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DEPOSITIONAL HISTORY AND CLAY MINERALS OF THE UPPER CRETACEOUS BASIN IN THE SOUTH-CENTRAL PYRENEES, SPAIN

BY

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SUMMARY

An ordered sequence of well-defined sedimentary environments reflects the deepening and shallowing stages in the depositional history of the Upper Cretaceous basin in the South-Central Pyrenees, Spain.

The sequence, which has a Santonian age at its base, starts with a calcarenite barrier system on which occurs a coral reef showing well-differentiated fore- and back-reef facies (Congost Limestone Formation). Marked deepening of the basin occurred during the deposition of the overlying glauconitic nodular limestones, as is evidenced by a predominantly planktonic fauna, consistent fine grain size, a very large planar slump scar, and frequent other slump structures (Anserola Formation). Sedimentation continued in a deep marine, carbonate turbidite facies (Vallcarga Formation). An olistostrome at the top of the turbidites contains dislocated masses of the same turbidites, an indication of the extreme mobility of the basin. The sequence is then regressive, via marls, which in part constitute the turbidite basin slope facies, into shallow-neritic to coastal cross-bedded calcarenites (Arén Sandstone Formation), and, via lagoonal and coastal-swamp deposits, ultimately into fluvio-lacustrine red beds (Tremp Formation; Upper Maastrichtian/Lower Paleocene). The total of added maximum thicknesses is approximately 3000 m.

The depositional history of the basin demonstrates the dominating influence of tectonic movements on the type of sedimentation. In the initial stages, the rate of subsidence exceeded the rate of sedimentation, and the basin consequently deepened (Anserola to Vallcarga formations). The regressive development is due to a slowing down of the general subsiding movement and continued sedimentation (Vallcarga to Arén formations). Many features in the turbidite sequence are directly related to tectonic movements.

The clay-mineral assemblages consist of illite, montmorillonite, chlorite, kaolinite, and mixed layers in varying proportions. Poorly crystallised, degraded illite occurs in the Congost and Anserola formations; well crystallised, fresh illite was introduced into the basin with the start of turbidity-current deposition. Montmorillonite is absent in the high-energy deposits (barrier calcarenites in Congost limestones, Arén sandstones), which is attributed to the relatively slow settling rates of this clay mineral. The absolute vitrinite reflectance as determined in the Vallcarga turbidite beds is 0.96% (fixed carbon 67%). Except for possible authigenesis of chlorite, no significant diagenetic clay-mineral transformations have taken place.

CONTENTS

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I. INTRODUCTION

The main object of this study is to trace the depositional history of the sedimentary fill of the Upper Cretaceous basin in the South-Central Pyrenees, Spain.

Because major parts of the basin fill have been eroded, little is known of its original size and shape. The present distribution of deposits suggests, however, that the basin was southeast-northwest oriented, some 80 km long, and relatively narrow. It was, at its deepest stages, subdivided into various lows and highs (van Hoorn, 1970). The fill of the basin, which totals approximately 3000 m, consists predominantly of a wide range of generally clay and quartz-sand bearing, clastic limestone types. Attention is directed in particular towards the sources of material, the direction and means of supply, and the degree of depth-related diagenesis as reflected by vitrinite reflectance and clay mineralogy.

Detailed 1:200 sections were drawn up in the field, recording and measuring, wherever possible, the sedimentary structures. The laboratory work consisted mainly of a petrographic study of thin sections and X-ray diffraction analyses of the clay fractions.

Following Mey et al. (1968), the basin fill is subdivided into five formations (ages after Souquet, 1967):

Tremp Formation – continental – (Upper Maasstrichtian/Lower Paleocene)
Arén Sandstone Formation – deep marine to neritic to lagoonal – (Upper Maastrichtian)
Vallcarga Formation – deep marine – (Campanian/Maastrichtian)
Anserola Formation – neritic to deep marine – (Santonian/Campanian)
Congost Limestone Formation – shallow neritic – (Santonian)

A composite section showing lithology, main sedimentary features and the subdivision of these formations is given in Figure 2.

Previous work
Dalloni (1910; 1930, p. 235) gave the first account of the development of the South-Pyrenean Upper Cretaceous basin. He correctly recognised the very shallow marine character of the Congost Limestone Formation, the transgressive nature of the Anserola Formation, the shallow marine environment of the Arén Sandstone Formation, and the fluvial and lacustrine environments of the Tremp Formation. The works of Misch (1934), Sanuy (1965), and Souquet (1967) have added a wealth of detail, as well as several corrections to a picture which in essence has remained unchanged. Nagtegaal (1963) and Mutti & Sanny (1968) reported turbidites from the Vallcarga Formation; van Hoorn (1970), who made a detailed study of the Vallcarga turbidites in the Esera region, demonstrated that they form the main fill of the basin at its deepest stages and that they vary in thickness, composition and direction of transport in sympathy with the paleo-submarine morphology.

II. GROSS RELATIONSHIPS

A geological map of the study area, only comprising a narrow north-south slice of the southeast-northwest oriented Upper Cretaceous basin, is shown in Figure 1. The area is characterised by generally south to southwest dipping strata, complicated by an anticlinal structure, the San Corneli anticline, plunging steeply to the west, just south of Aramunt. This anticline, which in the area shown consists of competent limestones (Congost and Anserola limestones), is immediately surrounded by highly incompetent marls. The structure is apparently 'smothered' in the marls, being hardly recognisable on the west side of the lake of Talarn; the marked change in strike of the Arén sandstones, however, is probably a structural adjustment to the presence of the anticlinal core underneath. In the southeastern part of the area, the tip of a second anticline, also plunging to the west, is visible. The whole southern part of the area in Figure 1 shows low dips in westerly directions. The gentle structure here forms the eastern extension of the large Tertiary basin located to the west of Tremp.

The Congost and Anserola formations, the two lowermost formations of the sequence studied, are exposed in the extreme northern part of the area. Both reappear in the San Corneli anticline. The Congost limestones, which in the north consist of a coral reef and associated shallow-neritic deposits, change in facies to the south only to the extent that no coral reefs are developed. The overlying Anserola nodular limestones have a similar aspect in the San Corneli anticline as in the north, except that no slump structures (frequent in the north) were observed and they are much thinner (less than 20 m against 230 m in the north).

The lower (Mascarell) member of the Vallcarga Formation, which consists of a turbidite sequence, shows a remarkable distribution on the map. In the northern part of the area, it is approximately 1100 m thick, but at the San Corneli anticline, where the turbidite sequence should reappear, only a few thin and fine-grained turbidite beds are found intercalated between marls which directly overlie the Anserola limestones. Because no faults that could explain the absence of the turbidites are found on the north side of the anticline, this shows that the thick turbidite sequence wedges out almost completely against what probably was a paleo-high.

The Salas marls are very uniformly developed, even far (tens of kilometres) outside the area. The Arén and Tremp formations, the first deep marine to neritic to lagoonal, the second continental, and both deposited under the direct influence of a west-southwest directed regional regression, also have a rather uniform general appearance, but have a complex pattern of minor facies differentiations.

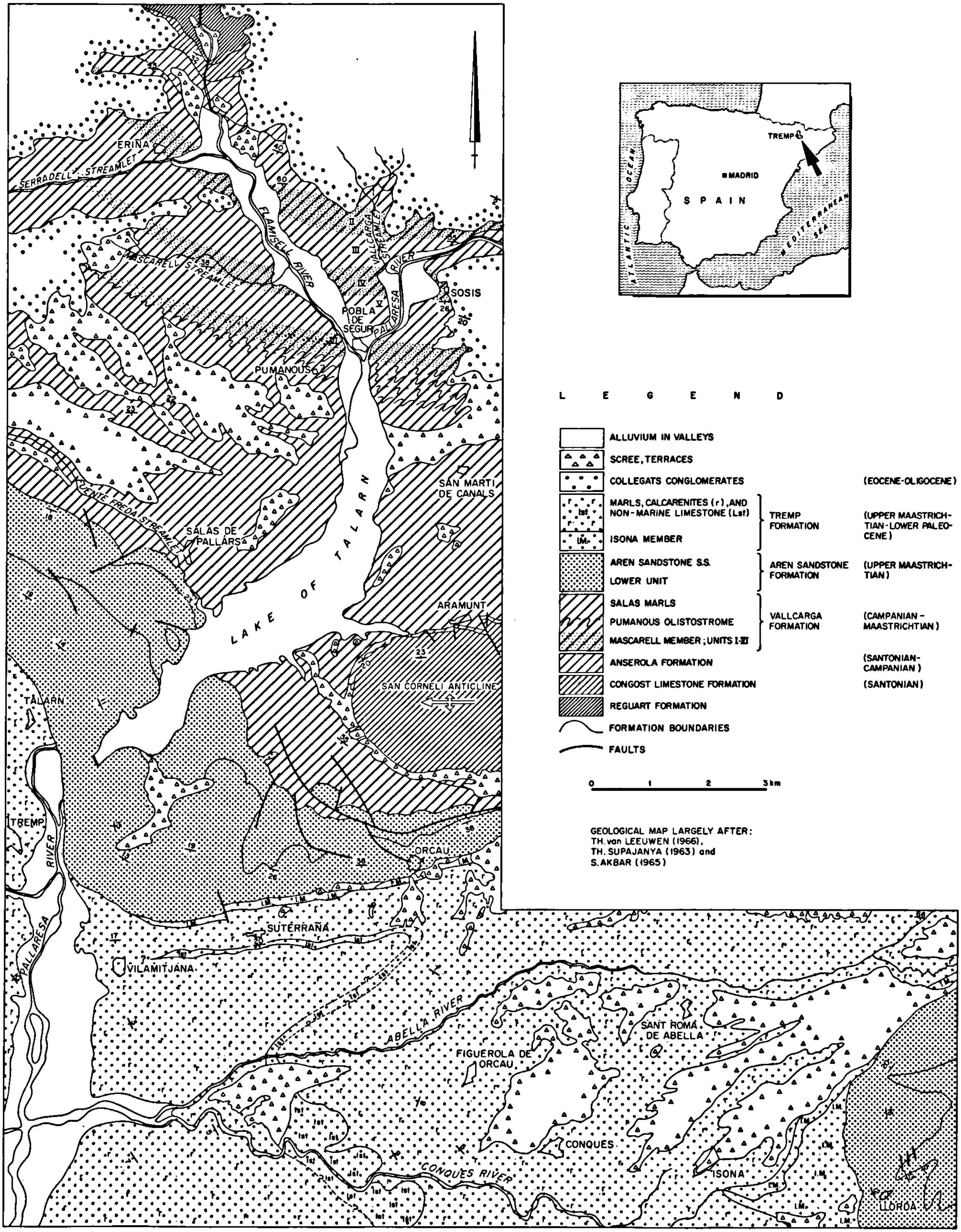


Fig. 1. Geological map of the Pobla de Segur – Tremp area, South-Central Pyrenees.

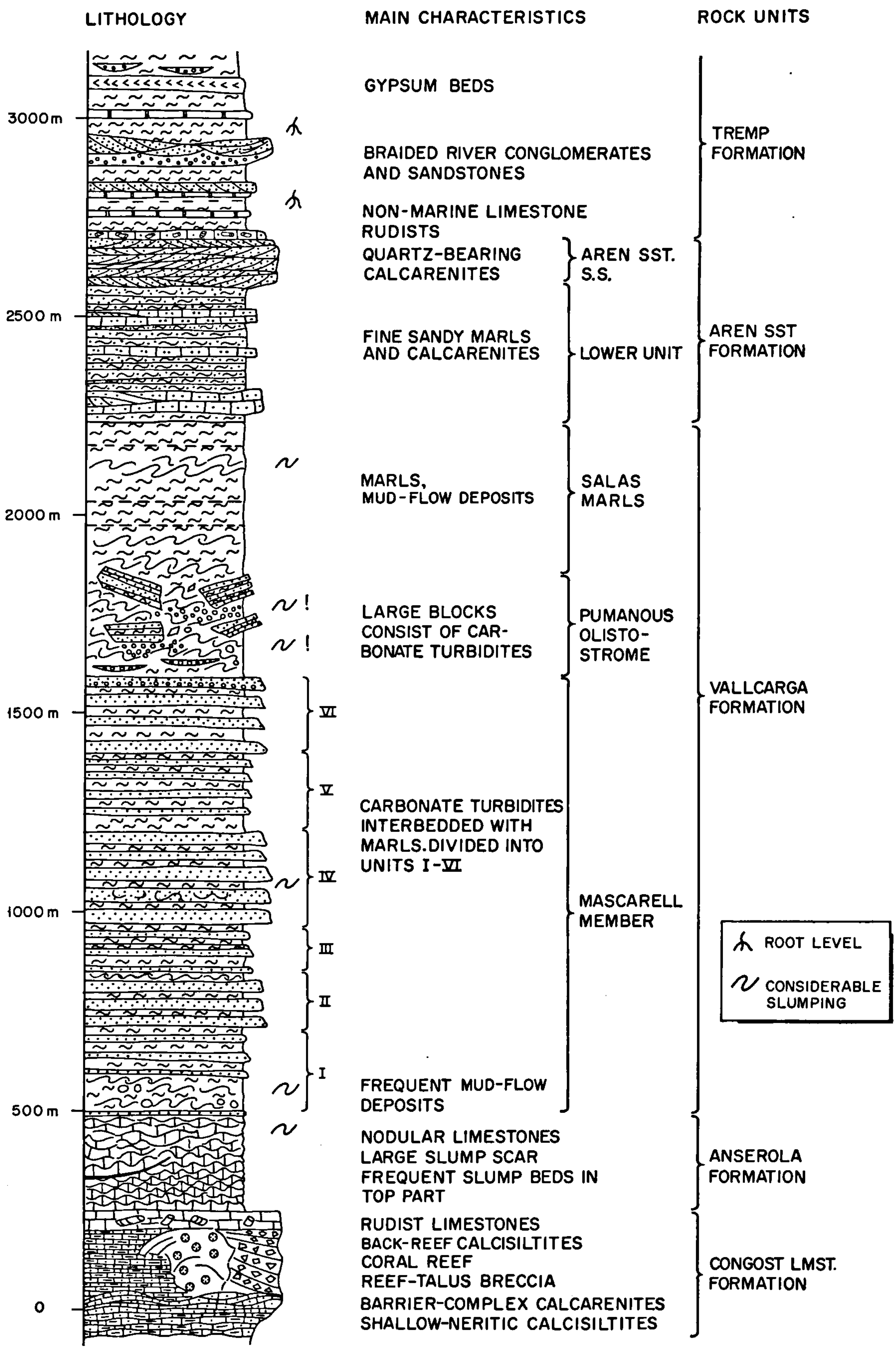


Fig. 2. Composite lithostratigraphic column Pobla de Segur – Tremp area.

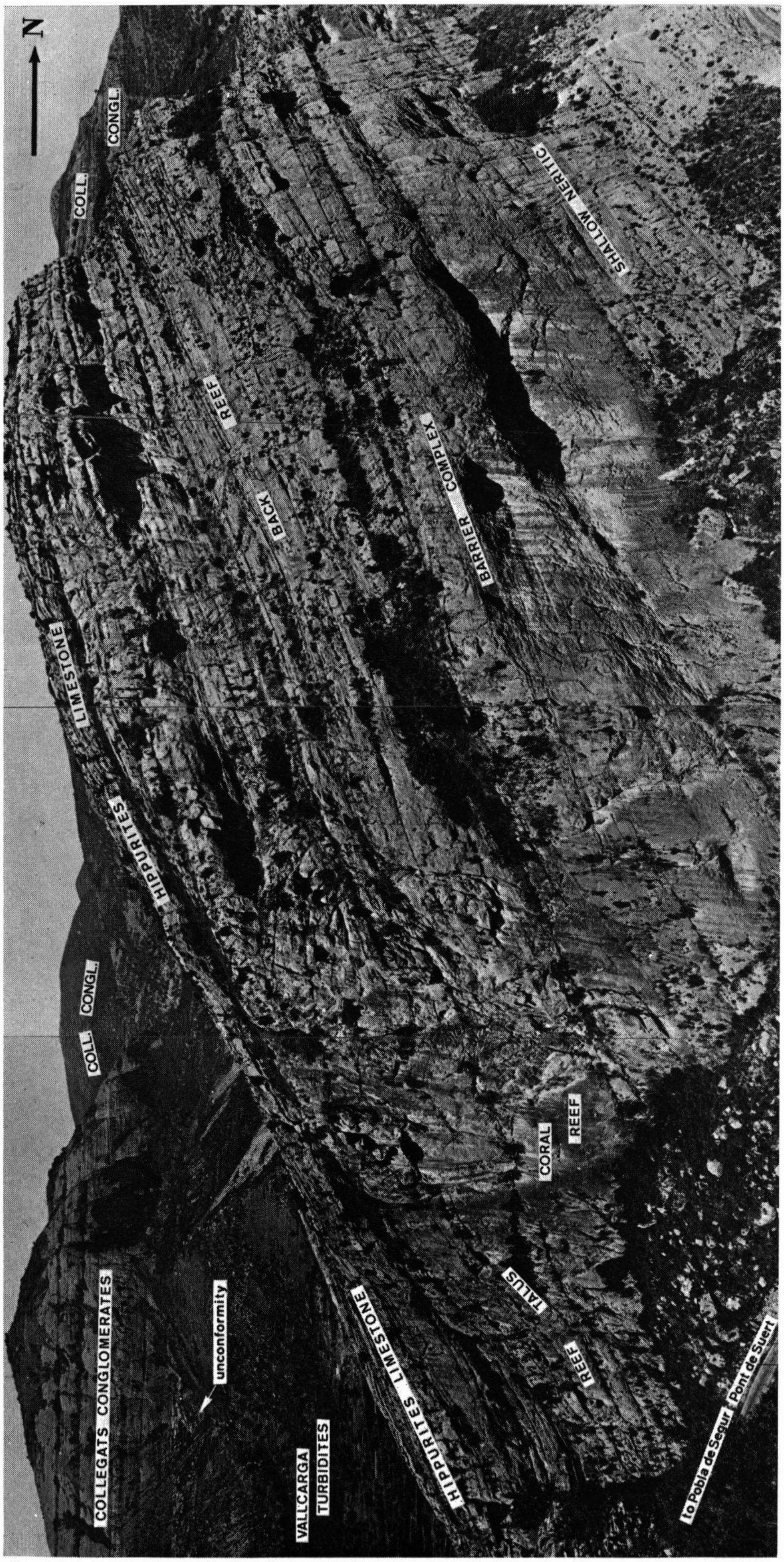


Fig. 3. Western flank of the Congost gorge. The main facies types of the Congost limestones are indicated.

III. CONGOST LIMESTONE FORMATION – CORAL-REEF AND ASSOCIATED SHALLOW-NERITIC DEPOSITS

The west flank of the 'Congost', the deep gorge cut out by the Flamisell River, offers an extraordinary view of the Congost limestones, allowing direct observation of the major facies types of the formation (Fig. 3).

A. *Underlying shallow-neritic and barrier deposits*

Grey, parallel-bedded, slightly nodular but otherwise homogeneous fine-grained limestones occur at the base of the sequence (Fig. 3, lower right). In thin section these limestones appear to be grain-supported and moderately sorted, consisting largely of pellets and small micritic fragments the origin of which cannot always be ascertained. Many of these fragments, however, are micritised fossil debris and algal grains. The grain sizes of the components are mainly confined to the 50–250 microns range. Other components are echinoderm, pelecypod, calcareous algal and sponge spicule fragments and Foraminifera (common Miliolidae), as well as angular corroded quartz grains (quartz: 4% by volume, 50–150 microns). Much of the intergranular space is occupied by lime mud, often recrystallised to a microsparry texture, though clear pore-filling calcite spar is also present. Within the grain sizes mentioned, there is a slight increase in an upward direction.

The upward increase in grain size is gradual into the unit overlying the parallel-bedded limestones. This next unit shows two thick accretionary lenses, flat at their bases and convex upwards (Fig. 3). Both consist of very clean, sparry calcite-cemented well-sorted and rounded calcarenites. The components, mainly micritic calcite fragments, calcareous algal, bryozoan and pelecypod fragments, fall largely in the 0.5 to 1 mm range; occasional fragments measure up to 2 mm.

The upward change from parallel bedding to accretionary lenticular structures and the accompanying gradual increase in grain size and sorting justify the conclusion that the sequence is regressive from a probably shallow-neritic, moderate-energy environment into an even shallower, high-energy barrier environment. Neither the general facies relationships, nor the composition of the sediment show whether these barriers are coastal or formed on offshore shoals.

B. *Coral reef, fore- and back-reef deposits*

Directly overlying the barrier deposits are three laterally transitional units; these are from right to left on Figure 3, (i) parallel thinly bedded fine-grained limestones, (ii) a seemingly homogeneous, large mass of limestone, and (iii) thick, slightly wedge-shaped very coarse-grained limestone beds. The central homogeneous mass of limestone was inaccessible and could not be sampled.

Thin sections of the parallel-bedded, fine-grained limestones, from samples collected near the top of the Congost cliff, are markedly different from the underlying barrier calcarenites. They are poorly sorted, containing large, fragile and angular pelecypod fragments as well as a large quantity (up to approx. 50%) of mud-sized calcite. However, all fossil fragments recorded from the underlying deposits are still present, though possibly in different proportions (no quantitative faunal analysis performed). In ordinary light, some of the thin sections show numerous dark patches consisting of organic matter. Evidently, the parallel-bedded limestones were laid down in a quiet-water environment, in open connection with the sea.

The wedge-shaped beds of limestone, which thin southwards (Fig. 3, lower left), are extremely poorly sorted. In the lower part they consist of limestone breccia, containing much angular coral debris in the 5–20 cm range, with occasional fragments up to 40 cm. Several fragments are partly coated with up to a few centimetres of thick encrusting calcareous Algae. Few pelecypod fragments occur, but echinoderm fragments (echinoids and crinoids), calcareous algal fragments and Foraminifera (Miliolidae), are abundant. The coarse fragments are set in a lime-mud matrix. The depositional environment is clearly one of quiet water, but with the supply of large angular fragments.

The three laterally transitional limestone units can be interpreted as the members in a coral-reef environment. This coral reef, the central homogeneous mass of limestone, was large and shallow enough to bring about a short-distance lateral facies differentiation. Behind it, on the north side, were the quiet conditions of a back-reef environment in open connection with the sea. At the seaward, south side of the coral reef a wedge, reef talus, accumulated in quiet and somewhat deeper water.

C. *Overlying rudist limestones*

The coral reef and its associated deposits are overlain by an approximately 30 m thick, well-stratified unit of limestone. The transition from the reef-talus deposits into this unit, as observable along the road, is gradual. The coarse coral debris decreases upwards in size and quantity, while the lime-mud content increases. The larger part of the overlying limestone unit consists of lime-mud deposits in which numerous large, up to 1 m long and 20 cm thick rudists (hippuritids) are embedded. Although very abundant at several levels, the rudists are suspended in the lime-mud deposits and are seldom in direct contact. Fossil debris includes many large pelecypod and echinoderm fragments, few calcareous algal fragments, and Foraminifera (predom. Miliolidae).

The high lime-mud content and very poor sorting point to quiet water; the gradual development from the reef talus suggests that the environment remained below wave base. As indicated by the lateral continuity of the rudist-bearing limestones, this somewhat deeper environment must have spread slowly northwards, across the coral reef. The conditions for reef growth became unfavourable probably as a result of the increased depth and the deposition of lime mud.

IV. ANSEROLA FORMATION – NODULAR FINE-GRAINED LIMESTONES; NERITIC, DEEPENING TO DEEP MARINE

A. *Sedimentary structure*

The Anserola Formation is, at the Congost, a 230 m thick deposit of nodular fine-grained limestones. Subdivision of the formation, on the basis of lithological characteristics, is difficult because the differences are minor and gradual. However, some large-scale sedimentary structures indicate that the depositional history of the formation was far from monotonous.

At 105 m above the contact of the formation with the underlying rudist limestones, an intraformational unconformity occurs (Fig. 2). It is clear from the total absence of slicken-sides or any other features indicative of tectonic movement that this plane, photographs of which can be found in Figure 4, is not a common fault plane. Slumped beds appear 70 m above the unconformity; these structures increase in frequency and intensity towards the top of the formation, which is defined by the appearance of the first turbidite beds of the overlying Vallcarga Formation.

The angle between the unconformity plane and the underlying beds is 12–15°; the strike of the plane could not be measured accurately, but is estimated to be roughly parallel to the general tectonic strike at the locality (110° SE 290° NW, Fig. 1). The beds under- and overlying the unconformity plane are all fully marine and of comparable facies (some compositional differences that nevertheless occur are pointed out below). From this fact, it is concluded that the unconformity is of a submarine origin. Evidently, at a stage when about half of the nodular limestones had been deposited, a large upper part of the sequence was suddenly removed, after which sedimentation continued under approximately the same conditions as before. It is thought that this sudden removal of a large mass of sediment is the result of detachment and downsliding on a submarine slope. The unconformity is then the detachment plane proper; it is not unlike the slump scars discussed by Wilson (1969), which occur in similarly fine-grained limestones. When the beds underlying the unconformity plane are rotated back to the horizontal position, the detachment plane attains a 12–15° north to northeast slope. It is not possible, however, to deduce a paleoslope direction from this one exposure.

The lowermost levels of slumped sediment, 70 m above the intraformational unconformity, are difficult to recognise and are only evidenced by sudden local changes in the regional dip of the beds. These deviating dips are restricted to masses of sediment, up to 4 m thick and 10 m wide, which are separated from the underlying deposits by curved and striated fault planes. These fault planes do not continue into the overlying beds, and are therefore explained as early sedimentary. The curved fault planes and the offset sediment masses are very similar to the rotational slumps and slump scars described by Laird (1968). 90 m above the unconformity, an 8 m thick interval consisting of fault-bounded, internally folded masses of sediment begins; higher up in the section, until the very top of the Anserola Formation, many more such levels are found. The general appearance of the deposits in the top part of the formation is very chaotic. These strongly disturbed beds are directly overlain by thinly bedded turbidites of the Vallcarga Formation (Fig. 5). The upward-increasing frequency and intensity of sediment failure are explained as the result of the formation of steeper slopes due to accelerated subsidence and deepening of the basin.

B. *Composition and depositional environment*

The Anserola nodular limestones are of a fairly constant composition throughout the 230 m section exposed in the Flamisell valley. Total carbonate content (including lime mud admixed with clay), ranges from 89.4 to 99.8% (thin-section point-counting analyses: 500 points counted), with a standard deviation of only 1.59%. Quartz and glauconite occur in subordinate amounts. The insoluble residues range from 7.2 to 44.8%, averaging 22.9% (Table I). The average clay content of the rocks is estimated to be in the order of 15–20%.

	Average %	Range %	Number of samples
Carbonate (pellets, fossil fragments, lime mud admixed with clay, microspar)	98.5	89.4–99.8	58
Glauconite (>10 microns)	2.1	0.2–4.2	in 7 out of 58
Quartz (silt-sized)	1.5	0.2–10.6	58
Insoluble residue (HCl, weight %)	22.9	7.2–44.8	8

Table I. Composition of Anserola limestones.

The carbonate part of the rocks consists of varying proportions of lithified lime mud admixed with clay, and large amounts of small faecal pellets and (often unidentifiable) fine-grained fossil debris (sponge spicules, bryozoan and pelecypod fragments; many echinoderm fragments, numerous Foraminifera). Fragments of calcareous Algae (rhodophyceans) are present, but nowhere common, in the lowermost few tens of metres of the section. There is no evidence of the presence of Algae at higher levels.

From approximately 60 m above the base of the Anserola section, the lime-mud content increases to generally more than 50%; below this level some beds that are almost free of lime mud occur, being grain-supported calcisiltites. The limestones are completely homogenised by burrowers. Dense patterns of burrowing tracks on bedding planes can be observed in the field; in the thin sections, rearrangement of the material into clouds and irregular patches of varying lime-mud content evidence the burrowing.

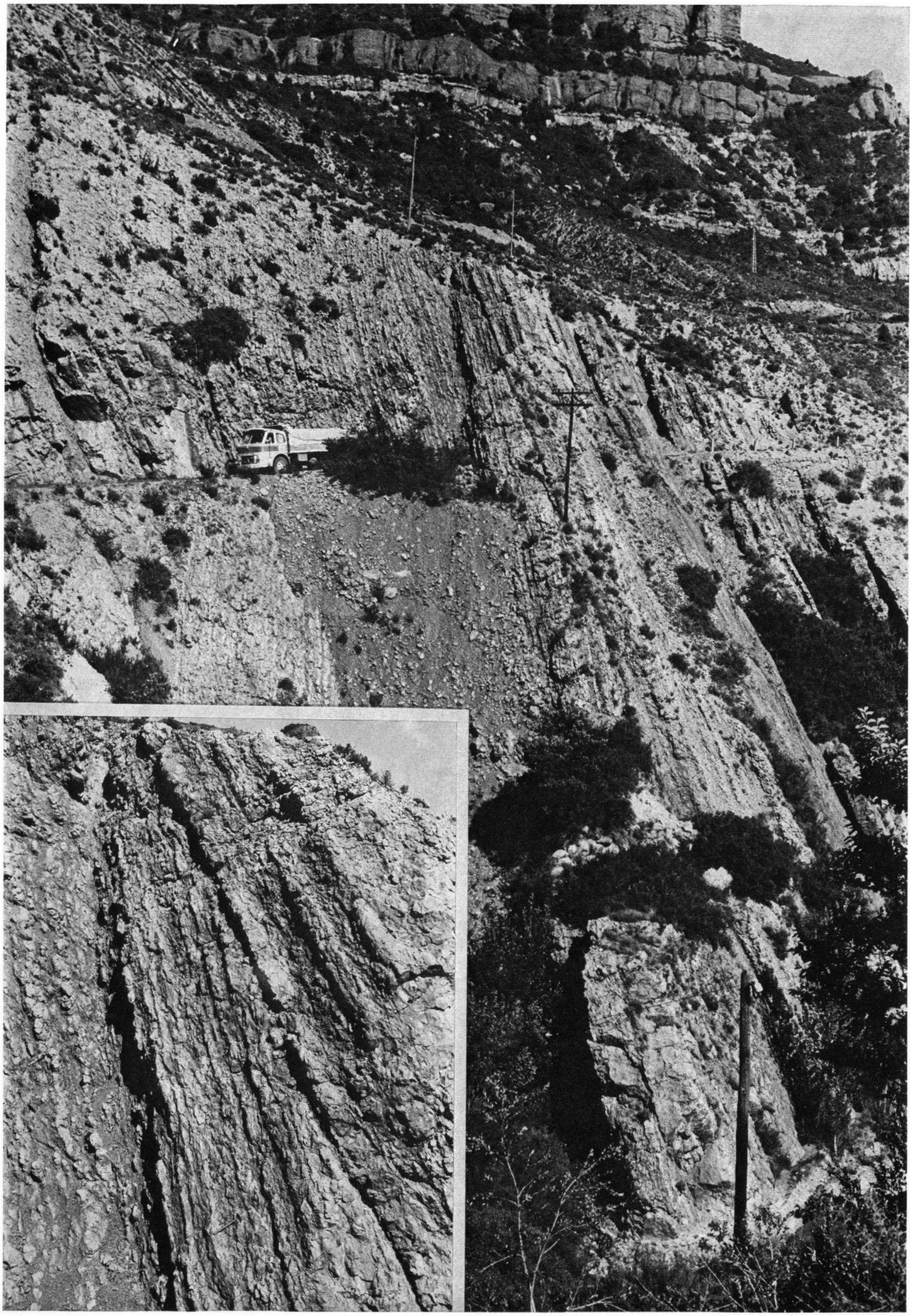


Fig. 5. Contact between the strongly slumped upper part of the Anserola Formation (large, offset mass of sediment in central part is Anserola sediment) and the Vallcarga Formation (turbidite beds in left hand part). Vallcarga streamlet; measuring staff is 1 m long.

With reference to the level of the large slump scar, some small differences in composition can be noted. The content of silt-sized clastic quartz shows a slight tendency to decrease in an upward direction; below the slump scar the quartz content averages 1.3% (25 samples), above it 0.8% (33 samples).

Glauconite is virtually absent in the samples from the upper 70 m of the section, but is generally present in the other samples in the form of small specks. The mineral shows a sharp upward increase in grain size and frequency in the first 30 m directly above the slump scar plane. Many of the limestone beds in this interval are light coloured, some showing reddish hues, in contrast to the other limestone beds which are generally medium grey. The high concentration of glauconite, and the light grey and reddish colours of the limestones above the slump scar are interpreted as being due to a decrease in the rate of deposition. It is generally accepted that glauconite content is in part a function of the rate of deposition (Cloud, 1955). A slow rate of deposition, given a generally oxidising environment, will have favoured thorough oxidation of the deposits, thus leading to the removal of organic matter and resulting in light and reddish tints.

Fig. 4. Eastern valley wall of the Flamisell River, showing the nodular fine-grained limestones of the Anserola Formation. Note the intra-formational unconformity, interpreted as a large slump scar. Inset shows the slump scar in detail, at the height of the road.

The glauconite occurs locally as fillings of foraminiferal tests, but it is mainly found in the form of up to 0.7 mm colloform, or angular to subangular, green microcrystalline aggregates. The typical well-rounded pellet form does not occur. This in part may be due to irregular replacement of glauconite by calcite, which is common. Some of the glauconite grains enclose ghost structures of pellets and small Foraminifera; others contain small, newly-formed dolomite rhombs.

In four samples collected just above the slump scar, the glauconite content was high enough to allow X-ray analysis (hand-picked grains; Guinier-de Wolff type camera). The glauconite in three samples is very similar, giving moderately diffuse bands at 10.2–10.5 Å and 3.3 Å, and a much better defined reflection at 4.53 Å. No (112) and (11$\bar{2}$) reflections were recorded. On this basis, the glauconite is considered to be of the Class 1[a] (disordered 1 Md glauconite) type (Burst, 1958; Bentor & Kastner, 1965). The other sample shows a high admixture with chlorite.

On the basis of the data presented, it is concluded that the Anserola limestones were deposited in a normal, open-marine environment. The near-constancy in composition and fine grain size, the absence of any size sorting outside the range of silt and mud, and an absence of current-scour structures are indicative of the quiet conditions found in rather deep waters. Sedimentation largely under aphotic conditions is suggested by the absence of calcareous algal fragments above the lower few tens of metres of the deposits. It is safe to assume that the depths, at the latitude of the Pyrenees, were in excess of 150 m. The lower limit of vegetation in the Gulf of Naples, for instance, is at approximately 160 m (Schimper, 1935, p. 1463). The near absence of glauconite in the upper 70 m of the formation (as against its general presence below that level) may also be related to the deepening of the basin. It could mean that the water depths in the upper part of the section increased beyond 700–800 m (Cloud, 1955). This is corroborated by the predominance of a planktonic fauna, consisting of Radiolaria and *Globotruncana*, in the upper part of the section.

V. VALLCARGA FORMATION – CARBONATE TURBIDITES, OLISTOSTROME, MARLS; DEEP MARINE

In the area studied, the Vallcarga Formation can be subdivided into three members: (i) a lower one, the Mascarell member, approximately 1100 m thick and characterised by a sheer endless repetition of quartz-rich calcarenite and marl beds. These deposits, developed in a typical flysch facies, are generally interpreted as turbidites. (ii) A middle one, consisting of an up to 300 m thick, highly chaotic mass of marls, locally pebble-bearing, and huge, detached blocks composed of the same turbidites as found in the underlying member. This member, which thins towards the west (Fig. 1) is here called the 'Pumanous olistostrome', and (iii) an upper (partly lateral) one, composed of mainly marls, but with a few turbidite beds in the lower part, referred to as the Salas marls. This member is estimated to be some 400 m thick (Fig. 2).

A. *Mascarell member*

The turbidites in the Mascarell member have been discussed by several authors (Farrés, 1963; Nagtegaal, 1963; Wiersma, 1965; Mutti & Sanuy, 1968, 1969; van Hoorn, 1970). We will concentrate on some points that have not yet been fully treated.

Slump structures, fluxoturbidites, and channel fills. – Mud-flow and slump deposits intercalated between undisturbed turbidite beds are very frequent in the lowermost part of the Mascarell member (Fig. 2), but are also found at several higher levels in the sequence. The mud-flow deposits are partly made up of rounded fragments of Anserola limestones, but this can only be shown conclusively wherever larger masses are involved (Fig. 5) (Mutti & Sanuy, 1969). The mud-flow deposits also contain strongly folded and fractured turbidite beds similar to the under- and overlying beds.

All transitional stages, from the initial small fault in a turbidite bed to levels composed entirely of slump-faulted and -folded turbidite beds, can be observed (Fig. 6). Quite often, thin, fine-grained turbidite beds carrying only b to e, or c to e divisions (Bouma, 1962; Walker, 1967) are included in the slump structures. The fact that these turbidite beds (which must have been deposited during the decelerating stage of turbidity currents, and hence on a near-horizontal bottom) are involved in the slumping indicates that tectonic generation of slopes within the basin was responsible for the sediment failure.

Thick, coarse-grained to microconglomeratic beds, which often contain large irregular slabs of marl, in many cases show strong internal slumping but have retained flat upper and lower surfaces. These beds are found in the upper half of the Mascarell member (Fig. 7). They are thought to have been very rapidly deposited, in the form of 'sand slurries', and are referred to as 'fluxoturbidites' (Dzulinski et al., 1959).

At several localities, undeformed, in situ channel fills are closely associated with the slump and mud-flow deposits. It is thought that this association results from the fact that up-slope parts of the active slump/mud-flow sediment developed into a turbidity current which eroded (and later filled) a channel into the mud-flow deposit proper (Fig. 8).

Turbidite-bed thickness and turbidite/marl ratios. – Observations in the outcrop areas of the Mascarell member have shown that approx. 250 m-thick 'units' (tentative subdivisions of the Mascarell member) can be distinguished and mapped. These units, six in number, are differentiated on the basis of their turbidite/marl ratios and coarseness of turbidite beds. Units II, IV and VI (Figs. 1 and 2) have turbidite/marl ratios of more than 1, and usually of the order of 1.5 to 2.5. In units I, III and V, the turbidite/marl ratios are less than 1, and mostly of the order of 0.7 to 0.3. The turbidite beds contained in units II, IV and VI are generally coarser than those in units I, III and V. The transitions from one unit to the next are gradual.

The results of detailed turbidite-bed-thickness analyses in units I, II and III (in each unit 500 turbidite/marl couplets measured) are shown in Figure 9. A comparison of the thickness distributions and percentages of turbidite beds in the three units shows that when the percentage of turbidite beds is high, the frequency of thicker (relatively coarse) beds is higher. Although measured only for units I, II and III, field observations suggest that this applies also to the three overlying units.

The six units were mapped by estimating the fre-

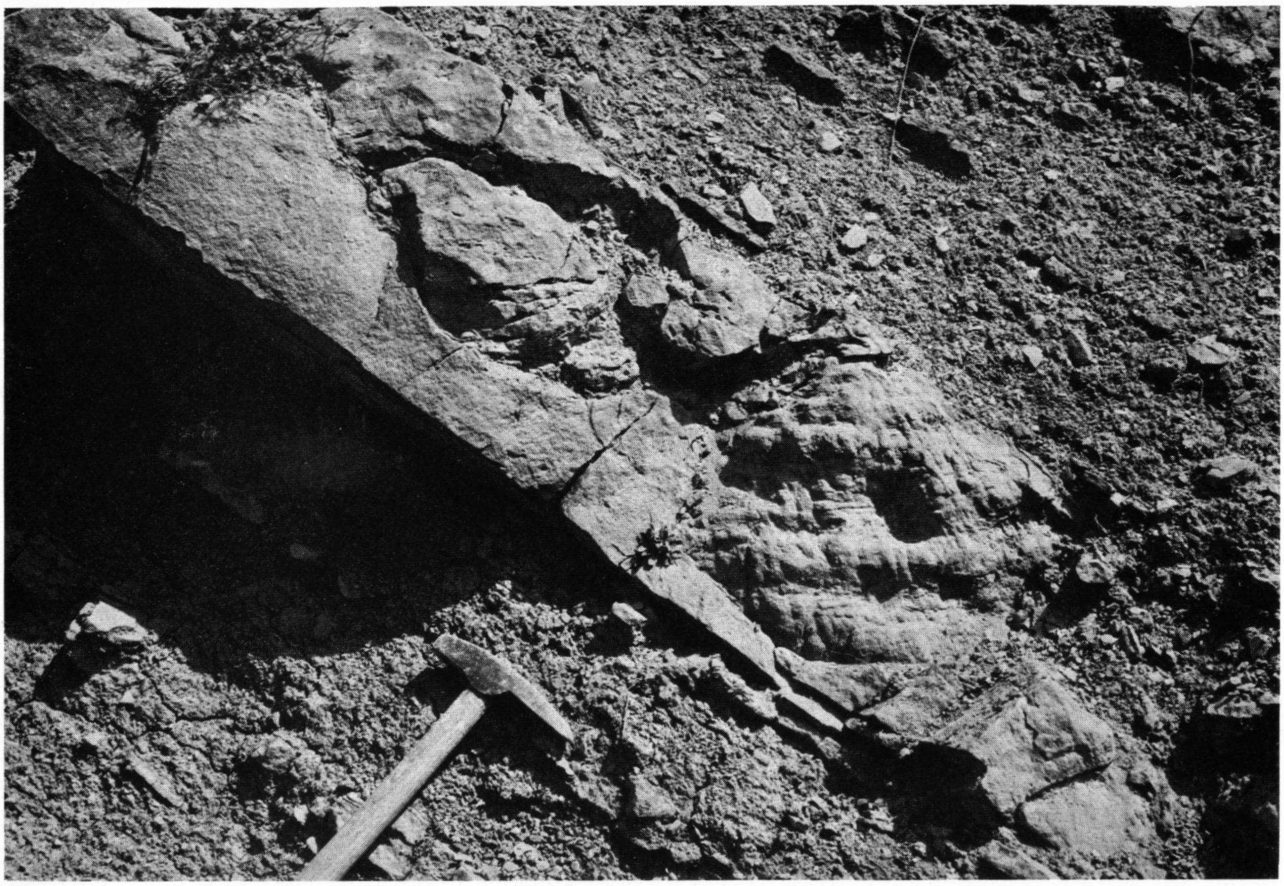
Fig. 6. Small overthrust in turbidite bed; lower part faulted, laminated upper part shows adjustment to fault but no rupture. The difference in behaviour is attributed to a difference in plasticity. Flamisell valley.

quency, thickness, and coarseness of the turbidite beds in the field. This revealed that units IV and VI in particular (and the overlying Pumanous olistostrome as well) wedge out towards the west. At the same time, the turbidite/marl ratios, average turbidite-bed thickness, and grain size (all estimated in the field) decrease (Fig. 1). In combination with the fact that the paleocurrent data indicate generally westward transport (p. 264), these trends suggest the structure of a clastic submarine wedge or fan that was built out from east to west.

The coarseness of the 'coarse units' increases upwards in the Mascarell member. The turbidite beds in unit II can be described as fine to coarse grained, many in unit IV are microconglomeratic, while lenses of conglomerates occur in unit VI (no conglomerates found in the other units). Units III and V are also significantly coarser than unit I. Fluxoturbidites and channel fills were only found in units IV and VI. This trend may reflect two things: (i) as one, in the Pobla de Segur area, moves upwards in the Mascarell member there is also a considerable l a t e r a l shift, the beds dipping 40–60° SW (Fig. 1). An apparent upward increase in coarseness and the appearance of fluxoturbidites and channel fills could therefore correspond to the approaching of the turbidite fan axis, where the coarsest material may be expected to have been deposited. (ii) Tectonic movements, probably step-like increases in basin slope, have to be invoked to explain the alternating character displayed by the six units. This increase in basin slope could also have been responsible for the upward increase in coarseness and appearance of fluxoturbidites and channel fills, because on steeper slopes coarser material can be transported and strong bottom erosion is more likely to occur. The most significant result of this tectonic development would have been the generation of the Pumanous olistostrome, which contains large detached turbidite masses derived from the underlying turbidite sequence.

Paleocurrents. – The strikes of small sedimentary thrust planes (Fig. 6) and of axial planes of slump folds, the orientations of channel axes and of elongate mud-pebble axes, and the current directions derived from flute casts are presented graphically in Figure 10. The measurements of flutes and channel axes were evenly spread over the area in the immediate vicinity of Pobla de Segur, along the Mascarell streamlet, and between the Pallaresa and Flamisell rivers. The mass movement directions come mainly from unit I in the Flamisell valley, and the mud-pebble orientations were measured in unit IV along the Pallaresa River (Fig. 1). In the reconstruction of the original orientations of sedimentary structures, simple tectonic tilt has been assumed.

It is seen that the small sedimentary thrust planes and axial planes of slump folds strike predominantly northwest to southeast. On the basis of a marked



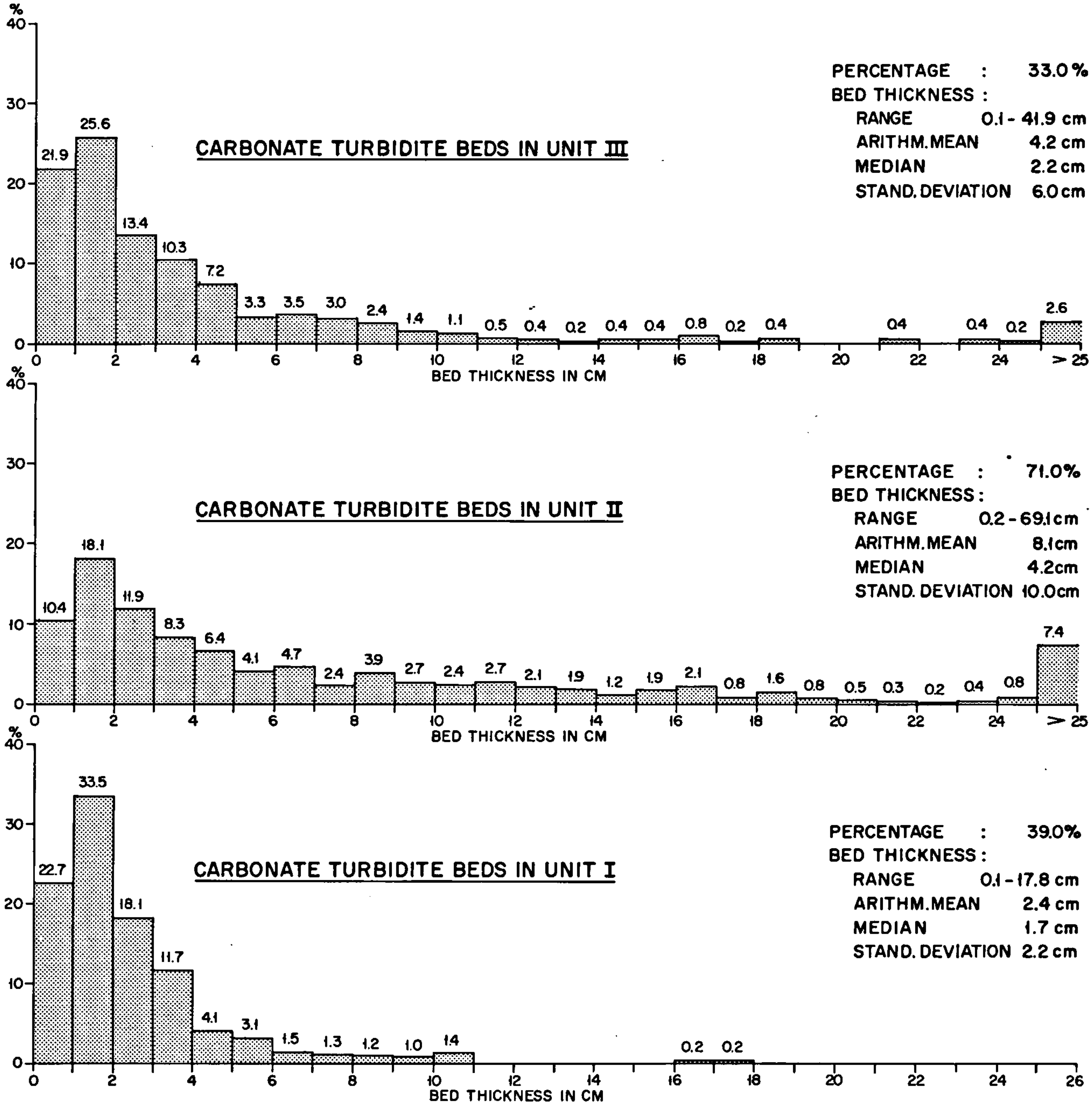


Fig. 9. Vallcarga Formation. Turbidite-bed-thickness frequency distributions in units I, II and III, Mascarell member.

predominance of northeast-dipping, small sedimentary thrust planes (overthrusting towards southwest; in Fig. 6: movement towards the left), it is assumed that

Fig. 7. Vallcarga deposits (Mascarell member). Internally deformed, thick bed interpreted as a fluxo-turbidite. Vallcarga streamlet, lower part of unit IV. Hammer is 40 cm long.

Fig. 8. Channel fill in Vallcarga deposits (Mascarell member). The channel must have cut into the directly underlying mass of sediment which was emplaced by slumping; the channel fill is overlain by in situ turbidite beds. Vallcarga streamlet, lower part of unit IV. Hammer is 40 cm long.

the mass movements generally occurred in a southwesterly direction.

The flute casts indicate predominant flow of the sediment suspensions in a westerly direction, parallel to the axis of the longitudinal basin. This was also found by Mutti & Sanuy (1968). One southeastward and one northward flow direction were encountered. This could indicate that periods during which there was an almost horizontal bottom occurred during the infilling of the basin. The major axes of large mud pebbles are generally parallel to the flute-cast directions although their orientations are spread over a much wider range. The axes of the channel fills also tend to

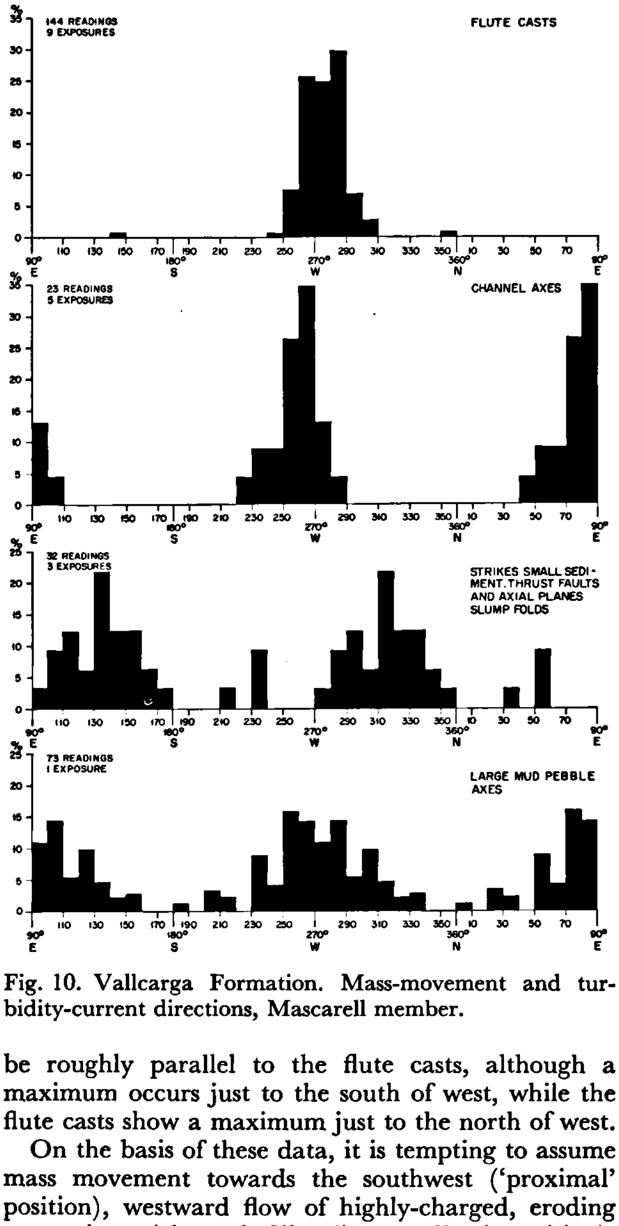


Fig. 10. Vallcarga Formation. Mass-movement and turbidity-current directions, Mascarell member.

be roughly parallel to the flute casts, although a maximum occurs just to the south of west, while the flute casts show a maximum just to the north of west.

On the basis of these data, it is tempting to assume mass movement towards the southwest ('proximal' position), westward flow of highly-charged, eroding suspensions (channel fills; 'intermediate' position), and continued flow of the suspensions turning to just north of west (flute casts, more 'distal' position). This interpretation of the data is highly hypothetical, however, because (i) the mass-movement directions were measured in unit I only and moreover they may represent only temporary, tectonically induced slopes, and (ii) the difference in orientation between the flute casts and the channel axes is considered too small to be of real significance (Fig. 10).

Composition of turbidite beds. – The turbidites in the Mascarell member vary considerably in composition, sorting and grain size from bed to bed, and in the vertical sense also within a single bed. The average compositions of 46 turbidite samples, collected at evenly spaced intervals in the section exposed in the Vallcarga streamlet (Fig. 1), are given in Table II.

In all samples examined the turbidites are grain-supported, quartz-bearing to quartz-rich calcarenites. Varying amounts of clay-bearing lime mud and sparry calcite occur between the grains; commonly the lime-mud content increases with the upward decrease in grain size in a bed. The coarse thick turbidites, among them the fluxoturbidites, commonly have low lime-mud contents. The carbonate grains include numerous fragments of calcareous Algae (commonly rhodophyceans), bryozoan, echinoderm and pelecypod fragments showing micritic rims and algal borings. These components indicate supply from shallow-marine environments. The Foraminifera in the turbidite beds include both shallow-water representatives (Miliolidae, *Nonionella troostae* (Visser), and *Pararotalia tuberculifera* (Reuss), and planktonics (a.o. *Globotruncana*, Heterohelicidae). Glauconite, mainly in the form of rounded pellets, was found in 40 of the 46 samples analysed. The mineralogical composition of the pellets has not been determined.

Among the extrabasinal constituents, which comprise approximately one quarter of the bulk volume of the rock, quartz is strongly predominant, but some rock fragments (quartzite, phyllite), chert and feldspars occur. Coalified plant fragments, not listed in Table II, are present as traces in most samples. Plant and wood fragments up to 1 cm in size are very abundant on the upper surfaces of some turbidite beds.

The petrographic data show that the turbidites have a complex origin; the hinterland was a constant source (ever-present quartz, clay and vegetational debris), but material produced in a neritic environment formed a predominant supply (type of fossil material, glauconite).

The marls, at levels just below turbidite beds, contain rich *Globotruncana* faunas, and benthonic assemblages consisting of *Bathysiphon latissimus* (Grzybowski), *Haplophragmoides walteri* (Grzybowski) and *Spiroplectammina* ex gr. *anceps* (Reuss). This indicates, according to J. Brouwer (pers. comm.), a deep marine (bathyal-shallow abyssal) environment.

	Average %	Range %
Carbonate (lime clasts, pellets, fossil fragments, lime mud admixed with clay, calcite spar)	71.2	42.2–91.8
Glauconite (>10 microns, mainly pellets)	0.7	0.0–2.8
Quartz (angular to well rounded, silt- to sand sized)	26.0	6.4–42.0
Feldspar	0.4	0.0–1.4
Chert	0.5	0.0–2.2
Quartzite, phyllite	1.1	0.0–0.8
Mica	0.1	0.0–0.8
Insoluble residue (HCl, weight perc.)	50.4	35.3–71.7

Table II. Composition Vallcarga turbidites.

B. *Pumanous olistostrome, Salas marls*

Pumanous olistostrome. – The Pumanous olistostrome represents a wedge of highly chaotic deposits, which diminishes in thickness from approximately 300 m at Pumanous to less than 20 m further west, where it is unconformably overlain by the Eo-Oligocene Collegats Conglomerates (Fig. 1). The various types of fragments found in the deposits decrease in size in the same direction.

The olistostrome contains (i) large angular blocks made up of turbidites, (ii) pebbly marls, (iii) numerous irregularly contorted slabs of marl, and (iv) some undeformed channel fills. The blocks composed of turbidites are up to 20 m long and 5 m high and are embedded in pebbly marls and masses of contorted slabs of marl. They can be observed in the direct vicinity of Pumanous, on the northwest side of the village (Fig. 1). The turbidites in the blocks are in all respects similar to those in the underlying Mascarell member; it is therefore concluded that they are derived from some level in this member. No internal deformation was observed in the larger blocks; it seems that they skidded, in the form of flat slabs, over a muddy substrate into their present position. However, thinner turbidite sequences (one or a few beds) occur in highly-complex slump folds.

Contorted levels and pockets of pebbly marls up to a few metres thick, as well as masses composed of contorted marl slabs, occur throughout the Pumanous deposit. In many places the marl is similar to that found between turbidite beds in the Mascarell member, but locally it is dark coloured and rich in vegetational debris. The pebbles, locally even boulders, are generally well rounded, and Palaeozoic schists and Devonian and Lower Cretaceous limestones have been encountered among them. They are similar in composition and roundness to the pebbles found in conglomerate levels in unit VI.

The channel fills, similar to those observed in units IV and VI of the Mascarell member (Fig. 8), occur at several levels in the Pumanous olistostrome. The presence of these channel fills, the origin of which is thought to be the same as suggested for those in the Mascarell member, implies that the olistostrome was built up in phases, and not in a single catastrophic event. Episodic build-up has also been suggested for the olistostromes of the Appennines (Görler & Reutter, 1968), and for the Campo olistostrome (van Hoorn, 1970).

The whole of the Pumanous olistostrome, and in particular the occurrence in it of the blocks made up of the directly underlying turbidite deposits, is interpreted as the result of relatively strong, tectonically-induced increases in basin slope in areas of previous turbidite deposition.

Salas marls. – The Salas marls, generally grey to dark grey and approximately 400 m thick, have a remarkably uniform development over a wide area. The lower part of the Salas marls, which still contains a few turbidite beds, is parallel, thinly, and well stratified; higher up in the marls several thick and extensive mudflow deposits composed of the same marls can be clearly identified. In most exposures the marl is apparently homogeneous, but in others thin and often quite vague bedding predominates. Several of these beds show upward-fining grading confined to the silt-clay size range, and probably represent mud turbidites. This is confirmed by the admixture of a shallow water fauna (*Orbitoides*, *Siderolites*) with the generally planktonic fauna (*Globotruncana*, *Rugoglobigerina*, *Heterohelix*) found in many Salas marl samples.

The sudden, almost complete cessation of turbidity-current deposition of sand-sized material is not fully understood. Three possibilities are considered: (i) a shrinking and/or retreating of the predominantly neritic carbonate-producing supply areas and an ultimate cut-off from the sources of sand-sized material. This process would have led to a gradually diminishing supply of sand-sized material and is therefore unlikely to account for the abrupt end observed; (ii) tectonic adjustment in the form of a regional uplift of the deeper parts of the basin, leading to a general decrease in basin slopes. Theoretically, this would create a setting unfavourable for the generation of turbidity currents carrying sand-sized material, and could therefore explain their near-absence in the Salas marls. However, some observations suggest the preference of a third possibility: (iii) the Salas marls were deposited, to a large extent, on a wide submarine slope and constituted an environment that was largely bypassed by the turbidity currents. The main evidence in favour of this solution is that in the direct vicinity of the San Corneli anticline, which is thought to have formed a submarine high, only Salas marls carrying a few turbidite beds are found directly overlying the Anserola limestones (p. 252). Evidence for at least the temporary existence of slopes lies in the extensive levels of marl-mudflow deposits in the Salas marls.

The suddenness of the change from turbidity-current deposition of sand-sized material to marl deposition can be related to the generation of the Pumanous olistostrome, which occurs on the transition of both. This large mass of detached sediment was explained as resulting from a marked increase in basin slope. The olistostrome material, where it is now exposed, may have come to rest partly on the newly formed slope, and thus in a position unfavourable for the continuation of turbidity-current deposition of sandsized material at this place.

If the Salas marls to a large extent form the turbidite-basin slope deposits, a fairly rapid lateral transition from turbidite beds to marls is implied.

VI. AREN SANDSTONE FORMATION – MARLS AND CALCARENITES; DEEP MARINE TO NERITIC TO COASTAL, LAGOONAL

The Arén Sandstone Formation is, for the purpose of the present study, subdivided into two main lithological units: (i) a lower Arén sandstone unit composed of silty to fine sandy marls and calcarenites, approximately 350 m thick, and (ii) an upper unit, the Arén sandstone sensu stricto, consisting predominantly of cross-bedded, quartz-bearing clean calcarenites, approximately 100 m thick (Figs. 1 & 2).

A. *Lower Arén sandstone unit*

The lower Arén sandstone unit is well exposed to the south and southwest of Salas de Pallars (Fig. 1); some observations were made in this area, but no detailed study has yet been conducted.

The contact between the lower Arén sandstone unit and the Salas marls, exposed on the first hill crest south of Salas de Pallars, is sharp and probably erosive. Just below the contact, the Salas marls retain their finely, vaguely bedded character and predominant planktonic fauna. Directly on top of it rests a wide lense composed of medium to coarse-grained, small-scale cross-bedded calcarenites. These calcarenites are grain-supported, well-sorted, and cemented with sparry calcite. They consist of large Foraminifera (neritic benthonic forms such as *Orbitoides, Siderolites*), rounded pelecypod, echinoderm and calcareous algal fragments.

In the lowermost 200 m of the lower Arén sandstone unit, many other wide (50 m to several hundreds of metres wide, and up to 10 m thick) lenses of calcarenites are found. The calcarenites, which are not all as well-sorted as those just described, are embedded in strongly bioturbated silty and fine sandy marls. Such marls, which at some levels show a nodular habit, predominate in the uppermost 150 m of the lower unit. A few tens of metres below the contact with the Arén sandstone proper, the marls are very rich in a neritic benthonic macrofauna, including levels with many well-preserved Bryozoa, pelecypods, gastropods, and echinoids.

The silty to fine sandy marls between the wide calcarenite lenses were only sampled in the very lowest part; they contain the same planktonic Foraminifera as the Salas marls, and a benthonic fauna interpreted as deep marine [*Allomorphina, Bathysiphon latissimus* (Grzybowski), *Gyroidina, Lenticulina, Nodosaria, Osangularia cordieriana* (d'Orbigny), *Quadrimorphina allomorphinoides, Recurvoides* and *Stensioina* ex gr. *exsculpta* (Reuss)]. This fauna is in marked contrast to the neritic assemblage in the calcarenites just described, which suggests that the lenses could represent wide channels that were filled with material derived from a neritic environment. Thus, after the deposition of the monotonous sequence of Salas marls, the contact with a neritic-carbonate producing supply area was reestablished. The mechanism of sedimentary infill of the presumed channels and the directions of transport are not known.

The regressive development of the basin, although not yet understood in detail, clearly sets in with the lower Arén sandstone unit. It is marked by a decrease in the rate of subsidence.

B. *Arén sandstone sensu stricto*

The Arén sandstone proper, approximately 100 m thick, forms a marked ridge in the landscape. Although at first glance the sandstones (quartz-bearing to quartz-rich calcarenites, in fact; p. 267) appear to be uniformly developed, detailed observation reveals a highly complex facies differentiation ranging between shallow neritic and coastal dune environments.

At Orcau (Fig. 1), an approximately 20 m thick regressive sequence can be observed. It is parallel-bedded and bioturbated in the lower part, cross-bedded in the middle part, and parallel-laminated in the upper part. Grain size increases from very fine to medium grained at the base, to generally medium to coarse grained in the upper part. This sequence is interpreted as a coastal barrier capped by beach sands (Visher, 1965). Parallel-laminated calcarenites are well-developed in the Arén sandstone sensu stricto in the northern exposures in the Ribagorzana valley (outside the area shown in Fig. 1; 15 km west of Tremp). These calcarenites, also interpreted as beach deposits, are overlain by fine-grained, well-sorted and cross-bedded calcarenites, which in turn are directly overlain by the fluviatile red beds of the Tremp Formation. Grain size, sedimentary structure, and position in the sedimentary sequence strongly suggest that the fine-grained calcarenites accumulated in the form of coastal dunes. This coastal dune facies, first recognised by E. Oomkens (pers. comm.), was also found at the same stratigraphic level in the Ribagorzana valley on the north side of the Montsech mountain range.

The characteristics of the sequences at Orcau and in the Ribagorzana valley form direct evidence of the extreme shallowness and even emergence of parts of the Arén depositional area.

A section of remarkable completeness is exposed 3 km northeast of Isona, along the road to Coll de Nargó (Figs. 1 & 11). Overlying strongly bioturbated silty to fine sandy, partly nodular marls and fine-grained calcarenites (upper part of lower unit; p. 266), the Arén sandstone sensu stricto is here cross-bedded from base to top. Both planar and wedge-shaped cross-bedded units are present. Thick units (up to 2.5 m) predominate in the middle part of the section, while less thick beds are found in the lower and upper parts. Grain size shows a tendency to increase in an upward direction from fine to medium grained in the basal part to coarse and locally pebbly already in the middle part. Some levels carrying well-rounded quartz pebbles up to 5 cm in diameter (most pebbles 1–2.5 cm) occur in the top part of the sequence. Although the dips of the foresets vary considerably, high-angle cross-bedding is common (Fig. 12).

All fossil debris in the Arén sandstones proper is derived from marine organisms (p. 268). The well-sorted nature of the sands and the high-angle cross-bedding, the dimensions of which in the middle part of the section correspond to high sand waves, indicate a high-energy environment as may occur in shallow parts of a neritic environment with strong tidal current activity (Houbolt, 1968; Hoyt, 1967). The upward increase in coarseness in the Arén sandstones proper could correspond to an increasing general shallowness, the pebble-bearing horizons in the top part probably

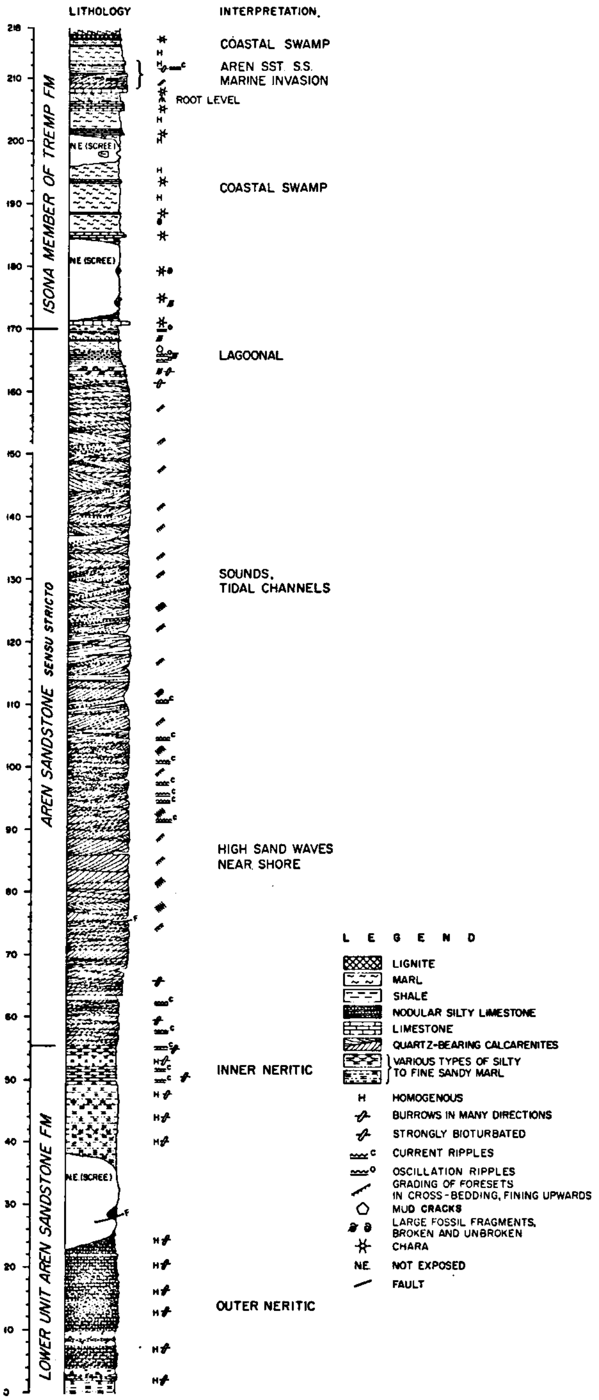


Fig. 11. Lithologic column and environmental interpretation of Arén Sandstone Formation and part of overlying member of Tremp Formation, road-side section Isona.

reflecting the high current velocities of tidal gullies.

The cross-bedding, measured in four intervals in the section, shows a full 180° reversal of predominant current direction from westwards in the lower part to eastwards in the upper part (Fig. 12). This change, which is thought to be due to the varying influences of ebb and flow tides, can only be fully understood when current data from neighbouring sections become available. The beds directly overlying the cross-bedded sandstones (Fig. 11) form the next phase in the regressive development. At the base of these deposits are a few silty, fine-grained calcarenite beds carrying in-situ rudists (hippuritids), pelecypod debris and corals, on top of which marls and fine sandy beds that have symmetric oscillation ripples are found. Some shallow channel fills and a mud-cracked horizon have also been noted. These deposits, some 8 m thick, must have accumulated in a quiet environment, which may well have been a shallow lagoon protected from the open sea by coastal barriers or possibly even barrier islands.

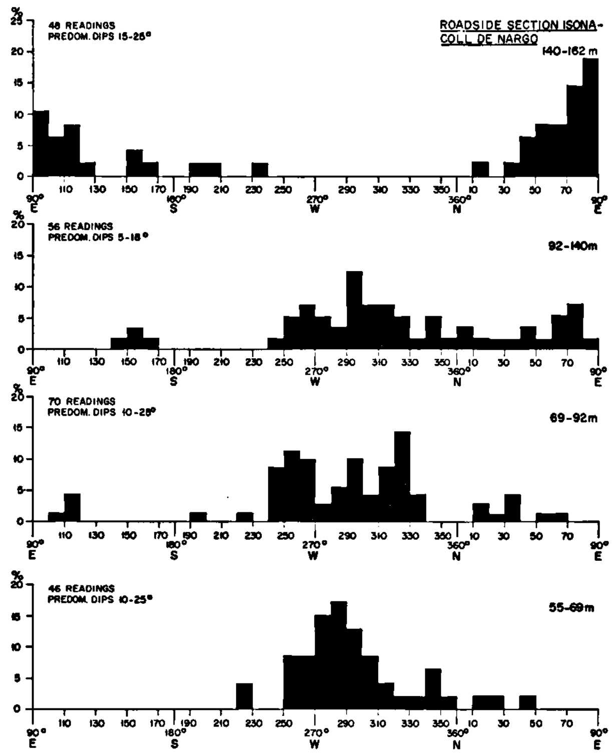


Fig. 12. Arén Sandstone Formation s.s. Directions of dip of foresets in cross-bedding.

Charophycophyta-bearing (non-marine) limestones, marls locally rich in vegetational debris and fluvial channel fills mark the completion of the regression with the establishment of a coastal-swamp environment. These coastal-swamp deposits, referred to as the Isona member, form the basal part of the Tremp Formation in this area.

Composition of the Arén sandstone sensu stricto. – A limited number of thin sections, four evenly spaced ones from the Arén sandstone proper and one from the lagoonal deposits at Isona, were analysed.

The quartz content in the calcarenite samples decreases regularly from 40.5% in the lowermost sample to 17.5% in the uppermost one. The quartz grains vary from angular to rounded; they are as well-sorted as the carbonate components. Among the carbonate grains, well-rounded calcareous algal fragments (rhodophyceans) and larger Foraminifera (*Orbitoides, Siderolites*) are particularly abundant. Other grains include mainly pelecypod and echinoderm fragments. Marly intercalations in the lower part of the sequence contain a neritic *Gavelinella-Cibicides* assemblage. The calcarenites are almost fully cemented with calcite that has a blocky texture.

The sample from the lagoonal deposits is a very fine-grained calcarenite containing 9.0% quartz; it is particularly rich in Ostracoda remains.

VII. TREMP FORMATION – COASTAL-SWAMP DEPOSITS AND CONTINENTAL RED BEDS

Three main facies types can be distinguished in the Tremp Formation: (i) coastal swamp deposits – the Isona member, (ii) lacustrine deposits (red beds), and (iii) fluvial channel fills and sheet-flood deposits (red beds).

The areal extent of the Isona member is well defined (Fig. 1). Facies types (ii) and (iii) show intricate interfingering; they have not been mapped separately.

A. *Coastal-swamp deposits – Isona member*

The gradual regressive development from the shallow-marine Arén sandstones sensu stricto, via lagoonal, into coastal-swamp deposits (the Isona member), was treated in the previous chapter. The coastal-swamp deposits, best developed near Isona where they are approximately 90 m thick, wedge out on the map towards the west; they are not found anymore at and to the west of Talarn (Fig. 1).

In colour, the Isona member is all shades of grey to black, while the other members in the Tremp Formation are generally yellowish grey to a distinct pale to moderate reddish brown (Rock-Color Chart, Geol. Soc. Am.). In the Isona member, fine-grained sediments – mainly carbonaceous shales and silty to fine sandy marls, usually rich in vegetational debris – predominate. Minable sapropelic lignite occurs at Suterraña and near Llorda (Fig. 1). No seat-earths directly below the lignites were encountered, although root levels occur at other horizons (Fig. 11). An interbedded relationship between the lignites and the Charophycophyta-bearing limestones, and a considerable lateral persistency suggest that the fine-grained organic matter accumulated, like the limestones, in wide and open brackish to fresh-water pools. Two channel fills containing coarse to fine sands, much coarse-grained vegetational debris and even logs, are found near Isona. They probably correspond to a tidal-creek system. One invasion of the sea into the swamp environment is recorded by a several metres thick cross-bedded unit of Arén sandstone proper, which directly overlies a Charophycophyta-bearing limestone (well exposed along the Isona-Coll de Nargó road; Fig. 11).

B. *Lacustrine deposits*

The Tremp deposits overlying the Isona member consist predominantly of homogeneous, generally pale to moderately reddish brown marls. Mottling with yellowish-grey spots is common. Small calcite concretions, present in large numbers throughout the deposits, are particularly abundant in fossil deep-root levels, several of which occur in the road-side section at Talarn. The distribution of calcite concretions locally reaches such a density that the horizons can be considered as fossil caliches (Nagtegaal 1969a, b). Charophycophyta-bearing limestones similar to those found in the Isona member are also encountered; one prominent horizon composed of several thick beds has been included on the map (Fig. 1). Common, though less abundant, are small gypsum concretions and up to 3 m thick gypsum beds (five such beds, two of which are being quarried, are well exposed some 5 km west of Tremp).

In the same marl sequence, just south of Orcau (Fig. 1) and at a level below the prominent horizon of limestone beds, an approximately 4 m thick unit of well-stratified, partly conglomeratic calcarenites is found (Fig. 14). The beds are of fairly constant thickness laterally. Some of them are graded, fining upwards, and show well-developed b-c-d divisions (Fig. 13, inset). Flute casts occur, but were only found on the lower surfaces of loose blocks. These beds are interpreted as turbidites and are regarded as the downstream lacustrine extension of fluvial channel fills and sheet-flood deposits (p. 270).

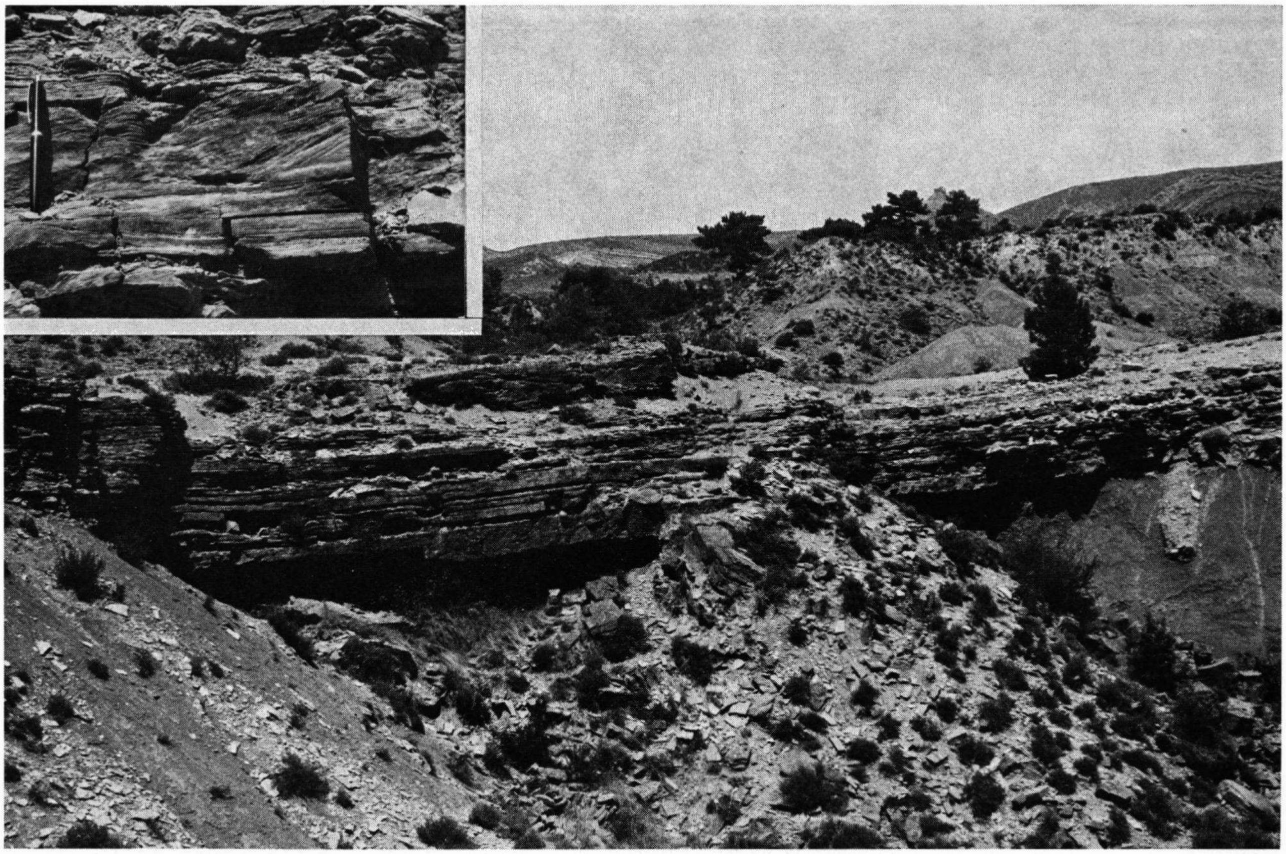
The environmental picture emerging from the observations is one of wide planar areas of fine-grained sediments which were sometimes dry (fossil root levels, caliche) and sometimes covered by, probably shallow, lakes (limestones, turbidites). The presence of gypsum concretions and of several gypsum beds is a direct indication of strong evaporation. However, considerable amounts of gypsum may have been carried in solution into the environment, derived from possibly exposed Triassic gypsum in the hinterland. Marked aridity may only have occurred temporarily, because no aeolian (dune) sands were found; the climate was probably semi-arid most of the time, as is evidenced by the levels of poorly to well-developed fossil caliche and common deep root levels.

C. *Fluvial channel fills and sheet-flood deposits*

Intercalated in the marls at Talarn, there is an approximately 20 m thick, reddish brown, coarse clastic unit with a strongly erosive base. The unit is made up of conglomeratic channel fills, cross-bedded calcarenite channel fills showing signs of limited

Fig. 13. Tremp Formation. Turbidite sequence intercalated between lacustrine homogeneous marls. Inset shows graded turbidite bed, having well-developed b, c and d divisions. Along road to Orcau.

Fig. 14. Tremp Formation. Sandy and conglomeratic braided stream deposits cutting into homogeneous lacustrine marls. Roadside section, Talarn. Hammer is 40 cm long.



lateral accretion, and paleosols developed mainly in calcarenites (Fig. 14). The channel fills cut into and are overlain by the paleosols, which are rich in calcite concretions and contain numerous still open, deep root traces. The fluvial system to which these deposits correspond was probably braided, with vegetated bars between the anastomosing gullies.

The orientations of erosion rills and scour-and-fill structures at the base of the unit, and the dips in pebble clusters showing imbricate arrangements, were measured, but they gave inconclusive results. The distribution of the erosion structures is narrow and unimodal; they are oriented NNW-SSE. The current directions deduced from imbricate pebble structure show a wider spread and an essentially bimodal distribution. Approximately one half of the readings show a maximum to the south (more or less coinciding with the orientation of the erosion structures); the other half is directed at right angles, to the northeast (Fig. 15).

South of Talarn and south of Orcau (Fig. 1), large exposures of 2–3 m thick and up to 200 m wide coarse-grained calcarenite sheets that have nonerosive bases are found. These sheets are parallel bedded; the calcarenite beds, usually two to four in number, are often homogeneous or laminated and contain 'floating' pebbles. They are separated by thin levels of fine sandy marls. The upper levels of the sheets commonly carry burrowing structures and root levels. Deposition is thought to have taken place by means of ephemeral streams flowing out over wide, flat areas; the wide sheets are consequently considered as sheet-flood deposits.

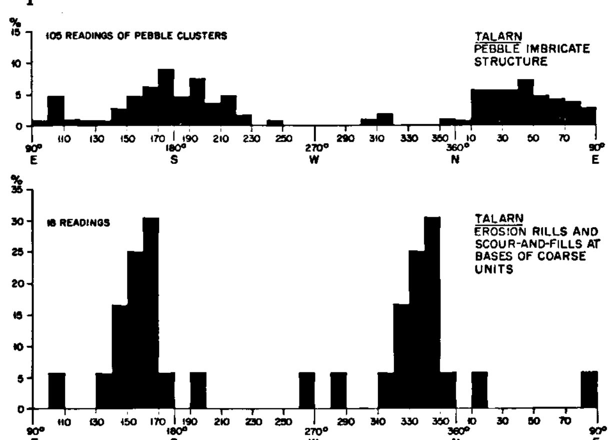


Fig. 15. Tremp Formation. Current directions in braided stream deposits, Talarn.

D. *Composition*

The sandstones in the Tremp Formation are largely made up of limestone fragments (60–90%) and are therefore referred to as calcarenites. Subordinate numbers of the grains consist of usually angular quartz, chert, and rounded Palaeozoic phyllites. The conglomerates consist almost entirely of Mesozoic limestone pebbles: a certain indication that at the time of deposition of the Tremp Formation areas of Mesozoic sedimentation had been uplifted sufficiently to act as supply areas. The sandstones are strongly cemented with various amounts of calcite.

VIII. CLAY MINERALS AND COAL RANK

A. *Clay minerals*

A number of samples were selected from each formation studied for the analysis of the clay fraction (Congost Limestone Formation: 16 samples; Anserola Formation: 15; Vallcarga turbidites: 16; Salas marls: 8; Arén sandstone sensu stricto: 12; Tremp Formation: 6). After crushing, treatment with 6% hot acetic acid, peroxyde, and separation of the <2 micron fraction, oriented mounts of the clays obtained were run on the diffractometer*.

On the basis of the diffractograms produced, smaller numbers of samples of each of the major rock units were analysed in more detail. From these first sets of diffractograms, variation in the clay mineral assemblages within each major rock unit appears to be fairly limited.

Continued treatment consisted of glycolation (2 hr vapour treatment at 70 °C; identification of montmorillonite and mixed layers), acid treatment (2 N HCl, 8 hr; removal of chlorite), and heating (2 hr at 550 °C; collapse of kaolinite, identification of type of chlorite). X-ray diffractograms of the samples considered to be representative are shown in Figure 16A-G.

Identification

The mineral swelling to 17 Å following glycolation and collapsing to 9–10 Å on heating is designated as montmorillonite. A non- to very slightly expandable mineral showing an integral series of (001) spacings from 14 Å is identified as chlorite. 'Well-crystallised' and 'poorly-crystallised' chlorites are tentatively distinguished from respectively, increased intensity or disappearance of the 14 Å peak following heating. The 10 Å reflection showing no shift on glycolation is taken to indicate the presence of illite. The 7 Å reflection remaining after HCl treatment and disappearing on heating is attributed to the presence of kaolinite. Direct evidence of the presence of both chlorite and kaolinite in many samples is the often good resolution of the 3.57 Å (kaolinite) and the 3.54 Å (chlorite) peaks. Different types of mixed layers are present, but no attempt has been made to identify them. For identifying the clay minerals mentioned, the indications given by Brown (1961) and Carroll (1970) have generally been followed.

In evaluating the relative proportions of particular clay minerals in the assemblages, the practice of Johns et al. (1954) and Biscaye (1965) has been followed. It consists of relating the 17 Å peak area (montmorillonite) to four times the 10 Å peak area (illite) and twice the 7 Å peak area (chlorite + kaolinite) in

* Philips diffractometer; working constants: Cu Kα radiation, Ni filter, KV 36, mA 20, scanning speed 2° 2θ min.

glycolated samples. The relative proportions of kaolinite and chlorite are found by comparing the HCl-treated and non-HCl treated samples, and the peak area ratios measured from the 3.5 Å reflections, the results of both procedures being in good agreement. However, no percentage values are given.

Because the clay minerals are generally fine grained and occur evenly disseminated in the marls and lime-mud matrices of calcarenites and calcisiltites, no direct optical observations could be made. Therefore, no definite ideas could be formed regarding possible authigenesis of some of the clay minerals.

Stereoscan photographs of the clay samples were also unsatisfactory in this respect, because the various clay minerals in the mixtures can not always be identified on the photographs.

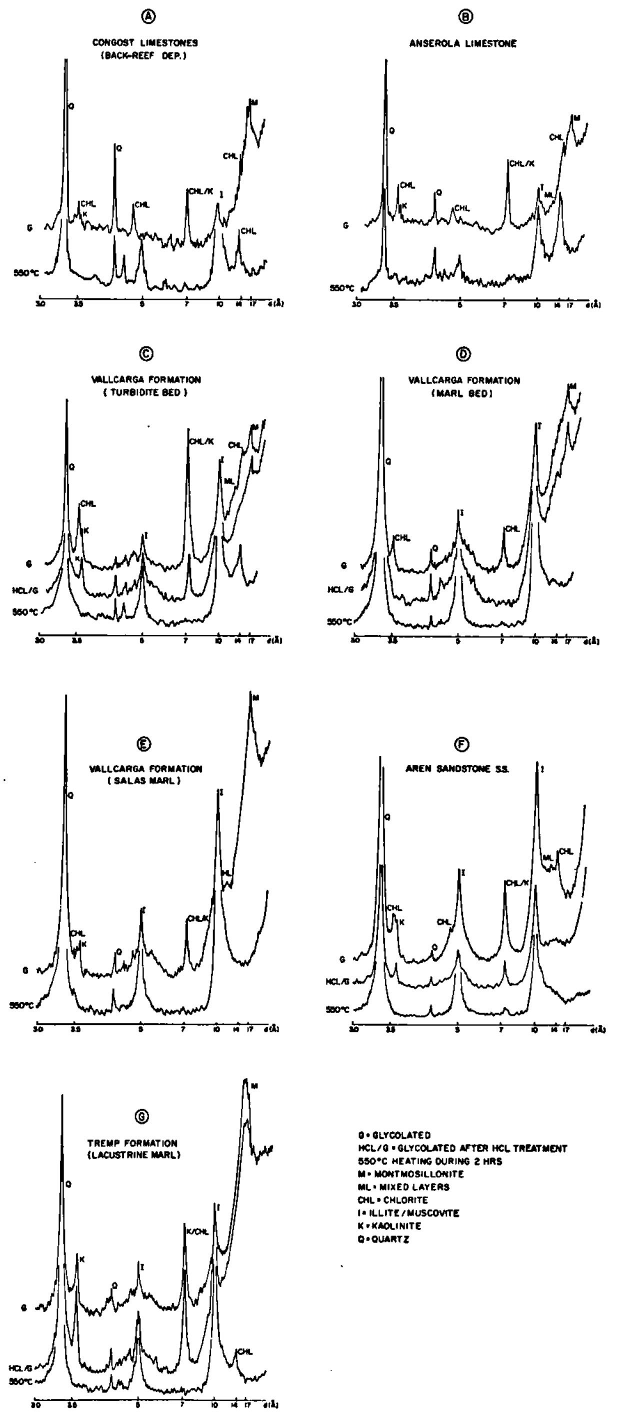


Fig. 16. X-ray diffractograms of the major rock units in the section studied.

Composition of the clay fractions

Congost and Anserola limestones (Fig. 16A & B). – The clay-mineral assemblages in these two formations are almost identical (except for the occurrence of glauconite in the Anserola limestones, p. 259). Montmorillonite and illite are the dominant clay minerals; chlorite, which makes up a significant part of the assemblages, tends to occur in somewhat higher amounts in the Anserola limestones. The chlorite, at least in part, is well crystallised and probably a high-Mg/low-Fe type, because the 14 Å peak is well expressed after heating and because it gives a full series of the first five basal spacings. Only traces of kaolinite occur in the Congost limestones, as is evident from a small and narrow residual 7 Å peak after HCl treatment (corresponding diffractogram not shown); slightly more kaolinite occurs in the Anserola limestones. In the Congost limestones montmorillonite is restricted to the back- and fore-reef deposits; it was not detected in the underlying barrier calcarenites. The illite gives rather broad and asymmetric 10 Å peaks and no, or only a minor, 5 Å reflection. It is therefore thought to be intensely weathered, and probably has a low K content (Weaver, 1965). This type of illite contrasts markedly with that found in the overlying formations (see below).

Vallcarga Formation, turbidite beds (Fig. 16C). – Chlorite plus kaolinite and illite are the dominant clay minerals, but montmorillonite, and some mixed layers are clearly evident. The major part of the 7 Å reflection in the glycolated sample can be attributed to chlorite, as comparison with the HCl-treated samples shows. The chlorite is well crystallised. Although the (003) reflection is less well expressed, the chlorite is probably still largely of the same type as found in the underlying Congost and Anserola limestones.

With the initiation of the turbidite facies, a 'new' type of illite is sedimented in the basin, up to the top of the sequence studied. It is characterised by narrow 10 Å reflections and low (001)/(002) peak ratios (Fig. 16C–G).

Vallcarga Formation, marls (including Salas marls) (Fig. 16D & E). – The clay-mineral assemblages in the marls of the Vallcarga Formation are of comparable composition. They consist predominantly of illite (all diffractograms showing sharp and symmetric 10 Å peaks) and montmorillonite. Chlorite is present in lesser quantities than in the deposits discussed previously and is also poorly crystallised. No kaolinite occurs in the marl sample shown in Figure 16D, but traces were found in some other marl samples from in between the turbidite beds. In the samples of the Salas marls, however, kaolinite is more frequent (Fig. 16E).

Arén sandstone sensu stricto (Fig. 16F). – Illite and chlorite plus kaolinite are the dominant clay minerals. Montmorillonite, an important constituent of the assemblages in all the other formations, appears to be hardly present in the Arén sandstones sensu stricto (traces of montmorillonite were found in one sample only). Chlorite (poorly crystallised) predominates over kaolinite. Mixed layers are present.

Tremp Formation (Fig. 16G). – Only the clay fraction of the red beds has been investigated. The diffractograms of the clays in the lacustrine marls and in the fluvial calcarenites are very similar. Illite, montmorillonite and kaolinite are the dominant constituents. The considerable amount of kaolinite, compared with the generally low contents in the previously-treated deposits, is particularly striking. The presence of some chlorite is apparent from the small 14 Å reflections after heating.

B. *Coal rank (vitrinite reflectance)*

Vitrinite reflectances have been determined for nine samples evenly spaced in a 500 m interval in the lower reaches of the Vallcarga turbidites (Fig. 2). The values obtained range between 0.56 and 1.45% absolute reflectance, but show a distinct maximum at 0.96% (total of 202 readings). The corresponding fixed-carbon content, read off from the nomogram published by Kötter (1960) is 67%.

C. *Discussion*

The main factors determining the composition of clay-mineral assemblages in marine sedimentary sequences are composition of source area, paleoclimate, differences in settling rates of clay minerals, halmyrolysis, and diagenesis. This complexity makes the interpretation of such assemblages a difficult undertaking, and there is, in the sedimentary sequence being discussed, often uncertainty with respect to the relative influence of each of these factors.

Clay minerals supplied to the basin

The composition of the Upper Cretaceous basin fill reflects a continuous, high production of bioclastic carbonate in neritic environments. From an actualistic point of view, this means that the climate was subtropical to tropical (Fairbridge, 1967). Montford argued for a tropical humid climate in the North Atlantic region during the Upper Cretaceous, and Triat & Médus (1970) concluded, mainly on the basis of palynological evidence, that the climate in Southeast France during the Lower Santonian was of a type in which laterite was formed.

A tropical humid or lateritic type of climate is not directly evident from the clay-mineral assemblages found in the South Pyrenean sequence. Although montmorillonite and kaolinite (both minerals that are preferentially formed in such climates) are generally present, chlorite predominates over kaolinite in most of the samples analysed (except in those from the Salas marls and in the Tremp Formation). The present-day distribution of clay minerals in the world's oceans shows not only that chlorite is concentrated near areas of limited chemical weathering (e.g. the polar regions), but also that influx appears to be from arid regions (Griffin et al., 1968, Rateev et al., 1969). The characteristic association of chlorite with kaolinite and montmorillonite as found in the present samples suggests that either: (i) the paleoclimate was subtropical to tropical, but arid to the extent that chlorite was not fully altered in the continental soil material, or (ii) the climate was subtropical to tropical and indeed humid, as suggested by the authors mentioned above. In this latter case the chlorite, at least in part, should be authigenic. It could theoretically have formed from gibbsite (halmyrolysis), as recorded from Pacific sediments by Swindale & Pow-Foong (1967), or from other clay minerals (montmorillonite) during any stage of diagenesis. It is difficult to show whether this actually took place.

The continental deposition in the top part of the section, the Tremp Formation, occurred under hot semi-arid to arid conditions, as is evidenced by the oxidised nature of the sediment (low water tables), and the occurrence of caliche, gypsum concretions and gypsum beds (p. 268). Although such an environment is not conducive to the formation of kaolinite, this mineral is nevertheless present in considerable amounts. In the case of the Tremp deposits, the kaolinite (as well as the ferric oxides in the red beds) could have been derived from previously formed lateritic soils. However, no direct evidence of laterite formation in the form of gibbsite has been detected, either in the underlying deposits or in the Tremp deposits.

The levels where montmorillonite is scarce or absent are those of high-energy deposits, the barrier calcarenites in the Congost limestones and the Arén sandstones sensu stricto. Slower settling rates for montmorillonite as compared to illite, kaolinite and chlorite are thought to offer an explanation for this montmorillonite distribution (Whitehouse et al., 1960; Porrenga, 1967).

The illite is thought to have been derived directly from low-grade metamorphic and sedimentary rocks in the continental source areas. Paleozoic rocks, which make up the core of the Pyrenees, are likely sources, but there is no independent evidence to show that these rocks were exposed. The occurrence of poorly-crystallised illite in the two lowermost formations, the Congost and Anserola limestones, could result from vertical sedimentation from highly diluted 'open-marine suspensions'. The finest, and therefore most strongly altered, clay fractions were deposited in this way. The introduction of the well-crystallised illite coincides with the beginning of turbidity-current deposition. The abundant supply of this 'new' type of illite may have resulted from negligible influence of settling rates due to the prevalent rapid lateral sedimentation, and from a new cycle of erosion in the continental source areas, which may have been initiated due to upheaval accompanying the formation of the turbidite trough.

Burial diagenesis

The fact that discrete clay minerals and the composition of clay-mineral assemblages show systematic changes with increasing depth of burial is now well documented (Dunoyer de Segonzac et al., 1968; Burst, 1969; Dunoyer de Segonzac, 1969; Kisch, 1969; Hayes, 1970; Perry & Hower, 1970). The earliest changes, most commonly noted and easily detected in standard-procedure-oriented clay mounts, are the conversion of montmorillonite, via illite/montmorillonite mixed layers, to illite, the increase in illite crystallinity (Kubler, 1968), and the disappearance of kaolinite.

The conversion of montmorillonite to illite/ montmorillonite mixed layers appears to set in before the disappearance of kaolinite. This is seen in the Upper Cretaceous Douala Basin (Cameroun), where montmorillonite strongly decreases and illite/montmorillonite mixed layers start to form at a depth of approximately 1500 m, while kaolinite is generally present up to some 3000 m (Dunoyer de Segonzac, 1964). The co-existence of kaolinite and illite/montmorillonite mixed layers showing a downward decrease in expandability is also evident in Gulf Coast wells (Pleistocene-Eocene; marked decrease in expandability of mixed layers from approximately 2000 m downwards, irregular mixed layers up to approximately 3000 m, and regular mixed layers and kaolinite up to more than some 5000 m; Perry & Hower, 1970). In deep wells, the illite crystallinity may show a considerable scatter, but a marked increase in crystallinity has been noted from 2000–3000 m downwards (Kubler, 1964; Dunoyer de Segonzac et al., 1967).

In the samples from the Upper Cretaceous section studied, none of these clay-mineralogical changes are observed. 17 Å montmorillonite is generally present and occurs in the lowermost formation, kaolinite is generally present, and the illite shows no downward increase in crystallinity (poorly-crystallised illite occurring, to the contrary, in the two lowermost formations). In combination with the fact that the reflectance measurements indicate only a limited degree of organic metamorphism (fixed carbon 67%, p. 272), and assuming a normal geothermal gradient, this suggests that the lowermost formations in the sequence were not buried much deeper than some 3000 m, which is the value obtained by adding the thicknesses of the formations in the sequence (Fig. 2). This confirms the assumption that the history of the basin, which started with accelerated subsidence in the Santonian (p. 257) ended in the Lower Paleocene, and that the area can never have been a site of appreciable Tertiary sedimentation.

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