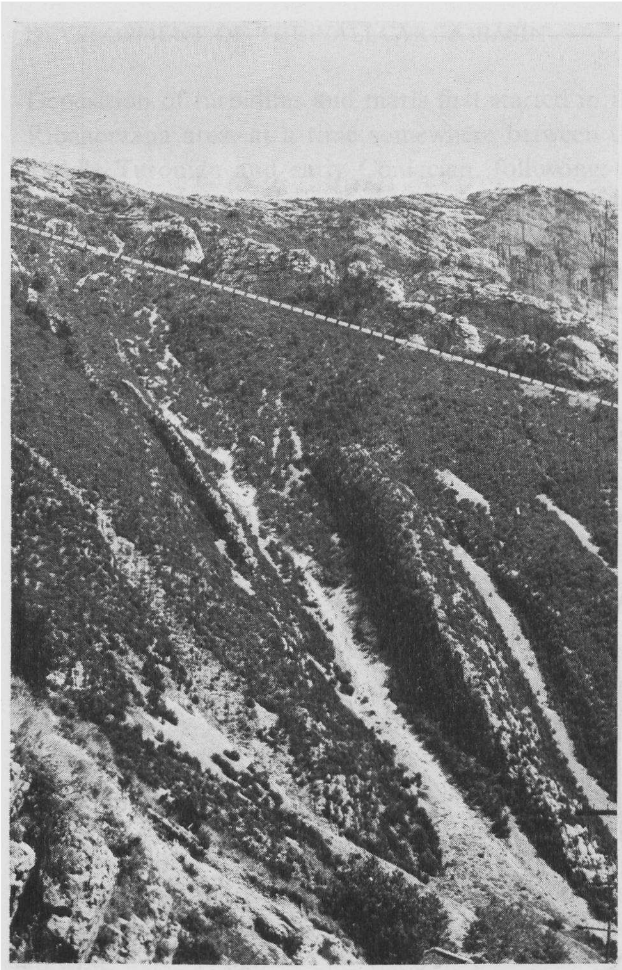


Fig. V-6. Southward dipping unconformity between the Campo Breccia Member (upper part of the photograph) and almost vertical pelagic limestones of the Aguas Salenz Formation. Exposure along the Esera Valley (Photograph by P. J. C. Nagtegaal).

rapidly subsiding area in which dark micritic limestones were deposited, and in an easterly direction by the subsiding area of the Isábena in which nodular limestones were formed during the lower Santonian, afterwards followed by turbidites and marls of the Vallcarga Formation. The presence of this ridge is possibly related to an old fault line in the basement as is suggested by a marked facies change, in the Devonian as well as in the Jurassic and Lower Cretaceous, along the strike of this elongated structure towards the north (Mey, pers. comm.).

During the Upper Cretaceous the old fault line was possibly reactivated leading to a diapire-like behaviour of Keuper sediments due to which overlying strata were tilted upwards. At present, the Keuper is outcropping in the core of the anticlinal Turbón structure. During the upper Santonian the deposition of the Aguas Salenz micritic limestones was suddenly broken off by relatively rapid subsiding movements leading to the emplacement of coarse limestone breccias which form the base of the Vallcarga sequence (Campo Breccia Member). The approach of these unstable conditions is already shown by the top of the Aguas Salenz Formation in which intraformational conglomerates, small slide planes and deposits which may have been deposited by turbidity currents are exposed, indicating that the rate of subsidence accelerated before turning into a catastrophic event. Locally the breccias attain a thickness of 385 m (section 2, Esera Valley) which decreases in an easterly direction (section 3, Llert area). They rest, partly unconformably, upon the Aguas Salenz Formation (Fig. V-6). In a previous

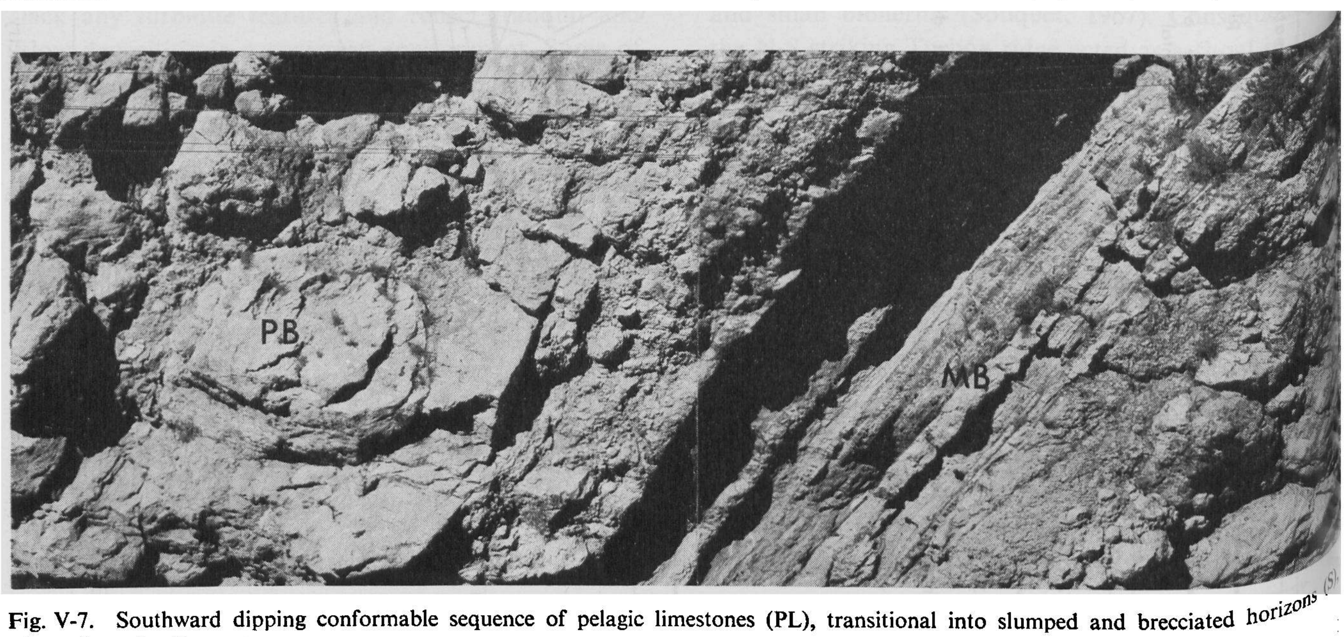
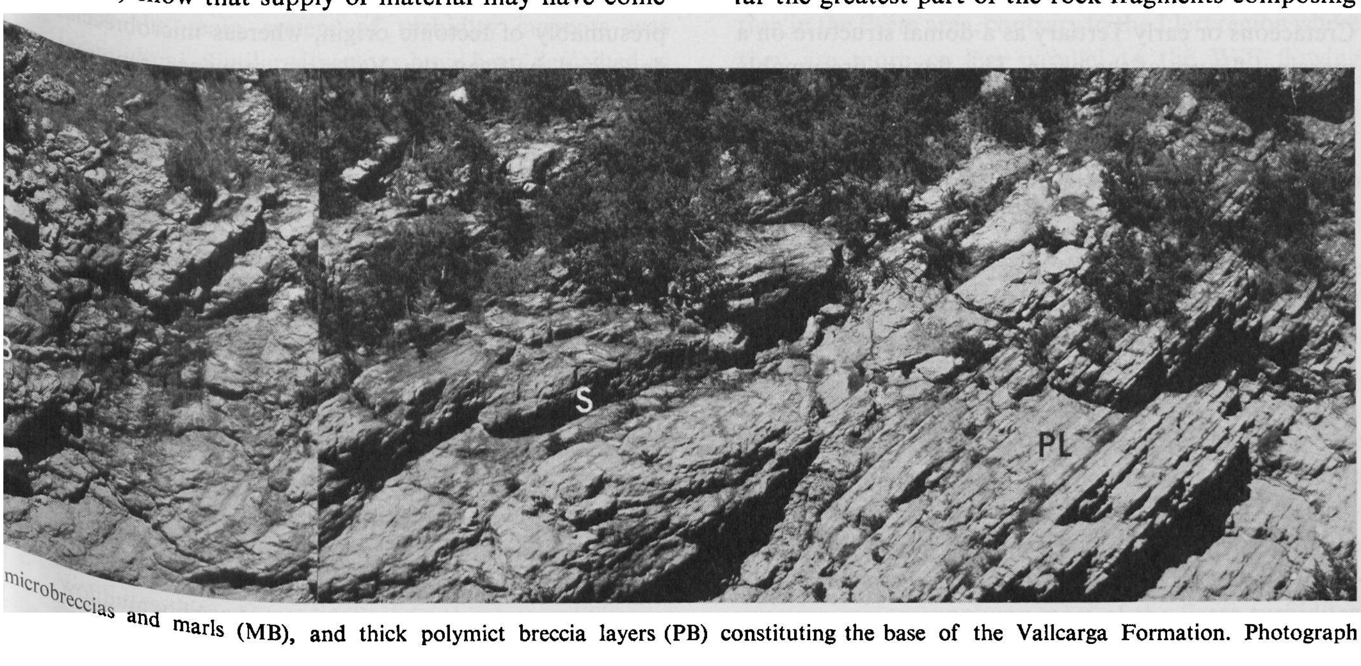


Fig. V-7. Southward dipping conformable sequence of pelagic limestones (PL), transitional into slumped and brecciated horizons (S) taken along the Esera river.

paper (1969), the present author suggested that the breccias were deposited in a submarine canyon with an approximately E-W striking axis (Fig. V-5). This canyon was probably formed as a small graben in the rapidly subsiding basin, bounded by more or less E-W striking fault planes, as is attested by mylonite zones and slickensides which may be observed at the top of the Aguas Salenz limestones unconformably underlying the Campo Breccia. These fault planes do not continue in this latter rock unit. The graben was possibly formed by a downwards sinking either of one block along a single set of faults, or of more blocks, leading to the formation of a complex system of step faults within the graben. This may explain other rupture planes (or faults) occurring at the top of the Aguas Salenz Formation, beneath the contact with the Campo Breccia Member. The conformable contact observed in the lowest part of the Esera Valley (Fig. V-7), and which does not correspond to the canyon axis, may possibly be explained as a deposition on a higher located block within the graben, after the lowest part had been filled. Another explanation may be that it simply represents a tributary of the main canyon, formed by erosion and afterwards filled when the lowest part of the canyon was filled up to the tributary level. Orientation of channel axes (Fig. V-5) shows that in one case (at the base of the Esera Valley sequence) a lateral supply direction may be assumed, whereas succeeding layers show channel axes orientated approximately parallel to the canyon. Measurements in the western part of the Campo Breccia exposures (NE of Víu) show that supply of material may have come either from the south-west or from the north-west; the canyon head was probably located in the vicinity. A western to southern provenance of the breccias was also indicated by Misch (1934), whereas Souquet (1967) and Papon and Souquet (1969) assumed a northern supply area. Our measurements as well as the composition of the limestone breccias (p. 108) do not support this assumption.

Fault-controlled submarine canyons indeed also occur at present, such as the Ganges trough which is bounded at least on one side by a straight slope suggesting a fault (Shepard and Dill, 1966, p. 266); several fault scarp gullies extending from the Californian coast into the sea (Shepard and Dill, 1966, p. 258–263); or canyons limited by faults at the Japanese coast (Shepard and Dill, 1966, p. 131–142). An interesting report on this latter area was written by Yamasaki (1926), who described how as a consequence of the great earthquake of 1923 large fault scarps were formed, along which a sudden subsidence of the sea floor occurred ranging from 50 to 200 m, and an elevation of up to 250 m of both sides of the trough, followed afterwards by deposition of debris with a thickness of up to 230 m. This debris was derived, by a submarine slide, from material which prior to the earthquake had formed the flanks of the troughs.

The lithology of the upfilling Campo Breccia material (p. 108), the current directions measured and the lateral transition from thick-bedded coarse breccias in the west into microbreccias and finer grained turbidites in the east (Llert area) suggest that the source area of by far the greatest part of the rock fragments composing



microbreccias and marls (MB), and thick polymict breccia layers (PB) constituting the base of the Vallcarga Formation. Photograph the breccia layers was located in a south-westerly direction. However, part of the material may also have been derived from the north-west. A nearby source must be assumed, as is indicated by the angularity of the rock fragments and the preservation of large slabs of limestones and Triassic sandstones retaining their original upper and lower stratification boundaries. These features also suggest that they were not derived from a land mass by erosion, neither did they undergo abrasion, but resulted from submarine erosion. Submarine erosion of a simple fault scarp does not seem very likely since in that case an inverse occurrence from youngest rock fragments in layers at the base of the sequence to oldest material at the top of the Campo Breccia Member would be expected; whereas our observations show that rock fragments of Albian to Santonian age are present in all layers and are mixed up throughout the entire sequence. Triassic components have also been encountered in a large part of the Campo Breccia.

The present author thinks that the rapid subsidence of the basin in the Esera area was accompanied by an upheaval at rather a short distance from the newly formed trough, which upheaval may possibly be considered as an anticlinal structure with a fault in its core (Fig. V-8), leading to mobilization of Keuper gypsum and salt unconformably underlying the Albian-Santonian sequence (p. 111). A tentative comparison may be drawn with the well-described anticlinal structure of Estella, Navarra (Ríos, 1948; Pflug, 1967), located at the southern border of the western Pyrenean trough. The Estella anticline was formed in the late Cretaceous or early Tertiary as a domal structure on a marginal flexure of the basin. This flexure presumably corresponded to a major fault in the subsurface cutting into the presaline basement. As a consequence of the continuous upfilling of the basin, Keuper material rose upwards. It contained inclusions of Paleozoic and

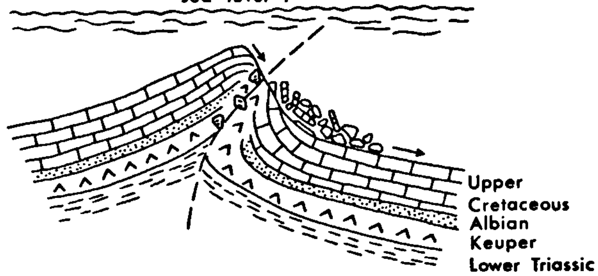


Fig. V-8. Possible origin of the breccia components of the Campo Breccia.

Lower Triassic age detached along the fault plane, and dragged younger material (Jurassic, Cretaceous) upward, reaching the surface when the anticline was faulted.

A similar mechanism may also be postulated for the origin of the Campo Breccia components (Fig. V-8), which probably derived from such a faulted anticline with Keuper in its core, containing blocks of Lower Triassic age and ophites. The presence of such a diapire-like structure in the Esera area does not seem unlikely since it would be located between the N-S striking Turbón in the east, which may have been previously activated by rising Keuper material (p. 138), and the large Peña Sestrales uparching (Selzer, 1934) in the west, located along the ancient basin border and displaying diapire-like structures (e.g. the Mediano diapiric anticline). This is in accordance with the findings of Gussow (1968), who stated that the formation of domes due to diapire-like activity becomes progressively younger from the inner parts of the subsiding basin towards the margins. Collapse of one of the flanks (e.g. leading to large slabs), sudden tectonic movements related to the fault plane activity, oversteepening of the structure, all may give rise to the formation of brecciated polymict material, which was afterwards rapidly carried to the site of deposition. The decreasing activity of these tectonic movements, related to a slowing down in the rate of subsidence, may be responsible for a gradual disappearance of coarse supply in the Vallcarga basin. The coarse breccias of the lower part of the formation in the Esera area are therefore directly attributable to a large trigger effect presumably of tectonic origin, whereas microbreccias deposited halfway the Vallcarga sequence represent deposition by currents triggered either by smaller tectonic movements or by simple overloading on a slope, also brought about by smaller movements in the source area.

It is, however, beyond all doubt that the emplacement of the Campo Breccia is related to a high degree of local subsidence of the Upper Cretaceous basin, constituting a beautiful example of tectonic sedimentation.

The latest part of the Vallcarga basin to subside was the Pobla de Segur area, in which during the upper Santonian-lower Campanian nodular limestones containing a characteristically high content of glauconite (Anserola Formation, Mey *et al.*, 1968) were formed first, to be followed by turbidites and pebbly mudflow deposits. Already in the upper part of the Anserola

Formation the increasingly unstable conditions are announced by various slump horizons.

Consequently, over a distance of about 80 km along the strike of the Pyrenees (NW-SE), a deeper turbidite basin was gradually formed during middle Turonian/ early Coniacian to lower Campanian times. It may be considered an area of strong local subsidence, brought about in the western part (Esera area) by fault-controlled movements, which in other parts (Isábena, Ribagorzana, Pobla de Segur) may be related to flexures.

FILLING UP OF THE VALLCARGA BASIN

Paleocurrents

Paleocurrent data are summarized in Fig. V-9, in which measurements of the direction of flute casts and of the dip of cross-stratification are directly plotted, whereas the orientations of channel axes, groove casts and slump axes are shown per region in the form of rose diagrams.

According to these data, three major regions may be distinguished in the Vallcarga basin:

a. an eastern region located at present between the Ribagorzana and Noguera Pallaresa Valleys, in which a persistent current direction is shown from east to west, with slump axes (in the Ribagorzana area) parallel to this direction indicating that the slump movements occurred down the sides of the basin in which the main system of turbidity currents was longitudinal. Similar situations have been described in other turbidite basins (Marschalko, 1961; Murphy and Schlanger, 1962).

b. The area located east of the Turbón ridge, in which measurements indicate currents flowing into a north-westerly direction. In the Vilas del Turbón area paleocurrents were also recorded showing an opposite direction, towards the south-east. Groove casts are chiefly orientated in the same direction as flutes, whereas channel axes may show a somewhat wider spreading. South of Vilas del Turbón they indicate N-S orientated currents. Slump axes do not show a significant mean orientation, varying in the Isabena Valley between values parallel to that of the main current direction to orientations at an angle of 60° to that of the flutes. Near Vilas del Turbón this latter difference was also commonly observed.

c. The area west of the Turbón ridge is characterized by sporadic occurrence of flute casts, which may be indicative of a relative proximality to the source area of turbidites (p. 79). This is in agreement with the coarseness of most deposits encountered here (see sections 1, 2, 3). Paleocurrent directions here have been inferred from measurements of cross-bedded strata such as relatively fine-grained microbreccias or calcirudites, orientation of channel axes of coarse-grained layers, groove casts and some flute casts. They all indicate a wide pattern of current directions: the currents may have flowed from SW to NE, from W to E or even from NW to SE in the Esera area, changing into a persistently easterly flow direction in the Llert area. Along the southern part of the Turbón ridge flute casts were observed indicating a current direction towards the north. Channel axis measurements were all carried out at the base of microbreccia layers showing a composition analogous to that of the thick breccias of the Campo Breccia Member, thus suggesting derivation from a same source area. This is in agreement with the transport direction recorded in both type of deposits. In the Llert region, too, flute casts were recorded, indicating currents in a south-easterly and north-westerly direction, the latter corresponding to those of the Vilas del Turbón area. It may therefore be concluded that the northern part of the Turbón ridge was the deepest part of the turbidite basin, to which most of the currents were directed. Consequently, this explains the presence of turbidites to the west and the east of the Turbón structure showing opposite current directions. Slump axes show a varied orientation in the Esera area, contrary to the Llert region where they are more or less parallel to the W-E flowing currents.

Lateral variations in sediment supply and deposition in the Vallcarga basin

In general, the directions of paleocurrents in a turbidite basin depend on the basin geometry and the location of areas supplying material for the turbidity currents. This is well illustrated in the Vallcarga basin, where, according to the paleocurrent data previously indicated, three major source areas must be considered: a source area supplying material for the turbidites deposited in the Ribagorzana and Pobla de Segur areas; a south-eastern source area from which the Isábena and Vilas del Turbón turbidites were derived; and a south-western region supplying most of the Esera turbidites.

Where current directions are longitudinal, as in a

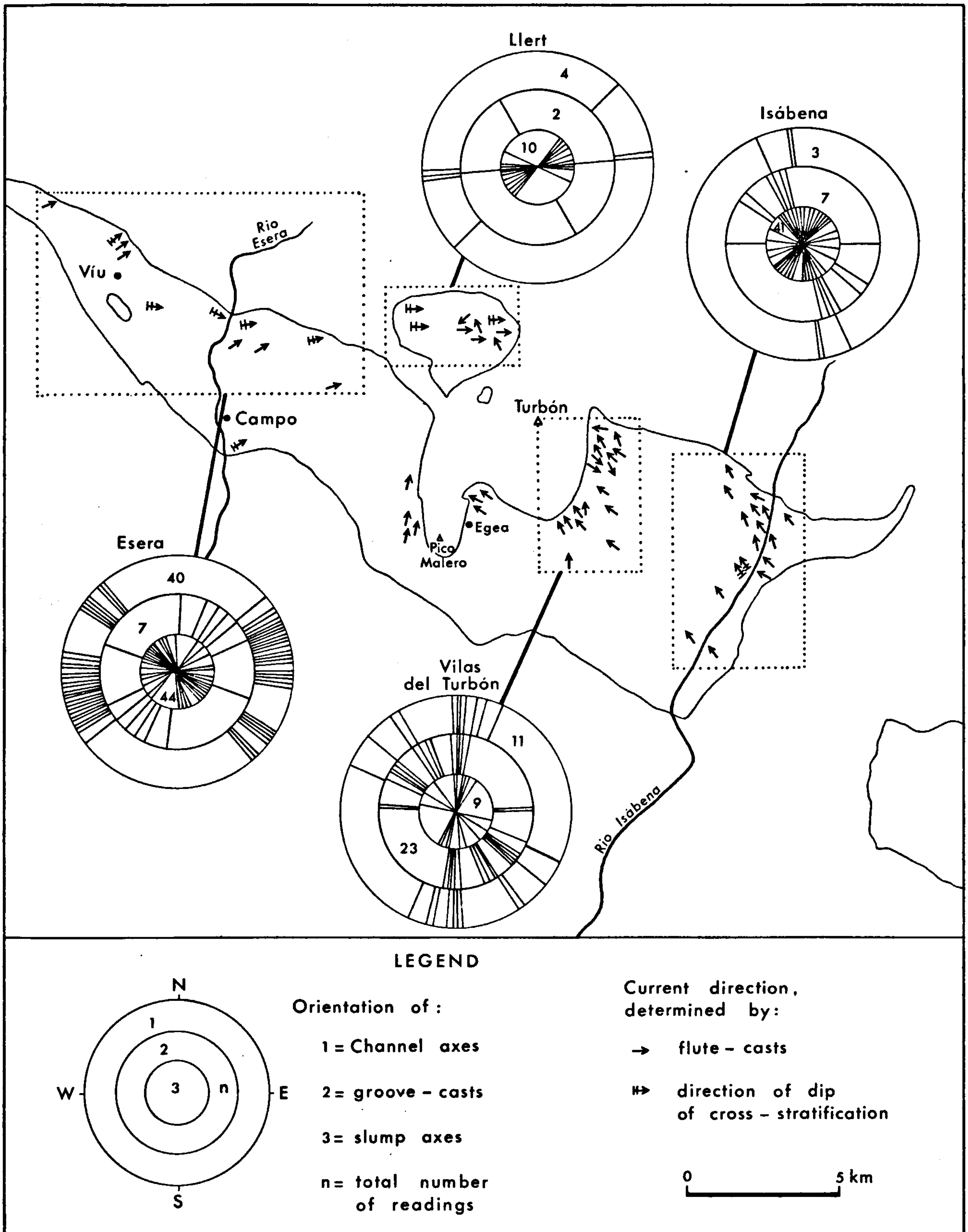
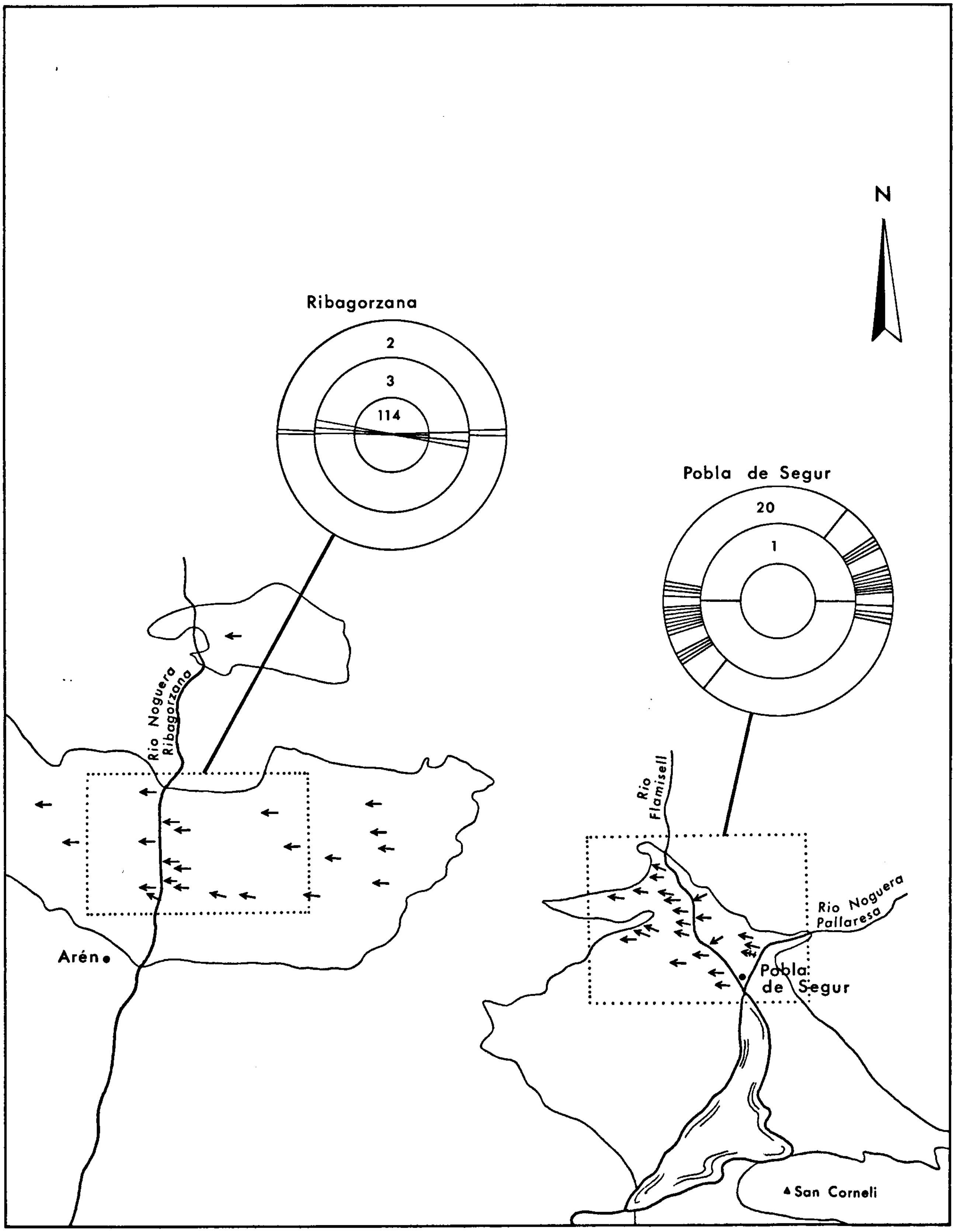


Fig. V-9. Paleocurrent map of the Vallcarga Formation. Measurements in the Pobla de Segur area after Wiersma (1965), Masselink (1965) and Nagtegaal (pers. comm.); readings in the Ribagorzana area partly after Ditzel (1966).


N
Ribagorzana
2
3
114
Pobla de Segur
20
1
Rio Noguera Ribagorzana
Arén
Rio Flamisell
Rio Noguera Pallaresa
Pobla de Segur
San Corneli

late stage those of the Ribagorzana and Pobla de Segur areas, it cannot be established whether the turbidites were supplied from a northerly, easterly or southerly located source area. In the Pobla de Segur region, however, a southern provenance does not seem very likely, since immediately south of Pobla de Segur an anticlinal structure (San Corneli) is situated which was already largely present during deposition of the Vallcarga sediments, as may be deduced by the absence of turbidites at its southern flank and the decrease in total thickness of the Vallcarga Formation from 1400 m near Pobla de Segur to 300 m of marls and nodular limestones south of the San Corneli (Rosell Sanuy, 1965). Consequently, in this area turbidity currents originated from a northern to eastern source area, changing their current direction to the direction observed at present after reaching the longitudinal axis of the basin.

Field observations indicate that turbidites in the eastern part of the basin, i.e. the Pobla de Segur area, are relatively more proximal than those found in the Ribagorzana region, and that in this direction pinching out of layers is quite common. Section 11 shows that in the Ribagorzana area the Vallcarga sequence is mainly composed of thin-layered fine-grained turbidites lacking those thick coarse-grained deposits occurring near Pobla de Segur (Wiersma, 1965).

Further to the west of the Ribagorzana Valley, it was noted that all turbidites decrease in thickness to extremely thin tails (less than 1 cm) or disappear altogether (Fig. V-4), the turbidite sequence being replaced there by a series of blue marls, attaining thicknesses of up to 1000 m. This feature indicates that in between the Isábena Valley in the west and the Ribagorzana Valley in the east there existed an approximately N-S striking submarine swell, dividing the Vallcarga basin into two parts and preventing turbidity currents coming from the east from continuing further on. This picture closely resembles the one found by Moore (1966) in marine basins off the coast of southern California in which sequences of turbidites were noted which terminated abruptly against topographic barriers.

West of this swell an area of maximum subsidence existed (Fig. V-4) in which a thick succession of relatively proximal coarse-grained deposits (section 12) was laid down, derived by turbidity currents from a southern or south-eastern source area. Turbidites in this region flowed towards the north-east, in the direction of the deepest part of the basin located north of the Turbón ridge.

The decrease in thickness of the formation towards the southern part of this structure (sections 4, 5, 6, 7, 8, 9, 10, 11) shows that a local height must have existed here, which partially or totally prevented the deposition of turbidites. This is demonstrated by the occurrence of a large cap-like non-deposition horizon composed of a Fe-Mn-oxyde crust on top of the Pico Malero, south of Egea, a satellite massif of the Turbón. The non-deposition horizon is directly overlaid by the blue marls (Salas Member) of the upper part of the formation. Consequently, the presence of this height, which may be considered a sea-mount, determined the by-passing of turbidity currents to its north. Only after the adjacent parts of the basin had been filled up to the level of the sea-mount did renewed deposition begin here. This may possibly be related to a decrease in the rate of subsidence at both sides of the sea-mount followed by slow subsidence involving this part of the basin as well.

The location of this height in the Vallcarga basin suggests that the ancient N-S striking ridge, which existed prior to the deposition of the Vallcarga Formation (p. 136), moved in a southerly direction at the time when the basin originated. Part of the sea-mount was subjected to submarine erosion, as is indicated by large pebbly mudflow deposits at both flanks containing lithified rock fragments similar to the limestones building up the height. In recent environments such submarine erosion has been reported, by Jones and Funnel (1968), of an Upper Cretaceous sea-mount in the Bay of Biscay where at its base eroded pebbles of Maastrichtian age are found within Quaternary turbidites and marls.

It may be concluded that the Vallcarga basin consisted of three parts, separated partially or totally by swells, each of these filled up by turbidites derived from distinct source areas. Its geometry may be compared with the bottom configuration of the sea off southern California (Emery, 1960; Moore, 1966), in which several large basins and troughs are known, separated by ridges and in the course of being filled up by turbidites.

Vertical filling up of the Vallcarga basin

The rather quick subsiding which caused the formation of the Vallcarga basin, also led to an initial supply of coarse material, deposited as breccias, slumps, pebbly mudflows or coarse-grained turbidites (sections 1, 2, 3, 6, 12) which built up the lowest part of the formation and which are followed by a succession of turbidites and marls alternating with pebbly mudflows. Several slump levels have also been recorded higher up in the formation, demonstrating the tectonic instability of the basin. This sequence gradually passes upwards into a thick series of bluish-grey marls (Salas Member) in which no recognizable turbidites occur.

At the base of this Salas Member a thick unit (about 200 m) of slumped turbidites, marls and olistoliths is observed in the easternmost area, between the Ribagorzana Valley and Pobla de Segur, which unit has been called the Pumanous Member (Wiersma, 1965; Ditzel, 1966). Even further to the west, in the Isábena Valley, several horizons occur composed of deformed nodular limestones and marls forming large olistoliths, with a Santonian or Campanian pelagic fauna which suggests that they were formed shortly before sliding down. The direction of oversteepening of deformed flanks of the slumps indicates a northern origin. Consequently, north of the basin these types of deposits must have occurred, having been deposited almost simultaneously with the Vallcarga turbidites in the south. They were possibly derived from a shallower part of the basin. Analogous deposits have been described in localities in the north (Seira section, p. 116). Tectonic activity, increasing locally, must be held responsible for the formation of these crumpled and slid down deposits.

The decreasing proportion of turbidite deposition, as well as the decrease in maximum grain size in these sediments towards the top of the Mascarell Member, has to be related, not to an increasing distance to the source areas, but to the gradual disappearance of the factors controlling the origin of turbidity currents. Such may be the slowing down of subsidence as well as the absence of material building up turbidity currents, both possibly related to a smaller degree of tectonic movements in the basin as well as in the hinterland. These conditions may have differed in various parts of the basin. For instance, incidental deposition of coarse-grained sandy turbidites continued during a longer period in the Isabena Valley (section 12) than in the area west of the Turbón ridge (sections 1, 2). These variable conditions may also have determined the supply of turbidites with a somewhat different composition and coarser grains sizes as, for instance, those building up the 770–980 m interval of the Ribagorzana section in which a larger supply of terrigenous material and limeclasts was noted than in turbidites previously deposited. However, these slight differences do not alter the general picture of the Vallcarga Formation, showing a gradual decrease in turbidity current deposition, followed by sedimentation of autochthonous marls.

The upper 100 m of the Vallcarga Formation lack evidence of turbidite deposition, with the exception of thin-layered sandy calcarenites in the Esera Valley, which may have been emplaced by small turbidity currents. In other areas this part of the formation is composed of nodular limestones (Ribagorzana and Isábena Valleys) or calcareous mudstones (Viú area), which were probably deposited in a rather shallow environment, since they are followed by thick-layered coarse-grained calcarenites sometimes showing large-scale cross-bedding (Arén Formation, Mey *et al.*, 1968). The fossil content of these deposits and their sedimentary features (such as crcss-bedding, scour-and-fills, clean-washed texture) suggest deposition rather close to a shore. During the deposition of the upper part of the Vallcarga Formation the subsidence may have slowed down, so that the sedimentation rate exceeded the rate of subsidence by which this regressive development was attained.

PALEOGEOGRAPHY

The stratigraphic work of Souquet (1967) and our own observations (see Chapter IV) provided sufficient data to evaluate the paleogeography of the Upper Cretace-

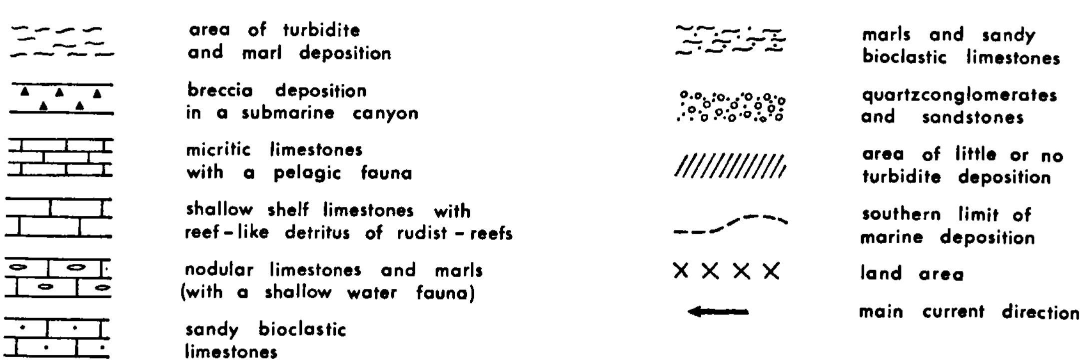


Fig. V-10. Legend to the lithofacies maps.

ous basin during the deposition of the Vallcarga Formation.

Coniacian and lower Santonian (Fig. V-11)

As was previously stated (p. 136), turbidites were first emplaced in the Ribagorzana area by currents flowing in a westerly direction and carrying intrabasinal material. In the other parts of the Upper Cretaceous basin, deposition continued on a shallow shelf, as is shown by the faunal content. The Turbon ridge was an area of little or no subsidence, in which limestones composed of bioherms and reefal detritus were deposited, flanked in the west and east by regions of greater subsidence in which thick series of pelagic limestones were laid down (Aguas Salenz Formation).

To the south, the basin was bordered by a land mass with Upper Triassic, Liassic and Upper Jurassic at the surface; this probably was an area of low relief, from which no material was derived. Terrigenous supply (quartz sand to quartz pebbles) came from the north-east and was carried along the coastline in a south-westerly direction by longshore currents. The basin was probably also closed in the west (by the Aragón swell) and in the east (Ampurdán swell) so that the only possible connection with the ocean may have been in a northerly direction.

Upper Santonian (Fig. V-12)

During upper Santonian times the turbidite basin extended in a westerly direction due to a rapid subsidence. In the Esera area this led to the formation of a fault-controlled submarine canyon, which was filled with extremely coarse material coming from directions between north-west and south-west, and transported inside the canyon in an easterly direction, interfingering there with turbidites provenient from a south-easterly direction. To the south of this deepest part of the basin, a submarine sea-mount (S of the Turbón) existed which prevented sedimentation by turbidity currents. Another submarine swell was present east of the Isábena Valley, due to which swell turbidity currents flowing from east to west in the Ribagorzana area did not continue further westwards.

The upper Santonian sea slowly transgressed in a southerly direction, over the Upper Triassic land mass. Terrigenous material was still carried in a south-westerly direction by longshore currents, being deposited even further in this direction than during the

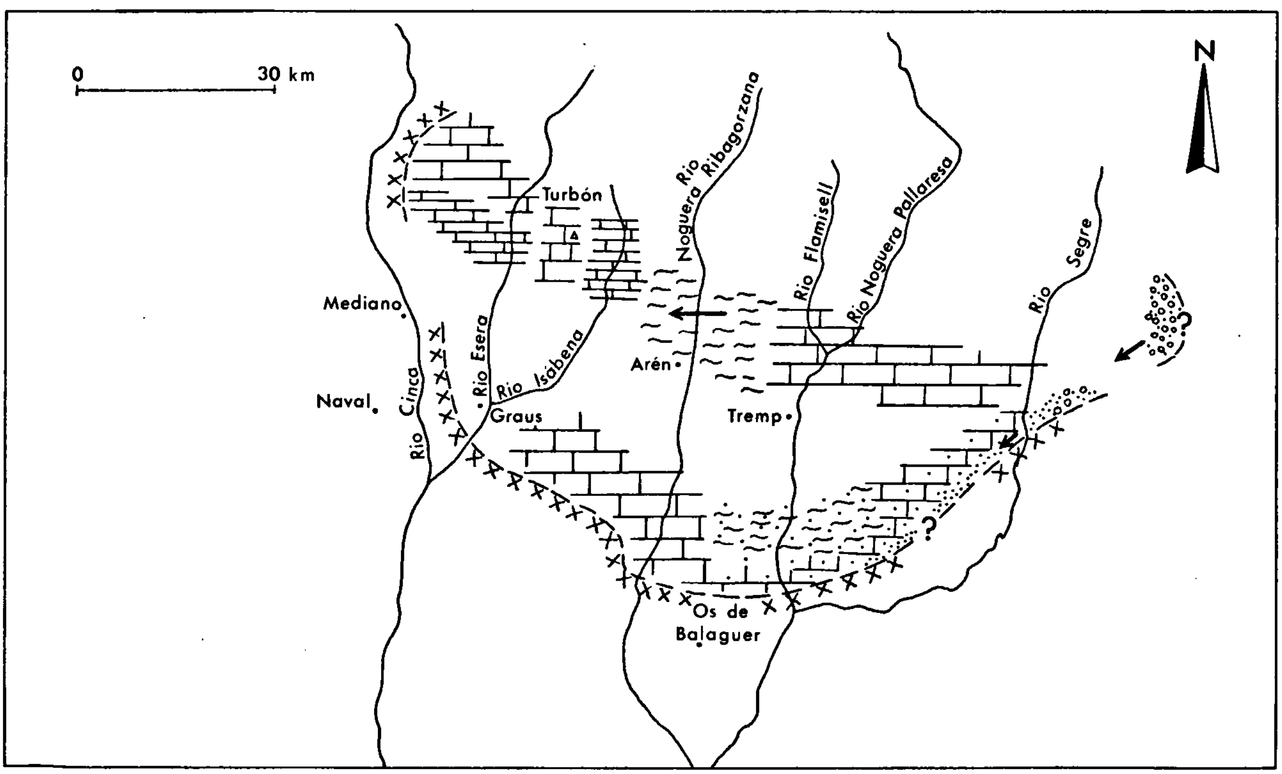


Fig. V-11. Lithofacies map of the lower Santonian.

lower Santonian. These longshore currents formed a sandy unit located between the coastline in the south and a shallow shelf on which nodular limestones and marls originated containing abundant reefal detritus. It seems likely that this calcareous material, together with terrigenous admixture, was carried northwards by turbidity currents into the Vallcarga basin, as is indicated by the mean paleocurrent direction in the Isábena area.

In a westerly direction, the upper Santonian sea transgressed the Aragón swell, thus establishing a connection with the marine basin present in the western Pyrenees.

Campanian (Fig. V-13)

During Campanian times the sea acquired its greatest extension in the south-central Pyrenees. The turbidite basin extended along the present strike of the mountain range, over a longitudinal distance of 80 km, and was surrounded at its eastern, southern and western sides by a shallow shelf on which calcareous deposition took place. Deposition of turbidites continued during the lower Campanian, these turbidites being emplaced by currents flowing approximately in the same directions as in previous times. The composition of these turbidites reflects the bioclastic deposits of the source areas.

Some terrigenous admixture also occurs. In the Esera area this may have been supplied by strong currents on a shallow shelf flowing from the north-eastern granite massifs (Cinca area) in a southerly direction, and being subsequently picked up by turbidity currents flowing in north-easterly directions. The terrigenous components of the Ribagorzana and Pobla de Segur turbidites were probably derived from northern or eastern Paleozoic rocks, whereas the quartz content of the Isábena turbidites and the locally associated phyllite grains suggest that far to the south, on the Upper Triassic land mass, exposures occurred (probably of Paleozoic age) which supplied these components. Unfortunately, these supposed outcrops are at present covered by a thick series of Cainozoic rocks.

During the upper Campanian, large turbidity currents ceased to exist, so that only a thick series of blue marls was deposited, whereas in shallower areas surrounding the deeper basin the sedimentation of bioclastic limestones on a shallow shelf continued.

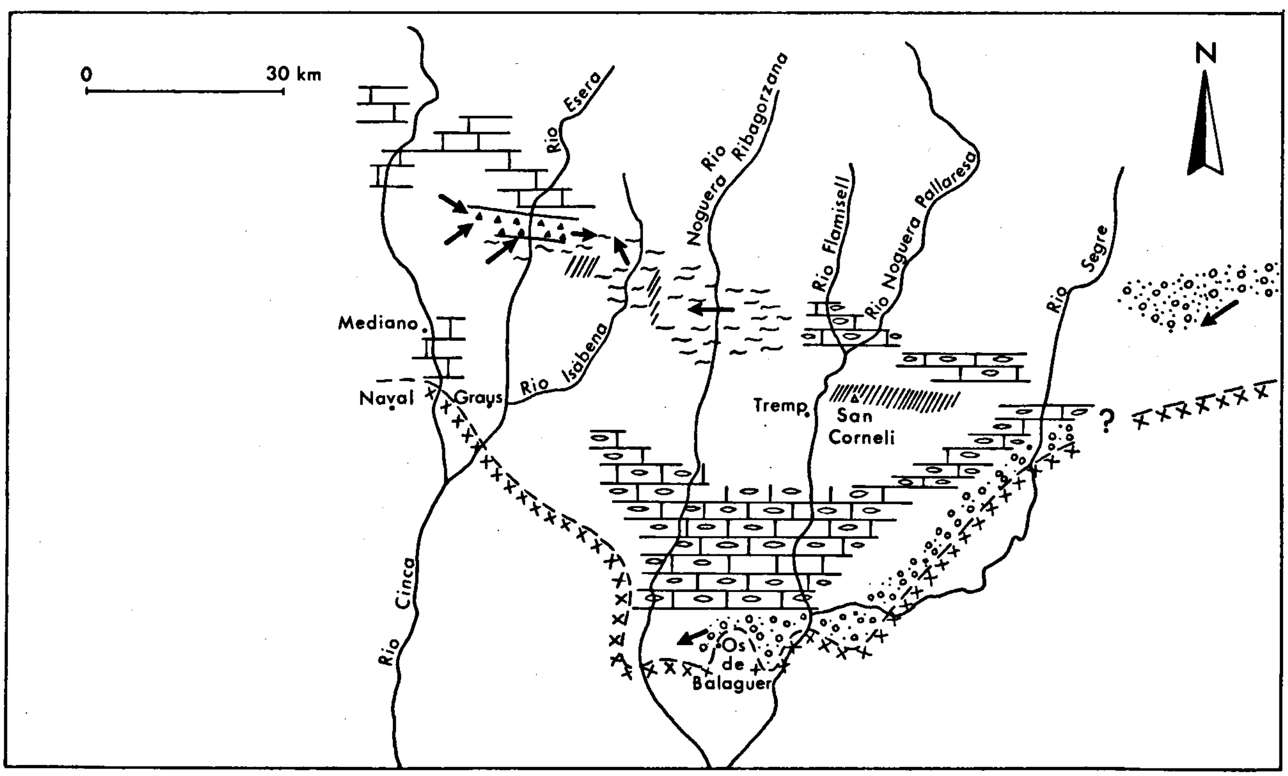


Fig. V-12. Lithofacies map of the upper Santonian.

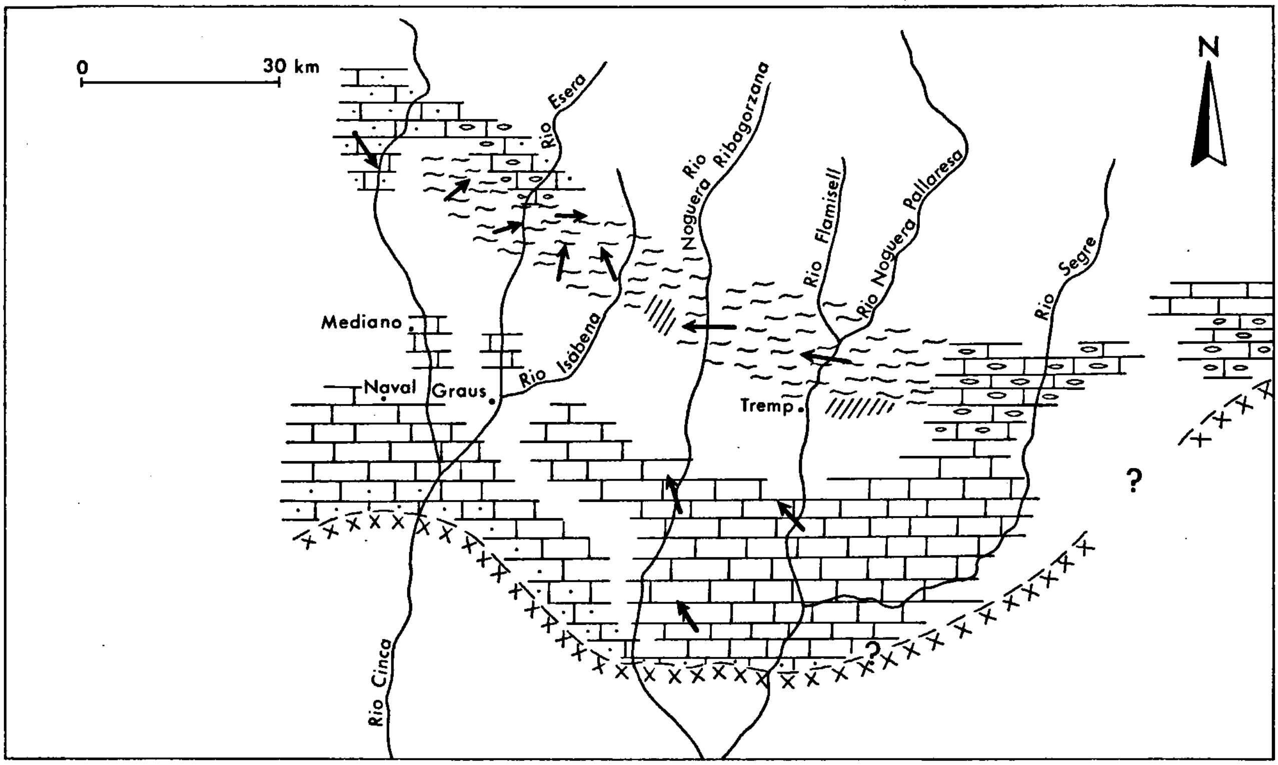


Fig. V-13. Lithofacies map of the Campanian.

Maastrichtian

During Maastrichtian times, the basin was characterized by a regressive development due to which the sea was restricted to the same area which it had occupied during lower Santonian times. A large supply of terrigenous material caused the deposition of thick-bedded sandstones and calcarenites all over the area, gradually following the bioclastic limestones in the south and east and the blue marls of the Vallcarga Formation in the north. Beyond the marine basin either no sediments were deposited or they were laid down in a continental environment, building up reddish, purple or violet marls and sandstones belonging to the Garumnian facies. Only in a westerly direction (near the Cinca Valley) were thick shallow marine deposits laid down, suggesting that the area of maximum subsidence slowly shifted to that area.

CONCLUDING REMARKS

This paper shows that a turbidite basin was formed in the south-central Pyrenees during the Upper Cretaceous times, as a consequence of a relatively sudden subsidence, related to a westwards shift of the Cretaceous basin. This subsidence may possibly be viewed as a sinking downwards of part of the Cretaceous basin floor along flexures, except in the Esera area where faulting led to the formation of a submarine canyon.

The turbidite basin did not exceed a longitudinal NW-SE distance of 80 km and a lateral extension of probably no more than 20 km (Esera region), being surrounded at its western, southern and eastern sides by a shallow shelf on which mainly bioclastic limestones were deposited. A transition into a northern shallow depositional area was only noted north of Campo (Seira area). Consequently, not much can be said regarding the question as to whether an emerged continuous axial zone existed in the central Pyrenees during the Upper Cretaceous.

Field observations show that only in the north-western (Cinca Valley) and north-eastern (Segre Valley) areas were crystalline massifs (granodiorites) exposed to erosion supplying material for the Upper Cretaceous basin, whereas in the region in between no flow directions of northern provenance were observed. Nevertheless, by far the greatest quantity of the most stable minerals such as quartz and muscovite in the turbidite basin may have come from older Mesozoic or Paleozoic rocks outcropping north of the Vallcarga basin. However this be, the diminishing sedimentation thickness in a northerly direction in the Cinca Valley, Seira area and Segre Valley indicates the presence of a zone in which the rate of deposition during the Upper Cretaceous was not so great as in the Vallcarga basin, and which possibly continued along the largest part of the central Pyrenees, parallel to the present strike of the axial zone.

Compositional analyses of the Vallcarga turbidites show that they are largely built up of intrabasinal calcareous material such as fossils, intraclasts, pellets and micrite. Rock fragments composing the breccia layers at the base of the formation were also derived from an intrabasinal erosion of Mesozoic formations.

If we consider the turbidites filling up the basin in terms of proximality or distality (Walker, 1967, p. 32), it becomes clear that only in the Pobla de Segur and Ribagorzana areas may a transition be noted from proximal to distal turbidites, whereas the features of the layers in the Esera and Isábena areas are typical of very proximal turbidites (features summarized by Walker, 1967).

The dimensions of the Vallcarga basin were evidently too small to permit of a great flow distance, so that all turbidites were deposited rather proximal to the source area.

A depth rating for the Vallcarga basin cannot be concluded, except in those parts of the basin where large-scale cross-bedded layers (e.g. Isábena sequence) were deposited, in that case proving a depth of up to 80 m (p. 89).

The Vallcarga Formation can by no means be compared with some other well-known ancient turbidite sequences such as in the Carpathians, Apennines, etc., since its deposition is not directly related to a major orogenic phase. This turbidite basin simply appears to be a small local area of high subsidence within the shallow Upper Cretaceous sea, which basin was filled up by longitudinal and lateral turbidity currents.

SAMENVATTING

Deze studie behandelt de primaire lithologie, sedimentaire strukturen, afzettingsgeschiedenis en paleogeografie van een uit het Boven-Krijt daterend turbidietbekken in het centrale deel van de Zuidelijke Pyreneeën en geeft bovendien een kort overzicht van de samenstelling en het milieu van afzetting van de omringende even oude sedimenten. Gedurende het Coniacien tot onder-Maastrichtien werd, over een afstand van ongeveer 80 km langs de huidige strekking van de Pyreneeën, een dik pakket van kalkturbidieten, modderstroomafzettingen, slumps, sedimentaire kalkbreccies en mergels afgezet. Zij worden samengevat onder de naam Vallcarga-Formatie. Deze formatie kan worden verdeeld in een onderste deel (Mascarell Member), bestaande uit geresedimenteerde afzettingen en autochtone mergels, en een bovenste deel (Salas Member), dat is opgebouwd uit blauw-grijze mergels waarin turbidieten schaars zijn.

De sedimentaire strukturen die voorkomen in de Vallcarga-Formatie, zoals flute- en groovecasts, geulen, loadcasts, gegradeerde gelaagdheid, parallele laminatie, stroomribbel-laminatie, convoluties en slumps worden beschreven in verband met milieu waarin zij werden gevormd. Zij geven aanwijzingen voor een relatief proximale plaats van afzetting in een troebelingsstroom, wat mede wordt aangetoond door met grof materiaal opgevulde geulen, dikgelaagde, niet-gegradeerde of gegradeerde kalkbreccies en microbreccies. Deze conclusie wordt ook aannemelijk gemaakt door de aanwezigheid van dikke lagen met een mega-scheve gelaagdheid, welke in sommige delen van de opeenvolging voorkomen. Afzettingen met het voorkomen van proximale turbidieten en van distale turbidieten komen soms echter samen voor in eenzelfde opeenvolging, hetgeen aantoont dat de afmeting van een troebelingsstroom de belangrijkste factor is bij de schijnbare proximaliteit of distaliteit van hun afzettingen. Er zijn vrijwel geen argumenten gevonden die wijzen op een transport door andere types bodemstromen dan troebelingsstromen. De troebelingsstroom-hypothese verklaart op de meest logische wijze de eigenschappen van de meeste geresedimenteerde lagen. Slumps komen overal in de formatie voor; hun assen zijn meestal evenwijdig aan de stroomrichting van de troebelingsstromen. Hierbij zijn alle mogelijke overgangsstadia van kleine breukjes tot chaotische massa's van meerdere turbidieten en mergels waargenomen.

Een nauwkeurig onderzoek van de oorspronkelijke samenstelling van de Vallcarga-afzettingen laat zien dat zij voornamelijk zijn opgebouwd uit kalkmateriaal afkomstig van een ondiepe shelf, zoals fossielen, kalkige gesteentefragmenten, intraklasten, pellets en micriet, met toevoeging van wat terrigene kwarts en muscoviet. De kalkfragmenten en de gesteentefragmenten van Triadische of Albien-ouderdom, evenals glauconietkorrels, zijn afkomstig van een erosie die binnen het bekken plaats vond. Het belangrijkste kleimineraal, waaruit de mergels zijn opgebouwd, is illiet, welke afkomstig kan zijn van oudere gesteenten of door nieuwvorming kan zijn ontstaan.

De diagenetische processen, die hebben geleid tot de lithificatie van de afzettingen, omvatten rekristallisatie van micriet, cementatie door spariet, en enige andere processen zoals kwarts-vervanging, kwarts-nieuwvorming en oplossing door druk.

Tevens werd een kort onderzoek verricht in gebieden rondom de Vallcarga-Formatie, waar gesteenten van dezelfde ouderdom ontsloten zijn. De hierin aanwezige fauna geeft aan dat deze sedimenten werden afgezet in een warme, goed-doorluchte, ondiepe zee. Terrigeen materiaal werd aangevoerd in gebieden in het NW en NE van het Boven-Krijt-bekken, afgeleid van granodioriet-massieven die hier ontsloten waren, terwijl in het zuidelijk deel van het bekken kalkmateriaal werd afgezet, voornamelijk opgebouwd door rifachtige detritus.

Het gebied van maximale sedimentdikte bewoog langzaam naar het westen gedurende het Onder-Krijt tot het Eoceen. Het turbidietbekken was een smalle trog, ontstaan door een locale snelle daling. Het ontstaan van deze trog vond het eerst plaats in de centrale sectie van de Zuidelijke Pyreneeën, en breidde zich later uit naar het westen en naar het oosten. In het meest westelijk deel leidde deze daling, die misschien verband houdt met een migratie van Keuper-evaporieten in de ondergrond, tot het ontstaan van een submariene canyon langs breuken. Deze werd opgevuld door sedimentaire breccies, vermoedelijk gevormd door afstorting en submariene erosie van een omhoog bewegende diapierachtige antiklinale structuur.

Zowel veldgegevens als metingen van stroomrichtingen laten zien dat het bekken bestond uit drie delen, gedeeltelijk of geheel gescheiden door submariene ruggen. Deze drie delen werden opgevuld door stromen die elk uit een verschillende richting kwamen; een overeenkomstige situatie vindt men in de recente bekkens voor de kust van Californië. Het laatste stadium van opvulling van het turbidietbekken wordt vertegenwoordigd door een dikke serie van mergels en het verdwijnen van turbidieten, hetgeen vermoedelijk te wijten is aan de afname in de dalingssnelheid en de afwezigheid van materiaal dat in troebelingsstromen kon worden opgenomen.

Deze afzettingen worden gevolgd door de grofkorrelige zandige kalkarenieten van de Arén-Formatie, die zijn afgezet in een ondiep marien milieu.

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