Olisthostromes and other delapsional deposits, and their occurrence in the region of Moratalla (Prov. of Murcia, Spain)

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In the first part of this publication an attempt is made to bring some order in the terminological chaos existing in the field of *delapsion,* which is the independent gravitating of rock material along sloping surfaces. The term *olisthon* is introduced for everything that has been delapsed. For the purpose of a rational and comprehensive classification of geological phenomena delapsion is introduced as a separate class between tectonization and sedimentation. Special emphasis is laid upon the concepts of *olisthostrome* and *olistholith,* which are redefined. The principles of the mechanics involved in delapsional processes are discussed.

In the second part emphasis is laid upon the description of the olisthostromes, olistholiths, reversely graded mud flow and muddy gravel avalanche deposits that invaded the Moratalla basin during the Miocene. Attention is also given to the olisthostrome occurring in the Lower Eocene, the slump packets and slip blocks in the Miocene, and to the extensive slip sheets of Cretaceous rocks that closed the depositional phase of the basin. The regional stratigraphy and tectonic structure are briefly discussed. The interpretation of the structures – the greater part of the Triassic rocks of the Subbetic zone is thought to be in autochthonous to parautochthonous position – differs from the current interpretation that the Triassic rocks form the basal part of the Subbetic nappe that would have thrust over some twenty kilometres.

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Preface

The investigations that led to this publication were twofold:
1) Geologic field work and the processing of the data obtained were done during the summers of the years 1963, 1964, 1967, and 1968. The field work during the first two summers was an examination requirement for the Dutch degree of 'doctorandus' in geology at the University of Amsterdam, and it merely had the aim of obtaining a stratigraphic and structural picture of the region. Then it appeared that olisthostromes and many other gravitationally emplaced rock masses occur in the investigated area. The field work during the last two summers had therefore the aim of investigating these olisthostromes and related phenomena. This was chosen as a subject for my dissertation.
2) A thorough investigation of the literature concerning olisthostromes and related phenomena was done. A large amount of facts was brought together in order to obtain a comprehensive picture of the action of gravitation in the rocks of the earth's crust.

These twofold investigation resulted in a dissertation consisting of two parts.

Acknowledgments

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Part 1. Terminology and concepts

> Desidero insistere in modo particulare sulla estrema necessità di non coniare termini nuovi quando esistono già quelli correspondenti. Le leggi della priorità nella terminologià scientifica non è qualcosa di facoltativo, né constituisce un riconoscimento di meriti. E' semplicamente condizione di vita per il linguaggio scientifico e per la possibilità di comunicazione che ne deriva. Senza il rispetto della priorità terminologica arriviamo rapidissimamente alla torre di Babele.
>
> L. Ogniben, 1967.
>
> The mixing of genetic criteria with observational criteria is a particularly common source of confusion.
>
> B. Bayly, 1969.

Introduction

Presumably the majority of the mechanical processes that occur or have occurred during the earth's history are the direct or indirect manifestation of gravitation. Since it is still doubtful to what extent the processes that geologists have called endogenic processs, and which among other things may result in the formation of a relief on the surface of the lithosphere, are the direct or indirect manifestation of gravitation, we will disregard these processes in the present study. When we henceforth refer to gravitational processes, we only mean those surficial processes which tend to smoothen the relief of the surface of the lithosphere into a plane that parallels the surface of the geoid.

These gravitational processes can be divided into 1) exogenic processes resulting in the displacement and the deposition of solid parts of the earth's crust without the primary help of tractional forces exerted by a mobile medium travelling in the same direction, and 2) those with the primary help of these tractional forces.

The processes leading to the formation of olisthostromes should be grouped into the first category of gravitational processes. In this part we will attempt to make clear what to understand by 'olisthostrome' and what position olisthostromes have in the whole of geologic phenomena. The essence of the controversies around the olisthostrome concept is connected with the question, where the boundary between tectonization and sedimentation should be drawn. The settling of this boundary became urgent only after the introduction of the olisthostrome concept. In order to answer this question it appears necessary to make an attempt to restore order in the terminological chaos in the field of geologic phenomena related to olisthostromes.

The concept of delapsion

ARGILLE SCAGLIOSE

The problem where to draw the boundary between tectonization and sedimentation can well be illustrated by going into the controversies concerning the nature of the argille scagliose of the Northern Appennines (Italy).

Concept of argille scagliose – The term 'argille scagliose' was introduced by Bianconi (1840) to designate a peculiar kind of clay which occurs in large masses in the Tuscan-Emilian Appennines. This clay characteristically falls apart into smooth, soft, lustrous scales, which are greasy on touch, and to which the rock owes its name (Appendix 1).

The concept of the term 'argille scagliose', originally merely a petrographic name for a particular kind of rock, has passed through the same regrettable development as the terms 'flysch' and 'wildflysch' in that a particular genesis has been added to the petrography. Since then the term has become controversial, because the supposed genesis did not cover completely its petrography, which was therefore repeatedly adapted to the genesis (and the reverse) which varies according to the different opinions of the various investigators. For instance, the argille scagliose masses have been considered as rather dismembered nappes, or as large landslides (Migliorini, 1933, 1949, "frane orogeniche"; Annelli, 1938, "frane tettoniche").

Ogniben (1953, 1954a, b) explains the scaly structure of the argille scagliose as the result of diapiric rising of clayey rocks from anticlinal cores due to tangential stresses. He considers the argille scagliose from a petrographical point of view as a special kind of tectonite (according to the definition of Sander, 1948), but of extremely superficial type, so that the mechanical deformation that gives rise to the scaly structure is not attended by a metamorphic change of the clay minerals.

Since the early fifties the concept of argille scagliose adhered to by many investigators has been that expressed by Merla (1953). He considers the argille scagliose as allochthonous formations generated by mud flowing and sliding on the slopes of orogenic ridges during the Late Cretaceous and Tertiary, and covering vast areas of the Appennines; they lie *tectonically transgressive* upon the autochthonous formations of the Appennines. According to Merla (1953) the argille scagliose are the result of tectonic instead of sedimentary phenomena (Appendix 2). In 1956, however, Merla suggests that the term 'argille scagliose' should only refer to the chaotic clays and that the complex should be designated as *allochthonous terrains*.

The allochthonous complex of the Northern Appennines can (Maxwell, 1953, 1959; Abbate et al., 1970a) briefly be described as a discontinuous sheet composed of eugeosynclinal material, which is a chaotic mixture of contorted and slickensided grey claystone with numerous more or less competent inclusions that are extremely diverse in size and lithology, and of very large masses of the same lithologies forming whole mountains, the largeest of which are some 40 to 60 km long. The smaller blocks embedded in the shaly matrix commonly are lenticular, rounded, boudinaged, crinkled, or of cracked surface. These facts suggest that the chaotic structure developed when the rocks were still imperfectly lithified. The larger slabs show their original stratification and are randomly distributed; the directions of dip of the strata in these slabs are highly variable. The slabs may

have a normal or reversed attitude or are steeply dipping; they may be plane, synclinoid, folded, or faulted; the bases of these slabs may be parallel to the internal stratification or may cut it obliquely.

Because of the abundance of greenstone fragments the allochthonous complex is usually called 'the ophiolite formation'. In addition to the Upper Jurassic greenstones (frequently associated with cherts), blocks and large masses of pre-Triassic granite, grey marlstones interbedded with fine-grained grey limestone beds (Albarese), and Upper Cretaceous sandstone (Pietraforte and Monte Antola Sandstone) lack representatives in the rock sequences of the substratum and must have been transported over a considerable distance. The ophiolite masses decrease markedly in size and abundance from southwest to northeast, suggesting a provenance from the Tyrrhenian sea area.

Next to these rock masses of unknown origin one finds in the allochthonous complex: Upper Triassic dolomites, uppermost Jurassic and lowermost Cretaceous cherty Calpionella limestone (Maiolica), Upper Cretaceous to Eocene Scisti Policromi, Oligocene and Miocene flysch formations (Macigno and Marnoso-arenacea). These rocks can be found in the rock sequences of the miogeosynclinal substratum (the Tuscan and Umbrian sequences).

In the southwestern and central ranges of the Northern Appennines the allochthonous complex overlies the Oligocene flysch of the Tuscan rock sequence with generally apparent conformity; in the northeastern ranges it overlies autochthonous Miocene flysch; south and west of Bologna and southwest of Rimini the complex locally overlies uppermost Miocene and even Pliocene deposits. Thus it has some of the ostensible characters of a sediment in that it is laid down upon a depositional surface and that it has been folded together with the formations on which it lies; on top of it, locally, Upper Miocene and Pliocene marine and lacustrine deposits have been laid down (Maxwell, 1953, 1959; Abbate et al., 1970a).

Concept of olisthostromes – Similar 'allochthonous terrains' are found in Sicily. On the occasion of the Fourth World Petroleum Congress in Rome (1955) Beneo mentions the presence of allochthonous intervals of "argille scagliose" (scaly clays like in the Northern Appennines) and "argille brecciate" (brecciated clays) intercalated in beds of normal deposition in the Tertiary and Quarternary successions of Sicily.

According to the published text of the discussion after this lecture, Flores (1955) adressed the meeting saying: "The presence of intervals with chaotic petrography intercalated with beds of normal deposition became evident. These intercalations have been found to be limited in both time and space . . . We have not found it practical to discriminate between "argille brecciate" and "argille scagliose" outside the realm of petrography. In fact, chaotic units of argille scagliose type, as a result of their mode of origin (large scale submarine slumps or slides) contain masses of argille brecciate type. Furthermore, as one may find in the same stratigraphic intervals all gradations of gravity deposited sediments, from fine-grained beds deposited by turbidity currents to conspicuous plastic masses with englobed rigid bodies, it is, again, difficult to differentiate between argille scagliose and argille brecciate in terms of stratigraphic units."

"Consequently, while working in Sicily and confronted with the same problems as Beneo, we have found it convenient to introduce a nomenclature for chaotic intervals occurring within the stratigraphic column. This is due primarily to the fact that the available terminology to indicate this type of sediments is poor or non-existent. Sediments of this kind are commonly referred to as "slumps" or "slides". While these names refer to the action of slumping or sliding they do not refer to the accumulated sediment. Further, the two terms are also used to designate some intraformational irregularities. As to the terms argille scagliose and argille brecciate, their use is necessarily restricted to the Appennines and Sicily, while phenomena of this kind are likely to have occurred elsewhere also. It is necessary to introduce a more general term of universal value. The terms we have introduced in our internal reports to designate

chaotic intervals within the stratigraphic column in Sicily are: Olistostrome: from the Greek words "olistomai" (to slide) and "stroma" (accumulation). By analogy with biostrome (accumulation due to life) olistostrome indicates accumulation due to sliding."

"By olistostrome we define: those sedimentary deposits occurring within normal geologic sequences that are sufficiently continuous to be mappable, and that are characterized by lithologically and (or) petrographically heterogenous materials, more or less intimately admixed, that were accumulated as a semifluid body. They show no true bedding (except for possible large inclusions of previously bedded materials)."

"In any olistostrome we distinguish a "binder" or "matrix" represented by prevalently pelitic, heterogenous material containing dispersed bodies of harder rocks. These latter may range in size from pebble to boulder and up to several cubic km. There is no constant ratio between the total volume of the inclusions and that of the bedding mass."

"The name olistolith from the Greek wordt "olistomai" (to slide) and "lithos" (rock) is applied to the masses included as individual elements within the binder. The masses were previously referred to as "exotics" or "erratics". In some olistostromes one or more types of accumulation due to flowage can be recognized, ranging from the chaotic deposition of coarser elements which were bodily detached from their original position to "graded bedding" due to turbidity currents."

It follows from these quotations that olisthostromes are considered as sedimentary deposits and that Flores (1955, 1956, 1958) regards the allochthonous complex of the Northern Appennines as olisthostromes.

Since the Greek word is ολισθαινω, the new terms should be spelled: olisthostrome and olistholith.

Concept of mélanges – In connection with the controversy whether the allochthonous complex of the Northern Appennines should be regarded as the products of tectonization or as sedimentary deposits, it is relevant to discuss the concept of mélange, meaning mixture.

In geology the term 'mélange' was first used by Greenly (1919) to describe an intricately mixed assemblage of isolated, relatively obdurate rock fragments of different lithology and size, floating in a more tractable fine-grained matrix in the Precambrian of Anglesey Island, Wales (Appendix 3). He has furnished the term 'mélange with the adjective 'autoclastic', because he thought it to have been generated by a tectonic crushing of a rock sequence originally constituted of alternately more and less obdurate beds (crush-conglomerates of Geikie, 1896). Shackleton 1969), however, has given decisive evidence that the Anglesey mélange is a Precambrian olisthostrome, hence a "sediment".

In the Ankara region (Turkey) a chaotic complex has been found showing close analogies with the Anglesey mélange and with the allochthonous complex in the Northern Appennines. It has been called mélange by Baily & McCallien (1950, 1953), Gansser (1959) and Boccaletti et al. (1966). Baily & McCallien interpreted the Ankara mélange as the result of a tectonic brecciation beneath a nappe of crystalline rocks which was pushed several hundred kilometres southward, bulldozing the Taurus Mountains in front. This tectonic interpretation is severely disputed. Gansser (1959), however, regards the Ankara mélange and the allochthonous complex of the Northern Appennines as olisthostromes in the sense of Flores (1955), thus as sedimentary deposits. Subsequently the allochthonous complex of the Northern Appennines has been advanced as one of the most important examples of mélanges (Baily & McCallien, 1963; Boccaletti et al., 1966; Hsü, 1968). Boccaletti et al. (1966), like Jacobacci et al. (1959), Jacobacci & Martelli (1962), and Jacobacci (1966), are of the opinion that olisthostromes should be separated from the "argille scagliose" of the Northern Appennines (which they refer to as mélanges)

on account of the fact that the former are sediments by definition, whereas the latter would be tectonic by definition. They refer, however, to the definition of "argille scagliose" given by Merla (1953, genetic concept) instead of that given by Bianconi (1840, descriptive term).

Hsü (1968), following the current tendency, has unfortunately restricted the concept of mélange to the products of tectonic fragmentation and mixing as a result of sliding under the force of gravity (Appendix 4). The Franciscan complex in California, U.S.A., has been put forward by him as an important example of mélanges (Hsü, 1968, 1971; Hsü & Ohrbom, 1969). Recently the Franciscan mélange is interpreted as the result of intricate fragmentation and mixing of parts of the oceanic crust and its sedimentary cover, possibly with parts of the continental crust and cover, in the so-called Benioff shear zone along which shear zone oceanic crust is underthrusting the continental crust. Mercier & Vergely (1972), who also think that the fragmentation in mélanges should be tectonic, give many examples of yet different types of mélanges.

So it appears that very different geologic processes may result in mélanges, which therefore had better be defined as *the products of deformational fragmentation and mixing of rocks, irrespective as to whether the deformation is of tectonic nature or not.*

Various interpretations – At present all investigators agree about the gravitational emplacement of the allochthonous complex of the Northern Appennines (Merla, 1953, 1964a, b; Maxwell, 1953, 1959; Trevisan, 1956; Signorini, 1956; Abbate et al., 1970a). The main point of controversy concerns the nature of the complex. The various opinions are: 1. The allochthonous complex consists of olisthostromes invaded into normal autochthonous sedimentary sequences. 2. The complex is a rather dismembered nappe (the Ligurid nappe), which is extensively chaotic in consequence of gravity sliding. 3. The complex is a tectonic mélange. 4. The complex is the result of orogenetic landslides forming a chaotic mass; the forward movement of the allochthonous cover would be mainly caused by repeated sliding down the slopes of the successively formed orographic ridges when still in a process of uplift.

In fact, the difference between the opinions 1) and 2) to 4) reflects the controversy regarding the question whether pure gravitational displacement should be considered as a sedimentary or as a tectonic process.

Beneo (1955, 1956a, b) stresses the similarity of the structural characteristics between the classical Tuscan-Emilian "argille scagliose" and the "argille scagliose" masses in Sicily which serve as the type of olisthostromes; he postulates an analogous origin.

Ogniben (1967) advocates an intermediate position for olisthostromes between the products of tectonization and those of sedimentation. According to him olisthostrome accumulations are too small to enter, from a systematic point of view, into the field of "normal" structures of tectonic transport, such as thrust-sheets and imbrications, and they cannot be grouped into the field of sedimentary mud flows, owing to the circumstance that they lack all evidence of "fractioning into sedimentation elements deposited in exogenic media that is characteristic for sedimentary rocks".

CRITERIA HITHERTO USED TO DISTINGUISH BETWEEN TECTONIZATION AND SEDIMENTATION

Hitherto the boundary between the genetic concepts of sedimentation and tectonization has been defined by non-genetic criteria instead of by genetic ones as they should be.

State of stratal disruption – Jacobacci (1966), for instance, uses the adjective 'tectonic' to indicate a series of processes in consequence of which pre-existing rock masses assume a new spatial distribution maintaining, however, wholly or partly their mutual stratal connections. He uses the adjective 'sedimentary' to indicate a series of processes of relatively slow deposition that gives rise to a new rock by the accumulation of grains derived from a more or less intense chemico-physical destruction of pre-existing rocks (this concept is, of course, restricted to clastic rocks). Jacobacci considers an olisthostrome as that particular sedimentary phenomenon consequent upon a sudden mechanical breakdown ("frana") of pre-existing rocks and upon so rapid an emplacement as to permit only a disordered mixture of the elements that constitute the olisthostrome itself.

In fact he uses merely an accidental and gradational state of stratal disruption of the rocks involved in the downsliding motion as a criterion for distinguishing between tectonic and sedimentary phenomena. But a distinction between Tuscan-Emilian "argille scagliose" ("tectonic") and olisthostromes ("sedimentary") on these grounds is merely a matter of degree and non-genetic.

State of induration – Hsü (1968), when discussing the tectonic nature of mélanges, uses the adjectives 'tectonic' and 'sedimentary' without specifying exactly what he means by them.

Regarding the distinction between mélanges and olisthostromes he writes however: " . . . two different sets of processes could lead to stratal disruption and to mixing. In mélanges, the fragmentation and mixing were tectonic. However, the disintegration of a rock-stratigraphic unit into large blocks and a mixing of such blocks with muds could also have resulted from resedimentary processes, such as submarine sliding, or a combination of erosion and sliding. A similarity between sedimentary and tectonic mixtures led to the designation of both types of mixtures by the same name (for example "wildflysch"). Yet they constitute very different types of map units. A distinction should be made and different terms should be used wherever genetic interpretations are possible. The term "mélange" should be restricted to designate a tectonic mixture. Where chaotic masses owed their mixing to a sedimentary process, they should be called olistostromes . . . Although both mélanges and olistostromes have been considered as having originated as a result of sliding under the force of gravity, conditions under which the sliding actually took place were very different. Mélanges resulted from the deformation of consolidated rock under an overburden. Olistostromes, on the other hand, resulted from the submarine sliding of unconsolidated sediments."

Thus the distinction between mélanges sensu Hsü (1968) and olisthostromes, in other words between tectonic and sedimentary phenomena, is merely reduced to a distinction in the accidental and gradational state of induration or competence of the rocks involved in the process of mixing. Moreover, Flores (1956), when repeating his definition of olisthostromes in a published letter to the chairman of a symposium on "argille scagliose", has omitted the phrase 'accumulated as a semi-fluid body' of the original definition, arguing that it was irrelevant with respect to the terminology of olisthostromes. Therefore a distinction between tectonization and sedimentation of the supposed state of induration or competence of the rocks

involved in the process is unsatisfactory, because it is based on non-genetic criteria.

Size – A generally applied criterion to establish whether olisthostromes are sedimentary or tectonic phenomena and to distinguish between sedimentation and tectonization in general, is the size of the rock masses involved.

According to Hsü (1968), who considers mélanges as tectono-stratigraphic units (which are mixtures of rock-stratigraphic units resulting from tectonic deformation), "the size of the blocks . . . may indicate whether a chaotic mass is an olistostrome or a mélange. Movements of blocks tens of hundreds of feet long under surficial conditions can still be envisioned as a sedimentary process. However, displacement of large slabs hundreds of square miles in extent and a few miles thick must be considered as tectonic movements."

According to Merla (1964a) there is a whole, conceivable scale of phenomena and of effects substantially similar with respect to their causes and their nature, but fundamentally different in their dimensions. He says that at the extreme of their major dimensions one may conceive slide masses, disconnected, but substantially internally ordered, of dimensions of some tens of kilometres and a thickness in the order of thousands of metres. These masses would move (in the case of these major masses next to gravity also a push is conceivable) more or less independently from each other and tend to disintegrate later on, being surrounded and immersed in minutely chaotic material. In these cases Merla does not hesitate to treat them as tectonic phenomena.

Merla continues that at the extreme of the minor dimensions flowed masses of a few metres thick can be found, in the argillaceous matrix of which are wrapped lithic olistholiths of the dimensions of pebbles; and the flowed mass, if its movement has occurred subaqueously, may be preceded or terminated by purely sedimentary and graded strata like turbidites due to the deposition of finer materials that had been put into suspension in the surrounding water. Merla writes that in this case (of minor mud flows with possible associated turbidites) one may also speak of sedimentary phenomena.

According to Merla (1964b) olisthostromes would represent gravity slides which are smaller than those of the allochthonous cover of the Northern Appennines and which entered the sedimentary sequence deposited ahead of the advancing slide front of the allochthonous cover. He is of the opinion that the large Pliocene olisthostromes of the fore trough are not different from the allochthonous cover, but merely represent the final stage of the gravity sliding wave of the Appennines.

So a distinction between tectonic and sedimentary phenomena on the basis of the size of the rock masses involved is also merely a matter of degree and depends entirely on an arbitrary choice of the investigator (Fairbridge, 1946; Dott, 1963; Maxwell, 1959; Hubbert & Rubey, 1959; Merla, 1964a).

DELAPSION AND OLISTHON

Definitions of delapsion and olisthon – From the foregoing discussions about the hitherto proposed boundaries between tectonization and sedimentation it can be concluded that all phenomena comprising purely gravitatively displaced rock masses along slopes are wholly transitional from one to the other and that they constitute a continuous series. All gradations exist between turbidites at the one extreme of the series to immense slide blocks; these blocks in turn continue to

range in size up to huge masses resembling tectonic thrust plates at the other extreme (Crowell, 1957). Therefore it is highly preferable to regard this continuous series as an inseparable unity instead of dividing it, on unsubstantial and non-genetic grounds, into two parts of which one is considered as tectonic (gravity tectonics) and the other as sedimentary (gravitationally displaced sediments, resedimentation).

Von Bülow (1954) unites all the processes that give rise to these phenomena under the term "Detraction" (Latin: detractio = the pulling down) defining it as "alle Massenbewegungen unter dem ausschieszlichen oder bestimmenden Einflusz der Schwerkraft, bei denen also ein anderes Transportmittel fehlt oder untergeordnet ist". He regards "Detraction" as a specific kind of exodynamic process. Voigt (1962) explicitly uses the term "Detraction" also for submarine slide masses. Unfortunately the term 'detraction' has another implication in English viz. the lessening of and damage to reputation, esteem, or credit by malicious misrepresentation or false charges (Webster's English Dictionary). A better term, therefore, is *delapsion* denoting the process of sliding down (from Latin: delabi = to slide down); this term is used in the present publication. *The term 'delapsion' covers the surficial processes 1) that manifest themselves as independent displacements of solid rock material, whether or not in suspension, along sloping surfaces, except along the lee sides of current-generated ripples and mega-ripples, after the detachment of this material from its source or lateral parent rock equivalent, and 2) that are effected only by the tangential pull of gravitation on the solids; these processes are neither primarily effected by any endogenic force except that providing a slope, nor by any exogenic tractive force generated by a mobile medium, gas, fluid, or solid moving in one mean direction.*

It is proposed here to designate these phenomena as 'delapsional' and the action or process as 'delapsion'. For the entire body of rock, which is the result of one event of delapsion (one depositional unit) the term *olisthon* is proposed. Olisthon means something that has slid, from Greek: ολισθον, the neuter of the participle of the second aorist. Although strictly taken the plural form of olisthon should be olisthonta, it is preferred here to apply the English rules of the formation of plural forms, olisthons. The term 'gravitite' is less specific and therefore less preferable. The portions of rock dislodged by the delapsing mass from its substratum and transported with it are considered to be parts of the olisthon. The rocks that are only passively transported on the back of a delapsing rock mass are also considered to be parts of the olisthon. Finally, also the rocks deposited on top of the rock mass that is in course of delapsion and which became involved in the further delapsion of the mass are regarded as parts of the olisthon (Lava flows are essentially down-flowing non-solidified fluids; they are therefore maintained in the category of igneous rocks and not incorporated in the concept of delapsion).

What position does delapsion occupy in the whole of geologic phenomena? This is a fundamental question which every investigator of olisthostromes seems to struggle with, for up to now olisthostromes seemed to occupy an intermediate position between tectonic and sedimentary phenomena. The need for a clearly defined boundary between tectonization and sedimentation only became urgent after the introduction of the olisthostrome concept. I have considered this problem exhaustively and came to the conviction that delapsional processes form one inseparable unity which must be kept unimpaired. The concept of delapsion can be bounded rather precisely on genetic grounds.

There are three possibilities:
1. Delapsion may be incorporated in the concept of tectonization.
2. Delapsion may be incorporated in the concept of sedimentation.
3. Delapsion may be instituted as an independent category between tectonization and sedimentation.

Before making a well-founded choice between these possibilities we should try to define the terms 'tectonization' and 'sedimentation'.

Tectonization – It has become customary among geologists to use the term tectonization almost as a synonym of 'deformation of rocks in the earth' irrespective as to whether the processes involved are endogenic or exogenic; deformed rocks are called tectonites. Though the adjective 'tectonic' originally is a non-genetic term pertaining to the architecture of the earth (Naumann, 1849, as quoted by Dennis, 1967) the conception is so firmly associated with mountain building and large-scale structures that, for the sake of a more rational classification of geologic phenomena, the term 'tectonization' had better be used as a genetic term (cf. Scheibner, 1972) and restricted to *the deformation of the rocks of the earth by endogenic processes.*

The detachment of rock masses from their parent rock equivalents and their subsequent independent downward movement along sloping surfaces which intersect the surface of the lithosphere at their upper ends is excluded from the concept of tectonization, because after separation the process is no longer an endogenic one, only gravitation is the driving force. The phenomena of deformation by endogenic processes are designated as 'tectonic' and the material that is or has been tectonized as 'tecton'. Deformation by compaction or by glacial ice is something quite different, for it is neither endogenic nor delapsive. The adjective 'structural' is preferred to 'tectonic' as a term non-committal to any genesis.

The basic idea of the olisthon concept is that the rock masses incorporated in the olisthon are displaced on their own along a sloping surface only by the tangential pull of gravitation. In a great many cases of tectonic deformation and displacement gravitation plays a prominent part and may even be the main force, for instance, in the movement of a nappe along the downsloping distal end of its thrust-plane. If the initial movement was tectonic, the thrust-plane should be plunging backwards at its proximal end. As the ratio between the magnitude of the force of gravity and that of the total force responsible for the deformation or displacement cannot possibly be established, it would be impractical to place the boundary between delapsion and tectonization there where 50% of the total force would be gravitation. However, a rock mass is undoubtedly emplaced by delapsion, when it can be shown that it has been separated from its parent rock equivalent and has subsequently moved down on its own along a sloping surface which intersects the surface of the lithosphere at its upper end (this movement should necessarily be gravitative), or when it can be shown that the gap, whenever present, between the rock mass in consideration and its parent rock equivalent has been wholly or at least partly produced by the independent downslope movement of this rock mass subsequent to detachment and not only by erosion. This observable criterion is preferable to the 50%-of-the-total-force criterion and it should, therefore, be the only one used for the distinction between delapsion and tectonization. Often the last movement of a nappe or parts of it can be proved to be undoubtedly delapsive, for instance, when it can be shown that the main mass of the thrust-sheet has become separated from its root, or that frontal parts of it have

broken loose and have moved downslope for some distance on their own along the same basal plane.

It would, however, be unrealistic to designate every nappe or part of it of which the last movement can be imputed only to gravitation subsequent to separation as olisthon, because this last part of the movement may have been very small in proportion to the total displacement of the rock body away from its original site of deposition. In other words: should every rock mass of which only the last movement can be shown to be delapsive be called olisthon? It seems preferable, at first sight, when attempting to establish whether the displacement of a rock mass would be due to tectonization or delapsion, to evaluate the ratio between the distance of the displacement of this rock mass before separation from its parent rock equivalent and the distance of independent downslope displacement along the same basal plane after its separation. If the former distance is greater, then the rock mass would be designated as tecton. If the latter distance is greater the rock mass would be designated as olisthon. In practice, however, this will pose many serious and often insuperable problems, for instance when no root or no original position of the rock mass before its delapsional displacement is known, or when the root-portion or parent rock mass also has moved after the separation and subsequent delapsion of its frontal parts, or when the erosive contribution to the widening of the gap cannot be estimated.

So in these cases we are confronted with the unsatisfactory circumstance that the boundary between olisthon and tecton often cannot be established. Therefore, after all, it is more feasible to designate a rock mass as olisthon whenever the last movement of it, however small it may be, can be shown to be delapsive. A boundary based on this criterion is therefore used in this paper (see also page 42).

Sedimentation – The many different definitions of the genetic concept of sediment (and sedimentation) can be grouped into two categories, viz. those which do not include delapsion as an agent of transportation and those which do. In the first category, *sediments are considered as solid material that – after having been suspended or dissolved in air, water, or ice and after having been transported or moved from its site of origin by these media – has come to rest upon the surface of the solid earth under normal surface conditions* (compare Oldham, 1879; Bryan, 1922; American Geological Institute, 1960; Krumbein & Sloss, 1963); the processes leading to the formation of sediment are called sedimentation. Strictly speaking the material that has not come to rest is not yet a sediment, so that the clause referring to this was omitted from the above definition. In the second category, sediments are regarded as solid material deposited upon the surface of the solid earth; the agents of transportation and deposition may be any of those operating upon the earth's surface including deposition by delapsion (Twenhofel, 1932; de Raaf, 1968). Accordingly sedimentation becomes synonymous with 'transportation and deposition of solid material'.

In order to maintain the classical threefold division of the rocks of the earth's crust one needs to unite all non-igneous and non-metamorphic rocks under one name. The name 'sedimentary rocks' has been chosen for this third category and the genetic concept of sediment has been extended and adapted so as to cover these exogenetic rocks. But if olisthons are regarded as sediments, then, for example, the giant slipped-down sheets of the allochthonous cover of the Northern Appennines, the Tellian nappes in Algeria, and possibly the Median Prealpes in Switzerland should also be considered as sediments. This is in fact advocated by

Perrodon (1961), who is inclined to think that the downsliding and accumulation of "nappes de charriage" should be considered, at least partly, as a particular sedimentary phenomenon. But then the terms 'sediment' and 'sedimentary rock' are obviously inadequate to comprise all the rocks of the third category. The terms 'deposit' and 'depositional rock' would be better to comprise them. Unfortunately the term 'deposit' is also used in connection with magmatic and hydrothermal accumulations, but the context in which this term has been used has never given rise to any ambiguity. The problem whether olisthons are sediments or not is now shifted to whether the concept of sediment should be extended to comprise olisthons or not.

On the other hand some geologists tend to restrict the concept of sediment. Bayly (1969), for instance, is of the opinion that "a geologist who calls a coral reef "sedimentary" is extending the normal idea of sediments for sake of making his classification comprehensive and that it is in his own interest to be aware of what he is doing". But coral reefs can also be considered as constituted of solid material which has come to rest upon the solid earth's surface by withdrawal of calcium carbonate from the sea water by biochemical processes. In fact this applies to most fossils and it is a matter of opinion whether we consider them as ordinary constituents of sedimentary rocks or as objects that should be distinguished from sediments and treated separately.

Most investigators who are concerned with olisthostromes, however, tend to distinguish between so-called "normal sediments" and the products of "resedimentation" (Beneo, 1956; this terminology has subsequently been adopted by Facca, 1956; Reutter, 1965; Jacobacci, 1966; Ogniben, 1967; Broquet et al., 1967; Remane, 1970; Broquet, 1970) or "gravity displaced sediments" (Sanders, 1956). According to Beneo (1956) "normal sediments" would, in contrast with the products of "resedimentation", have been deposited particle by particle and would, if fossiliferous, contain contemporaneous fossils. This concept of "normal sediments" fits in the first definition of sediment given in this paper. Beneo (1956) and most investigators of olisthons restrict the term 'resedimentation' to denote the general process of subaqueous downslope gravity movement of rock material. Broquet (1970) argues that perhaps one should consider the process and say that sedimentation only concerns the elements transported by a moving fluid (water, wind) or by ice, whereas "resedimentation" would be due to the action of gravity.

I am at one with Broquet, but the term "resedimentation" means more than only 'sedimentation due to the action of gravity' (see page 17) and in this context it had better be replaced by 'delapsion'. The introduction of the term 'olisthon' made it however possible to restrict the concept of sediment to the first definition given in this publication. The term 'deposit' is therefore introduced to make the term 'sediment' available to denote exclusively the rocks that satisfy the first definition of sediment given in this paper (which are the normal sediments of Beneo). As a consequence sediments and olisthons are regarded as equally ranked subdivisions of depositional rocks. The latter term is used in this paper to denote the rocks satisfying the second definition of sediment which includes all kinds of olisthons. Depositional rocks (or deposits) are rocks that came to rest upon a depositional surface. 'Depositional rocks' can also be more suitably applied to bioherms than the term 'sediment'. Delapsion can now be considered as an independent category of processes between the processes of tectonization and sedimentation.

In order to establish the boundary between olisthons and sediments we have

to take into consideration the phenomena designated as 'avalanche of grains', 'traction carpet', 'grain flow', 'inertia flow', and 'flowing grain bed (or -layer)'. Tractional forces give rise to traction carpets (Dzulynski & Sanders, 1962). The tangential components of gravitation give rise to avalanches of grains (Bagnold, 1941). The terms 'grain flow' (Bagnold, 1941, 1954, 1956; Stauffer, 1967b), 'inertia flow' (Sanders, 1965) and 'flowing grain bed' (Bagnold, 1956; = flowing grain layer of Sanders, 1965) are used to designate flows of inert grains due to tractional or tangentially directed gravitational forces or both.

The term 'traction carpet' denotes the way by which the bed-load part of the solid-particle load of a stream is transported. The avalanching of grains is, however, strictly speaking a delapsional phenomenon and accordingly its ultimate result in the geologic record should be designated as olisthon. Of course all transitions exist between traction carpets and grain avalanches. We may say that when the angle of repose is exceeded, the movement is without doubt delapsional. For most aggregates of loose, non-cohesive grains, however, (except quick-sand) the angle of repose is about 30°. Hence, when the slope is about 30° or steeper, we may as well conclude to grain avalanching. But the avalanching of grains at the lee side of ripples and mega-ripples in water, and of eolian ripples, poses a problem. These ripples are formed only by the tractional force exerted by a mobile medium over loose granular deposits. The avalanching of grains at their lee sides may therefore be regarded as the continuation of the transport of solid particles over the bed in the direction of the flow and thus as an integral part of the traction carpet.

If the flow is generated only by the tangential pull of gravitation on the non-solid medium (water, air), the avalanching at the lee side of ripples and mega-ripples is excluded in this publication from the concept of delapsion, for the formation of ripples is a constructive process and the ultimately resulting cross-laminated or cross-bedded structures are confined to the built-up sedimentation unit. But if the avalanching of grains is diastratal and breaks down existing sedimentation units to form 'inclined bedding' such as in talus, reef flanks, slip slopes of point bars, and in the tributaries of submarine canyons, it is included in the concept of delapsion, and its ultimate deposit called olisthon.

But also if the flow is generated by the tangential pull of gravitation on the solids instead of on the non-solid medium, the whole ultimate deposit formed by these solids is, according to the definition, designated as olisthon irrespective as to whether cross-laminated structures confined to the sedimentation unit are present or not. This is the case in turbidites.

Turbidity currents – Turbidity currents are a particular variety of density currents, viz. density currents in which the excess (or effective) density that drives the current is due to suspended solid material. The unique feature of a turbidity current is that the solid material drives the current and not vice versa. The current moves downslope along the bottom, because the tangential components of gravitation on the suspended particles are sufficient to keep the grains in turbulent suspension and to cause the mass to flow as a denser fluid. A turbidity current must carry its load almost entirely in suspension. Overloading by decrease in velocity as a result of friction, or of reduction in slope, will cause a part of the suspended material to drop out. The grains of all available sizes fall out simultaneously, but the larger grains relatively more. If the suspended material is dropped the driving force diminishes, and the flow ceases when the suspended load gives out (Sanders, 1965;

Kuenen, 1967).

The ultimately settled-down solids are commonly designated as sediment. But they are not true sediments, because no tractional force exerted by a medium on the solids has played a part in the displacement of the current, but rather the reverse, viz. traction exerted by the solids on the medium in which they are suspended. But decisive is that, after detachment from the source area, these solids are also primarily propelled by the tangential pull of gravitation. Accordingly this gravitating of solids should be included in the continuous series of delapsion. The solids of the turbidity current are, it is true, suspended in water, but that alone does not make it a sediment. For, if the suspensive medium had been mud instead of water, as for instance in a mud flow in which a settling of larger inclusions to the base had taken place resulting in a conglomeratic basal part and a muddy upper part, there would be no hesitation to designate the mud together with its basal conglomerate as an olisthon. It is merely an accidental circumstance that only the solids settled down from the suspension of the turbidity current are preserved in the olisthon.

We will therefore consider as olisthon also those solids settled down to the base of a delapsing mass composed of a volatile, fluid, or visco-plastic mobile medium surrounding these solids and in which the gravitational force is acting only on the solids. Hence turbidity currents are here considered as a particular kind of delapsion and turbidites as olisthons instead of sediments.

'Resedimentation', exotic and allochthonous – The term 'resedimentation' (Migliorini, 1950) has often been used in connection with olisthons. If used, it should be applied only to sedimentary rocks, for etymologically it means 'renewed sedimentation of elements that already have been sedimented previously'. According to ten Haaf's remarks to Kuenen (conveyed to Sanders via a letter from Kuenen dated december 21, 1964), Migliorini used the term only for deposits of turbidity currents (Sanders, 1965). As turbidites are not considered as sediments but as olisthons the term 'redeposition' would be preferable to 'resedimentation'. Kuenen (reply: in Sanders, 1965) states rightly that almost "all mechanical sediment particles . . . have been deposited many times before they came to the position where we find them. They may have rested in sandbars, levees, dunes, flood plains, terraces or beaches and may have been part of a consolidated deposit or not, before they set in motion again".

The term 'redeposition', therefore, remains an ambiguous one, because not all redeposited materials can be recognized as such. In this paper redeposition is defined as the deposition of organogenic and (or) polygranular inorganogenic elements that show by their lithology or fossil content that they are derived from rocks deposited in other time-space relationships than the deposit in which they are found; but also deposits that can be shown or reasonably be considered to be olisthons could be referred to as redeposited.

Closely connected with redeposition is the concept of exotics, a term frequently used in relation with olisthons but never exactly defined; it means 'foreign constituents'. Should they however be foreign with respect to the rock in which they are found or to the geologist by whom they are found? In accordance with the latter sense, for example, the adjective 'exotic' has been restricted only to rock elements of unknown origin (Brunschweiler, 1966; Feuillée, 1967). This is not preferable, because it introduces a subjective element in the concept which should be avoided. Moreover, many elements thus defined may be found to be "non-

exotic" after a detailed survey of the region. Thus exotics should be foreign with respect to the other constituents of the rock in which they are found (Murawski, 1972). Jacobacci (1966) is of the opinion that the adjective 'exotic' should be restricted to those blocks of uncertain origin that are foreign to the enclosing deposit, but which are not enclosed by olisthostromes, in order to distinguish them from olistholiths. This is, however, in disaccord with the current use of this adjective with respect to blocks enclosed in the olisthostromes of the alpine wildflysch.

Exotics are here defined as those allogenic constituents of which the original rock-type can be identified and that came into existence in other time-space relationships than the time-space relationships of the rock of which they now constitute a part. Igneous rocks may form exotics but are generally not redeposited.

The adjective 'allochthonous' (meaning from another region) is almost synonymous with 'exotic', but it stresses in particular the distance travelled by the elements, whereas 'exotic' stresses the difference in nature. 'Allochthonous' refers to organogenic or inorganogenic elements that can be shown to be derived from an area outside the area of accumulation, lithotope, or paleobiotope in which they are found (cf. Schieferdecker, 1959; American Geological Institute, 1960). 'Allochthonous' cannot be used as a synonym of 'exotic', for exotic may also be derived from rock units situated above or below the rock unit in which they are found. Also 'exogenic' cannot be used as a synonym of exotic, as Bissell & Chilingar (1967) do, for this adjective refers to geological processes originating at or near the surface of the earth, and to rocks, ore deposits, and landforms which owe their origin to such processes (Holmes, 1928; Schieferdecker, 1959; American Geological Institute, 1960).

Olisthons and stratigraphy – The delapsion of relatively *competent* rock masses generally takes place by slipping along discrete shear planes with minor internal deformation, so that their original shape remains relatively intact. This process would not result in strata (though these rock masses may be internally stratified) and is therefore called 'non-stratifacient delapsion' in contrast to 'stratifacient delapsion'. The latter term would refer to the delapsion of relatively *incompetent* rock masses involving largely internal deformation by relative movements of the constituent solid elements with respect to each other; these rock masses may have suffered thorough mixing and homogenization and would spread in a sheet-like manner forming strata.

The conventional conception that all stratified rocks (with the exception of stratified metamorphites or igneous rocks) should also be sedimentary rocks is the main reason why most olisthons forming strata fall within the group of subjects investigated by sedimentologists and stratigraphers, whereas most olisthons that do not form strata are or were previously the subject of study of structural geologists. This conception is likely to be the reason why most investigators endeavour a bipartition of the continuous series of delapsional phenomena and for the delay in the general recognition of delapsion as a third and independent category of processes between sedimentation and tectonization. But as 'delapsionists' do not exist nor will exist in the near future (if they will ever exist), both extremes of the series will continue to be studied by respectively sedimentologists/stratigraphers or structural geologists. The broad middle portion of the series, such as olisthostromes and closely related phenomena, will be studied by both by both categories of investigators.

Olisthons, irrespective as to whether they are stratiform or not, are com-

posed of the slid-down parts of one or more lithostratigraphic units and are regarded here as depositional rock bodies which occur in normal depositional sequences. The contacts of olisthons with the rocks surrounding them are not tectonic, but depositional. Since sedimentary deposits fall within the scope of stratigraphy, evidently also all "resedimented" deposits should fall within it. Consequently also olisthons should be classified as kinds of lithostratigraphic units.

As statigraphy means descriptive science of rock strata (International Subcommission on Stratigraphic Classification, 1972), no objections are expected in respect of stratiform olisthons in which the original lithologic units are mixed thoroughly and homogenized into a new rock-type, for instance olisthostromes and turbidites. The exotic clasts included in them are usually regarded as parts of the lithostratigraphic unit in which they are found.

But objections may be raised against the view that also non-stratiform olisthons, such as the huge exotic rock masses enclosed in the continuous flysch successions in Algeria and Sicily, the 'Tellian Nappes' and the 'Panormide Nappes' respectively, are lithostratigraphic units. Yet they are no more than large exotic blocks composed of one or more lithologic units and slipped down to become redeposited in a basin during the deposition of the flysch, so that they also fall within the scope of stratigraphy. The contacts of these blocks with the surrounding lithostratigraphic units are not tectonic, but certainly are structural. Broquet et al. (1967), aware of the intermediate position of the Sicilian examples between the conventional conceptions of tectonic and sedimentary phenomena, called them after Lamare (1947) 'klippes sédimentaires'. Because of the large size of these masses the rocks bordering the slip-planes may be crushed, brecciated and slickensided. Even large dislodged slices may be encountered along these slip-planes. These features need not necessarily be tectonic, but may then be merely a product of strong shear stresses generated during the delapsion.

Each of these slipped-down rock masses is redeposited in another time-space relationship than the parent rock units and therefore constitutes or forms part of one or more other lithostratigraphic units. However, a part of an Eocene limestone unit, for instance, redeposited in Miocene marlstones has not ceased to be an Eocene limestone, though forming part of the Miocene marlstone unit. This limestone body has, like all redeposited rock slabs, the dualistic nature that it is an Eocene limestone and that it simultaneously belongs to a Miocene marlstone unit of which it cannot possibly be a Member. In fact, no lithostratigraphic unit represented by a redeposited rock slab can possibly be given a rank with respect to the units of the receiving rock sequence. Moreover the boundaries of a large slip mass may cross the boundaries of several distinct lithostratigraphic units surrounding it, so that it may remain controversial which of the latter is the host unit. Therefore large redeposited rock slabs had better be separated from the host units, especially if their redeposited nature is not accepted or controversial; in that case they can be considered as structural units (which are not necessarily of tectonic nature, but may also be of delapsional origin).

It is a current mapping practice that structural units are outlined by heavy lines or red lines, possibly in combination with saw-teeth symbols, to express the structural nature of the contacts; they are given the colour(s) of the parent lithostratigraphic unit(s). However, the mappable blocks of a land-slide are not heavily outlined; nevertheless they are given the colour of the parent rock unit, but in addition they are marked with a V or U-symbol to express their land-slide origin.

The contacts between the lithostratigraphic units of a depositional sequence

(which may be of sedimentary as well as delapsional nature) are not thickened or reddened to express the non-structural nature of the contact. Possible redeposited rock slabs are usually regarded as being part of the lithostratigraphic unit in which they are found. No difficulties will arise when these inclusions are small, but how should they be indicated when they are of mappable size? Since the rock of an Eocene limestone mass does not convert into a Miocene rock in which it is redeposited, it should, if mapped, be given the same colour as the parent rock unit, the more so as a lithostratigraphic unit present in a redeposited rock slab cannot be given any rank with respect to the receiving lithostratigraphic unit to which it belongs, so that it had better be separated from the host unit. Some redeposited rock slabs may, according to the ideas of the investigator, be interpreted as structural units and in that case furnished with the corresponding symbols. But if they are not interpreted as such, but still are mapped, it is proposed here that they should, in analogy with the mappable blocks of land-slides, be marked with the symbols 'E', 'O', 'I', or 'R' to express their being exotics, olistholiths, inclusions or redeposited rock slabs respectively in order to avoid any possible confusion with outcrops of the parent unit.

Serious objections are however to be expected when the Ultrahelvetic complex in the Swiss Alps is regarded as the uppermost lithostratigraphic unit of the Helvetic sequence, or when the allochthonous complex in the Northern Appennines is considered as the uppermost lithostratigraphic unit of the Tuscan-Emilian sequence. But since all investigators agree about the delapsional emplacement of these complexes, they should, according to the ideas expressed in this paper, be regarded as depositional units which fall within the scope of stratigraphy.

With this a new and unexpected dimension is added to conventional stratigraphy, for all redeposited rocks should be incorporated in the concept. Italian geologists, who in particular are familiar with problems concerning large allochthonous influxes in geologic sequences, have already reached this stage in the development of geologic thinking. They use the terms 'autochthonous', 'meso-autochthonous' and 'neo-autochthonous', terms introduced by Marchetti (1957), to denote respectively the rock-stratigraphic units below, between and above the rock-stratigraphic units in which large allochthonous influxes are predominant. This terminology has been used, for instance, in connection with the olisthostromes in Sicily and southern Italy. Many rock-stratigraphic units previously designated as post-orogenic may turn out to be neo-autochthonous. These autochthonous, meso-, and neo-autochthonous units may however actually be composed of other types of olisthons, turbidites for instance, and may in many cases contain as many allochthonous constituents as the olisthostromes. Hence the use of the terms autochthonous, meso-, and neo-autochthonous for the non-olisthostromes in contrast to the allochthonous olisthostromes in-between may evoke a false suggestion.

Mechanics of delapsion

MECHANICAL PRINCIPLES

The cause of gravitational sliding is always that the cohesion between the different constituents of the mass of rock or the friction between the sliding rock mass and

the substratum do not suffice to resist gravitation (Hadding, 1931). But the reasons why cohesion and friction cannot resist the force of gravity are extremely complex. Many factors are involved which are still unknown or poorly understood. If we want to obtain a well-founded picture of the various types of delapsion we must enter into the mechanical principles involved in downslope mass-movements.

Shear strength – The mechanical principles can be well demonstrated by examining the shear strength characteristics of an over-consolidated, non-cemented, cohesive rock, clay for instance, as shown in figure 1a (Skempton, 1964). This figure typifies the results obtained by carrying out slow drained tests (= without the development of excess pore pressures) in a shear box apparatus, in which the clay is subjected to displacements amounting to several inches.

When the clay is strained under a given effective pressure, the shear strength passes a peak value, τ. If the displacement is increased, the strength of the clay decreases until a certain "residual strength", τ_r, is reached which the clay maintains even when subjected to large displacements. If the test is repeated several times under different effective pressures, the results would be of the same type as those already described. The peak and residual strengths, when plotted against effective pressure (figure 1b), would show a relationship which approximately obeys the equation

$$\tau = \tau_o + \sigma \tan \varphi \qquad (1)$$

for the peak strengths in which τ_o represents the initial shear strength of the material, that is the shear strength when the normal effective stress is zero. It is often called cohesion intercept or, briefly, cohesion. σ Represents the effective stress normal to the slip surface. φ Represents the effective angle of shearing resistance. This equation is known as Coulomb's equation. The curve which is the locus of the points representing the peak values for every magnitude of σ is called the 'strength envelope'.

The residual strengths can be expressed by the equation $\tau_r = \tau_{or} + \sigma\tan\varphi_r$ (figure 1b). According to Skempton (1964) the test-results show almost invariably that τ_{or} is very small, and probably not significantly different from zero. Hence for the residual strength we may write $\tau_r = \sigma\tan\varphi_r$. Thus, if for any reason the clay is forced to pass the peak at some particular point within its mass, the strength at that point will decrease and the initial shear strength will become negligible. This action will throw additional stress on to the clay at some neighbouring point, causing the peak to be passed also at that point. In this way a progressive failure can be initiated and, in the limit, the strength along the entire length of the slip

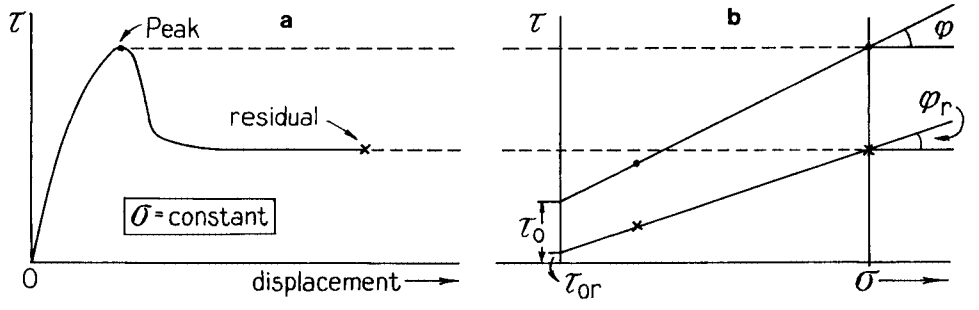


Fig. 1. Shear-strain curve (a), and relationship between shear strength and effective stress (b), after Skempton (1964).

face will fall to the residual value. Simultaneously τ_{or} along the slip surface will become negligible.

When applying these shear strength data to explain the mechanics of gravitational sliding of rock masses, it is obvious that the resistance to shear, τ, of equation (1) on a potential sliding plane must be overcome by the total shear stress along that plane before sliding can occur. This can be illustrated in figure 2. The vertical stress on a potential sliding plane in a given rock is represented by OS_1; it is at the slope angle ϑ to the normal stress OA, and has the corresponding shear stress component AS_1. Also the strength envelope T_oS of a given rock is plotted. The shear strength of the rock is represented by AP. It can be seen that with a vertical stress of less than OS the shear stress component will be less than the shear strength and the slope will be stable. No accumulation beyond that causing a vertical pressure OS is possible, because at this point the shear stress BS equals the shear strength. It is apparent from this diagram that, if vertical homogenity is assumed and the effective angle of shearing resistance, φ, is greater than the slope, ϑ, the two lines will never intersect and infinite sediment accumulation is theoretically possible.

From laboratory experiments in the triaxial compression test apparatus it appears that the representative mean value of the magnitude of the angle φ would be about 30° (Hubbert & Rubey, 1959; Handin & Hager, 1957; Handin et al., 1963). From field data it appears, however, that the slope angle ϑ of the slip-plane in many cases of sliding has been only a few degrees. Hence in these cases we are confronted with the logical necessity of somehow reducing the frictional resistance to sliding, τ, to a small fraction of that which we have assumed.

In the following 3 paragraphs we will discuss various possibilities by which the resistance to sliding can be small.

Abnormal pore-fluid pressure and metastable fabric – The total normal stress S on a horizontal plane in a water-saturated rock will be the sum of the pressure produced by the water and that produced by the solids of the rock above that plane. In dry soil the stress S produces a change in all of the other mechanical properties of the soil such as its shearing resistance. For this reason it is known as an effective pressure (σ). In a saturated soil the increase in pressure due to the weight of the water does not have a measurable influence on any mechanical property of the soil. Therefore the pressure produced by the water load is called a neutral pressure (p) (Terzaghi & Peck, 1948). We may say that the effective pressure in a rock is the pressure transmitted by the points of contact between the solid grains. Thus $S = \sigma + p$.

A change in the pore-fluid pressure, p, in a rock of high permeability has no

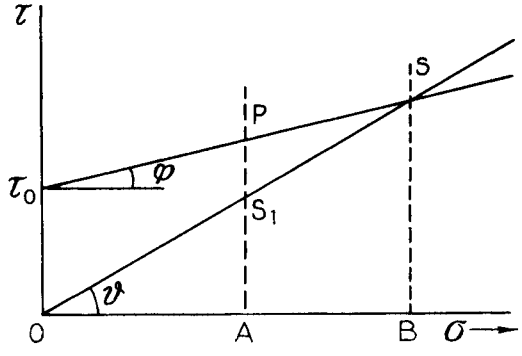


Fig. 2. Diagram showing the relationship between slope angle and the effective angle of shearing resistance, after Moore (1961).

direct influence on the magnitude of the effective stress needed for failure of the rock. This is shown by the unjacketed compression tests done by McHenry (1948)*), Bishop & Eldin (1950-1951)*), Bishop & Henkel (1957)*), and Handin et al. (1963).

From the jacketed confining triaxial compression tests carried out on sandstone by Handin et al. (1963), however, it can be concluded that failure occurs at ever lower incremental stresses with increasing pore-fluid pressure. In these tests the pore-fluid pressure counteracts the confining pressure owing to the impermeable jacket. In nature this jacketed condition can be only realized, if a stressed porous rock is enveloped in a rock of very low permeability, or if the stressed rock itself is virtually impermeable such as clay.

The effect of low permeability on the compression of a rock body under load can be demonstrated experimentally by the undrained triaxial tests (= without permitting any changes in the water content during the tests) on water-saturated clays by Bishop & Eldin (1950-1951)*). It appears that when the total stress is increased, the pore pressure increases by the same amount that counterbalances the increment of the total stress. Thus $S = p$ and $\sigma = 0$, hence $\tau = \tau_0$. So "if a readily compressible rock of large porosity and low permeability, initially filled with water at normal hydrostatic pressure, were to have a mechanical loading suddenly imposed upon it, the water, because of its slowness of leakage would initially assume a corresponding increase of pressure almost equal to the entire increment of load" (Hubbert & Rubey, 1959).

As $\sigma = S-p$, the Coulomb's equation can also be written $\tau = \tau_0 + (S-p)\tan\varphi$. However, once fracture is started τ_0 can be eliminated and the magnitude of φ reduced to the residual magnitude φ_r. Then further slippage results when $\tau_r = (S-p)\tan\varphi_r$, which indicates that the resistance to shear along the slip-plane (down which the block will be able to slide further) can theoretically be made to approach 0 as the fluid pressure p approaches the total normal stress S at that plane.

Unless the rocks are ideally impermeable the pore fluid must be flowing away from any region of greater than 'normal' pore-fluid pressures and must continue to do so until the excess pressure is dissipated. Because of this continuous leakage, abnormal pore-fluid pressures are thus "transient" phenomena and require some mechanical activity to bring them into existence and to maintain them, for instance loading or lateral compression. Mechanical loading of a porous and saturated body of rock of low permeability can be produced, for instance, by so fast an accumulation of sediment on top of the rock body that the process of consolidation by reduction of pore water can not keep pace with it and that an excess pore-water pressure within the burried rock body is set up. Also a slab of rock falling or sliding upon a porous and saturated rock body of low permeability will cause anomalous pore-fluid pressures within that rock body.

An advantageous condition for failure is created by the metastable fabric of some clayey rocks.

According to Terzaghi & Peck (1948) "most clays consist of a mixture of more or less equidimensional particles that constitute the coarsest fraction, and flaky (or exceptionally needle-like) particles that make up the balance. If a mixture of such grains is deposited by sedimentation, the uppermost layer of sediment is composed of a honeycombed extremely compressible aggregate of clay particles. This layer contains the coarser grains in a random distribution. If such a sediment is subsequently compressed, the coarse equidimensional grains come in contact with each other like the particles of a sand in the process of sedimentation. Their arrangement

*) as cited by Hubbert & Rubey (1959).

is inevitably very unstable, because the clay which occupies the interstices prevents the grains from tipping and rolling into stable positions with reference to each other. After they touch each other, they constitute a more or less self-supporting skeleton, embedded in a clay matrix. As long as the sediment is laterally confined, the skeleton is able to carry all or at least the major part of the overburden pressure, as does a very loose sand or even a true quicksand. However a slight disturbance, such as a gentle deformation, destroys the equilibrium of the grains composing the skeleton. It collapses and at that instant all the stresses previously carried by the skeleton are transferred onto the matrix." This causes a sudden excess pore fluid pressure, and in tilted position the collapsed system will slide down rapidly (a so-called collapse slide).

As a matter of fact, soft clay strata with such properties are by no means uncommon. Some of them possess abnormal laboratory consolidation characteristics involving excessive compressibility under the first increments of load greater than the overburden pressure. The high initial compressibility is supposed to be due to a high degree of instability of the load-carrying skeleton. Such clays are designated as "extra-sensitive" clays. The term 'sensitivity' indicates the effect of remolding on the consistence of a clay, regardless of the physical nature of the causes of the change. The degree of sensitivity is different for different clays and it may also be different for the same clay at different water contents.

Angle of shearing resistance and anisotropic fabric – The angle φ is an experimentally determined constant for a given material. The magnitude of angle φ of cemented sand or of moist sand is approximately equal to that of the same sand, at the same packing, in cohesionless state (Terzaghi & Peck, 1948). In many rocks it also appears to be approximately equal at low or at higher confining stresses (at very high stresses φ may become smaller, presumably because of a change in the nature of the grain contacts). However, angle φ is greater for dry sand composed of angular grains with a gradation in grain diameter than for dry sand composed of rounded grains with approximately the same diameter (Terzaghi & Peck, 1948). This implies that the magnitude of angle φ is primarily uninfluenced by the circumstance whether the grain contacts are cohesive or not, or whether the stresses transferred through these contacts are low or high, but that it rather depends on the number of grain contacts.

The shear-strain curves of dry, dense sand show peak values for τ at a certain strain (figure 1). This must be due only to variations of the apparent value of tan φ. These peak values can be explained as follows: during the first increments of strain in a horizontal direction the grains shift in position in the zone of the potential sliding plane, but in such a way that it tends to resist further displacement. Originally effective contacts (which are contacts that prevent the grains to yield by shifting to the force exerted upon them) are partially replaced by other effective contacts; originally ineffective contacts become effective and the reverse. The result will generally be an increment of the effectiveness of the grain contacts and in many cases also of the total number of the effective grain contacts. This shifting in position proceeds with growing strain till the grains are driven out of their stable position (decreasing number of effective contacts) and failure takes place at a lower (= residual) value of tan φ than that reached at the peak value of τ. This implies that the value of tan φ also depends on the effectiveness of the grain contacts.
Simultaneously with the reaching of the peak value, however, the void ratio (which is the ratio of the total volume of voids to the volume of the solid substance: $\frac{V_v}{(V-V_v)}$) increases as has been shown experimentally (Terzaghi & Peck, 1948). On

the other hand, for a given sand the magnitude of angle φ increases with increasing relative density ($Dr = \frac{N_o - N}{N_o - N_{min}}$ in which N_o represents the void ratio of the sediment in its loosest state, N_{min} the void ratio in the densest state obtained in the laboratory, and N the void ratio of the sediment in the field), thus in general with increasing number of grain contacts, but with decreasing void ratio. This implies that in granular rocks tan φ is also primarily independent of the void ratio. Generally, however, the number of effective contacts will be smaller at high void ratios; but this need not be so, for example, if the grains are flaky.

So we may say that the shearing resistance in rocks is effective through the points of contact between the solid grains, and that, if the stresses are not too high, it appears to be dependent only on the number of effective grain contacts, on their effectiveness, and on the nature of these contacts (for instance the hardness of the various grains).

An example of the importance of the effectiveness of grain contacts can be found in clays. If a rock in which the potential sliding plane is situated consists of clay, a shear stress tends to orient the platy or elongate particles parallel to the plane of movement (Gillott, 1968). "There is decisive evidence for the presence of continuous bands of almost perfectly oriented particles in clays subjected to large strains . . . It is probable that the formation of orientation domains begins at relatively small strains" (Goldstein et al., 1961, as quoted by Skempton, 1964). The relative resistance to such reorientation varies with the fabric of the clay, the strength of the interparticle bonds, and the orientation of the shear plane. If the clay minerals have a random flocculated arrangement considerable energy will be required to reorient the particles regardless of the orientation of the shear plane. But when the clay structure yields, the particles become oriented in the direction of movement, so that the shear strength in that direction will be considerably smaller. This must mainly be due to the grain contacts being considerably less effective in the direction parallel to the flakes in comparision to those perpendicular to the flakes (anisotropic rock fabric). Shearing in soils with a clay content greater than about 35% induces preferred orientation of the clay particles in the shear zone and causes a reduction of φ to φ_r. In some clays the reduction is only 1° or 2°, but in others as much as 10° (Skempton, 1964).

Also in compacted clays the particles are approximately parallel in position (O'Brien, 1970; Heling, 1970), so that the shear strength in a direction parallel to the reoriented clay particles is smaller than that in any other direction. If the overburden above the compacted clay body is wholly or partly removed, by erosion for instance, some clays (especially those with a high amount of montmorillonite or other clay mineral with considerable swelling property) will swell osmotically by sucking up water, while the particles keep their parallel position. Some cross-links may be formed during the swelling process, but only in a relatively small number (van Olphen, 1963). It will be clear that in that case the value of tan φ parallel to the particles, especially when they are wrapped in a film of fresh water, will be considerably smaller than that perpendicular to the particles.

Colloidal behaviour and Bingham plastic flow – In dry cohesionless sand $\tau_0 = 0$, but in non-cemented sediments which are entirely or for an important part composed of particles showing colloidal behaviour (hydrophobic colloids), clays for instance, τ_0 has a certain value and is often briefly called cohesion. The electrical charge carried by the clay particles is compensated by the accumulation of an equi-

valent amount of ions of opposite sign (= counter ions) in the liquid immediately surrounding the particle, keeping the whole assembly electroneutral. The concentration of the counter ions decreases with increasing distance from the particle surface. Simultaneously there is a deficiency of ions of the same sign in the neighbourhood of the surface, since these ions are electrostatically repelled by the particle. Therefore, when two particles approach each other in a suspension owing to their Brownian motion, their diffuse counter-ion "atmospheres" begin to interfere, resulting in a repulsion between the particles. The general Van der Waals attractive forces, however, counteract the repulsive forces. At increasing electrolyte concentrations in the liquid surrounding the particles the diffuse counter-ion "atmospheres" are compressed by which the range of the repulsion is considerably reduced. By summation of the repulsive and attractive forces, the net interaction forces between two particles as a function of particle separation can be calculated for increasing electrolyte concentrations and represented by the curves of fig. 3.

At high electrolyte concentration the potential curve shows a deep minimum at close separation, and attraction predominates at any particle distance. If two particles approach each other, they will agglomerate to form "flocs" when they reach the position at which the deep attraction minimum occurs. The system is flocculated. At medium and low electrolyte concentrations repulsion dominates part of the way during mutual approach of the particles. Only if the collisions due to Brownian motion are strong enough can this repulsive energy barrier be "jumped over". The rate of agglomeration will decrease.

Clay flakes appear to have a negative charge along the flat surfaces, while the edges of the flakes may well carry a positive charge. When a suspension of platelike clay particles flocculates, three different modes of particle association may occur: face-to-face, edge-to-face, and edge-to-edge. Face-to-face association merely leads to thicker and probably larger composite flakes, whereas edge-to-face and edge-to-edge associations will lead to three dimensional voluminous linked cardhouse structures. The viscosity of dilute clay suspensions increases when agglomerates are formed by edge-to-edge and edge-to-face associations; the viscosity decreases when the particles become thicker and larger by face-to-face association. The rheological properties of the three types of association in the flocculated suspensions will therefore be quite different.

The rheological behaviour of a material under flowing conditions differs from that under rigid conditions in that shear does not take place along specific shear planes, but intergranularly (or intermolecularly), i.e. that every particle within the material moves relative to its neighbours and the shear is general. Under flowing conditions the coefficient of viscosity or briefly viscosity (η) is used instead of tan φ and the rate of shear (cm.sec.$^{-1}$. cm $^{-1}$, dU/dy) instead of σ. The shear stress is

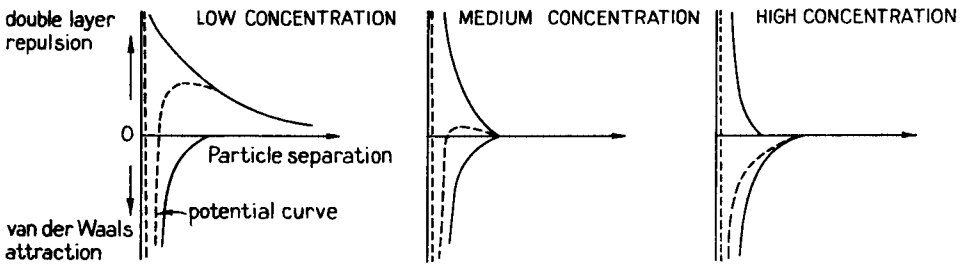


Fig. 3. The potential curves for low, medium, and high electrolyte concentrations, after van Olphen (1963).

represented by τ. Many kinds of flowing can be distinguished, of which the most important ones are: Newtonian flowing, dilatant flowing, and Bingham plastic flowing. We will confine ourselves to the latter.

If sediment particles are of colloidal size, the addition of an electrolyte, sea water for instance, converts the system into a Bingham plastic system, which "has a property of permanently supporting a shearing stress less than a certain critical yield value with only slight total deformation, but for greater shearing stresses plastic flow takes place at a continuous uniform rate as for a highly viscous fluid" (Bingham, 1922, as quoted by Strahler, 1952). Bingham plastic flowing obeys the equation $\tau = \tau_0 + \eta \, dU/d_y$ in which τ_0 represents the yield stress.

The magnitude of the yield stress in a Bingham plastic system appears to be dependent upon:

1. The amount of electrolyte in the system.
2. The system being stirred or in rest.
3. The void ratio of the system.
4. The amount of non-colloidal particles in the system.

Ad. 1. In concentrated suspensions in which an electrolyte is added the colloidal particles may link to build a rigid network developing a certain yield stress, of which the magnitude is related to the force required to break a link between two particles and to the number of such links per unit volume. If face-to-face association occurs simultaneously, for instance at very high salt concentrations, the number of units building the card-house structure is reduced and the Bingham yield stress decreases (van Olphen, 1963). It should be noted that the difference in yield stress of the same clay in sea water and fresh water is small compared to the difference in yield stress between various types of clay minerals (Kuenen, 1951). Upon peptization the particle links are weakened and finally broken as soon as repulsion between the particles predominates. Consequently the yield stress reduces and finally disappears. The originally stiff suspension obtains the character of a rather freely flowing liquid.

Ad. 2. Clay suspensions which are more or less flocculated show thixotropic behaviour. Thixotropy (Greek $\vartheta\iota\gamma\gamma\alpha\nu\omega$ = I touch, and $\tau\rho o\pi\eta$ = change) is a property of colloidal materials, being the reversible gel-sol-gel transformation, induced by shaking and subsequent rest. The system becomes more fluid when stirred, but slowly recovers its original consistence when left at rest. When the sol state is not fully achieved, the phenomenon is called false-body thixotropy (Pryce-Jones, 1936; Boswell, 1961). The transition is then from gel to very-fluid-gel to gel. False-body-thixotropic systems, therefore, exhibit a very rapid rise in consistence after shaking has ceased (Röder, 1939). Benthonite plus water is an example of a system which shows false-body thixotropy. The changes in consistence appear to be due mainly to changes in the yield stress. The reduction of the yield stress in a stirred system indicates that shear causes a breakdown of the particle links in the skeleton structure of the flocculated system (van Olphen, 1963). The restoration of the original yield stress indicates that during a period of rest the links are re-established when Brownian motion brings the particles together. When the salt concentration increases, the size of the residual energy barrier in the particle interaction potential curve is reduced and the rate of particle linking and stiffening of the suspension increases (van Olphen, 1963).

In nature, shaking of a thixotropic clayey rock can be brought about by an

earthquake which would momentarily lower the value of τ_0. If the sol state or the state of a very fluid gel (in false-body systems) is reached, a slope angle slightly greater than 0° would be sufficient to initiate flowing. Stirring of such a clayey rock may also occur when a rock mass is falling or sliding down upon it, for instance from a cliff. The general breakdown of the particle links in the clay below the site of impact may strongly reduce the value of τ_0. The propagating shock wave incited by the falling block might be able to spread this reduction over a large area. Simultaneously an excess pore fluid pressure may be generated below the site of impact. This pressure may assume a value almost equal to that of the weight per unit area of the block, so that its falling hardly causes any incremental effective stress in the substratum. If the block is large it might be able to fluidize a large part of the substratum, which is set in motion. The block may then become involved in the downward flowing of the clayey substratum in which it could even serve as an additional impetus.

Ad 3. The water content of a granular material can increase to a critical value beyond which the material can no longer behave as a solid, but as a fluid (Terzaghi & Peck, 1948; Dott, 1963); the yield stress of the material reduces to a negligible value. In non-colloidal material this critical value is usually called the liquid limit, which is determined as the water content (grammes of liquid per 100 grammes of solid) in percent of the dry weight at which two sections of a pat of soil, having the dimensions shown in figure 4, barely touch each other, but do not flow together when subjected in a cup to the impact of sharp blows from below (Terzaghi & Peck, 1948). In colloidal materials the critical value is called the thixotropic limit beyond which the colloid behaves like a sol. In the laboratory the thixotropic limit is determined by the ratio of the volume of liquid to the volume of solid (thus the void ratio) in a mixture which would just not flow when the containing tube was turned upside down after one minute's rest, but which flows at once if a slight tap is given to the tube (Boswell, 1949). The thixotropic limit varies when tubes of different sizes are used. These values are therefore entirely arbitrary and only suitable to evaluate the relative degree of thixotropy of a certain material.

High void ratios can be achieved by rapid deposition. The grains and (or) flocs do not have the opportunity to reach their stablest positions. Such rapid deposition can be found in marine environments off large rivers carrying predominantly fine-grained silt and clay sized materials (Terzaghi, 1956; Moore, 1962). But also strong erosion of coasts will cause rapid deposition. The peptizing effect of fresh river water may in certain circumstances be able to lower τ_0 to a still lower value. Moreover, rapid deposition may generate an abnormal pore fluid pressure which reduces the effective normal stress, σ. A high void ratio also reduces the

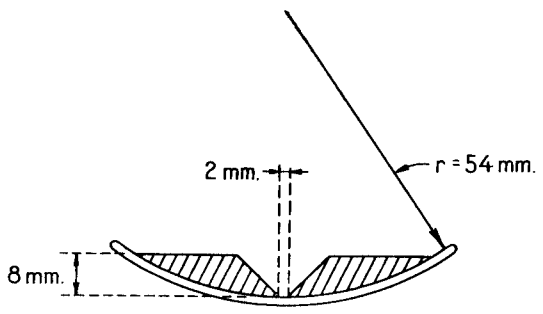


Fig. 4. The dimensions of the cup used to determine the liquid limit, after Terzaghi & Peck (1948).

submerged weight of the sediment and generally the magnitude of the angle φ. All these circumstances may lead to a low magnitude of the slope angle necessary for the deposit to slide down.

The void ratio of an uncemented sediment is directly proportional to the density of that sediment. In this respect Kuenen (1951) has published some results carried out for him by the Royal Shell Laboratory of Amsterdam on the viscosity and Bingham yield stress of clay suspensions of different densities. Dangeard et al. (1965) have extended these investigations to much higher densities with Aiguillon clay. They have shown that with the ascending density and consequently with the ascending viscosity of the clay suspension the delapsional process was respectively turbidity current action, mud flowing, and slumping. Two thresholds were passed with increasing density, viz. one at the boundary between turbidity current action and mud flowing, and one at the boundary between mud flowing and slumping. The first threshold may correspond to a more natural thixotropic limit for the Aiguillon clay. It occurs at the concentration of about 250 g/l (for industrial kaolin this threshold appears to be situated at a concentration of about 375 g/l.) The second threshold probably corresponds to the yield limit of the Bingham plastic system.

Ad 4. If clay is mixed with a certain amount of sand-sized particles, the thixotropic limit is lowered, as Boswell (1949) has demonstrated. This is probably due to the fact that sand-sized particles, which do not show colloidal behaviour, prevent a large number of possible links between colloidal particles to be formed. Moreover, they would be able to break these delicate links by their relatively large momentum when the mixture is in movement. The rate of lowering of the thixotropic limit or yield stress depends on the number of prevented links. It seems likely that this depends on the one hand on the total surface area of the non-colloidal particles, and consequently on their relative number, shape, and size; on the other it is likely that, when the sediment mixture is in motion, the particle links in the clay between the non-colloidal grains are broken by the relatively large momentum of the latter, depending on their velocity and mass, and consequently on their volume and specific weight. However, as the surface area increases with the second power of the diameter and the volume with the third power, the effect of the total momentum of a certain volume of large non-colloidal particles, beyond a certain upper boundary-size, exceeds the effect of the larger total particle surface of the same volume of smaller non-colloidal particles. It is therefore clear that the maximum effect is achieved at a certain combination of clay, sand and gravel.

This reasoning is confirmed by some experiments made by Kuenen (1951) who has observed that "when releasing a viscous clay suspension with or without additional sand the mass slides down the slope some distance but soon comes to rest. The movement in many ways recalls that of a glacier although much swifter and numerous cracks form in the surface. A slight dilution of such a paste, *especially when containing much sand,* results in a combination of a highly turbulent turbidity current with a thin wedge of more viscous consistency following close behind the head of the current . . . Again, a slight addition of water or *reduction of the relative amount of clay* is sufficient to cause the entire suspension to flow turbulently without development of the bottom wedge . . . Further dilution reduces the velocity without altering the nature of the flow, but the competency to carry sand is reduced . . . Although turbulent flow in pure clay suspensions is not possible at higher values of density than about 1,5, sand suspensions with densities of 1,8 can form turbidity currents and with a suitable combination of clay, sand and gravel the highest limit is about 2,0" (my italics).

At a certain point in the addition of sand to a clay suspension the mobility sud-

denly becomes too small for turbulent flowing in a turbidity current. Beyond that point the only way for downslope movement is by slumping or avalanching.

Toe, tail, progressive and retrogressive slumping – In several cases of slipping down of rock sheets the sheets are unobstructed at their downslope ends, for instance when their fronts are determined by fault scarps or erosional escarpments. In many other cases the moving rock sheet has to shear off somewhere along its length and ride up a stationary rock mass. In still other cases there is no riding of a toe over a stationary rock mass, but a frontal crumpling of strata. In the latter two situations the downslope motion of the rock sheet will be opposed by friction at its base plus a tangential force, either required to move the toe uphill or to crumple the strata at its lower end. Raleigh & Griggs (1963) have calculated that in the case of a toe the magnitude of the slope angle required for sliding to occur would be only a few degrees more than that required for the unobstructed case. But they have not taken into account the conditions at the upper end of the rock sheet which is to slide. In many cases tension cracks at the hinges of a crustal upwarp cross the rock sheet perpendicular to the stratification. In other cases, however, the sliding plane intersects the strata obliquely, resulting in a tail. The effect of this tail would be a small push from the rear which to a certain extent reduces the extra resistance of the toe.

The rock body overridden by the toe may assume an anomalous pore fluid pressure and a consequent decrease in shear strength. This may give rise to the phenomenon of progressive slumping.

Moore (1961) describes this process as follows (figure 5): "Assume that in a slope AD_1 is the sediment surface. The portion ACC_1 is in a state of limited equilibrium while portion CDD_1C_1 has a considerable margin of stability. A decrease in the shearing strength within part ABB_1 as a result of a rise in pore water pressure would increase the pressure on BB_1 by some value Δ F. This would result in an equal and opposit reaction Δ F in the stable part of the slope and a compression of the portion BCC_1B_1. As a result part ABB_1 would slide down somewhat. This sliding, even though only a few inches, would cause remolding and destroy most of the shearing resistance along AB and in the adjoining part of BC. Thus successive failure could continue down an extensive, otherwise stable, slope. The initial rise of the pore water pressure in a submarine deposit might be brought on by overriding of the upper section ABB_1 by a local slump from a high angle slope which might be triggered by an earthquake."

When a rock sheet has slipped down, a steep scarp is formed at the upslope end. This furnishes the opportunity for subsequent slumping along curved slip planes progressing in a headward direction (Grant-Mackie & Lowry, 1964), called retrogressive slumping (Andresen & Bjerrum, 1967).

Speculations on the hydromechanics of mud flows – In moving Bingham plastic sediment mixtures we are confronted with the mechanics of flowing materials. Hydrodynamics and hydraulics were developed mathematically and empirically largely in terms of Newtonian fluids. The hydromechanical theories developed,

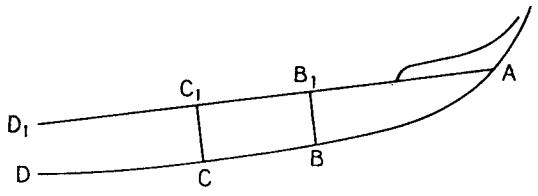


Fig. 5. Diagram showing the initial stage of progressive slumping, after Moore (1961).

therefore, can only be applied with certain modifications to Bingham plastic materials. In clay suspensions, for instance, the apparent viscosity at a certain temperature decreases with increasing rate of shear instead of being constant as in hydromechanics. Nevertheless some speculations with respect to flowing Bingham plastic materials will be given in the following.

Since downgliding motion of plastic material does not take place along specific shear planes there must exist a surface of zero shear velocity which acts as a base. Moreover, subaqueously sliding material meets with "more friction along its base than its surface, and because of uniform density, gravity acts equally on the whole thickness. Hence it will show its highest velocity at its surface" (Kuenen, 1956). This implies that a velocity gradient, dU/dy, exists within the flowing sheet and a parabolic velocity profile can be assumed as starting point. The flow at the upper surface is retarded somewhat by the friction with the overlying water.

If we assume the material below the base to be approximately the same as that in the moving sheet there will be a gradual increase of the rate of shear from the base in the y-direction. If we further assume a laminar flowing at the outset, the rate of shear will not be very large owing to the high apparent viscosity. In the zone, however, where the rate of shear is greatest the apparent viscosity and τ_0 will decrease owing to the clay flakes tending to be parallel to the flow lines and due to thixotropy. This will increase again the rate of shear up to a limit at which turbulence starts. This turbulence extends progressively in the y-direction. On the other hand, the rate of shear decreases above the zone in which turbulence is going to start, so that thixotropic stiffening may set in, gradually progressing in a downward direction. The result could be a gradual damping of turbulence in the y-direction, so that in some cases the relatively stiff upper part of the sheet may travel downslope riding on the turbulently flowing lower part. Along the base a laminar sublayer may be maintained. Consequently the velocity profile is altered somewhat (figure 6).

The type of turbulence depends on the apparent bulk viscosity of the material involved. In more viscously flowing material a slow eddying motion is generated. One can therefore say that the flowing is turbulent, but the high viscosity renders the motion rather different from the turbulence of a river (Kuenen, 1956). The lifting effect of this kind of "turbulence" might be greater than that of the sinking of possibly included pebbles; the slow upward currents might be able to carry many pebbles that cannot sink fast enough. The pebbles near the centres of eddies could have experienced several revolutions. I do not share the opinion of Görler &

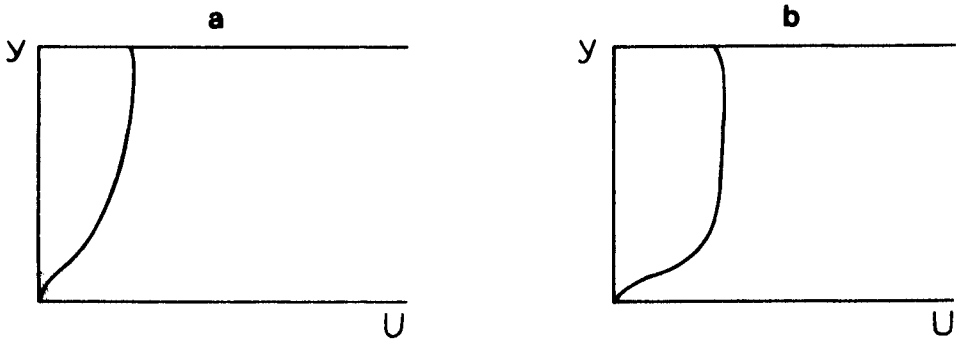


Fig. 6. Presumed velocity profile of a laminarly flowing sheet of Bingham plastic material (a), and its velocity profile after turbulence has set in (b).

Reutter (1968) that inclusions in upside-down position have been gone through only one rotation of 180°. After stopping of the motion and subsequent thixotropic stiffening, the upward and downward currents are frozen displaying themselves as curved strings of disoriented pebbles close together and as threads of mud enclosing only few pebbles.

In less viscously flowing material the sinking of pebbles will not be balanced by the lifting effect of turbulence. Here the type of turbulence approaches that of a river and sinking of pebbles will continue until the mean size of the remaining pebbles is diminished to such an extent that they can be succesfully kept suspended due to the turbulence. It seems likely that a concentration of pebbles is set up at a level at which the thixotropic strength is such that sinking comes to an end; here a graded sequence of pebbles may be built up. This level will generally be at or near the base.

The size distribution of the inclusions in a mud flow deposit is therefore a measure for estimating the bulk viscosity of the mud flow in relation to that of other mud flows. However, it should be kept in mind that flowing muds of equal viscosity but with different flow velocities will obtain different size distributions. The sinking of inclusions will reasonably be preceded by the acquisition of a preferred orientation. In stiff pasty mud flows only the larger inclusions will become oriented, but with decreasing order of viscosity ever smaller inclusions acquire a preferred orientation. Thus also by measuring the orientations of the inclusions for every dimensional category the relative viscosity of a mud flow can be estimated . The preferred orientation would also be a measure to what extent the flowing of the delapsing mass had been turbulent or laminar. In the ideal case a the pebbles carried in a laminar flow have their longer axes parallel to the flow laminae with their thin ends pointing upflow and their longest ribs perpendicular to the flow direction (Ruznak, 1957). Pebbles carried in a turbulent flow are aligned with their longer axes in broad girdles dipping upflow and have their thin ends pointing upflow as well (Lindsay, 1966). Also it can be argued that an inclusion of a particular consistence may be deformed in a visco-plastic way or be brecciated when enclosed by thick pasty delapsing muds, or may remain undeformed in thinly flowing muds. Lindsay (1966) has used this criterion to prove that an amount of sand in the mud does considerably lower the thixotropic strength and the viscosity of the delapsing mud.

If a large amount of sand is thoroughly mixed with mud of low bulk viscosity into a homogeneous sediment mixture, the behaviour of that mixture may, when flowing, approach that of a Newtonian fluid. The velocity of downslope movement is relatively high. If the velocity or the amount of sand is high enough the thixotropic limit might be exceeded and the sediment mixture could go into suspension forming a turbidity current. If the amount of sand is relatively low a higher velocity is needed for the thixotropic limit to be exceeded. This can for instance be achieved, when a certain stretch of the slope has a greater inclination or when the flow is falling from a scarp.

In the relatively stiff upper parts of some moving sheets of Bingham plastic material relatively large pebbles may remain floating. Moreover, if by turbulence a large pebble is conveyed into this upper part the chance that it may remain there is great especially when the flow is waning. This may cause a reverse grading in the pebbles floating in the mud sheet. This relatively stiff upper part may break into chunks in which pebbles may be enclosed. These chunks may become involved in the more fluid flowing mud of the lower part. Also in the centres of the

eddies a thixotropic stiffening may set in, which may give rise to spherical, rather stiff mud lumps; these may show contorted internal structures and may also enclose pebbles. During further delapsion the mud balls as well as the chunks of the broken-up top-layer may behave as independent inclusions and they may be shielded from subsequent destruction by the acquisition of an armour of pebbles that become stuck to the outer surface (armoured mud balls).

Due to turbulence also a mixing of the mud of the upper part of the mud flow with the overlying water may occur. This would give rise to a gradual lowering of the density and thixotropic strength of the mud towards the top of the flow. The inclusions suspended in the mud of the upper part will start to sink to a level where the density and the thixotropic strength of the mud remains high enough to prevent further sinking; the heavier inclusions will sink to a lower density level than the lighter ones. In the lower part of the mud flow, which will now be protected from liquefaction by the blanket of serried inclusions on top of it, the inclusions remain floating in the mud. All this would give rise to (1) a concentration of larger inclusions at an intermediate level of the flowing sheet and generally a gradual upward transition from mud-supported conglomeratic mudstone to grain-supported conglomerate, and to (2) a normal grading in the conglomerate towards the top.

The process of liquefaction enhances the turbulence in the upper part and the mud may be thrown into suspension together with the finest fraction of the clasts of the upper part. The latter may settle down in a normally graded fashion on top of the conglomerate and even far ahead of it. As this topmost deposit has been emplaced by the same depositional event as the remainder of the mud flow deposit below, it should not be separated from the mud flow deposit as a turbidite.

DISPERSIVE GRAIN PRESSURE AND REVERSE GRADING

Mechanics of dispersive grain pressure – If a rock merely consists of non-colloidal particles that are not cemented together, it may avalanche downslope, provided that the magnitude of the slope angle equals that of the angle of repose.

Bagnold (1954) begins his theoretical approach to the dynamics involved as follows: "When dry sand of fairly uniform size is supplied at the top of a slope of the same sand at an angle very near that of collapse, the supplied sand moves down the slope as a flow, i.e. every grain moves relatively to its neighbours and the shearing is general. The moving mass can be distorted round obstacles placed in its path. The flow has a well defined front and advances at a constant speed, depending only on the grain size and depth of the flow. This phenomenon has hitherto never been explained quantitatively, though it can be seen on any steep dune slope... The phenomena observed here . . . cannot be explained unless a dispersive grain pressure is postulated, of such a magnitude that an appreciable part of the moving grains is in equilibrium between this dispersive pressure and the force of gravity, whether or not fluid or air turbulence also plays a part."

The motion of the grains consists, in addition to a drift in the x-direction, of oscillations in all three directions, involving approaches to, and recessions from, neighbouring grains. In the case when the effects of grain inertia dominate, the oscillations are supposed due to successions of glancing collisions as the grains of one layer overtake those of the next slower layer. At each collision each grain of the overtaking layer experiences a change of momentum resulting in a dispersive grain pressure within the avalanching sheet of sand. Bagnold (1954) has given a formula

(whose derivation is not repeated here) for this pressure normal to the direction of shear.

$$P_y = a_i \rho_s (\lambda D)^2 (dU/dy)^2 \cos \alpha_i \quad (2)$$

in which a_i is a constant, ρ_s the density of the grains, λ the linear concentration (=linear ratio between solid and non-solid), D the diameter of the grains, dU/dy the shear strain, and α_i some unknown angle comparable with the angle φ of Coulomb's equation. There should also be pressure components in other directions, but their magnitudes are uncertain.

In the case when the effect of fluid viscosity dominates, for example in mud flows, the theoretical approach given by Bagnold (1954) is less simple. The parameter η, the viscosity of the fluid, enters the equation, while the parameter ρ_s, the density of the solid grains, does not. The equations of dispersive pressure, as developed by Bagnold, are only valid for elastic grains of uniform size and density.

Mechanics leading to reverse grading

Bagnold (1954) argues that "in the inertia region of grain flow the dispersive pressure P should vary as D^2, for a given shear stress dU/dy. This suggests that when grains of mixed size are sheared together, the larger grains should tend to drift towards the zone of the least shear strain, e.g. towards the free surface of a gravity flow, and the smaller grains towards that of greatest shear strain, e.g. the bed. This certainly conforms with experience. The effect seems very analogous to that of partial pressures". According to Middleton (1970), however, the acceptance of this argument by geologists to explain reverse grading "seems to be erroneous, because it is based on using a concept which has only a statistical (continuum) validity, and applying it to individual particles. An analogy would be to attempt to apply the concept of gas pressure to individual molecules in the gas". He argues that the correct explanation of this process would be that the smaller particles tend to fall into the spaces between the larger particles and thus displace the larger particles towards the surface.

Reverse grading can be achieved in the laboratory by a simple experiment in which medium sand and pebbles are rocked back and forth under water in a shaker mechanism; the pebbles quickly work their way up to the top of the sandy layer (Wellman, unpublished results in Sanders, 1965). This is, however, the exact opposite behaviour of fine sand and pebbles, where the pebbles sink into the fine sand (Sorby, 1859, p. 669-670). I am inclined to explain this seeming inconsistency of behaviour as follows: the reverse grading brought about by rocking of coarse sand and pebbles is, like dispersive grain pressure, also the effect of mutual grain collisions transferring momentum from grain to grain in the y-direction, which depends on the impulses of the grains involved. As these impulses depend on the grain masses which vary with the third power of the grain diameter, it will be clear that at a certain lower limit of spherical diameter the sum of the impulses of the small grains is not able to compete with the impulses of the large pebbles. The pebbles will no longer be carried by the impulses of these very tiny grains and consequently will sink in response to the force of gravity. This sinking will start earlier for spherical pebbles than for non-spherical ones, because the surface of the latter is larger than that of the former; the non-spherical pebbles are carried by more impulses of the fine grains than the spherical ones. So reverse grading cannot be explained without a dispersive grain pressure being active; mutual grain collisions may give rise to dispersive grain pressure as well as reverse grading.

Continuing this line of reasoning we might explain the sinking of pebbles in a fluid matrix of clay-sized particles in an analogous way. If, however, the clay is rather stiff with a high viscosity, η, the pebbles will not sink but remain suspended in the clay. In that case the effect of mutual collisions of pebbles is suppressed

by the inability of the pebbles to move with respect to the surrounding mud. As there is a gradual transition between diluted and stiff clay systems, there will be a gradual transition between sinking and not-sinking of the enclosed pebbles according to their sizes, and consequently between grain collisions being effective and suppressed. Thus for every apparent viscosity there is a critical size below which no sinking takes place and no grain collisions can be effective to bring about dispersive grain pressure or reverse grading.

The way in which reverse grading in mud flows come about – If a concentration of larger grains is set up at the base or at some higher level within a moving mixture of mud, sand and gravel, frequent collisions are likely to occur between the non-colloidal grains. This may give rise to dispersive grain pressure and to the tendency to form a reverse grading, provided that a certain rate of shear is maintained. In hydraulics this provision is fulfilled in the upper flow regime in which plane beds are formed. A large part of the transported material is then confined to a relatively thin layer near the base, the so-called flowing grain layer (Sanders, 1965). In fact, whether or not a plane bed occurs appears to be intimately related to the fall velocity of the grains in the fluid and the tangential tractive force of the current. A dune bed could be changed into a plane bed, or a plane bed into antidunes only by decreasing the fall velocity (Simons et al., 1965). Therefore the conditions for the formation of plane beds and flowing grain layers, even for large grain sizes, are reached in flowing muds at relatively low velocities, because of the low fall velocity and the high tractive force due to the high apparent viscosity (Allen, 1972). The tractive force may be sustained by an additional tangential component of gravitation acting upon the moving mixture.

In the flowing grain layer at the base of this sediment mixture the larger grains are more strongly urged towards the zone of least shear strain than the smaller ones. The larger the grains the more rapid they will be drifting upward, overtaking the less coarser grains. The inevitable effect must be that the spaces between the largest grains will gradually be filled with ever finer material, while, on the other hand, but simultaneously, the neighbouring grains become ever less different in size. Since the action of dispersive grain pressure is restricted to those grain sizes that are able to move with respect to the surrounding mud, there is a lower grain-size limit. The grains of this limit-size will tend to concentrate along the base of the flowing mud stream. So the median diameter found there, should be a measure for the consistence of the mud during movement.

The result is a very peculiar type of reverse grading. In the thin laminae of sand produced in flume experiments the peculiarity of this type of grading cannot be observed. But I had the opportunity to study this type of grading on a large scale in the conglomeratic basal parts of mud flow deposits and in conglomeratic

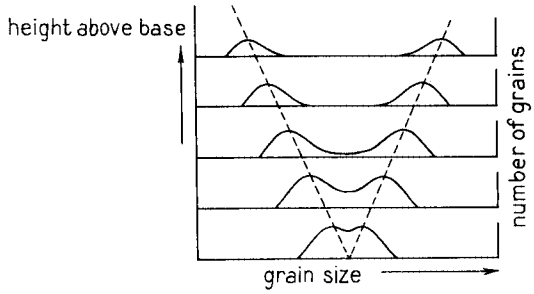


Fig. 7. Scheme showing the divergence of the modes in successive higher levels above the base of the reversely graded unit.

fan-deposits in the Moratalla region (Southeastern Spain). It is in fact an upward gradation from mono or moderate bimodality to extreem bimodality concomitant with a divergence of the constituent modes (figure 7). If we consider the basal conglomerate to be built up of "matrix" and "clasts", we may describe the grading by saying that the modal diameter of the clasts becomes progressively larger, whereas that of the particles in the matrix becomes progressively smaller from the base to the top of the basal conglomerate. The grading is from badly sorted conglomerate to a very badly sorted conglomerate. Apparently because of turbulence in the flowing mud above the basal conglomerate, large clasts had repeatedly been picked up from the conglomerate; due to thixotropic stiffening after the flow has stopped, the clasts remained floating within the mud forming a conglomeratic mudstone. This conglomeratic mudstone may be considered to represent the climax of unsortedness. The unexpectedly large number of small clasts found in some of the basal conglomerates in the Moratalla region may be explained by assuming that they were dragged along in the wakes of sinking larger clasts, or were pushed downward in their bow-waves.

During the course of the differentiation the upper conglomeratic mudstone and the lower conglomeratic part of the mud flow tend to move gradually more independently from each other. The boundary between both parts may become sharp giving the suggestion of two beds that seem to have resulted from two independent events.

Limiting slope angles for flows of gravel-mud mixtures – With respect to mud flows with a very large amount of pebbles and (or) boulders, Bagnold (1968) has calculated the minimum slope needed for the flowing to maintain, which would be about 3° for subaerial mud flows and 6° for subaqueous mud flows. Also Middleton (1970) has calculated the minimum slope angle necessary for the maintainance of the downward flowing of gravel with interstitial mud in subaqueous conditions. He has shown that this slope angle for the system being in the inertial regime is considerably less (4.5°) than that for the system which is in the viscous regime (10.5°). The flowing could be shifted from the viscous to the inertial regime if the rate of shear, or the grain size is increased, or if the viscosity of the interstitial fluid is reduced. For no interstitial fines his calculation gives a minimum slope angle of 18°. Dill (as quoted by Moore, 1966) has proved the effectiveness of slow creep and local relatively small-scale slumping in moving sand slopes in the upper parts of submarine canyons with slopes of about 10° or more. Very steep slopes of 25° or greater are shown to be subject to sand flowing similar to the movement of a river.

Processes of delapsion and their results

The terms 'slumping' and 'sliding' have often been used interchangeably, referring to downslope movement of rigid as well as Bingham plastic rock masses (Jones, 1937, 1940, 1947; Fairbridge, 1946; Kuenen, 1956; Dzulynski, 1963; Potter & Pettijohn, 1963; Dzulynski & Walton, 1965; Morgenstern, 1967; Görler & Reutter, 1968). According to Dott (1963), however, sliding should be restricted to mass movements of either rigid or semi-consolidated masses along discrete shear planes

with relatively minor flowing; slumping should refer only to rotational sliding of masses along discrete shear planes, following recommendation of the National Research Council Landslide Committee (Varnes, 1958). The downslope movement of rock masses involving largely distributed internal shear between particles when stress exceeds the yield limit is referred to by him as "plastic mass flow" (= Bingham plastic flowing); they would have retained most of their internal stratification, although much contorted. "Viscous fluid flow" would occur when the liquid limit is exceeded and is, according to him, characteristic of sediment-laden density currents.

Dott (1963) continues that in colloidal material "viscous fluid flow" exists without the material going into suspension. "If rate of shear and turbulence increase sufficiently to destroy bulk cohesion and if dilution by in-mixing of water increases the void ratio sufficiently, then viscous flow ensues and a subaqueous mudflow or "mudstream" may be formed. In these cases the material passes beyond a wholly plastic state and becomes sufficiently fluid for turbulence to destroy internal stratification and mix together all size fractions in a wholly unsorted fashion. Such masses just begin to behave according to fluid flow, but are arrested before all of the material can form true turbidity currents. They represent false-body thixotropy. Consistency increased again during flow due to decreased rate of shear, or decrease in void ratio accompanying gelation, or both. Cohesion in thixotropic materials may be restored just as suddenly as it breaks down under shock". Elliston (1968) subdivides this latter category of Dott, according to its supposed viscosity during delapsion, into sediment that "flows as aqueous slurry to deposit as mud flake rock or tilloidal mudflow conglomerate" and sediment that "moves as a thick pasty mud glacier to deposit as "mudstream rock" or rheopelite."

Another type of delapsional process is that which results in brecciated clays (argille brecciate). The downward movement of these clays is supposed to take place by continuous and progressive brecciation of the clayey rocks caused by the weight of the clay mass itself. All size grades of breccia elements can be found in these brecciated clays representing the frozen evidence of all stages of pulverization to which these clays had been subjected. The pasty groundmass represents the finest disintegration products of the breccia elements. The same process may take place in soft calcilutites.

Preference should be given to the opinion of Sharpe (1938), Strahler (1952), Varnes (1958), Marchant & Black (1960), Dott (1963), and Laird (1968) to restrict the term 'slumping' to the slipping of a rigid to semi-consolidated rock mass moving as a unit or as several subsidiary units along a discrete concave-up shear plane with relatively minor internal visco-plastic flowing (which may give rise to slump folds) involving backward rotation on a more or less horizontal axis parallel to the cliff or slope from which it descends. Good examples of slumping are given by Reiche (1937), Jones (1947), Newell et al. (1953), Rigby (1958), Grant-Mackie & Lowry (1964), Williams et al. (1965), Laird (1968), and Sturm (1971). The slumped masses are called slump blocks when about isodiametric in shape (Newell et al., 1953; Williams et al., 1965), or slump packets when flat lenticular in shape (Jones, 1947; Grant-Mackie & Lowry, 1964; Laird, 1968). Reiche (1937) has introduced the term 'toreva blocks' for slump blocks that occurred in subaerial conditions (cliff recession; Felsstürze, Heim, 1882). Since a difference in medium does not effect the concept in its essence, Rigby (1958) has used it for subaqueous slump masses. The boundaries between slump masses or between slump masses and non-slumped rock may look like angular disconformities or thrust faults. Particularly indicative are the "inversed angular disconformities" in which the stratification of the upper segment is truncated by that of the lower one. The strata near these slip planes may be intricately crumpled (dragfolding). The idea of Potter

& Pettijohn (1963), Pettijohn & Potter (1964), and Pettijohn et al. (1972) to apply the term 'slumping' to gravity generated movements of small lateral displacements, and 'sliding' to similar movements of greater lateral extent is rejected here.

Neither preferable is the opinion of Dott (1963) to restrict the term 'sliding' to movements of rock masses along discrete shear planes. Here it is proposed to change the term 'sliding' sensu Dott (= including slumping) into 'slipping', whereas the term 'sliding' should cover at least both slipping and delapsion by "plastic mass flow" sensu Dott, for in the latter the displacement generally takes place along discrete lubricating horizons and in this respect it does not differ much from slipping along shear planes. In fact, Mary & Dangeard (1970) unite the result of slipping and delapsion by "plastic mass flow" into "glissements". This would also be more in accordance with the current use of the term 'sliding' among geologists. The terms 'slide', 'slump' and 'flow' have been used to denote both the process of sliding, slumping or flowing and the material involved in these processes. Preferably geologists should use the terms 'slump' (or slump deposit), 'slide' (or side deposit) and 'flow' (or flow deposit) only with reference to the material during or resulting from the process of slumping, sliding or flowing respectively.

The continuous series of delapsional processes has been subdivided on physical grounds according to two natural breaks, viz. whether the yield limit or the liquid limit of the rocks have been exceeded. But as geologists are only concerned with the result of delapsion, a geologic subdivision should necessarily be based on observable criteria in the olisthon. It has repeatedly been attempted to co-ordinate both subdivisions, but they never match, because they are based on different criteria.

The magnitude of the limiting stress beyond which the rock yields varies with different confining pressures. In many cases of delapsion the rock yields only in restricted parts of the delapsing mass, only along the slip plane for instance. Dott (1963) apparently refers to "plastic mass flow" when the yield limit is exceeded in almost the entire rock mass involved; for the criterion used by him to distinguish between "sliding" and "plastic mass flow" is the degree of internal contortion in the olisthon. It is therefore correct to distinguish between sliding without or with minor internal contortions (slipping) and sliding with strong internal contortions but in which the internal stratification more or less retains its coherence. Of course all transitions exist between the two. The term "plastic mass flow" is not suitable for a geological subdivision of the processes of delapsion, the more so as Dott (1963) ranges for the sake of a comprehensive subdivision the internal folding in non-cohesive sandstone beds under "plastic mass flow" which is strictly speaking unacceptable.

Similar objections exist with respect to the liquid limit. This limit is reached when the cohesion of the material is reduced to a negligible value. It frequently occurs that the liquid limit is reached only in a small part of the delapsing rock mass. One of the criteria used by Dott (1963) to determine whether the liquid limit of an olisthon had been exceeded is that the material had gone into suspension. Nevertheless he considers "subaqueous mud flows" as just beginning to behave according to "viscous fluid flow", although the material generally does not go into suspension. In olisthons that are composed of mud with inclusions and that apparently are emplaced by flowing without going into suspension, it is generally impossible to determine to what extent the bulk cohesion in the mud had been destroyed. On the other hand, "flowing" pure sand, in which the liquid limit is exceeded, will not go into suspension. It is therefore preferable to distinguish between

delapsion in which the greater part of the material is thrown into suspension to settle down in graded fashion (turbidity currents) and delapsion in which it is not. Of course again all transitions exist between the two. So also the term "viscous fluid flow" is not suitable for a geological subdivision of delapsion.

In olisthons composed of mud with inclusions also the criteria used by Dott (1963) to distinguish whether "plastic mass flow" or "subaqueous mud flow" had been responsible for the emplacement of the olisthon do not hold. In the Tuscan-Emilian "argille scagliose", for instance, the innumerable slip planes in the scaly clays indicate visco-plastic flowing, though the original stratification is entirely destroyed and all size fractions of the indurated inclusions are mixed together in a wholly unsorted fashion which would be a recognizing criterion for viscous fluid flowing of the muds.

In order to cope with these difficulties it is more practical to distinguish between incoherent and coherent sliding, a terminology borrowed from Dzulynski (1963) and Dzulynski & Walton (1965) (they used the terms "incoherent" and "coherent slumps"). In incoherent sliding "extensive mixing of broken up layers of indurated sediment in a mass of sand, silt or mud" takes place. In incoherent sliding the "indurated" parts may also behave in a visco-plastic way and may show a variety of shapes and internal contortions. The examples given of incoherent slides are always intraformational, but this should not be an impediment to call similar slides in which exotics are present also incoherent slides. The progressive brecciation which gives rise to argille brecciate may be incorporated in the concept. In coherent sliding (of which "plastic gliding" is a common type of movement) less extensive mixing occurs and beds, though strongly folded and sometimes thrust over one another, largely retain their identity. All examples of coherent slides suggest a weak competence of the rocks involved. Though slip blocks, slip sheets, slump blocks, and slump packets also retain their coherence during delapsion, they are not incorporated in the concept of coherent slides. Slipping and coherent sliding may however be united into coherent delapsion. Of course all transitions exist between slipping, coherent sliding, incoherent sliding, and sediment flowing.

According to Dzulynski & Walton (1965) "sediment flows involve the downslope cascading of loose sediment and, depending on the clay/sand ratio, the mass can be referred to as a mudflow or sandflow." The two types are treated together by them for convenience, but it will be realized that differences of cohesion lead to different types of movement. The extreme case of the pure "sand flow" involves movement of the individual grains, whereas increased mud content leads to a more and more cohesive mass moving as a unit. Apparently they have separated incoherent sliding and mud flowing on account of the presumed difference in the consistence of the material involved. But as it is often very difficult to distinguish incoherent slides from mud flow deposits, which both consist of muddy rocks within which fragments are floating which do not show any coherence, we might better unite both incoherent sliding and mud flowing into incoherent delapsion. Examples or subaqueous "flows" of pure sand, such as the spectacular "rivers of sand" and "sand falls", have been observed by Dill (1964) in the tributaries of the submarine San Lucas Canyon.

It should be noted that the scope of "sediment flow" of Dzulynski & Walton (1965) is not the same as that of sedifluction as defined by Richter (1952). The latter term has been proposed to comprise "hypothesenfrei die fliesende Bewegungen unverfestigter Sedimente, der Subaquatischen wie Subaerischen" as distinct from bioturbation, which is defined as movement due to the action of organisms,

and from solifluction, which is defined as movement of ground subsequent to consolidation of sediments. Sedifluction can be divided into endostratic and diastratic sedifluction. The endostratic sedifluction is considered as a "physiologische Begleiterscheinung der Sedimentbildung" and is limited within the layer, for instance load-casting and convolute lamination (which Dott, 1963, includes in his "plastic mass flow"). Diastratic sedifluction disturbes the stratification, as in subaquatic sliding.

Instead of the term "sand flow" of Dzulynski & Walton (1965) the term 'sand avalanching' is proposed in order to make a clear distinction between the flowing of a mud slurry or any other liquid and the downward travel of non-cohesive sand grains. From a mechanical point of view the avalanching of gravel or rubble can be approached with the same parameters as that of avalanching sand, so that the terms 'gravel avalanching' and 'rubble avalanching' do fit well. 'Grain avalanching' would be suitable as a general term for the delapsion of predominantly non-cohesive grains that were not cemented together and did not go into suspension while delapsing; the dynamical contribution of the colloidally behaving mud is negligible. The term 'grain avalanching' is preferred here to the usually employed terms namely 'debris avalanches', 'grain flow', 'inertia flow', or even 'mass flow'. The term 'grain avalanche deposit' is borrowed from 'avalanched encroachment deposit" of Bagnold (1941), which refers to deposits of sand on the slip face of dunes, and from "avalanche deposits" of Imbrie & Buchanan (1965).

Gravel avalanche deposits are commonly non-graded, but normal grading, especially at their very top, reverse grading and recurrent grading have frequently been found. Locally cross-bedding can be detected. They show all kinds of sorting and roundness of their often plurigene constituents. They have a grain-supported framework, displaying the aspect of orthoconglomerates. Mud, silt, and sand may be present in various percentages in the interstices. They are commonly inconspicuously bedded. The shape of the conglomerate body usually resembles a very flattened lentil or it may vary in thickness. They are usually associated with deposits which are supposed to be of delapsional origin. In many cases clasts are present which can be taken as being too large to be emplaced by water or air currents. Some examples of gravel or rubble avalanche deposits are: breccias of the Nappe de la Brêche (Switzerland, Caron, 1966; Hendry, 1972), certain non-graded conglomerates of the Nappe du Niesen, conglomerate of Vouan, part of the Mocausa conglomerate, (all in Switzerland, Caron, 1966), "brêches du Flysch" (Switzerland, "brêches d'écroulement margineux", Tercier, 1948), Bolgenconglomerate (Austria, Cornelius, 1924), San Onofre Breccia (California, Woodford, 1925; Dott 1963; Fisher, 1971), polymictic conglomerates of the Istebna Beds (Poland, Unrug, 1963), Permian reef front breccias (U.S.A., Newell et al., 1953; Rigby, 1958; Bucher, 1959; Dott, 1963), Baganza Conglomerate (Italy, Sames, 1965), conglomerate of Castell'Azarra (Sicily, Jacobacci et al., 1959), conglomerates of the Repetto Member (California, Natland & Kuenen, 1951), boulder beds of Sutherland (England, Crowell 1961). Gravel avalanche deposits have differently been explained as: slide breccias or conglomerates (Kuenen & Carozzi, 1953; Kuenen & Sanders, 1956; Stanley, 1961; Walker, 1967), as deposits from turbidity currents (Newell et al., 1953; Sames, 1965), as rock fall deposits (Dott, 1963), as wildflysch (Ogniben, 1963), and as olisthostromes (Bucher, 1959; Bortolotti, 1961-1963; Passerini, 1962, 1965; Boccaletti et al., 1966; Jacobacci, 1966). Unrug (1963) explains them as watery slides of sand and gravel and includes them in his concept of "fluxoturbidites".

Grain avalanching (and also slipping) may pass into rock falling when the slope approaches the vertical. Rock falling comprises the relatively free falling of detached segments of bed-rock of any size from a cliff, steep slope, cave, or arch (Sharpe, 1938). Also the process of the rolling down of individual fragments can be incorporated in the concept of rock falling. The same process may occur subaqueously. The material accumulates at the feet of the scarps as screes. If the angle of repose is exceeded these screes may give rise to rubble avalanchings in which the fragments collectively roll and sliver downslope. This subsequent avalanching in the screes should not be incorporated in the concept of rock falling as is done by Dott (1963).

If the mud content is large the delapsing mixture of sand and mud may go into suspension forming a turbidity current. Turbidity currents are supposed to be generated by large gravitating sediment masses which lose their bulk cohesion because their thixotropic (or liquid) limit is exceeded. Morgenstern (1967), van der Knaap & Eijpe (1969), and Middleton (1970) suggest that liquefaction due to the breakdown of structurally metastable, porous sediment may be important in the generation of turbidity currents. According to them the breakdown of the unstable sediment structure may cause a temporary production of excess pore-fluid pressure which greatly reduces the shearing resistance of the liquified sediment, which may flow even on very small slopes. Van der Knaap & Eijpe have shown that the time taken for the dissipation of the excess pore-fluid pressure would be sufficient to permit acceleration, and mixing with the overlying water might lead to the delapsing mass going wholly or partly into suspension. This depends on the velocity of the delapsing mass and on its mud content which determines the capacity of the surrounding water to take up non-cohesive grains into suspension. If this suspensive capacity of the clay-laden, turbulent water is too low, a part of the non-cohesive grains is dropped out instantaneously, or better, the greater part of the mud merely escapes from the interstices between these grains. This non-suspended part may travel downslope in a manner which differs mechanically from that of the suspended part. If so, the delapsing mass then consists of a suspended part and a bed-load part. Generally the suspended part outdistances the bed-load part. As both parts are due to one delapsional event, the entire delapsing mass should be designated as turbidity current, provided that at least the greater part of it is in suspension. The whole ultimately resulting bed (one depositional unit) has to be called turbidite irrespective as to whether a bed-load part is present or not. This is in agreement with the opinion of Kuenen (1965), but in contradiction to that of Sanders (1965) who designates only those materials settled down from this suspended part as turbidite. Also in a mud flow deposit with a conglomeratic basal part and a muddy upper part, both parts, though differing considerably in composition and in mechanical behaviour during downward travel, are the results of one depositional event and the entire bed should be called mud flow deposit.

The most comprehensive subdivision of delapsion with respect to the processes is that of Dangeard et al. (1965) and Mary & Dangeard (1970). They distinguish between:

1. "glissements" (subdivided into "glissements intraformationels" and "glissements de bancs, de blocs, ou de nappes")
2. "coulées boueuses"
3. "coulées sableuses (fluxoturbidites)"
4. "écroulements et avalanches (avalanchites)"
5. "courants de turbidité (turbidites)"

The criteria, however, by which the resulting deposits could according to them be recognized in geologic sequences are unsatisfactory. They are of the opinion that the distance travelled is an important distinguishing character and consequently their being allochthonous or autochthonous. This is the reason why they make a distinction between "glissements intraformationels" and "glissements de bancs, de blocs, ou de nappes", and between "écroulements" and "avalanches"; the second and fourth mentioned slidings would have travelled over a larger distance than the first and third mentioned ones. The difference between "écroulements" and "avalanches" is rather one of slope angle. Here "glissements intraformationels" and "glissements de bancs, de blocs, ou de nappes" are treated together, as their distinction is rather a matter of size than of genesis; for the same reason also "coulées sableuses" and "avalanches" are treated together. Consequently the following five extremes, between which all transitions exist, were distinguished among the processes of delapsion:

1. Coherent delapsion: a) incoherent sliding (including progressive brecciation).
   b) coherent sliding.
2. Incoherent delapsion: a) incoherent sliding (including progressive brecciation).
   b) mud flowing.
3. Turbidity current action.
4. Grain avalanching.
5. Rock falling.

Non-stratiform olisthons

Geologists have also introduced various genetic terms for olisthons indicating their supposed mode of generation by geologic processes other than the processes of delapsion discussed in the preceding chapter. The terminology based on these geologic processes, however, does not match with the one based on the delapsional processes. This "geological" series of olisthons is here subdivided into non-stratiform and stratiform olisthons which nearly correspond to the products of coherent delapsion (by sliding) and of delapsion by "flowing" or avalanching respectively. The non-stratiform olisthons are in turn subdivided into those generally studied by structural geologists and those generally studied by sedimentologists, because this reflects the controversy where to draw the boundary between tectonization and sedimentation. A current tendency among geologists is to distinguish between olisthons composed of exotic (or allochthonous) rocks and those that are not. But as the kind of process is not affected by the rock mass being exotic (allochtonous) or not, the distinction should only find expression through appropriate prefixes. Descriptive criteria such as numerical quantities, for example dimensions, percentages etc., must be omitted from the definitions of genetic concepts.

NON-STRATIFORM OLISTHONS GENERALLY STUDIED BY STRUCTURAL GEOLOGISTS

Delapsional outliers – Those parts of a thrust sheet – generally the marginal parts – that owe the circumstance that they are outliers only to delapsion along a basal plane which is the lateral extension of the thrust plane of the original thrust sheet,

should be called delapsional outliers in contrast to tectonic outliers, which owe the circumstance that they are outliers to the fact that they are remnants preserved by the erosion of a previously continuous thrust sheet. The circumstance that a rock mass can be designated as an outlier, delapsional or tectonic, does not depend on the distance travelled by the rock mass, but only on the fact that it has no connection with the parent rock mass. This is the main reason why the boundary between tectonization and delapsion was chosen (see page 14) whenever the last movement of a rock mass, however small it may be, can be shown to be delapsive.

Slip-sheets and gravity collapse structures

The phenomenon called slip-sheet is described by Harrison & Falcon (1934, 1936) as "a slab of rock which has become detached from its original place on a dipslope and has slipped down remaining more or less intact" and in normal position. "A slip-sheet is the product of movement in the upper limb of a knee-fold". A knee-fold is a structure "produced from a sine curve fold when the crest is notched by erosion and the two sheets so produced slip down and outwards bending at first". According to Harrison & Falcon (1934, 1936) a knee-fold may evolve into a roof-and-wall structure, when the knee breaks. "The lower limb remains nearly vertical, whilst the upper rests upon it with a sudden change in dip, and separated from it by a low-dipping thrust-fault". If this limb becomes overturned, the structure is referred to as a cascade. They describe a cascade as "developed by one or more subsidiary wrinkles forming upon a dipslope. Sometimes only the upper sheets of a formation are affected, sometimes the whole formation." Such a knee-fold with overturned lower limb may give rise to a roof-and-wall structure with an overturned lower limb (wall). Subsequently the upper limb may develop into a slip-sheet and the lower limb into a flap which is "a part of a (limestone) sheet bent over backwards without breaking. The basal beds are turned upwards while the top beds lie below, resting upon any of the higher strata left in the syncline which the flap covers". If the reversed parts of such flaps become separated from their non-overturned parts the relicts present themselves as reversed slip-sheets. These so-called gravity collapse structures are explained as the products of deformation and displacement of rock masses due to gravity after their cover and support have been eroded away.

It may be gathered from the description of Harrison & Falcon that their gravity collapse structures were formed mainly subaerially, but this should not be an impediment to designate similar phenomena that took place subaqueously also as gravity collapse structures. The requirement that slip-sheets should be a product of movement in the upper limb of a knee-fold is here abandoned as unnecessary with respect to their formation.

In the foregoing chapter we have already defined slip blocks, slump packets, and slump blocks which in geologic literature are generally employed with respect to non-exotic rock masses. But it would be unwarranted to restrict these purely genetic concepts solely to non-exotic rock masses. The adjectives 'exotic' or 'allochthonous' may be added when the slip masses cannot be referred to as non-exotic.

Diverticules – Another type of slip mass studied by structural geologists is the so-called diverticule, a term introduced by Lugeon (1943) to explain the disordered complexity of the Ultrahelvetic nappes (Appendix 5). It seems likely that the allochthonous complex of the Northern Appennines originated largely by diverticulation like the Ultrahelvetids. Diverticules can be defined as allochthonous slabs of rock that slipped into the adjacent depression, by which process a normal but tilted rock sequence became successively dismantled, in some cases even down to its basement; their emplacement closes the depositional phase in the geologic development of the area. Diverticules are slip sheets, slip blocks, slump packets, slump blocks, or other slip masses of irregular shape, which may have been subject to deformation by folding or faulting. If strongly folded the term 'slip sheet' cannot

be used, for it has lost its sheet-like appearance. The term 'diverticule' is therefore more comprehensive than 'slip sheet' etc.; the two terms cannot be considered as synonyms.

The Ultrahelvetic diverticules are emplaced subaqueously upon the very rocks of which the Helvetic nappes proper are built of, but before these nappes came into being (Trümpy, 1960; Badoux, 1973). Also the diverticules that gave rise to the allochthonous cover of the Appennines are emplaced upon the Tuscan and Umbrian sequences before the Tuscan nappes proper were formed. Badoux (1963, 1967, 1973) has endeavoured to reconstruct the sequence of events that led to the diverticulation in the Ultrahelvetids starting from the original situation before sliding to the emplacement of the various diverticules upon each other (which occurred in such a way that the diverticules, when mounting the series, successively contain ever older rocks) and their subsequent deformation. Because of their emplacement after the "sedimentary phase" the diverticules have hitherto been referred to as "tectonic". But according to the concepts of tectonization and delapsion advocated in this paper, diverticules are not tectonic phenomena but olisthons which are depositional rocks.

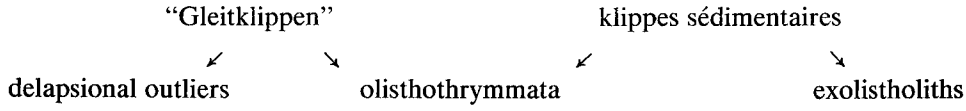
NON-STRATIFORM OLISTHONS GENERALLY STUDIED BY SEDIMENTOLOGISTS

The olisthons that do not form strata (though they may be internally stratified), but are nevertheless generally studied by sedimentologists, will be discussed in this chapter.

Klippes sédimentaires, "Gleitklippen", and olisthothrymmata - Closely related to the diverticules are the so-called klippes sédimentaires, a name originally given by Lamare (1947) to large blocks or lentils of exotic rocks occurring in certain levels within Cretaceous breccias and conglomerates (which are according to Feuillée, 1967, deposited by "écroulements", rubble avalanching) covering transgressively the Hercynian massive of Mendibelza (Pyrenees, France) (Appendix 6). Klippes sédimentaires can be defined as indurated exotic rock slabs that have slipped into the adjacent basin after their detachment from coastal or submarine cliffs; they were emplaced during the depositional phase in the development of the basin. The principle difference between klippes sédimentaires and diverticules is their time of emplacement in relation to the depositional phase in the development of the basin. This is the main reason why the former have hitherto been considered as sedimentary and the latter as tectonic phenomena and why the two have always been rigorously separated.

I do not share the opinion that exotic slip masses terminating a stratigraphic rock succession should necessarily be things that are entirely different from similar exotic slip masses intercalated within that rock sequence. Also Merla (1964a) is of the opinion that the olisthostrome closing a rock sequence is not something that is different from the intercalated ones. It is illogical to designate a large delapsed rock mass enclosed by depositional rocks as a sediment, whereas a similar rock mass on top of them would be tectonic. Delapsional deposits such as diverticules, klippes sédimentaires, and olisthostromes mark an episode of unrest in the diastrophic development of a basin. In the case of delapsional deposits concluding the depositional phase of a basin, it marks the initiation of a diastrophic phase so violent that the tectonic movements and the possible uplifting enhances the dis-

turbance and the denudation of the records of the possible further augmentation of depositional rocks, so that these are generally absent or unrecognizable. Obviously there is no essential difference between diverticules and klippes sédimentaires, for it is just an accidental circumstance whether exotic slip masses become klippes sédimentaires or diverticules. It would depend solely on the diastrophic stage of development of the basin at the time of emplacement. It should be noted that olisthostrome-like rocks also occur between the giant allochthonous rock slabs (diverticules) in the Northern Appennines (Abbate et al., 1970b), so that many of the latter may also be regarded as klippes sédimentaires.
Badoux (1963) and Lombard (1971) used the term "lames exotiques" to designate the klippes sédimentaires of Lamare. Broquet et al. (1967) used the term 'klippes sédimentaires' to designate those rock masses for which Jacobacci (1966) has restricted the term exotics, viz. exotic rock slabs enclosed by non-olisthostromic deposits (page 17). But this restriction of klippes sédimentaires by Broquet et al. is not justified, because olisthostromes are also depositional rocks and it would imply a discrimination between different depositional rocks. Olisthostromes are not so extraordinary among olisthons that they require a special treatment. Moreover, many olistholiths may be exotic rock slabs that have slipped down into the basin (thus klippes sédimentaires) and that only by accident became involved in olisthostromization. The proposition of Broquet et al. would also imply a discrimination between different olisthons, for many of the exotic rock masses of Jacobacci are enclosed in the Numidian turbidite sequence, whereas the types of klippes sédimentaires are enclosed by rubble avalanche deposits. Görler & Reutter (1968) proposed the term "Gleitklippe" to designate exotic rock masses not enclosed by olisthostromes but by other kinds of depositional rocks (Appendix 7). But according to them "Gleitklippen" partly originate by the detachment of smaller portions from the front of an approaching thrust sheet, so that the delapsional outliers that are embedded within non-olisthostromic deposits should be considered as a special type of "Gleitklippe". Fortunately Richter & Mariolakos (1973) proposed the term olisthothrymmata (sing. olisthothrymma) to designate the exotic slip masses that are not enclosed by olisthostromes and that are not delapsional outliers, but broke loose from steep slopes or cliffs. It does not alter the concepts of olisthotrymma and klippe sédimentaire (or diverticule) in their essence if the steep slope or cliff appears to be the front of an advancing nappe. The term "Gleitklippe" can be retained for all slip masses that are not enclosed by olisthostromes. Klippes sédimentaires can now be properly subdivided into olisthothrymmata and exolistholiths (defined op page 48).



The term 'klippe sédimentaire' is an unfortunate one, because it is neither a klippe (neither in its original, nor in its present-day sense), nor a true sediment. Moreover, this term is introduced to designate comparatively voluminous rock masses. Also the term 'diverticule' is commonly used for rather large rock slabs. Large dimensions are also suggested by the term "Gleitklippe". Since geologists should avoid to put dimensional quantities into their genetic concepts, every exotic rock mass, no matter how small, that has slipped down instead of rolled down should be considered as a small diverticule, klippe sédimentaire, "Geitklippe", or olisthothrymma.

Olistholiths, original definition – Olistholiths, generally studied by sedimentologists, are also olisthons that do not form strata, though they are frequently internally stratified. Although occurring within olisthons that actually do form strata (olisthostromes), they will be discussed in this chapter, because many of them can also be considered as klippes sédimentaires which by accident have become included in olisthostromes.

Olistholiths are, according to the original definition of Flores (1955), rock masses included as individual elements within the binder of an olisthostrome. Further he states about olistholiths that they are dispersed bodies of harder rocks, ranging in size from pebble to boulder and up to several cubic kilometres, that the masses have previously been referred to as "exotics" or "erratics", and that there is no constant ratio between the total volume of the inclusions and that of the bedding mass.

The definition and its amplification may be explained in two ways: 1. Only the exotic harder rock parts dispersed as individual elements within the binder of an olisthostrome should be designated as olistholiths, which would imply that the non-exotic ones should not be called olistholiths. 2. All the dispersed bodies of harder rocks included as individual elements within the binder of an olisthostrome may be designated as olistholiths. Only Beneo (1956a, b) and Klemme (1958) explained the definition in the first way, whereas all other investigators, including Flores himself (1959), use the olistholith-concept in accordance with the second explanation which therefore seems to be the intended one.

Olistholiths versus "pezzame" (= *debris*) – Beneo (1956a, b) restricts the designation 'olistholith' for the large exotic rock masses, whereas the pebbles and small blocks within the olisthostrome are designated as "pezzame". Hence he introduced the factor of dimension into the genetic concept of olistholith.

Jacobacci (1966) partly returns to the definition of Beneo (1956), for he considers still valid the denomination of "pezzame" to indicate the minute material mixed with the matrix of the olisthostrome (Appendix 8). From his statements can be gathered that "pezzame" are the usually small macroclastic elements uniformly distributed in the binder of an olisthostrome which are considered as the broken up fractions of the main lithological constituents of the parent rock of the olisthostrome. Olistholiths would be the lithic elements in an olisthostrome that are in whatever way liable to be differentiated from the binder and the "pezzame" (genetically, dimensionally, lithologically or paleontologically).

Abbate el al. (1970b) extend the concept of olistholith, according to the ideas of Jacobacci (1966), to cover also rock masses that are "locally independent from an olisthostrome, provided there are elements to suggest their emplacement by sliding." This is not justified, for it is entirely in disaccord with the original concept, and there already exist terms for exotic rock masses enclosed by non-olisthostromes, namely "Gleitklippe" (Görler & Reutter, 1968) and olisthothrymma (Richter & Mariolakos, 1973). Abbate el al. (1970b) also follow Jacobacci in retaining the term "pezzame", but they use this term for all clasts within an olisthostrome that are smaller than 4 m, thus defining an upper size-limit. They state that in their mapping practice in the Northern Appennines "it is usually convenient to consider any slid mass up to 200-300 m thick and 1-2 km wide to be an olistolith. Allochthonous sheets some tens of kilometres wide, have not been considered olistholiths." For intermediate sizes they cannot give any definite rule.

However, the introduction of size as a limiting criterion into the definition

of any genetic concept in geology is absolutely unacceptable; genetic concepts should be bounded by genetic criteria. Therefore the term "pezzame" must be rejected and replaced by microlistholith or endolistholith (p. 48), and the concept of olistholith must not be confined to any size-interval.

Subdivision of olistholiths – Rigo de Righi (1956) uses the term 'olistholith' to define all the diverse bodies included in the olisthostrome, which present a great variety, from entire formations or packs of strata having dimensions in the order of hundreds of thousands of cubic metres to nuclei of clay that have conserved the original sedimentary structure and texture and, finally, also to each single constituent element of the brecciated and conglomeratic clays (argille pudingoide). For a better definition of the entity of olistholiths and for greater clearness of terminology, the following prefixes are proposed by him: macro- for olistholiths in the order of size of 100 000 m^3 and more, meso- for olistholiths in the order of size of thousands of m^3, and micro- for olistholiths in the order of size up to ten m^3. The subdivision of olistholiths according to their size is a useful one when describing olisthostromes, but the subdivision of Rigo de Righi is somewhat impractical, because in most cases the volumes of olistholiths cannot be estimated from the dimensions of their exposed parts. More practical would be a subdivision according to the maximum apparent diameter. The limit volumes of Rigo de Righi correspond to spherical diameters of approximately 46.4 m and 2.1 m respectively. For practical use it is preferable to convert these limit-volumes into longest apparent diameters of respectively 50 and 5 m.

Görler & Reutter (1968) have extended the series by proposing the term megolistholiths for olistholiths having a maximum diameter between 100 and 1000 m. They also propose to call rock slabs with a maximum diameter larger than 1000 m "Gleitdecken" (Appendix 9). The latter, however, may be enclosed by or situated on top of olisthostromes as well as non-olisthostromic rocks (thus partly the same as "Gleitklippen"). Just before the publication of Görler & Reutter (1968), however, Richter (1968) has defined under the heading "tectonic features" the term "Gleitdecke" as "Schweregleitung groszer Gesteinsmassen", like the Préalpes Romandes and the Helvetic nappes. The extension of the concept of "Gleitdecken" so as to comprise also large rock slabs originating by dismembering of competent rocks intercalated with or situated upon the original clayey materials of an olisthostrome, as Görler & Reutter did, is unjustified. The need of a separate name for olistholiths with a longest apparent diameter larger than 1 kilometre may be satisfied by introducing the term 'gigantolistholiths'. The subdivision of olistholiths according to their longest apparent diameters would then become:

microlistholiths : less than 5 m,
mesolistholiths : 5 to 50 m,
macrolistholiths : 50 to 100 m,
megolistholiths : 100 to 1000 m,
gigantolistholiths : more than 1000 m.

Moreover there exist already names for olistholiths derived from rocks intercalated with or overlying the clayey sediments from which the olisthostromes originate. These are proposed by Marchetti (1957), who writes that the term 'olistholith' is proposed to indicate the blocks included in, or floating on, an olisthostrome. He subdivides the olistholiths according to their site of origin.

He is of the opinion that "these blocks may have three different origins . . . namely younger,

contemporaneous and older than the shales in which they are included or . . . floating." He makes a distinction between:
1 . . ." hard rocks sedimented on top of the shale and caused by their overburden to break into isolated blocks during gravitational movement of the olistostrome and, consequently, the resulting olistoliths are more frequent at the top of the olistostromes. These blocks . . . may be formed by hard neo-autochtonous formations and may be called neo-olistoliths."
2 . . ." hard intercalations of the shales broken into pieces during the gravitional movement, and which originated blocks mostly included in the olistostrome. They may be called meso-olistoliths."
3 . . ." the "humps" of the hard formation of the substratum originated by uplifts, thrust faults and thrust folds which were cut by the movement of the shaly body and included in the olistostrome. These olistoliths may be called hypo-olistoliths."

Borello (1970) extends this series of Marchetti by adding the term "téléolistolithe" denoting olistholiths derived from the rocks of the substratum that are much older than those which the hypo-olistholiths came from (Appendix 10). The term 'tele-olistholith' is here considered as an un-necessary extension of the already large number of prefixes for olistholiths.

The term "meso-olistolith" of Marchetti (1957), however, has been used before by Rigo de Righi (1956) in another sense. The Marchettian term should therefore be regarded as a junior homonym. It is proposed here to rename the "meso-olistoliths" of Marchetti 'endolistholiths', which are equivalents of the "native blocks" in mélanges (Hsü, 1968).

Marchetti thought of hypolistholiths as rock parts that were scraped and torn away from the substratum, but not all olistholiths constituted of rocks older than the shales in which they are embedded may have originated in this way. For instance, the original rocks of an olisthostrome may, before they slide down, already contain exotics which are older than the enclosing rocks and which may, for example, have fallen from a cliff bordering the area of origin of the olisthostrome. The resultant olisthostrome will then contain olistholiths composed of rocks that are older than the enclosing shales, but are not derived from the substratum. These olistholiths may only be detected if the source area contains or is composed of rock types different from those of the substratum, from those overlying the olisthostrome, and from those of the parent rocks of the olisthostrome. It is proposed here to call these olistholiths derived from a source outside the realm of the olisthostrome 'exolistholiths'. These exolistholiths can be considered as diverticules or klippes sédimentaires that have become involved by accident in olisthostromes. Moreover, not all olistholiths derived from the substratum have to be constituted of rocks older than the shales in which they are included. They may be contemporaneous (for instance, if the rocks of the substratum are time or time-rock equivalents of the parent rock of the olisthostrome) or even younger than the binder shales (for example: Valedolmo olisthostrome unit in Sicily: binder shales: Cenomanian to Late Eocene; underlying rock unit: Early Miocene). Consequently not all olistholiths composed of rocks younger than the enclosing shales could possibly be designated as neolistholiths.

So the Marchettian prefixes neo- and hypo- by no means always coincide with younger and older rocks, as is suggested by Marchetti (1957). He has apparently made the mistake to link a specific derivation with the age of the rock of which the olistholith is composed. It is therefore preferable to use separate prefixes to indicate the derivation and the age of the olistholiths. The prefixes caeno-, syn-, and palaeo- may be used to designate olistholiths composed of rocks which are younger, contemporaneous, or older respectively than the rocks in which they are enclosed. The age of the latter in this respect should not be the time of the

emplacement as olisthostrome, but the time of deposition before olisthostromization.

Olistholiths may be only furnished with the prefixes hypo-, neo-, exo-, and endo- when an exact investigation of the nature of the constituent rocks, comprising both organogene and inorganogene constituents, can determine these olistholiths as being derived from the substratum of the olisthostrome, from rocks overlying those of the olisthostrome before or during the redepositional process, from rocks outside the realm of the olisthostrome, or from the original rocks before their redeposition as an olisthostrome respectively. This presupposes a thorough knowledge of the rock types constituting or having constituted these possible source areas. The mere establishment of these rocks as older, contemporaneous, or younger than the enclosing shales is insufficient. Also the mere establishment of these rocks as exactly alike those that constitute the present substratum or overlying rocks is insufficient for employing these prefixes. This implies that in most olisthostromes a more or less large number of olistholiths of unknown derivation will be present.

It is therefore far more practical to use the terms 'idiolistholith' and 'anidiolistholith', terms also used by Borello (1970), which apparently denote olistholiths constituted of rocks similar to the rocks of the lithostratigraphic unit in which they became included and those constituted of rocks dissimilar to those of the enclosing lithostratigraphic unit.

So there are fifteen prefixes for olistholiths indicating the apparent longest diameter of the olistholiths, their lithologic difference with the host rock, their age in relation to that of the host rock, and their derivation:

diameter	*lithology*	*age*	*derivation*
micro-	idio-	caeno-	neo-
meso-	anidio-	syn-	endo-
macro-		palaeo-	hypo-
mega-			exo-
giganto-			teleo-(rejected here)

The concept of the term "Scherben" (Richter, 1968) appears to be a junior synonym of neolistholiths. Richter, however, considers "Scherben" as tectonic phenomena (Appendix 11). The rocks constituting neolistholiths and "Scherben" are always younger than the rocks directly surrounding them.

New definition of olistholith - Yet the olistholith concept is inexact. It should be defined in connection with the concepts of matrix and binder.

In olisthostromes of the argille brecciate type, for instance, the pelites have a brecciated structure. The small angular clasts, which are generally in mutual contact, show all stages of diminution. Their dimensions grade continually down to a relatively small amount of pasty ground-mass which is the finest disintegration product of the clasts. Rigo de Righi (1956) explicitly includes these angular clasts in his olistholith concept, but it may be doubted whether Flores (1955) has done so. Nevertheless, these small clasts, though pelitic, are indeed individual bodies of harder rock included within a pasty ground-mass.

In olisthostromes of the argille pudingoide type, on the other hand, one may find large, incompetent, pelitic inclusions which are widely dispersed in, and as soft as the visco-plastic matrix, and which can be differentiated from the latter by their distinct outline. These inclusions may or may not differ from the matrix in lithology and fossil content and may have any shape (threads, whirls, balls, ar-

moured mud balls, etc.). These isolated inclusions are also incorporated in the olistholith concept of Rigo de Righi (1956), though it may be doubted whether Flores (1955) has done so. Nevertheless, these inclusions, though soft and pelitic, are indeed rock masses enclosed as widely dispersed individual elements by the surrounding pelitic mass.

It may be argued that the essence of a matrix-olistholiths system would be that it is composed of an "enclosing" and an "enclosed" part which are mutually exclusive. The logical consequence of this would be that the latter part is separated from the enclosing part by discontinuities, and that both parts show some kind of internal continuity. For the matrix this continuity would have been effected by thorough mixing and diminution of its constituents. The olistholiths on the other hand would not have been subject to such mixing. On account of this view the matrix may be considered as the enclosing fine-grained continuum in an olisthostrome, the olistholiths as the inclusions enclosed by this continuum.

From this point of view the soft, pelitic inclusions in the argille-pudingoide-type olisthostromes that are separated from the matrix by a distinct outline should, rightly, be regarded as olistholiths. But the pelitic breccia clasts of the argille-brecciate-type olisthostrome (except, according to Jacobacci, those in whatever way liable to be differentiated from them) should, in contrast with current practice, be considered as parts of the matrix, because of the continuous gradation between these clasts and the pasty ground-mass. The opinion of Reutter (1965), Görler (1967), and Görler & Reutter (1968) to extend the scala of olistholith-dimensions to those of the smallest clasts is preferred here, because there is no reason to exclude isolated rock parts smaller than pebbles from the concept of olistholiths; the more so as in the other extreme of the scala the sizes of olistholiths are not restricted to any dimensional boundary. But on account of this also the smallest particles of the matrix would be olistholiths and in that case the system of mutually exclusive matrix and olistholiths would be fictitious. So it appears that olistholiths cannot be exactly defined in connection with the matrix of an olisthostrome; essentially this reflects the old and not yet satisfactorily solved "matrix problem" (Dott, 1964).

Moreover, it is conceivable that an olisthostrome has no matrix, that it is composed exclusively of competent and less competent olistholiths, and that it moves by virtue of the visco-plastic flowing or of the continuous cataclasis of the less competent olistholiths. Ultimately the latter will be converted into a homogeneously mixed mass, but this ultimate stage might not be reached.

On account of this it may be argued that an olisthostrome can be considered as a system composed of a phase that provides an active contribution to the movement of the olisthostrome and a phase that is only passively transported by the moving phase, or, briefly, of an "active" and a "passive" phase. The mobile rock medium that constitutes the "active" phase may well be called binder, since this term only refers to its function of holding together instead of its composition or texture. If the binder is converted into a disordered mixture of monocrystalline particles, some of the soft or friable inclusions in the olisthostrome may be separated from this binder by a transition zone in which the constituents of both binder and olistholith are mingled. The nuclei of these inclusions, irrespective as to whether they differ in lithology or fossil content from the binder or not, can be differentiated from the binder, because they have not been mixed and, consequently, have retained their original cohesion and identity (in which the original structures and textures can still be recognized). It is essentially this retention by which the

nuclei can be differentiated from the binder of which the components in this case have lost their original coherence and identity.

The binder of an olisthostrome may now be defined as the medium of mobile rock through which the movement of the olisthostrome has exclusively taken place, either by continuous and progressive cataclasis or by continued visco-plastic flow. The composing parts usually have lost their identity, the original stratification is totally obliterated, and the original sedimentary structures are entirely destroyed. Locally, however, faint relicts of the original structures may be discerned.

Olistholiths may be defined as those parts of an olisthostrome that have, at least partly, retained their coherence so that their original identity can be recognized. They are generally separated from the surrounding mobile medium by more or less sharp discontinuities in lithology, fossil content, colour, hardness, consistence, fissility, or wheathering. The competent ones are passively transported by the moving medium in which they are enclosed. Olistholiths may have suffered some attrition, brecciation, or visco-plastic deformation. They may have every possible dimension and shape, and show their original stratification, sedimentary structures and textures.

The idea should be dropped that binder and olistholiths are mutually exclusive, for their definitions are based on different genetic criteria. The binder may be constituted of olistholiths and, consequently, the breccia elements of the argille brecciate must be considered as olistholiths. The concept of olistholith, as advocated in this publication, is similar to that of Rigo de Righi (1956).

Tectonic features looking like non-stratiform olisthons and the transition of the latter to stratiform olisthons - The non-stratiform olisthons (slip sheets, slip blocks, slump packets, slump blocks, delapsional outliers, gravity collapse structures, diverticules, klippes sédimentaires, "Gleitklippen", olisthothrymmata, and many olistholiths) may at first sight look like some tectonic phenomena such as tectonis outliers, dislodged slices along thrust faults, "Gleitbretter", and "autochthone Klippen", with which they must not be confounded. "Gleitbretter" are defined by Richter (1968) as "auf Schrägflächen durch eine jüngere Schichtfolge flacher oder steiler aufwärts wanderende kompetente Gesteinsmassen verschiedener Groszordnung". "Autochthone Klippen" are defined by him as "diapirisch aufgedrungene und ohne Verbindung nach Unten abgepreszte kompetente Gesteine, die stratigraphisch älter als ihre jetzige Umgebung sind" (in order to distinguish them from "Scherben" which are in fact neolistholiths).

A link between stratiform and non-stratiform olisthons is formed by many not too incoherent slide deposits, which are commonly referred to as "slump sheets" (Kuenen, 1948). They generally form sheet-like intervals in a stratigraphic sequence, though their delapsion takes place by slipping along lubricating layers. They are a transition between slumps and very incoherent slides which will be included in the olisthostrome concept discussed next. Coherent slides may be regarded as gravity collapse structures composed of incompetent rocks, and therefore of smaller amplitude.

Stratiform olisthons

OLISTHOSTROMES

In a review of the olisthons that form strata we have to start with olisthostromes, since they show the closest relation to the previously discussed non-stratiform olisthons. We shall not repeat the definition, which is already given on page 8. Flores' obvious intention of the term olisthostrome is to give a necessary and general applicable designation of universal value for chaotic intervals within the stratigraphic column (like those that can be found in Sicily) resulting from actions of "slumping or sliding", but in such a way that it does not merely bring about "intraformational irregularities" (Flores, 1955). Unfortunately this intention is not entirely realized, because of the fact that in the definition the genesis of the deposit is linked with the description of the particular lithology of the deposit. In this chapter it will be attempted to arrive at a wholly genetic definition for olisthostromes. We would now confine ourselves to analyse the implications of some of the terms, phrases, and clauses of the original definition of Flores (1955) and discuss them.

Implications of some phrases of the original definition

1. The term: *sedimentary deposit* can, according to the ideas advocated in this publication, be changed into olisthons without further comment.

2. The phrase: *occurring within normal geologic sequences* means that an olisthostrome should be overlain by non-olisthostromic deposits. But this implies that those olisthostrome-like olisthons that terminate a normal depositional sequence would be excluded from the olisthostrome concept. According to Merla (1964a) and Abbate et al. (1970b) this limiting condition represents an imperfection of the definition. This opinion is endorsed here, for the phenomenon would be the same if the material of the slide does not become intercalated in a depositional sequence, and the circumstance that olisthostromes are not covered by non-olisthostromic deposits is merely accidental. So olisthostromes may be intercalated within as well as terminate a depositional sequence. Not so much the intercalated condition as the flattened shape of an olisthostrome, which is the implication of this intercalated condition, seems important in this phrase.

3. The phrase: *they show no true bedding* first implies that the original strata, if present, should be entirely disrupted into isolated chunks (incoherent slides). Those slid-down masses of rock that have retained their stratification, even if strongly folded and faulted (coherent slides), should be excluded from olisthostromes. Secondly it implies that no differentiation should have occurred during the movement of the olisthostrome which would result in any layering. No settling of the heavier constituents should have taken place nor any gradation in size.

4. The clause: *that were accumulated as a semifluid body* was deliberately omitted by Flores himself, when he repeated his definition in 1956, as being irrelevant with respect to the terminology. Moreover, in 1959 Flores separates olisthostromes from mud flow deposits, although both types of olisthons are often associated. However, it is not clear what Flores exactly understands by mud flow deposits.

5. The sentence: *we distinguish a "binder" . . . containing dispersed bodies of harder rock* implies that an olisthostrome should contain bodies of harder rock. But what would be the minimum volumetric percentage of these competent rock masses that should be present in an olisthostrome? Ten percent, one percent, or might it be zero percent? It would be an inconsistency to exclude from the olisthostrome concept those rock bodies that only consist of mud that apparently have suffered incoherent delapsion. For olisthostromes without indurated inclusions should theoretically be possible and should be included in the genetic olisthostrome concept. In the Moratalla area (Southeastern Spain) olisthostromes were found in which the olistholiths are so sparsely disseminated that in outcrops of several thousands of square metres not even one small olistholith was observed.

'To disperse' usually implies a wider separation of the units than 'to scatter' and a completely breaking up of the mass or assemblage (Webster's dictionary of synonyms, 1968). It also usually implies that the units have lost their mutual contacts and are widely distributed throughout the medium. Moreover, in 1959 Flores stresses the prevalence of pelitic rocks in olisthostrome masses, and also uses the phrase "diffused with no apparent order" with respect to the distribution of the blocks included in the argillaceous matrix of an olisthostrome. This means that the inclusions of hard rock in an olisthostrome should be isolated and mud-supported. This implies in turn that also the maximum volumetric percentage of these competent olistholiths is set to a limit, for the rock of an olisthostrome should just not be grain-supported. This is in contradiction with the ideas of Bucher (1959), Bortolotti (1961-1963), Passerini (1962, 1965), Boccaletti et al. (1966), Jacobacci et al. (1959), Jacobacci (1966), and Abbate et al. (1970b), who include grain-supported masses, for example the ophiolithic breccias, in their concept of olisthostromes.

This sentence also restricts the scope of olisthostromes to those olisthons in which the original stratification is in an advanced state of disruption; it excludes deposits of coherent delapsion. As all transitions exist between coherent slides and incoherent slides it may be difficult to establish whether an olisthon should be called olisthostrome or not.

6. The sentence: *there is no constant ratio between the total volume of the inclusions and that of the bedding mass* does not imply that the ratio may amount to every possible value, because at least a rather large amount of binder should be present. In this respect it should be noted that also fossils are to be considered as olistholiths and that an olisthostrome may thus be composed only of pelites with fossils.

7. The sentence: *in some olistostromes one or more types of accumulation due to flowage can be recognized, ranging from the chaotic depostion of coarser elements which were bodily detached from their original position, to "graded bedding" due to turbidity currents* has caused many controversies. But, as olisthostromes do not show true bedding, except for possible large inclusions of previously bedded materials, these bedded intercalations should not be called olisthostromes. Nevertheless Beneo (1956b) and Abbate et al. (1970b) have interpreted this sentence in such a way that turbidites were a special kind of olisthostromes. But in 1959 Flores definitely made a distinction between turbidites and olisthostromes; he never regarded turbidites as a special kind of olisthostromes, so that the extension of the olisthostrome concept of Beneo and Abbate et al. is unacceptable.

However, Flores (1959) also states that "the Tertiary sequences in Sicily are

intercalated with thick olistostromes which are to be considered, for practical purpose, as stratigraphic units." These units, depicted as intervals in the stratigraphic columns of Sicily figured by Flores (1959) and Marchetti (1957) – for example those called Valledolmo unit and Marianopoli subunit – are, it is true, predominantly composed of olisthostromes, but they also contain bedded intercalations due to turbidity currents. Marchetti (1957), for instance, reports that in a few places the rocks of the Valledolmo unit have a regular bedding which is well indicated by the regularity of the thin intercalations of fine grained sandstones. These stratigraphic units now, though not explicitly given a rank, are treated as formations and members. So it appears that Flores too is not quite consistent in the use of the term olisthostrome, because it is used by him to denote (1) a chaotic, non-bedded, heterogeneous deposit due to sliding (genetic concept), but simultaneously (2) an interval in the stratigraphic column which contains next to chaotic, non-bedded deposits also intercalations of graded bedded deposits presumably emplaced by turbidity currents (descriptive concept). This inconsistency in the use of the concept throws another light upon the idea behind this last sentence of the definition, for instead of "olisthostromes" we should read "stratigraphic units containing olisthostromes" (or wildflysch). It explains also the phrase: *sufficiently continuous to be mappable,* because this was a requirement for establishing formations. In connection with this we can better understand Flores (1955) when stating in the discussion with Beneo that "one may find in the same stratigraphic intervals all gradations of gravity deposited sediments from fine grained beds deposited by turbidity currents to conspicuous plastic masses with englobed rigid bodies".

The allegation that olisthostromes may reach a thickness of more than 1000 m (Görler & Reutter, 1968) apparently refers to intervals in the stratigraphic column in which olisthostromes predominate instead of to separate olisthostrome sheets. The requirement "sufficiently continuous to be mappable" should only be applied to lithostratigraphic units that are mainly composed of olisthostromes and be omitted from the definition of the genetic concept of olisthostrome.

So every olisthostrome "invasion" (term used by Rigo de Righi, 1956, Flores, 1959, and Görler & Reutter, 1968) consists of many separate influxes of olisthostromes, often even of different types, for argille scagliose or argille pudingoide-like materials may alternate or gradually pass into argille brecciate-type materials (Flores, 1955, Marchetti, 1957, Marianopoli subunit in Sicily; Rigo de Righi, 1956).

8. The phrase: *characterized by lithologically and/or petrographically heterogenous materials* signifies in its strictest sense that an incoherent olisthon composed of one single rock type may not be designated as olisthostrome. The sentence: *in any olistostrome we distinguish a "binder" or "matrix" represented by prevalently pelitic, heterogenous materials* implies (1) that olisthons which do not have a pelitic binder should not be designated as olisthostromes, and (2) that at least two different kinds of pelites should be present.

The fact that Flores (1955) excludes "intraformational irregularities" from olisthostromes implies that olisthostromes should be interformational. In other words, olisthostromes should contain components from at least two different formations of which one is exotic with respect to the other. Many students on olisthostromes, therefore, have interpreted the ambiguous term "heterogenous" as "containing exotics". But this requirement contains a hidden numerical quantity which must be omitted from genetic concepts. For, what percentage of exotics (or of the second formation) should be present to render the olisthostrome heterogeneous,

10%, 1%, or less? It appears unrealistic that the accidental presence of just one exotic inclusion in an incoherent olisthon would make it an olisthostrome. Therefore, and in order to make the concept a really genetic one, also those incoherent olisthons that do not contain exotics should be included in the olisthostrome concept. This, however, is an important deviation from the original intention of the concept. Jacobacci (1966) and Abbate et al. (1970b) also favour the opinion that olisthostromes may be monogenic and monolithologic.

In 1959 Flores described the Sicilian olisthostromes as chaotic accumulations consisting of "large masses of shales of different ages (Upper Cretaceous to Pliocene) with included blocks of variable size (a few cubic centimeters to several hundred cubic meters) and belonging to formations of different ages (Paleozoic to Tertiary) diffused in the argillaceous mass with no apparent order." Hence Flores has replaced the ambiguous term "heterogenous" with "of different ages". It should be noted, however, that the olisthostromes in Sicily were detected in borings by anomalies in the anticipated normal sequence of ascending ages. The circumstance, now, that the various constituents of the Sicilian olisthostromes differ in age, has been introduced as a necessary requirement for recognizing olisthostromes (Marchetti, 1957; Badoux, 1967; Görler & Reutter, 1968). Although most Sicilian olisthostromes certainly would fulfil this requirement, the introduction of it would imply an unwarranted restriction of the genetic concept of olisthostrome, for it would exclude incoherent delapsion of a rock sequence of one particular lithologic composition comprising only one biozone. This case should be included in the genetic olisthostrome concept, irrespective as to whether this rock sequence differs in age from the rocks of the receiving basin or not, because the process of olisthostromization is not affected by the ages of the rocks involved. Also Jacobacci (1966) and Abbate et al. (1970b) are of the opinion that olisthostromes may be monogenic and may originate almost contemporaneously with the receiving deposits. Consequently, neither the allegation that olisthostromes should have originated from, or should dominantly be composed of, formations which are older than those upon which it slides (Marchetti, 1957; Badoux, 1967; Görler & Reutter, 1968) is warranted. Differences in age or derivation of the rocks, which the olisthostrome is dominantly composed of, should at best find expression by appropriate prefixes.

In this respect Elter & Raggi (1965) prefer, when working in the Northern Appennines, to distinguish two types of olisthostromes which differ from each other either by the provenance of the materials or by their palaeogeographic significance. They define the first type as characterized by the fact that the materials, dislocated by sliding, belong to the same succession of formations as the one in which they are included. The second type would be characterized by the presence of materials belonging to a different succession of formations that can be referred to a separate basin of sedimentation. The first type, which they propose to call endolisthostromes, has been interpreted as due to precocious tectonic movements that have taken place in the internal part of a sedimentary basin with consequent sliding. It should be noted that endolisthostromes may contain many exotics, as defined in this paper. The second type, which they call allolisthostromes, has been interpreted as the result of slides and slabs of material detached from the front of a moving allochthonous unit and slid down into the adjacent basin of sedimentation. The latter olisthostromes would characterize the emplacement of nappes that has taken place in the sea during the course of sedimentation and are immediately preceding the very emplacement of these nappes. Finally they mention the possi-

bility of distinguishing a third type which would be constituted of olisthostromes intercalated in post-orogenic formations (semi-allochthonous or neo-autochthonous) and which would correspond to late orogenic movements that took place while the molassic formations were deposited at the conclusion of the orogenic cycle.

In this connection the prefixes endo- and allo- refer to a difference in palaeogeography. Consequently these prefixes can only be applied when the palaegeographic history of the region has been established. The distinction stands or falls with the recognition of the limits of the various "sedimentary basins"; these vary with time. If an area is uplifted and becomes denuded by delapsion, it ceases to be a basin of sedimentation. It depends on the angle from which the matter is looked at, whether the rocks of the uplifted area are referred to "a separate basin" of sedimentation or not. Nevertheless there remains the undeniable distinction between two main types of olisthostromes.

Therefore it is preferable to redefine endolisthostrome as an olisthostrome resulting from the incoherent delapsion of the rocks of the same lithosome as the one in which it is intercalated, or which it directly overlies; exotic rock fragments may be admixed with it. An allolisthostrome – which Broquet (1970) called exolisthostrome – may be defined as an olisthostrome resulting from the incoherent delapsion of the rocks of another lithosome than the one in which it is intercalated, or which it lies on. Of course all transitions can be found between the two types. Transitional cases are, for instance: (1) the olisthostrome resulting from the incoherent delapsion of the rocks of the same lithosome as the one in which it is intercalated, but only in a slightly different development, or (2) when two or more lithosomes in superposition become involved in one delapsional event.

Genetic definition of olisthostrome - The wording of the original definition of the genetic concept of olisthostrome (page 8) appears to be an unfortunate one because it links a certain mode of deposition (genetic criteria) with a particular lithologic constitution (descriptive criteria), and because it contains hidden numerical quantities, which should be omitted from any definition of genetic concepts.

It would be better to define an olisthostrome as *an olisthon resulting from the incoherent delapsion of cohesive rocks of such bulk consistence (1) that they were able to form a mobile medium through which the delapsional movement exclusively took place (binder), and (2) that, if present, the elements that have retained, at least partly, their coherence so that their original identity can be recognized (olistholiths) could not become subject to any kind of mechanical differentiation that would give rise to internal layering or grading.*

All transitions exist from olisthostromes to 1) incoherent delapsion deposits that do show the result of mechanical differentiation of the inclusions into an internal layering or grading (mud flow deposits pro parte), to 2) muddy olisthons in which the hard inclusions are so numerous that the movement of the system does not take place exclusively through the binder, because dispersive grain pressure is becoming active (muddy grain avalanche deposits), to 3) coherent slides, and to 4) diverticules and klippes sédimentaires.

Thus all incoherent slides may also be called olisthostromes, but during the delapsion of some olisthostromes the thixotropic mud may have been in a fairly fluid state, so that the result may be equally well designated as mud flow deposit. There exist, however, mud flow deposits in which the consistence of the mud apparently was so low that many of the inclusions would sink to the base where they may

become concentrated, while in the upper part a smaller or greater number of inclusions may remain floating in the mud. The basal conglomeratic layer may or may not be graded in a normal, reversed, or even recurrent fashion. As the inclusions in such mud flow deposits had been subject to some kind of mechanical differentiation which gave rise to internal layering, they should not be designated as olisthostromes. So the concepts of olisthostrome and mud flow deposit partly overlap each other. The term "differentiated mud flow deposit" is now introduced here to cover the mud flow deposits that are not olisthostromes. They may be defined as olisthons resulting from the incoherent delapsion of thixotropic rocks of such low bulk consistence that a secondary differentiation of inclusions may have taken place. The easy distinction between incoherent delapsion without and with internal differentiation of the inclusions is far more practical than the distinction based on the presumed viscosity of the mud during delapsion, like mud flow deposit and rheopelite (Elliston, 1968). Examples of mud flow deposits with a conglomeratic base are some of the pebbly mudstone beds within the Lago Sofía conglomerate in Chili (Dott, 1963; Scott, 1966), the "olisthostrome" at Bürnük in Turkey (Sestini & Canuti, 1968), and some of the conglomeratic marlstone beds near Moratalla in Spain (this article). However, it should be kept in mind that if the inclusions are small enough no sinking will take place, even at high fluidity of the enclosing mud, and that according to the genetic definition of olisthostrome such an olisthon should be designated as olisthostrome.

In an olisthostrome a large amount of binder should at least be present, for if there would not be a minimum percentage for the cohesive binder, then sand avalanche deposits, gravel avalanche deposits, rubble avalanche deposits, and "fluxoturbidites" would be included in the olisthostrome concept, which would then almost cover the olisthostrome concept of Jacobacci (1966). Even turbidites could then possibly be regarded as olisthostromes and the olisthostrome concept would then be equivalent to that advocated by Beneo (1956a, b) and all stratiform olisthons would then be olisthostromes. This extension of the scope of olisthostromes is unwarranted, as it is in contradiction to the original intention and idea of Flores (1955, 1959), who explicitly excludes turbidites, mud flow deposits, and other internally layered olisthons from his olisthostrome concept. The minimum amount of binder can however be defined without inserting numerical quantities into the definition by the requirement that the delapsional movement should exclusively take place through the binder. For this implies that no collisions between hard inclusions should occur that would complicate the mechanics of the movement. If the hard inclusions are so numerous that during the movement the effect of mutual collisions cannot be ignored anymore, we enter the field of grain avalanching. In practice this means that the competent olistholiths in an olisthostrome should be mud-supported, or better, entirely surrounded by the binder. Though the concepts of binder and olistholiths are not mutually exclusive, those of binder and competent olistholiths are so.

Olisthostrome concepts of other investigators - Beneo (1956) has availed himself gratefully of the new term 'olisthostrome' in order to institute a separate designation for the independent series of delapsional phenomena. But as a result he has extended the concept of olisthostrome unwarrantedly to all olisthons that are commonly studied by sedimentologists and has placed the term in synonymy with "resedimentation". He defines olisthostromes as those chaotic sedimentary accumulations that are intercalated with sequences of normal deposits. He includes turbi-

dites in his olisthostrome concept. His view is not accepted here, because it largely exceeds the original intention of Flores (1955).

Facca (1956) comprises in the olisthostrome concept the same phenomena as Beneo has done, but according to him the concept should be accepted in a different sense. It would not imply any stratigraphic concept, but would designate only a particular mode of emplacement. This view is quite acceptable. According to him an olisthostrome should be defined as a plastic series in anomalous stratigraphic position by virtue of gravitative movements that have taken place after its deposition. He is of the opinion that the concept of olisthostrome is incomplete, if we do not incorporate in its ranks three particular modes of "sedimentation": orogenic landslides, resedimentation, and turbidity currents. In contrast with Beneo, who classifies these delapsional processes under sedimentation, Facca considers these processes as an extension of gravitational tectonics, although he is well aware of their intermediate position between tectonization and sedimentation.

Marchetti (1957) (Appendix 12) extends the concept of olisthostrome to include subaerial deposits of the same type and genesis. This opinion is endorsed here, since difference in medium does not fundamentally affect the concept. Also Ogniben (1954b) and Jacobacci (1966) do not exclude a subaerial start of some olisthostromes of the argille brecciate type. In 1955 Flores did not say anything about olisthostromes being subaqueous or not, but in 1959 he wrote that "the evidence collected indicates that the olistostrome emplacement must have occurred in an aqueous medium as shown by the associated turbidity current and mud flow depositional phenomena observed. The sediments involved in an olistostrome could hardly have had, in a subaerial medium, the necessary buoyance and fluidity to be widespread by sliding over such large areas." Of course subaerial incoherent slides generally are less extensive, but that does not prevent them from being olisthostromes. On the other hand, Marchetti restricts the concept to those delapsed rock bodies derived from formations „which are older than those were they slide on". This is an unjustifiable modification, because it is based on a non-genetic criterion. In many cases olisthostromes do contain rock masses which are older than those of the substratum, but this does not imply that all parent rocks have to be older than the immediately underlying rocks.

Also according to Badoux (1967) and Görler & Reutter (1968) olisthostromes should predominantly be composed of exotic rocks of older age than the directly underlying rocks. According to them, non-exotic rocks, if present, would have got mixed in only during their downslope movement. These authors, therefore, erroneously distinguish olisthostromes from those pebbly mudstone deposits as described by Crowell (1957), which consist predominantly of material of the underlying rocks, though they may contain exotic rocks (endolisthostromes). They introduce again non-genetic criteria and almost restrict their olisthostrome concept to the allolisthostromes of Elter & Raggi (1965). Also the binder shales of the Marianopoli subunit in Sicily are "over large areas formed by brown shales, very similar to the underlying . . . units" (Marchetti, 1957).

Reutter (1965), Görler (1967), and Görler & Reutter (1968) consider olisthostromes as the products of "mud flows" or "mud streams" sensu Dott (1963). This would imply that the terms 'olisthostrome' and 'mud flow deposit' would be synonymous. But this is not in accordance with the terminology of Dott, as they suggest, because Dott restricts the terms 'mud flow' and 'mud stream' to the flowing of thixotropic material in a state near that of its liquid limit, whereas the scaly clays, which the binder of many olisthostromes is composed of, are not supposed to have

been formed in this way. It is assumed that they were formed by gravitational displacement along innumerable internal slip-planes along which clay flakes become oriented in a parallel manner giving rise to lustrous scales (cf. Ogniben, 1953, 1954a, 1954b). Moreover, the olisthostromes of the argille brecciate type have moved down by the continuous and progressive brecciation of the pelitic rocks. It should be borne in mind, however, that an argille scagliose-like texture may also be produced when a particular mud flow becomes tectonized after deposition and consolidation. The olisthostromes described by Görler and Reutter presumably have flowed in a rather stiff visco-plastic way, but that does not imply that every olisthostrome should have been emplaced in the same way. The concept of olisthostrome should not primarily be restricted to a particular consistence of its constituent rocks.

Richter (1968) makes a distinction between tectonic ("Gleitdecken") and sedimentary phenomena (olisthostromes). On the other hand, he unites olisthostromes, olistholiths, and "Gleitdecken" as "synorogene Resedimentation groszer Umfangs und in geschlossenen Gesteinsverbanden." The restriction of olisthostromes only to the results of slow submarine slidings, as Richter does (Appendix 13), is not justified, because it introduces a numerical quantity into a genetic concept.

Jacobacci (1966) writes about olisthostromes that they are deposits or accumulations of materials derived from sliding (or mud flowing). According to him this would be, though due to various causes, a sedimentary phenomenon and the results should therefore be intercalated, at least in some cases, within the ambit of a region or sedimentation area, with sediments of normal deposition or with turbidites, in order not to be confounded with the results of other phenomena (here Jacobacci probably meant the allochthonous cover of the Northern Appennines, which according to him would be tectonic by definition). His opinion that the entire allochthonous cover of the Northern Appennines should be excluded from the olisthostrome concept is not accepted here, for both the allochthonous cover and the olistostromes are deposits and it seems likely that the allochthonous complex consists of diverticules (many of the giant rock slabs) as well as olisthostromes (the so-called "chaotic complex"); diverticulation and olisthostromization gradually pass into each other. We should not lump all parts of the allochthonous cover together ignoring the distinctions. According to Jacobacci the body of every separate slide (or mud flow) should not show any stratification nor, of necessity, should it be delimited by subparallel surfaces. He is the first person who argues that an olisthostrome may be polygenic as well as monogenic and also may originate almost contemporaneously with the sediment which it contains. In this respect I fully agree with him. But Jacobacci does not consider the requirement of the deposit being prevalently pelitic as diagnostic for an olisthostrome. This opinion is not desirable, because it implies that fluxoturbidites and the deposits from sand and gravel avalanches would be included in the olisthostrome concept. In fact Jacobacci follows Beneo's ideas about olisthostromes; he merely separates the two extremes, the "argille scagliose" of the Northern Appennines and the turbidites, from the series of phenomena that Beneo (1956) incorporates in his olisthostrome concept, retaining the middle portion as a unity within the concept.

De Raaf (1968), on the contrary, proposes to restrict the term 'olisthostrome' to the gigantic subaqueous slide deposits like those known from Italy, whereas the smaller slide deposits of similar character are incorporated in his olisthanite concept (page 63).

Broquet (1970), also apparently influenced by the current conceptions of tectonization and sedimentation, is of the opinion that the olisthostrome concept must by no means be substituted by that of "nappe de glissement" and that it should be applied to particular "sedimentary" phenomena. He stresses that any intervention of tectonic nature should be excluded from the emplacement of an olisthostrome. He is right if he has the same idea about tectonics as that advocated in this paper, and if "sedimentary" can be replaced by "depositional".

Abbate et al. (1970b) incline, on the one hand, towards a genetic-sedimentological connotation of the term 'olisthostrome', and they reject the requirement that an olisthostrome should be so large that it can be mapped on a topographic map of 1/25 000 scale. On the other hand, when trying to delimit sedimentation from tectonization, they introduce size as a distinguishing criterion. They usually regard as olisthostromes chaotic units up to 100-200 m thick, but they think that the size criterion must be integrated with the analysis of the internal structure. They write that "when the bulk of the unit is composed of finely disrupted material, suggesting movement in a semifluid state, the upper size limit for olistostromes may rise considerably. On the contrary, when the structure is relatively undisturbed, or very large disrupted masses are piled up," they favour a tectonic interpretation, even for units of smaller thickness. However, the boundaries between genetic concepts, such as that of olisthostromes and „tectonic phenomena", must never be based on size or texture, but only on genetic criteria. Moreover, most of their "tectonic rock masses" may be olisthons (olisthothrymmata, diverticules or klippes sédimentaires).

OTHER STRATIFORM OLISTHONS

Herpoliths - Almost at the same time that the olisthostrome concept was introduced, Lotze proposed the term 'herpolith' from Greek: $\varepsilon\rho\pi\omega$ = I creep, and $\lambda\iota\vartheta o\sigma$ = rock) for extensive sheet-like subaqueous slide masses, originating from unindurated, thixotropically behaving deposits, and intercalated in normal stratigraphic rock sequences. Excluded explicitly from herpoliths are the deposits from "Schuttströmen" and of thixotropically deformed material, and the slides of indurated rock masses. Lotze assumes earthquake shocks to be the cause of the slidings. The term 'herpolith' was published for the first time in May, 1956, in an abstract of a lecture which Lotze held on the 17th of January, 1955. For the second time this term was published by Lotze in 1957, which represents the unabridged text of a lecture held on the 9th of December 1953, thus predating the lecture referred to in the first publication of the term 'herpolith' and even that of 'olisthostrome' (Appendix 14). The term has further been used by Voigt (1962), Voigt & Häntzschel (1964), Arnold (1964), and Richter (1965).

According to the figures given by Lotze (1957) of the herpolith occurrence in the Baumbergen (Westphalen, Germany) and according to the example given by Lotze (1957) of the disharmonically folded rocks at the Devonian-Carboniferous boundary in Sauerland, Germany (Lotze, 1951; Päckelmann & Richter, 1938) it can be concluded that slid-down rock masses that have retained their stratification, though more or less contorted, should be included in the herpolith concept. On the other hand, the examples of herpoliths given by Lotze in the Beckumer Bergen (Westphalen, Germany), the so-called "Bärsteine" (Giers, 1958), have totally lost their original stratification. The binder and the far greater majority of the harder blocks enclosed by the binder are derived from one and the same formation iden-

tical to the directly underlying and overlying rocks; exotic inclusions have been found only sporadically. According to Lotze (1957) an example of herpolith occurs at the boundary between Ordovician and Silurian rocks in middle Spain. Probably the so-called Orea Slates from the Sierra Morena are meant, which contain many exotic inclusions (Henke, 1926; Simon, 1951; Greiling, 1967). Only later (Voigt & Häntzschel, 1964) were the so-called "Flockenmergel" explained as the distal extensions of the "Bärsteine". In many of these "Flockenmergel" beds a concentration of "Flocken" (white limestone inclusions in the grey marlstones) can be found at the base, so that they may be designated as differentiated mud flow deposits. It may be doubted whether the "Flockenmergel" should be incorporated in the herpolith concept. So herpoliths, which are generally studied by sedimentologists, comprise at least coherent and incoherent slides, irrespective as to whether these deposits contain exotics or not. So olisthostromes should be considered as special types of herpoliths.

Richter (1965) describes from the West-Pyrenees slid-down masses which he designates as herpoliths and which may be the same as mentioned by Lotze (1957). Also these masses are derived from the same material as that occurring in the directly underlying and overlying rocks. Richter, now, even goes so far as to restrict herpoliths to those olisthons of thixotropic rocks that are chiefly characterized by the fact that they are under- and overlain by normally and parallel stratified sediments, and that they consist of the same material as the under- and overlying layers. Moreover, Richter separates "subaqueous slide structures" from herpoliths, because he supposes the former not to be the result of earthquake shocks (note on page 183). This distinction is incorrect and highly impractical, first because it is wholly in contradiction with the concept of Lotze (1956, 1957) who places 'subaqueous slide masses of thixotropic rocks' and 'herpoliths' in synonymy, and secondly because it cannot be settled whether the sliding was caused by gradual tilting, by earthquake shocks, or by whatever other occasion. Also the rocks constituting the herpoliths described by Voigt (1962) consist of the same material as those of the enclosing formation. Neither this nor the fact that they are intercalated within other deposits should become necessary requirements in the genetic concept of herpolith.

Slide conglomerates - This term (Hadding, 1931) originally refers to a sheet-like rock mass enclosing more or less distinctly tabular pebbles which as a rule occupy every imaginable position (edgewise conglomerate) and which may or may not be in mutual contact. According to Hadding it is formed by incoherent delapsion of a stratified deposit of which only some of the strata were consolidated before sliding. As both blocks and binder are derived from the same parent rock mass, they must have the same age, but this does not make them necessarily intraformational, as Hadding thought, though they often are. Also the blocks should not necessarily be of tabular shape. According to Hadding intraformational slide conglomerates could be distinguished from other intraformational conglomerates in that they may merge laterally into sharply folded and much contorted strata, that they would be relatively widespread, that they would have a muddy binder, and that they would be associated with other olisthons. Obviously olisthostromes of the argille brecciate type can also be designated as slide conglomerates. Also soft calcilutites may be brecciated and may slide down in a similar way as the clays of the argille brecciate. If it can be shown that the brecciation and sliding of the calcilutites took place before lithification, the resulting deposit can also be called an olisthostrome. Also

the slide conglomerates in which the indurated fragments are mud-supported are special types of olisthostromes. However, slide conglomerates in which the fragments, indurated before sliding, are grain-supported should be incorporated in the concept of gravel avalanche deposits.

The concept of slide conglomerate should remain restricted to conglomerates whose texture is *formed by sliding*. Kuenen & Carozzi (1953), Kuenen & Sanders (1956), Stanley (1961), and Walker (1967), however, have used this term to designate conglomeratic material *that have slid down*. The use of the term 'slide conglomerate' in the latter sense should be rejected, because it is a junior homonym. It should be replaced by gravel avalanche deposit.

Fluxoturbidites versus proximal turbidites - The concept of fluxoturbidite, first used by Kuenen (1958), has been defined by Dzulynski et al. (1959) (Appendix 15). They used the term 'fluxoturbidite' for gravitationally emplaced sandstone deposits with a certain combination of observable lithologic characteristics; thus again the unacceptable combination of genetic and descriptive criteria. But the description is not sufficient to identify each bed unambiguously as a fluxoturbidite. Moreover, the term is unfortunately chosen, first because it is the combination of two genetic terms that cannot be combined, and secondly because they have probably not been deposited by true turbidity currents. Finally it is not clear to what extent proximal turbidites (Walker, 1967) should be incorporated in the concept of fluxoturbidite and vice versa.

Walker (1967) considers proximal turbidites as deposits from turbidity currents that initially flow down-slope with all of its sediment homogeneously mixed in suspension. Fluxoturbidites are not considered by him as deposits from suspension currents, but, like Dzulynski et al. (1959), as deposited by delapsional processes in which most of the sand and gravel moves in a watery slide along the base. Kuenen (1969) is of the opinion that fluxoturbidites are "beds that were presumably deposited by a turbidity current unable to carry all sediment in suspension, so that a watery slide produces a stratum showing less obvious grading or sole marking, but plenty of shale pebbles and also coarser grains than the accompanying turbidites." As many deposits comprised in fluxoturbidites and proximal turbidites are "apparently similar" (Walker, 1967), it may be inferred that in these cases the mechanism of transport of these deposits may be more or less similar too, whatever they may be.

De Raaf (1968) restricts the concept of fluxoturbidite to those deposits that are partly reminiscent of a glided mass and partly of a turbidite, such as a thick mass of heterogeneous sand that passes upward into a thin, graded and laminated layer; but also a layer that starts with a coarse-grained and often gravelly base, whereas the main mass assumes the aspect of a normal turbidite, graded and possibly laminated. But Walker would call such deposits proximal turbidites, which comprise delapsionally emplaced sand beds with a certain combination of characteristics close to that of fluxoturbidites, but to which the numerical techniques of turbidite facies analysis (based on the Bouma-model) still can be applied. The latter aspect is used by Walker (1970) as the distinction between fluxoturbidites and proximal turbidites (although often quoted as if he had replaced the term 'fluxoturbidite' by 'proximal turbidite'). But this is a non-genetic distinction. The term 'proximal turbidite' should be only applied when it can be convincingly argued that it has been deposited by a turbidity current. Characteristic structures and textures only, such as the Bouma-divisions, do not suffice, for there may exist depo-

sits of turbidity currents that do not show these characteristics and that consequently have not yet been recognized as turbidites. Conversely there are sand deposits to which the numerical techniques of turbidite facies analysis can be applied, but which may not be turbidites (de Raaf, 1964).

Stauffer (1967b) has given some characteristics of supposed sand avalanche deposits (he uses the term grain flow deposits). Many characteristics appear to be similar to those of fluxoturbidites and proximal turbidites, but this does not imply that all fluxoturbidites and proximal turbidites are sand avalanche deposits. Supposed sand avalanche deposits have also been described by Imbrie & Buchanan (1965) and Sanders (1965). The concept of fluxoturbidite has been extended to slid-down conglomeratic sandstones and conglomerates (Unrug, 1963; Stanley, 1961), but have always been distinguished from slid-down conglomeratic mudstones. Of course all transitions exist between sand avalanche deposits and gravel avalanche deposits.

Rightly Walker (1967) states that the term 'fluxoturbidite' should be abandoned as a genetic term. It may be replaced by several other genetic terms indicating the various supposed processes of delapsional emplacement of these deposits, for instance 'sand avalanche deposits', or 'grain avalanche deposit', or by terms indicating all the possible, yet unknown, processes responsible for the remaining olisthons composed of non-cohesive grains and that are not turbidites. It would be far preferable, however, only to describe the deposits and after that cautiously add the supposed depositional mechanism.

According to Stanley & Unrug (1972) many deposits described in the literature as fluxoturbidites appear to be lenticular or tongue-like. This geometry and the assemblage of structures displayed by these deposits suggest a submarine channel origin. In order to "minimize to some degree the objectional mixing of descriptive characteristics and genetic (process) implications [. .] that admittedly are still poorly understood", Stanley & Unrug (1972) propose that "thick coarse tongues of this type on outer submarine margins be referred to as submarine valley or submarine channel deposits", terms that emphasize "the descriptive (mappable) aspect of these deposits rather than the transport mechanism." However, on the one hand they do not imply that all units described as fluxoturbidites are so clearly channelized, on the other hand many submarine channel deposits contain slump masses, pebbly mudstones, and even sand and finer-grained material that moved downslope by traction or by suspension displacement and that are not fluxoturbidites. Therefore, 'submarine channel deposits' should be regarded as a kind of facies term, which cannot be used as a descriptive equivalent of fluxoturbidites.

Olisthanites - In order to get around the difficulties of establishing the exact genesis of many olisthons, de Raaf (1968) (Appendix 16) has felt the necessity of introducing the genetic term 'olisthanite', comprising all subaqueous olisthons that are generally studied by sedimentologists (with the exception of rock fall deposits) that, while delapsing, do not form a strongly turbulent suspension (such as proximal and distal turbidites). It should be taken into account that his concept of fluxoturbidites comprises deposits that Walker (1967, 1970) would call proximal turbidites and which de Raaf therefore excludes from olisthanites. Fluxoturbidites sensu Walker are included in the olisthanite concept. De Raaf also proposes to restrict the term olisthostrome to the gigantic olisthanitic deposits like those well known from Italy. Also these deposits, apparently because of their large size, are separated from his

olisthanites. The term olisthanite has subsequently been used by Stanley (1970) and Stanley & Unrug (1972).

Non-genetic rock names applicable to olisthons

Before discussing the implications of the purely descriptive names applicable to both olisthons and sediments, we will discuss wildflysch, one of those terms which originally had a purely descriptive meaning (as petrographic name), but to which subsequently a genetic implication has been added, as with argille scagliose, mélange, and flysch. The terms argille scagliose and mélange have already been discussed in a previous chapter, but it is beyond the scope of the present publication to discuss the controversies around the term 'flysch'. For discussions of the flysch concept the reader is referred to Tercier (1947), Kuenen (1951), Trümpy (1961), Dzulynski & Walton (1965), de Raaf (1968), and Hsü (1970).

Wildflysch

The name 'wildflysch' is introduced by Kaufmann in a letter to Studer in 1870, of which Cadisch (1953) gives the following quotation: "Dunkelgrauer bis schwärzlicher, weicher Schiefer, dünngeschichtet bis blättrig, oft krummschalig mit glänzenden Flächen, auch wohl von weiszen Spatadern durchzogen. Meist sehr unregelmäszig gelagert, Blöcke oder harte Knauer umwickelnd, abwechselnd mit Fucoïdenschiefer und Macigno. Mit vielen Mulden, Sätteln und Krümmungen aller Art." In 1872 Studer for the first time published this petrographic term and its denotation.

The concept of wildflysch and its supposed origin have, like that of argille scagliose, suffered many changes. Kaufmann (1871), for example, thought that the rocks which he named wildflysch were the parent rocks of neptunistically formed granitic blocks and that the attendant molecular mass movements had caused the stratal disturbances. Beck (1911, 1912) considers wildflysch as the product of tectonic remodeling of flysch shales. Arnold Heim (1908) was the first who did not exclude the possibility that wildflysch deposits had originated by submarine sliding. Lugeon (1916) was the first to consider the exotic blocks in the wildflysch as transported by sliding along a sloping, muddy marine bottom not far from the coast of origin. Since the nineteen twenties wildflysch was generally assumed to be largely composed of submarine slides (Arbenz, 1919; Häfner, 1924; Kraus, 1932; Tercier, 1928, 1945, 1948; Soder, 1949; Gabus, 1958; Trümpy, 1960; Herb, 1962; Caron, 1965, 1966; Badoux, 1967; Leupold, 1942).

The chaotic structure of wildflysch has been attributed, partially to sliding, partially to posterior tectonic deformation (Arbenz, 1919; Leupold, 1942; Soder, 1949; Herb, 1962; Badoux, 1967; de Raaf, 1968). The largest inclusions were still regarded as tectonic slabs (Häfner, 1924; Tercier, 1928, 1945, 1948; Cadisch, 1953; Gabus, 1958; Caron, 1965, 1966). But others, for example Soder (1949), Trümpy (1960), and Badoux (1967) regard even these very large inclusions as slipped masses, which they call diverticules. Trümpy (1960), Sames (1965), Richter (1968), Caron (1966), Lomize (1966), Görler (1967), Reutter (1965), and Görler & Reutter (1968) are of the opinion that wildflysch deposits are olisthostromes. Badoux (1967), however, separates them from olisthostromes on the ground that such rocks should by definition only be called wildflysch, if they had been subject

to subsequent tectonization, whereas olisthostromes would not meet this requirement. For the same reason he criticizes the much quoted definition of wildflysch of Cadisch (1953), which runs; "Eine kretazische bis tertiäre Gesteinsvergesellschaftung orogener Facies, die aus einem mergelig bis tonigen Grundgestein besteht, welchem meist Bänke kalkigen, sandigen oder quarzitischen Materiale, sowie Gerölle verschiedener Art eingelagert sind." But a separation of wildflysch and olisthostromes on these grounds seems unrealistic. The only grounds for separation are the connotations of the terms being non-genetic and genetic respectively. Wildflysch deposits are now mostly regarded as a particular facies of the alpine flysch (Cadisch, 1953; Trümpy, 1960; Herb, 1962; Caron, 1966) which contains olisthostromes, but there is a tendency to use the term to all consanguineous rock-associations comparable, also in geological setting, with the flysch-wildflysch suite in the Alps. Leupold (1966), however, favours the opinion to use the designation 'wildflysch' exclusively for the intervals of black, scaly, crumpled-bedded marly clays in a flysch sequence, but in that case the term conglomeratic shale seems preferable.

Excellent descriptions of wildflysch deposits in Switzerland are given by Soder (1949) and Beck (1911) in the Habkern area, by Tercier (1928) in the Berra area, and by Heim (1910) and Herb (1962) in the Amden area. It appears that horizons of conglomeratic mudstones alternate with horizons of non-conglomeratic-mudstones (among which pelagic sediments) and that commonly intercalations of conglomerates and breccias (probably grain avalanche deposits) occur. The generally scanty exposures in wildflysch areas have for a long time severely hampered the understanding of the essence of wildflysch deposits.

With respect to the genesis of wildflysch Kraus (1932), Tercier (1945, 1948), and Soder (1949) think of emerged or submerged ridges with steep submarine slopes constituted of various rock types. These ridges became subject to very intensive erosion. The very coarse to very fine erosion products accumulated along the shore, along which there might have been a narrow platform. On these platforms sublittoral conditions (limestones with *Lithothamnium* and (or) *Nummulites,* sandstones with oysters etc.) prevailed. Steepening of the slopes owing to accumulation on the narrow platforms or tectonic movements caused the accumulated erosion products to slide down into deeper water. Caron (1966) thinks of a „démantèlement du front d'une nappe en cours de déplacement sur un fond marin."

Mandelschiefer - The term „Mandelschiefer" has been proposed by Seidlitz (1906) for a rock type that occurs in the Madrisjochgrat in Switzerland. He describes it as follows: "In einer schiefrigen oder sandigen Grundmasse schwimmen, in groszen Abständen von einander – selten dicht gedrängt – kuglige bis mandelförmige Brocken von triassischen Kalken und Dolomiten, seltener von Malm und kristallinen Gesteinen. Oftmals sind die Zwischenräume noch mit kleineren meist ähnlichen Komponenten erfüllt, gewöhnlich aber ist die Grundmasse ein homogener dunkler Schiefer." Häfner (1924) also used this term for the same lithostratigraphic unit. Dimensions of the components: few centimetres up to a few decimetres. Thickness of the bed: 7 m.

Argille brecciate – The designation 'argille brecciate' ('brecce argillose', Jacobacci, 1953) has been given to sheet-like intercalations of clays with a brecciated structure in the Tertiary rocks in Sicily. Their thickness varies from a few metres to some hundreds of metres (?). The various clayey and marly elements, varying in

dimensions from fractions of millimetres up to a few decimetres, are immersed in a ground mass consisting of a dark clayey paste with some sand-sized grains. Next to these clayey elements frequently lithic inclusions up to a few decimetres in cross section can be found, whereas less frequently large inclusions of several metres occur, and occasionally inclusions of stratified marls reaching up to hundreds of cubic metres may be encountered. The elements are angular to subangular. Intercalations of clays with a fluidal structure or of clays having retained their stratification, are occasionally found in the midst of the brecciated formation. Generally the argille brecciate are micro-paleontologically characterized by the presence of a rich microfauna. Mixed faunas due to reworking are only found in the fine ground mass. Generally the oldest elements are Cretaceous and the most recent ones have the age of the stratified formation on which the argille brecciate rest (Ogniben, 1953, 1954a, b; Rigo de Righi, 1953, 1956; Beneo, 1955; Marchetti, 1957; Flores 1959). A comparable structure is shown by the so-called clay pebble bed of Ancon (Ecuador) (Barrington Brown & Baldry, 1925; Sheppard, 1927; Busk, 1931; Sheppard & Bushnell, 1932; Baldry, 1932; Barrington Brown, 1938; Baldry, 1938; Marchant & Black, 1960; Marchant, 1965; Audley-Charles, 1965b).

According to Rigo de Righi and Ogniben the brecciated structure of the argille brecciate could be due to superficial disintegration of clay or clayey marl by the action of water, and to a relatively short superficial gravitational transport attended by very rapid deposition, so that a more complete disintegration is prevented. This action of water could have taken place in a submarine as well as in a subaerial environment.

Morris' disturbed bedding types - Morris (1971) has made an attempt to describe and classify various disturbed bedding types which he has found in the Ouachita Mountains, U.S.A. His classification, which is not intended to be world-wide and all-inclusive, is based (1) on the degree of disturbance of the original rocks which is thought to be a qualitative measure for the distance of delapsive movement, and (2) on the kind and state of rock involved. The terms given to the distinct, clear-cut end-members of the otherwise gradational bedding types are meant to be descriptive, though in this respect not always luckily chosen. It appears logical that, if the disturbance is such that the original rock units are mixed so thoroughly that they are homogenized into a new rock type, the ultimate deposit should be given one of the available descriptive sediment-petrographic names, instead of introducing another "descriptive" term for the same rock type only because it has been delapsed instead of sedimented. The terms "blocky mudstone", "rubble bedding", and "breccia" used by Morris, are therefore considered superfluous and may be replaced for instance by massive mudstone, conglomeratic mudstone, and breccial sandstone respectively. The other bedding types disturbed by delapsion recognized by Morris are: flowage shales, contorted bedding, chaotic bedding (Walthall, 1967), and slurried bedding (Wood & Smith, 1959).

Flowage shales are olisthons that are defined by Morris as consisting of "visibly folded, buckled, squeezed, ruptured, and sheared mudrock in which original bedding planes are still visible. Fold axes are chaotically arranged." Perhaps 'crumpled shales' would be a better descriptive term. Morris uses the term 'contorted bedding' for olisthons of interbedded sandstones and shales in which the competent sandstone beds generally have failed by folding with minor fracturing and shearing ("soft sediment deformation") and are rarely pulled apart to the ex-

tent that one cannot observe how the separated ends originally fitted together. Of course the term can equally be used for any other contorted alternation of competent and incompetent beds. Chaotic bedding refers to a more disrupted variety of contorted bedding and can be described as a mixture of elongated, angular, cohesive sandstone blocks in which the two long axes are oriented parallel to the bedding and each block is completely separated by attenuated shales or mudstones that have obviously undergone considerable shearing or failure. Slurried bedding is used by Morris in the sense of a general term for all disturbed masses that consist of a mixture of friable sand and varying amounts of mud. Slurried bedding ranges in appearance depending on the amount of mud within the olisthon as well as the degree of disturbance and mixing of sand and mud and the degree of desaggregation of the sand. At one end it may have the appearance of some sand blocks floating within more mixed aggregates of friable sand and mud, at the other it may have the appearance of "a jumbled, hummocky mass of sandstone blocks, friable sand, and scattered streaks, blocks, and pods of shale." An all-inclusive really descriptive term is difficult to give; perhaps 'mixed bedding' or 'desaggregated bedding' may be used.

Other sediment-petrographic names - Pettijohn (1957) makes a distinction between orthoconglomerates and paraconglomerates (or conglomeratic mudstones). The first type has an intact framework (grain-supported, having the principle mode in the gravel range and a lesser mode in the sand grade), whereas the second type has a disrupted framework (mud-supported, having the principle mode in the silt or clay range and the other in the gravel class, bimodal). Paraconglomerates contain more matrix than clasts and are in reality mudstones and (or) siltstones with a sparse to liberal sprinkling of pebbles, and (or) cobbles, and (or) boulders. Pettijohn believes them to be the products of subaqueous "turbidity flows" and "slides", or of glacial ice or of another mode of mass transport. They are subdivided into paraconglomerates with a stratified matrix (obviously produced by dropping of coarse blocks into still bottom waters in which the finest silts and muds were accumulating) and paraconglomerates with an unstratified matrix. The latter are genetically subdivided into (1) tills and tillites (glacial) and (2) "tilloids" or geröllton (non-glacial). However, the restriction of the term geröllton to conglomeratic mudstones of non-glacial origin is unjustified, first because it would link again a non-genetic concept with a particular genesis, secondly because Kalkowsky (1893), who introduced the term (Geröll-Thonschiefer) to describe particular rocks intercalated in the Kulm of Frankenwald (Thüringen, Germany), believed these rocks to have been deposited by ice rafting and not by mud flowing, so that the term geröllton must not be restricted to the products of the latter only. Although Kalkowsky thought that generation by mud flowing was quite well possible, he considered the outcrops too far away from the hypothetical coast and consequently the slope angle too small to enable the generation of mud flows. However, both the fact that they are intercalated in thick sequences of interbedded greywacke and slates, and the absence of glacial striae on the inclusions, make it probable that the "Geröll-Thonschiefer" represent mud flow deposits after all.

Crowell (1957) recognized the possibly diverse origin of rocks that are called paraconglomerates by Pettijohn, and advocated the use of the term 'pebbly mudstone' as a descriptive name for a rock composed of dispersed pebbles in a mudstone matrix without regard to the manner of origin. But Crowell also used this term for mudstones (which may be sandy) including cobbles and boulders. Flint et

al. (1960 a, b) therefore thought it more appropriate to call the pebbly mudstones sensu Crowell gravelly mudstones, so as to include a larger range of grain sizes than is suggested by pebble. But even this term is somewhat unfortunately chosen, for the term 'conglomeratic mudstone' should be used for the indurated equivalents of gravelly mud. Rigo de Righi (1956) introduced the Italian term 'argille puddingoide' for similar rocks.

Also the term 'clastopelite' is used with reference to till-like deposits (Sames, 1965). This term, however, is somewhat enigmatic in that pelites are often clastic rocks themselves or may show a brecciated structure. Enigmatic are also the terms 'boulder bed' (Dixon & Hudson, 1931; Greenly, 1938; Kugler, 1953; Renz et al., 1955; King, 1937, 1958; McBride, 1966, 1970) and 'rubble bed' (Kugler, 1938; Doreen, 1951; Kugler, 1953) used with reference to till-like deposits. These terms are inadequate, as they merely indicate the presence of boulders or angular fragments in a rock and do not inform us about the composition of the rock between these fragments. The term 'boulder clay', which is commonly used as a somewhat more descriptive equivalent of the genetic term 'till', is in fact also restricted to a particular size range of the inclusions, like the term 'boulder shale' used by Moore (1934) for the till-like submarine slide deposits of the Johns-valley shale in the Ouachita Mountains (U.S.A.). The term 'tilloid', as used by Pettijohn (1957), Schermerhorn & Stanton (1963), and Lindsay (1966), should be abandoned, because any mental association with genesis of any kind should be avoided when describing rocks. For the suffix -oid added to a purely genetic term does not convert it into a descriptive one.

Flint et al. (1960 a, b) propose the term 'diamicton' (from the Greek: $\delta\iota\alpha\mu\iota\gamma\nu\upsilon\mu\iota$ = to mingle thoroughly), which they define as "any non-sorted or poorly sorted terrigenous sediment that consists of sand and/or larger particles dispersed through a muddy matrix. Size distribution is commonly bimodal or polymodal, with one or more modes in the coarse-grain range and one or more in the silt-clay range. A diamictite is a lithified diamicton. Both terms are noncommittal as to genesis." Included explicitly are deposits that contain abundant rock-fragments of older limestone. Also included are greywackes, on the understanding that this term is used for sandstones with muddy matrix. They instituted these terms to avoid such lengthy names as 'sandy gravelly mud' or 'sandy conglomeratic mudstone' (Miller, 1953). But nevertheless geologists appear to prefer the lengthy names on account of their simplicity and clearness. But these lengthy terms cannot be applied to the names of lithostratigraphic units where one word is recommended to denote the lithology of the unit. In the names of lithostratigraphic units, therefore, the words 'diamicton', 'diamictite', and 'geröllton', which is a special kind of diamictite, are preferable. Flint et al. prefer the term 'diamictite' to 'paraconglomerate' because the prefixes para- and ortho- have been used in rather different ways in rock nomenclature. The application of these prefixes to conglomerates with still different meaning seems to invite confusion. The term 'diamictite' has been used by Frakes & Crowell (1967) for deposits which, according to them, were formed partly beneath grounded ice and partly beneath floating shelf-ice, and which had been subject to subsequent delapsion; by Unrug (1963) for mud flow deposits; and by Sestini (1968) for olisthostromes.

General aspects of olisthons

The only certain paleogeographic conclusion that can be drawn from the occurrence of fossil olisthons is the existence of a paleoslope, at least in the source area. The slope inclinations mentioned in the literature for herpoliths are generally very small amounting to only a few degrees or even less:

Heim (1908)	3°26' and 2°31'	Waterhouse & Bradley (1957)	max. 5°
Arkhanguelsky (1930)	1°-3°	Pierce (1963)	max. 2½°
Rettger (1935)	7°	Grant-Mackie & Lowry (1964)	8°
Natland & Kuenen (1951)	5°	Lomize (1966)	2°-3°
Hills (1953)	1°	Lewis (1971)	1°-4°
Sheppard (1955)	½°		

Most of these data are obtained by measuring directly or estimating the difference in altitude between the source area and the area of deposition, and by dividing this vertical distance by the horizontal one from the source area to that of deposition. Thus the tangent of the slope angle is obtained. Grant-Mackie & Lowry have measured the difference in dip between the rocks above and below the slid-down mass.

Olisthons occur, nearly without exception, in regions of rapid deposition and are commonly associated with other olisthons of the same or of other types. The only exception encountered in the literature seems to be some of the olisthostromes of the argille brecciate type in Sicily, which alternate and are associated with white marlstones and gypsum horizons (Trubi Formation). Also sandstone dykes and sills are always associated with rapidly deposited rocks and commonly with subaqueous slumps and slides (Fairbridge, 1946; Barrington Brown, 1938; Dorreen, 1951; Marchant & Black, 1960; Potter & Pettijohn, 1963; Laubscher, 1961). Mud-volcanoes are also often associated with slumps and slides (Kugler, 1953; Beneo 1957; Brunschweiler, 1966; Higgins & Saunders, 1967).

Two phases can be distinguished in olisthon successions (Görler & Reutter, 1968): a progressive phase and a regressive phase. In the former phase the successive olisthons tend to have an ever further basinward extension. The latter phase represents the reverse development. A progressive phase followed by a regressive one is here considered to represent one delapsional megacycle which is generally composed of many micro-cycles. Of course compound olisthon successions exist composed of a sequence of alternating progressive and regressive phases, possibly intercalated with rock units composed of true sediments instead of olisthons. Not always has the entire cycle been developed: the initial or final stages may be lacking or some intermediate stages may be passed over.

It seems as if olisthons become more and more "proximal" in the progressive phase and more "distal" in the regressive one, even though the source area remains in the same place. Hence, for turbidites the ABC-index of Walker (1967, 1970) indicates relative progressivity when used in a direction perpendicular to the stratification, and relative proximality when used in a direction parallel to the stratification. Progressivity may coincide with proximality when the source area is also moving towards the basin axis.

A progressive phase may be accompanied by an increasing intensity of delapsion (which can be defined as the volume of delapsed material per unit of time), but this need not be necessary. According to Perrodon (1961) the general increase in the intensity of delapsion in a basin of deposition is in close response to the

increase in the intensity of diastrophism in the area. Examples of increasing intensity in olisthon successions with only minor intermissions are: 1) the flysch–wildflysch–diverticule–nappe sequence in the Alps; 2) the flysch–olisthostrome–diverticule sequence in the Northern Appennines; 3) the turbidite–olisthostrome–diverticule–delapsional outlier sequence near Moratalla (Southeastern Spain). Examples of decreasing intensity in olisthon succession can be found in several sequences that are in some way comparable to "molasse" sequences. These may start with olisthostromes, for instance: 1) the Ranzano–Bismantova sequence in the Northern Appennines (Sestini, 1970); 2) the overthrust slice – scaly clay – mud flow – "greywacke" ("molasse") sequence in Timor (Audley-Charles, 1965a, 1967). An example of a fluctuating intensity in olisthon successions is the alternation of intervals with olisthostromes or "Gleitklippe" and intervals with turbidites in Sicily.

Every delapsional event, which can be considered as geologically instantaneous (Marchetti, 1957), reduces the depth of the basin. If the olisthon is a thick sheet, an olisthostrome for instance, the reduction of depth may materialized by the occurrence of shallow water deposits directly above the olisthon (Flores, 1959). The reduction of depth may be compensated by subsequent subsidence of the basin floor. Rapid deposition of successive olisthons from one source area generally results in a rapid reduction of depth on one side of the basin which causes a shift of the basin axis away from the source area.

A common feature in olisthon successions is the so-called inversion: stratigraphically higher olisthons may contain or may be composed of older rocks than the lower ones. The cause is that a rising source area or advancing thrust mass would first shed its upper and younger rocks and subsequently the lower and older ones (Kugler, 1953). Successive delapsion may involve several successive lithostratigraphic units. In relatively undisturbed sequences of normal attitude the marked stratigraphic anomaly that older beds overlie younger ones is explained in this way (Fairbridge, 1946). Inversion has been reported by Lugeon (1943), Badoux (1967), Rigo de Righi (1956), and Görler (1967).

As far as has been established the mud in muddy slides contains a notable percentage of clay minerals with highly thixotropic and swelling properties, which are known to be commonly associated with the most plastic clays (Grim, 1953, 1962; van Olphen, 1963; Gillott, 1968), for instance montmorillonite (in the scaly clays in Timor, Audley-Charles, 1965a) or clay minerals of the chlorite-vermiculite-saponite type (in the argille scagliose of the Northern Appennines, Gallitelli, 1956). Also the marlstones of the Represa Formation (Upper Albian), which have supplied most of the material for the olisthostromes near Moratalla (Southeastern Spain), contain a notable percentage of montmorillonite (van Veen, 1969). An onlap relationship is indicated for the base of the Turonian Venado Formation (Sacramento valley, California) truncating Cenomanian and Albian strata. The latter contain many bentonitic rocks (largely montmorillonite). The olisthostromes at the base of the Venado Formation in which large quantities of the Albian rocks are reworked (Brown & Rich, 1960; Peterson, 1965; Ojakangas, 1968) probably also contain much montmorillonite. The shales or mudstones within the conglomeratic mudstones of the Jackfork Group in the Ouachita Mountains (U.S.A.) have a clay mineral assemblage of 54.5% illite, 23.7% mixed layer material, 15.3% chlorite, and 6.5% kaolinite (Morris, 1971). The "flowage shales" of Morris (1971), however, contain up to 30% kaolinite, whereas the associated non-disturbed fissile shales contain almost no kaolinite (Jackfork Group). Owing to the high osmotic swelling properties of these minerals and the high plasticity index of clays contain-

ing such minerals, they may obtain very small angles of internal friction, even when heavily preconsolidated. Clayey rocks with a high percentage of montmorillonite of mixed layer material may therefore give rise to excessive diverticulation or olisthostromization when the overburden is eroded away. Also the clays of wildflysch seems to contain these types of clay minerals, because it vitrifies before the blow pipe (Beck, 1911).

Potter & Pettijohn (1963) write: "A slide may become a subaqueous mudflow and travel even greater distances or grade insensibly into slurries and turbidity currents which are reported to travel hundreds of miles."

This sentence may not be interpreted as if one can expect to find a bed of which one end consists of pebbly mudstone and the other of a turbidite gradually merging into each other. This idea, however, is strongly invoked from the observation that mud flow deposits are laterally replaced by turbidite sequences and that a complete spectrum of deposits exists between slumps and turbidites (Walker, 1967, 1970, with extensive bibliography). However, a conglomeratic mudstone bed has never been reported to merge laterally into a turbidite bed. On the contrary, many pebbly mudstone beds have been observed to wedge out, for instance the Lago Sofía Conglomerates in Chili (Scott, 1966). The marked difference in velocity and mechanics between mud flowing and turbidity currents (according to their traditional conceptual model) renders an insensible merging in lateral direction mechanically impossible. A mud flow stops its movement by an increase in consistence due to a decreased rate of shear (lesser bottom inclination), or to a decrease in void ratio accompanying gelation, or both (Dott, 1963). The movement starts to decrease at the nose of the mud flow and rapidly extends in the direction of the tail. If the material in the nose goes into suspension by increasing velocity (for instance due to the increase of the slope angle) or by the presence of a large amount of noncohesive sand grains, the stopping process may fail to start and the entire flowing mass may go into suspension. In that case no record is left of the initial mud flow. Moreover, the parent rock of a turbidity current should be a very sandy mud which may contain pebbles. On account of the large quantity of the material involved in the turbidity currents that are thought to be responsible for the turbidites in flysch basins, repeated liquefaction of large volumes of metastable sand-mud mixtures in one over large areas of the continental slope or shelf seems more appropriate as a cause of turbidity currents than local mud flowing or muddy sand-avalanching reported from submarine canyons. When going into suspension the latter two may give rise to some abortive or aberrant kind of turbidite.

On the other hand, however, "sliding or flowing mud masses would throw some debris into suspension directly above them, forming a slurry of which the solids would settle in graded fashion directly down onto the top of the displaced mass itself without necessarily forming a true turbidity current as normally conceived" (Dott, 1963). This probably accounts for some of the crudely graded pebble and sand units on top of some unsorted Squantum pebbly mudstone zones near Boston, U.S.A. (Dott, 1961); and for the graded beds welded to the upper surface of some plastic slides near Denbighshire, Wales (Jones, 1937, 1940; Barrington Brown, 1938); and also for the laminated sandstones filling the irregularities of the surface of the olisthostromes in the Pievepelago Formation, Monte Modino, Italy (Reutter, 1965). Also grain avalanching may produce a turbid layer of water directly above the avalanching material from which the suspended solids may settle, forming the normally graded top of the deposit (Sanders, 1965; Kuenen & Sanders, 1956; Kuenen & Carozzi, 1953). De Raaf (1968) restricts the term fluxoturbidite

to this type of ambivalent deposits. When the mud flow or grain avalanche stops, the suspended solids may travel further basin inward due to their inertia, in some cases forming a normally graded deposit like a turbidite. But these moving solids would have a velocity equal or less than that of the mud flow or grain avalanche. Consequently the resulting bed would not have a great extension, since the suspension would quickly dump the larger solids.

Summary and conclusions

In this part an attempt has been made to bring some order in the terminological chaos existing in the field of independent gravitative displacements of detached rock material along sloping surfaces, in order to make clear what to understand by olisthostrome and what position olisthostromes have in the whole of geologic phenomena. All these gravitational phenomena are wholly gradational from one to the other, constituting a continuous series which ought to be regarded as an inseparable unity called delapsion. The field of delapsion covers processes hitherto studied by structural geologists as well as sedimentologists and has therefore been cut on non-genetic grounds into gravity tectonics and gravitationally displaced sediments. But in this publication the term 'delapsion' was meant to comprise the exogenic processes resulting in the displacement and the deposition of solid parts of the earth's crust without the primary help of tractional forces exerted by a mobile medium travelling in the same direction. Every delapsional event produces an olisthon. Delapsion was separated from sedimentation, which was restricted to comprise the exogenic processes resulting in the displacement and the deposition of solid parts of the earth's crust, but with the primary help of tractional forces exerted by a mobile medium travelling in the same direction. The process of sedimentation produces sediments. Delapsion was also separated from tectonizaton, which was restricted to comprise the deformation of the rocks of the earth by endogenic processes. Hence delapsion was considered as a separate and independent category or processes between tectonization on the one side and sedimentation on the other. However, both sediments and olisthons are considered as equally ranked subdivisions of depositonal rocks, or briefly deposits, both of which may occur together in normal depositional sequences. Since sedimentary rocks fall within the scope of stratigraphy, also "resedimented", or better, redeposited rocks should fall within it, so that all olisthons are to be considered as rock-stratigraphic units.

From the mechanics involved in delapsion it can be gathered that the mode of delapsion and the type of olisthon is determined by (1) the nature of the parent rock (for instance the composition, texture such as the ratio cohesive to non-cohesive material, fabric which may be metastable or anisotropic, and its physical properties such as consistence, fluidity, plasticity, rigidity, compaction, consolidation, lithification etc.), (2) the nature of the slope along which the rock material delapses (for instance slope angle and roughness), and (3) the nature of the medium in which the delapsion takes place (liquid or volatile). As the variations in the nature of the parent rock and of the slope are wholly gradational, merging imperceptibly into each other, also the modes of delapsion and, consequently, the types of olisthons are wholly gradational showing a complete spectrum. Therefore, I would like to stress the little sense of instituting too many different names for arbitrary

intercepts in the continuous series of delapsional processes and olisthons. Any subdivision had better be restricted to "end-members".

Delapsional processes have largely been subdivided on the base of particular physical properties of the material during its down-slope movement, namely its rigidity, plasticity, or fluidity, or whether the yield limit, thixotropic limit, or liquid limit is exceeded. But a distinction on these grounds is useless in geology, because geologists can never be sure about the exact physical conditions of rocks before their lithification. Preferable, therefore, would be a subdivision of delapsional processes based on criteria observable in the resulting olisthons, such as the degree of contortion and (or) disruption of the original stratification, the ratio of cohesive to non-cohesive material, the degree of differentiation of the constituents (sinking of inclusions, Bouma-divisions) etc. But this would require an entirely new subdivision with brand-new names for delapsional processes as well as olisthons; this would merely add to the terminological chaos (admittedly a step in this direction was made by introducing the terms 'coherent and incoherent delapsion', 'stratifacient and non-stratifacient delapsion' and the term 'olisthostromization'). Therefore we better stick as far as possible to the available genetic names for delapsional processes and olisthons. But this has the disadvantage that many olisthons cannot readily be named, because the genetic concepts behind most of the available terms for olisthons cannot be defined by the non-genetic characteristics, such as the textural and structural features, by which they are generally recognized: not every conglomeratic mudstone is a mud flow deposit, nor every graded sandstone a turbidite. It should be emphasized that genetic concepts, also those of tectonization, sedimentation, and delapsion, must be defined on genetic criteria. Large scale structures are not necessarily tectonic, nor the small scale ones delapsional; not every sandstone deposit may be a sediment, though it certainly is a deposit.

A tentative genetic classification will be given for the various olisthons which is schematically represented in table 1.

Table 1. Subdivision of depositional rocks.

A. Sediments,

B. Olisthons:

Mode of delapsion	Olisthons: genetic names refer to the process of delapsion	Olisthons: names refer to the geological genesis	
1. Coherent delapsion: delapsion of rock segments with the preservation of internal coherence mainly along slip planes or lubricating horizons. It comprises: slipping (including slumping) and coherent sliding.			non-stratiform olisthons
a. along pre-existing thrust-fault planes		delapsional outliers	
b. along slip planes other than pre-existing thrust-fault planes	slump blocks slump packets slip blocks slip sheets	diverticules; "Gleitklippen"; klippes sédimentaires: olisthothrymmata, exolistholiths	

c. by folding in the distal end of the delapsing rock sequence	coherent slides	gravity collapse structures: kneefolds, roof-and-wall structures, cascades, flaps	olisthons
2. Incoherent delapsion: delapsion by the overall mobile behaviour of an incompetent mass of cohesive material that loses its original coherence. It comprises: incoherent sliding (including progressive brecciation) and mud flowing	"slump sheets"		stratiform
a. olisthostromization: without the inclusions being able to sink to the base	incoherent slides	olisthostromes with olistholiths: allolisthostromes, endolisthostromes, mud-supported slide-conglomerates	olisthons
b. in which the inclusions sink to the base	differentiated mud flow deposits		
3. Action of turbidity currents: delapsion of rock material that goes into suspension in the surrounding medium during the delapsional movement	turbidites: distal turbites, proximal turbidites		
4. Grain avalanching: delapsion by overall intergranular movement of many loose, non-cohesive grains together	grain avalanche deposits: gravel, rubble, and sand avalanche deposits	grain-supported slide conglomerates	
5. Rock falling: delapsion of rock masses by falling and individual rolling or tumbling along steep slopes	rock fall deposits	screes	

'Herpoliths' (Lotze, 1956) comprise at least coherent and incoherent slides,
'Olisthanites' (de Raaf, 1968) comprise all subaqueous olisthons generally studied by sedimentologists, including subaqueous olisthostromes (personal communication, appendix 16), except turbidites, rock fall deposits and those olisthons that are generally studied by structural geologists, such as gravity collapse structures, delapsional outliers, large diverticules, etc.

Appendices

1. Bianconi, P., 1840 - [Argille scagliose are rocks in which] una superficie levigatissima, dolce, ontuosa al tatto, lucente, ceròide e metalloide, si presenta andando a seconda delle scaglie di cui è constantemente composta questa sorta di argille . . . È questo un carattere talmente proprio di queste argille che crediamo doverle chiamare provvisoriamente argille scagliose. Appariscono infatti come un aggregato di tante squame o lenti di varie dimensioni, che si légano e s'innèstano vicendevolmente, ora adagiandosi piane e distese l'una sull'altra, ora abbracciandosi mutualmente, cuive e ravvolte. Disgiungonsi facilmente spogliàndosi e come scivolando l'una sull'atra, avendo ciascuna lenticella la sua superficie liscia e lucente, come la massa. La loro frattura, però, è terrosa. L'unione delle piccole scaglie produce scaglie di maggiore dimensione . . .

2. Merla, G., 1953 - [Definition of argille scagliose:] Ma una definizione sarà pur sempre

necessaria, e allora darei la preferenza a quella formulata de Penta (1950, p.9) press'a poco testualmente cosi: le "argille scagliose" sono la formazione alloctona nata per frane e scivolamenti sui pendii delle rughe orogeniche nel periodo che va dal Cretaceo superiore al Pliocene, e ricoprente vaste area dell'Appennino; il complesso, cioè, argilloso, scaglioso, spesso variegato, a struttura caotica, con inclusi diversi e di svariatissime dimensioni scompaginati nella massa argillosa che il avvolge o sorregge, complesso di varia età (essenzialmente Giura-Miocene) che si trova trasgressivo (sub. intruso) tettonicamente sulle (sub. nelle) formazioni autoctone dell' Appennino. Continua Penta osservando che "il termine, così inteso, comprenderebbe e sostituirebbe quelli di Liguride, formazione ofiolitica od ofiolitifera, flysch caotico eomiocenico, flysch alloctono ecc."

In somma, le argille scagliose sono il risultato di un fenomeno tettonico, cioè delle frane e slittamenti sui pendii orogenici, e non di un fenomeno sedimentario.

3. Greenly, E., 1919 - The essential characters of an autoclastic mélange may be said to be the general destruction of original junctions, whether igneous or sedimentary, especially of bedding, and the shearing-down of the more tractable material until it functions as a schistose matrix in which the fragments of the more obdurate rocks float as isolated lenticles or phacoids . . . It is developed on a great scale . . . In a country of this kind, the larger masses are found to be arranged as trains of lenticles overlapping *en echelon* . . . Upon a map the masses that can be separated out appear as if interbedded in a homogeneous and structureless matrix. But this country-rock is itself built up of interdigitating lenticular bodies; and, could we take in the whole region at a glance, it would present itself to us as a mèlange of torn and sheared lenticular masses of all sizes, from such as are two of three miles in length to the smallest that the eye can see, spilitic lava, diabase, quarzite, limestone, jasper and grit, floating in an undifferentiated but schistose body that is a weft of all the more easily deformable elements, itself pervaded throughout by the same lenticular structure.

4. Hsü, K. J., 1968 - [He defines mélanges as] mappable bodies of deformed rocks characterized by the inclusion of tectonically mixed fragments or blocks, which may range up to several miles long, in a pervasively sheared, fine grained, and commonly pelitic matrix. Each mélange includes both exotic and native blocks and matrix. Native blocks are disrupted brittle layers which were once interbedded with the ductiley deformed matrix. Exotic blocks are tectonic inclusions detached from some rockstratigraphic units foreign to the main body of the mélange . . .

5. Lugeon, M., 1943 - On est donc obligé d'admettre que des décollements, parallèles à des couches primitivement empilées et inclinées en succession normales, ont dû présider au départ des différentes nappes. Seule la gravité peut donc expliquer le phénomène. A ces nappes, d'un genre tout à fait nouveau et insoupçonné, je donne le nom de diverticules. Dans leur empilement, ces diverticules ne doivent pas présenter de flancs renversés; ils se présentent, en effet, comme des écailles ou des lames. Plus tard, ces diverticules, repris par l'orogénèse alpine, ont évidemment pu se replier sur euxmêmes, mais à ce moment ils sont exclusivement passifs. Ces diverticules, ce sont donc gigantesques glissements, qui se sont exécutés dant l'avant fosse alpine, sur le flanc de la cordillère frontale, soit dans le géosynclinal qui se trouvait au sud de la plateforme òu se sont déposés les sédiments helvétiques proprement dits, et cela pendant que se sédimentait le flysch tertiaire.

Il peut arriver que la diverticulation prenne un développement excessif. C'est ainsi que l'on appelle la nappe de la Plaine Morte est formé par des diverticules de second ordre, qui se présentent comme des lames ou lentilles de Turonien ou de Maestrichtien, ces deux terrains étant toujours sans rélations entre eux, lames perdues dans le Flysch et la plupart du temps dans le Wildflysch, soit le produit de glissements.

Il peut arriver que la diverticulation détermine une désordonnance dans la situation réciproque des diverticules si l'un d'eux a glissé plus loin que celui sur lequel il devrait normalement reposer. Saute-mouton.

Quand on essaie de déterminer les lieux de départ, disont ce que seraient des niches d'arrachement dans des glissements du versant d'une colline argileuse, on ne peut toutefois échapper à la nécessité de faire intervenir tout d'abord des poussées, selon la tectonique classique de compression. Il se forme comme une accumulation de matière en avant et sur le front d'un géanticlinal, puis, lorsque l'empilement s'exagère, se produisent les glissements, soit les diverticulations du trop plein . . . Ce sont évidemment les éléments les plus élevés qui glissent en premier.

6. *Lamare, P., 1947* - (p. 284) On a vu . . . les formations bréchiques se charger de matériaux de dimensions anomales: ce ne sont pas seulement des blocs gigantesques, dont la forme semi-arrondie reste malgré tout compatible avec la genèse des accumulations détritiques de ce type, mais des paquets beaucoup plus allongés que larges, des masses lenticulaires. Il en existe de plus étendues encore: certaines mesurent plus de 500 m. de longuer sur 4-5 m. d'épaisseur. C'est du problème de la mise en place de ces copeaux à allure de klippes que nous allons nous occuper maintenant.

(p. 285-286) On remarque, entre les manières d'être des brèches et celles des klippes une différence fondamentale: si les gros éléments des brèches accusent une allure chaotique, les klippes, elles sont orientées. Je ne dis pas qu'il s'agisse là de directions définies sur de grands espaces, ni d'une façon rigoureuse: mais localement, les klippes d'un même ensemble s'alignent parallèlement, et cela, dans des azimuts sans rapport avec ceux des accidents tectoniques.

(p. 287) Quelle origine attribuons-nous à ces empilements anormaux? Tout contredit l'hypothèse d'accidents tectoniques, même d'écailles de style cassant. 1) Allure nettement transgressive du contact du Crétacé, qui tire ses matériaux les plus gros du substratum immédiat; - 2) Indépendance des lames par rapport aux cassures majeures; - 3) Répétition du phénomène à une cadence incompatible avec celle de lames de charriage: trois lames distinctes sur 25 m.

Ainsi donc, au milieu de brèches incontestablement sédimentaires, ces lames représentent des intercalations sédimentaires: ce ne peuvent être que des paquets exotiques mis en place au cours de la sédimentation, et dont la présence est étroitement liée à la genèse des brèches.

(p. 306) . . . des secousses se produisent, détachent des pans entiers de falaises côtières ou sous-marines, en font glisser les matériaux sur la marge continentale, sans toutefois les entraîner dans la fosse elle-même.

(P. 311) . . . la transgression crétacée engendre dépôts extrêmement puissants, dont les caractères aberrants peuvent s'expliquer par des secousses sismiques. Ainsi se seraient détachés d'énormes paquets de roches déjà consolidées: ceux-ci, incorporés aux sédiments en cours de dépôts sous la forme de copeaux, de "lames", dont la mise en place ne saurait être due à des charriages, constituent ce que nous avons appelé klippes sédimentaires.

7. *Görler, K., & Reutter, K.-J., 1968* - ["Gleitklippen" are] . . . Blöcke von den Ausmaszen groszer Olistholithe und Megolistholithe die, ohne von Olisthostrome-Matrix umgeben zu sein, in jüngerem sediment stecken. Ist klar erwiesen, dasz hier kein Schlammstrom-Transport wirksam war, so darf konsequenterweise die Bezeichnung Olistholith nicht verwendet werden . . Derartige Gleitklippen enstehen zum Teil dadurch, dass sich von der Stirn einer näherrückenden Decke kleinere Partien lösen und in das Sedimentationsbecken voraneilen.

8. *Jacobacci, A., 1966* - Il termine *olistolite* dovrebbe essere riservato per quegli esotic . . . di grandezza eccelzionale rispetto alle dimensioni medie del pezzame oppure a quegli elementi litoidi comunque geneticamente ben differenziati rispetto alla massa dell'olistotroma e del pezzame.

9. *Görler, K., & Reutter, K.-J., 1968* - Sie [= megolistholiths and 'Gleitdecken'] liegen meist . . als Platten an der Oberfläche eines Olisthostromes. Für den Ursprung der mit Olisthostromen zusammen auftretenden groszen Olistholithen und Gleitdecken gibt es 3 Möglichkeiten: 1) Sie bildeten das bei der Aufwölbung der Liefergebiete zerreiszene kompetentere Hangende dieser tonigen Serien. 2) Sie sind aus jüngeren Gesteinen hervorgegangenen, die auf einen Olisthostrom während einer längeren Periode tektonischer Ruhe abgelagert wurden, aber später bei einem Wiedereinsetsen der Orogenese als grosze Schollen abwanderten. 3) Sie waren als kompaktere Horizonte in das tonige Ausgangsmaterial der Schlammströme eingeschaltet.

10. *Borello, A. V. 1970* - Intercalé dans celui-ci [= flysch cadocien au facies d'orthoflysch sur le flanc gauche de la Quebrada de San Isidro], on trouve un grand corps olistolitique de calcaires cambriens (hypo-olistolithes) et d'autres, plus petits, de roches granitiques du substratum précambrien qui ont ainsi le sens de "téléolistolithe" (nom employé ici par l'auteur pour la première fois).

11. *Richter, M., 1968* - Unter "Scherbentektonik" verstehe ich das Zerreiszen einer kompetenten jüngeren Schichtserie über einer mächtigen tonigen Serie beim olisthostromatischen Flieszen oder auch bei diapirischen Vorgängen dieser Serie. Die zerrissenen Schollen (Scherben) der

hangenden Schichtserie werden umflossen, vielfach sogar eingeschlossen und liegen als mehr oder weniger grosze Scherben auf und in der ursprünglich liegenden inkompetente Serie, mit der zusammen sie sogar weiter gleiten können.

12. Marchetti, M. P., 1957 - This term [olisthostrome] is proposed to indicate a shaly body, more or less chaotic, crumpled, and brecciated, containing rigid blocks of age younger, contemporaneous, or older than that of the shaly matrix, slided and flowed down by gravity into a low area (mostly into a basin of sedimentation) and originated by formations which are older than those where it slides on. The olistostromes are settled mostly in marine environment but they can be also subaerial.

13. Richter, M., 1968 - [His definition of olisthostromes is] mehr oder weniger tonige Massen in langsamen submarinen Rutschungen bis zu sehr groszen Ausmaszen. Vielfach auch mit turbiditischen Erscheinungen zusammen. Olisthostrome können mehrere hundert Meter Dicke erreichen und viele Kilometer weit gleiten. Der Gleitvorgang kann stratigraphisch und zeitlich erfaszt werden, chaotische Strukturen treten meistens auf. Während des Gleitens können jüngere Ablagerungen auf dem Olisthostrom sedimentiert werden.

14. Lotze, F., 1956 - Berichtet wurde zunächst über "synorogene Sedimente" in den Westpyrenäen. Es handelt sich dabei um mächtige, weit sich hinziehende subaquatische Rutschungen ("Herpolithe") innerhalb der höchsten Oberkreide; sie werden mit der laramische Gebirgsbildung in zeitlichen und ursächlichen Zusammenhang gebracht.

Lotze, F., 1957: In den Westpyrenäen, zum Beispiel, gibt es Diskordanzen zwischen Unterkreide und höherer Oberkreide, zwischen Oberkreide und mittlerem Alttertiär; sie sind durchweg mit umfangreichen Schichtlücken verbunden. Wenig weiter westlich sind diese Faltungen jedoch nicht mehr eingetreten, und es fehlen die Diskordanzen; das Meer blieb dort, und es lagerten sich auf seinem Boden auch während der orogenen Phasen Gesteinsschichten ab. Gerade diese synorogenen Sedimente zeichen sich durch besondere Eigentümlichkeiten aus: Vielfach sind ihnen in mehrfacher Wiederhohlung Bänke aus subaquatischen Rutschmassen (sog. "Herpolithe"), aus Schuttströmen, aus thixotrop deformiertem Material und so weiter eingeschaltet, alles Bildungen auszerordentlicher Bodenunruhe, schwerer erdbebenartiger Bodenerschütterungen . . .

Ein anderer Beispiel solcher synorogenen Sedimente bietet Westfalen . . . Die in nicht allzu groszer Entfernung vom Teutoburger Wald im Münsterland vorhandenen Schichten der höheren Oberkreide, die selbst eine geschlossene Folge ohne Diskordanzen darstellen, zeigen ganz ähnliche synorogene Erscheinungen ("Herpolith-Bildungen" und so weiter) wie die etwa gleichaltrigen Schichten in Nordspanien. In ganz groszem Still treten sie, flachenhaft weit ausgedehnt, zwischen den Baumbergen und den Beckumer Bergen auf, wobei zahlreiche solche Horizonte übereinandergestapelt sind, jeweils durch "normale" Sedimente getrennt. Ohne die Annahme heftiger Bodenschwingungen, welche die noch unverfestigten, thixotrop reagierenden Ablagerungen zum Flieszen und Schwimmen auch bei practisch horizontalem Boden brachten, sind derartige Phänomene nicht denkbar.

15. Dzulynski, S., Ksiazkiewicz, M., & Kuenen, Ph. H., 1959 - A different type of sedimentation is encountered amidst normal turbidites in many places. In this type the grain size is large, and the beds tend to be less muddy. The bedding is thicker and rather irregular, and the shales between are silty to sandy and thin or even absent. Current sole markings are scarce, load casting is more common, and coarse current bedding of somewhat varaible direction is encountered. Indications of slumping are found, and grading is absent, repetitive, irregular, or even inverted, and irregular lenses of coarser grains occur in a few places inside the beds. These sandstones may occur in large lenses between normal flysch and shales. In other cases the material or the directions of supply contrast with those of the normal surrounding flysch of the same age. Because characteristics of deposition from turbidity currents appear to be mixed with evidence for sliding, we propose to call this kind of bed a fluxoturbidite.

The abnormal properties appear to indicate:

1) a powerfull supply forming thick beds and leaving no time for accumulation of clay, and/or
2) a steep slope producing slides and not allowing the deposition of mud only of sandy or silty clays,
3) a short distance of origin, so that the material was not spread out into thin beds covering wide areas, coarse material has not been dropped on the way, and turbidity currents had not succeeded in sorting the materials to produce regular grading,

4) instead of the usual longitudinal supply this material can have been introduced from the side.

Whatever the cause, the mode of transport has changed. Instead of a well-mixed turbulent turbidity current carrying almost the entire load in suspension, one can imagine a turbidity current in which most of the sand and gravel moves in a watery slide along the base. The current is too poor in clay to raise this load in suspension, and the slope is too steep for the load to come to rest until it has spread out in a layer. A true turbidity current can be likened to a dry avalanche, and a normal slump is the equivalent of a wet avalanche remaining in contact with the ground. We believe that an intermediate type of movement is also possible and suggest that the resulting deposit be called a fluxoturbidite.

16. Raaf, J. F. M. de, 1968 - On trouve assez souvent, liés à des turbidites, des sédiments conglomératiques, sableux, argileux etc. sans granoclassement et moins fréquemment des slumps à couches individuelles deformées mais encore reconnaissables. Je les réunirai sous le nom d'olisthanites. Ce sont, comme les turbidites, des sédiments gravitatifs (mis en mouvement par les forces de gravité), mais le matériel qui les compose, reste, en dévalant la pente, incapable de former une suspension à forte turbulence. Ils se déposent sans granoclassement ni triage, comme une masse glisée. On pense que la grande majorité des courants de turbidité sont issus de pareils phénomènes de glissement.

Les olisthanites sont d'une grande diversité. Le terme comprend entre autre des conglomérats glissés non triés, des paquets de sédiment dilacéré ou plissé (à "slumpfolds"), des couches argileuses ou silteuses à fragment de roche ou blocs, lambeaux d'argile ou cailloux roulés, enfin des couches sableuses èpaisses généralement hétérogènes, contenant souvent des inclusions diverses ou de nids de sable grossier.

[In a footnote de Raaf continues:] J'ai remplacé le mot olisthostrome, utilisé "sensu lato" (cf definition Geol. Nomenclature Royal Geol. Mining Soc. Netherlands, 1959, qui permet d'inclure des dépôts sans dominance argileuse) dans le texte hollandais (1965c) par un nouveau terme: olisthanite (ολισθαινω=glisser). Il comprend les dépôts marins résultant de "subaqueous slumps and massflows" de Dott (1963), de "sédiment flows and slumps" de Dzulynski et Walton (1965) et les "massflow deposits" de Stauffer (1967a), sans comprendre des masses écroulées. Ce sont des sédiments sans granodécroissance vers le haut, qui peuvent être fins ou grossiers, hétérogènes ou (parfois) homogènes. Je propose de restreindre l'usage du nom "olisthostrome" aux gigantesques dépôts olisthanitiques, comme ceux bien connus de l'Italie et d'autres pays (Badoux, 1967b. Görler and Reutter, 1968).

On aurait alors cinq types de sédiments marins étrangers à la sédimentation pélagique, liés à des phénomènes gravitatifs, dont trois sans mise en suspension du matériel: 1) masses écroulées (rockfall deposits) 2) olisthanites, 3) olisthostromes, 4) fluxoturbidites, 5) turbidites.

De Raaf (verbal communication) would presently like to amplify his olisthostrome concept as follows: "of course only subaqueous types of olisthostromes of any composition–sometimes containing huge blocks – enter into this classification, which deals with marine sediments. The former are devoid of regular grading, though the distribution of inclusions is not always entirely random. This holds true also for the smaller-sized equivalents (olisthanites)."

Part 2. Description of the olisthons near Moratalla

> Das Beobachtungsmaterial zeichnet sich einerseits durch auszerordentlich zahlreiche Einzelheiten aus, dasz man weithin sozusagen jeden Quadratmeter einzeln besprechen sollte, und anderseits herrscht doch eine grosze Einformigkeit, die ermüdend wirkt und den Uberblick über die Verschiedenheiten auszerordentlich erschwert.
>
> P. Beck, 1911 (when describing wildflysch).
>
> Olisthostromes resort to the greatest efforts of imagination to explain the utterly haphazardous positions of the "outcrops" considered.
>
> G. Flores, 1959.

Introduction

The area investigated in the summers of 1963, 1964, 1967 and 1968 is situated around the village of Moratalla, province of Murcia, southeastern Spain. The concerning parts of the topographic map sheets Moratalla and Caravaca numbered 889 and 910 (scale 1 : 50 000), were magnified five times and considerably improved with the help of aerial photographs. Also several geographic names in use by the local population were added to these maps (encl. 1 and 2). For example it appeared that the Río Alárabe (from where it enters the Moratalla basin the river is called the Río de Moratalla) had been incorrectly designated as Río Benamor on map sheet 889, whereas the real Río Benamor had been mistakenly annotated as Río de Caravaca.

In order to avoid confusion, the thickness of bedding is indicated according to the system proposed by Ingram (1954), the grain size according to the scale of Wentworth (1922); the classification of carbonate rocks is based on that of Dunham (1962).

The complicated stratigraphic relations of the region can be understood best in terms of named Groups, Formations, Members, and Beds. It is however beyond the scope of the present publication to fulfil all requirements desired by the International Subcommission on Stratigraphic Classification (1972) with respect to the establishing of formal lithostratigraphic units. But as it is the intention that the names given here to the various lithostratigraphic units will henceforward be used by geologists working in the Betic Cordilleras, they are nevertheless formally proposed.

Geological setting

In the alpine mountain chain of the Betic Cordilleras (encl. 1 and 4) several geological zones have been distinguished. The area investigated is situated along and north of the border of two of these zones, which are known as the Prebetic zone, mainly deposited in a neritic environment, and the Subbetic zone mainly of pelagic deposits (with the exception of the Triassic and Lower Jurassic). In the area concerned, the rocks of the Prebetic zone are overthrust from the south by those of the Subbetic zone. Between the Prebetic and the Subbetic a rock sequence crops out, a part of which has Subbetic affinities, the other part Prebetic affinities. Similar rock sequences are referred to in literature as "unité intermédiair", "séries intermédiaires", "prébétique profond" or „prébétique méridional". The latter name, South-Prebetic zone, is preferred here, since this rock sequence is considered as a part of the Prebetic zone. Several Subbetic outliers are situated on top of Prebetic and South-Prebetic rocks.

Synopsis of the lithostratigraphic succession of the Subbetic zone near Caravaca

The Subbetic zone in the Caravaca area (encl. 4) can be divided into two tectonic units. The southern Loma de Solana Unit lies in imbricated position over the southern part of the Buitre Unit, which in turn has been thrust northward over the rocks of the South-Prebetic and Prebetic zones. From the Buitre Unit the Peñuelas, Gavilán, Tollo, Miravetes, Argos and Represa Formations have been formally proposed by van Veen (1969); from the Loma de Solana Unit he has formally proposed the Quipar, Jorquera and Gredero Formations. Large quantities of rocks from these Formations were found reworked in the Middle Miocene of the Prebetic zone in the north. Here follows a summary of the descriptions by van Veen (1969) of these Formations.

Peñuelas Formation

Lithology – Well-bedded, heterogeneous Formation of shales, sandstones and carbonate rocks predominantly of brown-red colour, though also purple, violet, orange, yellow, green, black, and white colours have been found. Characteristic is the presence of gypsum in beds and in large extruded masses. Some intervals contain abundant limonitized pyrite crystals of euhedral shape.
Thickness – Several hundreds of metres at least (Paquet, 1969); the lower boundary is not exposed and the upper boundary is apparently a fault contact.
Age - The very scarce fossils are indicative for a Triassic age.
Depositional environment - Partly terrestrial, partly shallow marine with restricted circulation (germanic facies).

Gavilán Formation

Lithology - About 300 m of light-grey, massive to ill-bedded, medium to fine

crystalline dolostones in which locally ghost structures of well-rounded, coarse sand-sized carbonate grains have been found. The dolostones are overlain by hard, pale-brown, thick-bedded to very thick-bedded, pseudo-oolitic bioclastic lime-grainstones with variable thicknesses of 25 to 90 m. The pseudo-oolitic limestones are covered by about 12 m of pale-brown, medium-bedded, crinoidal lime-packstone. The sequence ends with a highly fossiliferous bed, covered by rust-coloured encrustations ("hardground"). All rocks are virtually devoid of siliciclastic detritus.
Thickness - The lower boundary appears to be a fault contact. The upper boundary has been chosen at the top of the iron-encrusted bed. From aerial photographs it is apparent that the thickness of 600 m mentioned by van Veen (1969) is too high, due to repetition by faulting. The total reconstructed thickness is about 350 m.
Age - Late Domerian brachiopods and ammonites have been collected from the topmost bed. Usually an Early Liassic age has been attributed to the Gavilán Formation, though Upper Triassic rocks may be present.
Depositional environment - Presumably deposited above wave-base on a fully marine, shallow carbonate platform.

Tollo Formation

Lithology - Well-bedded alternation of light olive-grey, bioclastic to pelletal lime-mudstones, lime-wackestones and lime-packstones with, mainly in the upper half, a few red-coloured, nodular intervals with many ammonites ("Ammonitico Rosso"). At the base a 7 m thick unit of soft, white limestone occurs in which a serrate bedding fissility is developed. Chert concretions occur in the upper two thirds of the Formation. Characteristic microfossils are *Globochaetes,* "filaments" (cross-sections of minute thin-shelled pelecypods) and *Saccocoma.* The amount of silt-sized quartz grains is at the most 1 percent.
Thickness - The upper boundary has been chosen halfway the several metres thick interval in which the lithologic transition occurs from the Tollo Formation to the overlying Miravetes Formation. The thickness is about 210 m.
Age - In view of the ammonites, brachiopods and tintinnids (abundant in the topmost part of the Formation) an Early Toarcian to Late Tithonian age is attributed to the Tollo Formation.
Depositional environment - Presumably deposited for the greater part below wave-base on a fully marine, rather shallow platform.

Miravetes Formation

Lithology - Well-bedded alternation of more and less resistant, greyish, thin to thick-bedded lime-mudstones. The less thans 1% of silt-sized quartz grains and mica flakes observed in the upper half of the Formation are a precursory symptom of the overlying Argos Formation. In the lower half many tintinnids occur, whereas *Nannoconus* has been found throughout the Formation.
Thickness - 985 m.
Age - Tintinnids and ammonites are indicative for a latest Tithonian to Late Barremian age. About 270 m below the top of the Miravetes Formation a rich ammonite fauna suggests a latest Hauterivian age.
Depositional environment - presumably low energy marine; fossils of pelagic and nectonic type are predominant.

Argos Formation

Lithology - light grey lime-mudstones and wackestones with many graded, laminated, mica-bearing, calcareous sandstone interbeds of pale red-brown colour which are generally grading upward into dark grey marlstone. The calcareous material, mixed with the angular, silt to sand-sized siliciclasts that constitutes 50% of the total rock, consists of bioclasts and sparite, in several samples with a notable percentage of true ooliths and orbitolinids. The bases of some of the sandstone beds show somewhat loaded flute marks. Many interbeds show the almost complete sequence of divisions characteristic for turbidites.
Thickness - van Veen (1969) has chosen the boundaries at the base of the lowermost sandstone bed and at the top of the uppermost one; the total thickness is 120 m.
Age - On account of ammonites and planktonic foraminifera a Late Barremian to Late Albian age is assumed. A hiatus in the succession of biozones has been discovered: those biozones representing approximately the Clansayesian, Early and Middle Albian are absent (van Veen, 1969; Kuhry, 1971).
Depositional environment – Probably low energy marine below the sublittoral zone with repeated influxes of turbidity currents.

Represa Formation

Lithology - Alternation of harder and softer beds of thin to medium thickness consisting of grey bioclastic lime-wackestones with *Pithonella, Stomiosphaera* and single-keeled globotruncanids. The percentage of silt-sized quartz grains amounts to 1 percent.
Thickness – 315 m.
Age - The planktonic foraminifera and ammonites of the Represa Formation are indicative for a Late Albian and Vraconian age.
Depositional environment – Probably low energy marine below the sublittoral zone.

Quipar Formation

Lithology - Light yellowish-grey to white, bioclastic lime-wackestones with in the lower half an interval with black chert concretions and in the upper half an interval of brick-red colour. Non-calcareous detritus is virtually absent. Double keeled globotruncanids and heterohelicids are abundantly present, particularly in the middle and upper part. Rudist fragments have been found near the base of the Formation.
Thickness - The lower boundary has been chosen halfway the few metres thick interval in which the colour of the rock passes from grey into white. The Formation is badly exposed in the Buitre Unit, but in the northern part of the Loma de Solana Unit, only 2 km south of Caravaca, it amounts to 190 m.
Age - The planktonic foraminifera of the Quipar Formation suggest an earliest Cenomanian to Late Campanian age.
Depositional environment - Presumably low energy marine; fossils of pelagic type are predominant. The presence of the rudist fragments near the base, however, is incompatible with the presumed depositional environment. A tentative explanation of their presence is the assumption of a slight angular unconformity (see Ubacas Dolostone Formation, page 106).

Jorquera Formation

Lithology – The lower and middle part (members A, B, and C of van Veen) are composed of light yellowish-grey to white, bioclastic lime-packstones with many laminated, calcarenitic interbeds of pale orange wheatering-colour. Double-keeled globotruncanids and heterohelicids are abundantly present in the lime-packstones. The upper part (27 m, member D) is composed of olive-grey, bioclastic lime-wackestones with several greyish, laminated, calcarenitic interbeds. Globorotalids, globigerinids and a few *Discocyclina* fragments have been found here. Some of the calcarenitic interbeds show the almost complete succession of divisions characteristic for turbidites.
Thickness - The lower boundary has been chosen at the base of the lowermost calcarenitic interbed. The thickness amounts to 225 m.
Age - In view of the planktonic foraminifera an Early Maastrichtian to middle Early Eocene age has been attributed. Member D had better been separated from the Jorquera Formation and incorporated in the Moro Marlstone Formation which can be recognized in the Sierra de la Puerta (page 86).
Depositional environment - Presumably low energy marine, probably below the sublittoral zone with repeated influxes of turbidity currents.

Gredero Formation

Lithology - Soft, yellowish-grey, globigerinal lime-packstones with extremely abundant planktonic foraminifera alternating with more resistant laminated, calcarenitic beds, some of which have a calciruditic basal part predominantly composed of larger foraminifera. In the basal 16 m of the Formation the resistant calcarenite beds are commonly amalgamated and predominate above the soft globigerinal packstones. Moreover, these calcarenite beds are distinguished from those in the other parts of the Formation by their exceptionally high percentage (15 to 20%) of coarse, subrounded to well-rounded siliciclastic detritus. Some of these calcarenitic interbeds show the complete succession of divisions characteristic for turbidites. The percentage of siliciclastic detritus in the remaining part of the Formation does not exceed 3 percent.
Thickness - Van Veen has chosen the lower boundary of the Formation at the base of this 16 m thick sequence of amalgamated calcarenite beds. The Gredero Formation represents the youngest Formation that has taken part in the folding of the Subbetic zone. The next younger Formation known in this area, the Upper Miocene (?) *Alcores Formation,* has a horizontal stratification truncating the folded structures. The exposed thickness of the Gredero Formation is 420 m.
Age - The planktonic and the larger foraminifera (*Nummulites* in the lower half and lepidocyclinids in the upper half) are indicative for a middle Early Eocene to Aquitanian-Burdigalian age. But the basal 16 m had better been separated from the Gredero Formation and included in the Rasa Sandstone Formation which has been recognized in the Sierra de la Puerta (page 92).
Depositional environment - probably low energy marine below the sublittoral zone, with many influxes of turbidity currents.

Outline of the lithostratigraphic succession of the South-Prebetic zone in the Sierra de la Puerta

A large part of the following description is an abstract of the one given by Paquet (1969, p. 83-85, 108, 133, and 153; Magné et al., 1969) of the lithostratigraphic succession along, and in the vicinity of, the Barranco de la Puerta, about 10 km east of the boundary of the area investigated (encl. 4 en 5). The stratigraphic rock units, recognized are in stratigraphic order:

1. Canara Formation (the oldest rock-stratigraphic unit exposed).
2. Represa Formation.
3. Quipár Formation.
4. Jorquera Formation.
5. Moro Marlstone Formation.
6. Rasa Sandstone Formation.
7. Miñano Limestone Formation.
8. Roble Limestone Formation } presumably lateral equivalents
9. Cuervo Sandstone Formation }
10. Trieta Formation.
    Puerta Limestone Member with the Asares Conglomerate Bed at its base.
    Moratalla Marlstone Member (youngest).

Paquet's descriptions of the Moro, Rasa, Miñano Formations, and the Puerta Limestone Member will be amplified with more detailed descriptions of their development in the westernmost extension of the South-Prebetic zone, which is situated in the southeastern corner of the investigated area. Rocks identical to those of the South-Prebetic have been reworked in large quantities in the Middle Miocene of the Prebetic zone immediately to the north. The entire succession of the Sierra de la Puerta has a vertical to inverted attitude, dipping some 60° S.

Canara Formation

Derivatio nominis - After the hamlet of Canara situated at the outcrop boundary (encl. 4). The Formation is defined by the rock units a up to and including g of Pacquet's (1969, p. 83, fig. 38B) "Coupe del Horno", (type section), and by the units a, b and c of his (p. 82) "Coupe de la Casa de Sillen" (reference section).
Lithology – Near the Barranco de la Puerta it is rather heterogeneous and composed of alternating sandy marlstones and marly limestones with a notable amount of mica. About halfway the sequence two limestone intervals crop out (15 and 20 m thick respectively) containing many ooliths, orbitolinids, echinoids and brachiopods. Near Casa de Sillen, about 5 km to the southwest, the sequence is more monotonous and consists of blueish-grey, sandy marlstone, also with a notable amount of mica. Sandy marly limestone with ooliths and with the same fossils are the equivalent of the limestone intervals in the east. Six kilometres more to the west, at the eastern boundary of the investigated area, at least a large part of the sequence is replaced by the Represa Formation, a monotonous alternation of more and less indurated grey marlstones, which can be distinguished from the Canara Formation by its negligible percentage of sand-sized siliciclastic detritus, but cannot be distinguished from those of the Represa Formation in the Buitre Unit.

Thickness - The basal part of the Canara Formation cannot be seen because of a thrust fault, but near the Casa de Sillen 275 m of the Formation are exposed.
Age - On account of the presence of *Hedbergella washitaensis* and *H. planispira* in the lower exposed part, and because of the additional presence of *Globigerinelloides benthonensis* higher in the sequence, and of *Rotalipora ticinensis* in the top part, a (?Middle and) Late Albian age has been attributed to this Formation.

Represa Formation

Lithology - Alternation of yellowish-grey marlstones and marly limestones.
Thickness - In the vicinity of the Barranco de la Puerta the Formation is about 33 m thick (rock units 2, 3 and 4 of Paquet's "Coupe K 7,4 Rte de Calasparra").
Age - In a 3 m thick interval halfway through the sequence ammonites have been found indicative for a Vraconian age (this substage is here considered as the uppermost part of the Albian Stage and distinguished from the Upper Albian s.s.). This is in accordance with the planktonic foraminifera found in the lower part. The planktonic foraminifera from the upper part are indicative for a latest Vraconian to earliest Cenomanian age. In the southeastern corner of the investigated area the Represa Formation is much thicker and comprises rocks which are time-equivalents of the rocks of the Canara Formation.

Quipar Formation

Lithology - The lower 55 m of this Formation (units 5 up to and including 11 of Paquet's "Coupe K 7,4 Rte de Calasparra") is mainly composed of white conglomeratic marlstones (Paquet, 1969, adds the phrase "du type olistostrome.") and intraformational limestone conglomerates of which many are normally graded with a several metres thick basal part, and two 1 m thick intercalations of white, alternately soft and harder limestone. The enormous blocks floating in the marlstones of the lower one of the two 20 m thick conglomeratic marlstone units contain planktonic foraminifera indicative for a Vraconian age, while the marlstone binder yielded *Globotruncana helvetica, Gt. schneegansi, Gt. stephani, Gt.* cf. *sigali,* and *Gt.* ex gr. *lapparenti.* This association is indicative for an Early Turonian age. The conglomeratic marlstones are overlain by a 10 m thick sequence of alternately brick-red and white calcilutites ("couches rouges", unit 12 of Paquet) which in turn is followed by a 20 m thick alternation of grey marlstones and marly limestones (unit 13 of Paquet). In the Sierra de la Puerta the Quipar Formation differs from its Subbetic equivalent by its conglomeratic development. The intraformational limestone conglomerates were interpreted as the calcilutitic variety of olistho-stromes of the argille brecciate type.
Thickness - In total it is 85 m, half as thick as in the Subbetic zone.
Age - On account of the planktonic foraminifera this Formation is of Early Cenomanian to latest Campanian age; the "couches rouges" are of Santonian age.

In the westernmost part of the South-Prebetic zone (situated in the investigated area) the brick-red "couches rouges" are in direct contact with the rocks of the Represa Formation. This is due to a fault, which is at a small angle to the stratification.

Jorquera Formation

Lithology - Alternation of sandy mica-bearing limestone beds and grey or greenish marlstones. *Inoceramus* has frequently been found in this Formation.
Thickness - About 25 m (the upper 20 m of the Quipar Formation plus the 25 m of the Jorquera Formation are reported in Magné et al., 1969, to be 30 m thick). It comprises the rock units 14 and 15 of Paquet's "Coupe de la Sierra de la Puerta" (1969, p. 84, fig. 39).
Age - Maastrichtian planktonic foraminifera occur in the lower and middle part, whereas the marlstones at the top contain Lower Paleocene planktonic foraminifera.

This Formation can be correlated with the members A, B and C of the Jorquera Formation in the Loma de Solana Unit (van Veen, 1969). The rock-unit that can be correlated with member D has been separated from the Jorquera Formation and is called Moro Marlstone Formation.

Moro Marlstone Formation

In the Barranco de la Puerta this Formation, which is defined by Pacquet's (1969, p. 108, fig. 48) rock unit 1 of the "Coupe de Barranco de la Puerta" (type section), has been described by him as "marnes vertes, á galets de calcaires finement gréseux maestrichtiens, alternant avec des sables roux et des grès en plaquettes. Les éléments maestrichtiens peuvent atteindre plusieurs mètres cubes. Les marnes rélèvent une microfaune du Paléocène supérieur".

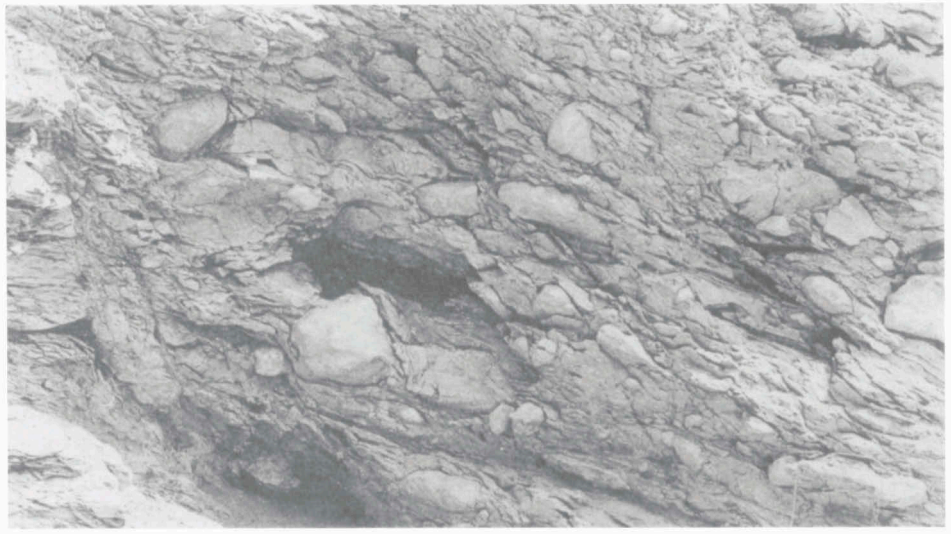
Derivatio nominis - The Moro Marlstone Formation is well exposed in the Barranco del Moro (reference section) at the western boundary of the investigated area (encl. 1 and 4), although the basal part is lacking here due to a thrust fault and the topmost part is covered by rock debris.
Lithology - Here the stratigraphic lower 10 m of the exposed part of the Formation consist of non-bedded grey oligomictic conglomeratic marlstone, the stratigraphic upper 25 m are composed of a bedded alternation of grey fine-grained limestones and grey marlstones. Within the upper part three more intercalations of conglomeratic marlstone occur of 1 to 1.5 m thickness containing only a few limestone inclusions of pebble and cobble size and in one intercalation also many dispersed lager foraminifera.

Many limestone beds do not show internal structures, others are parallel-laminated, whereas still others do show internal deformations which are assumed to be metadepositional. Some of the limestone beds have a calciruditic or calcarenitic basal part containing abundant larger foraminifera, and a fine-grained laminated upper part. Large burrows were discovered at the base of one of these beds. Two beds are even entirely composed of larger foraminifera. Some beds show the almost complete turbidite succession and are therefore interpreted as calciturbidites.

The conglomeratic marlstone, which is the most complex part of the Formation, consists of a non-stratified grey marlstone in which fragments, highly varying in shape, of different rocks are randomly distributed. Almost all these inclusions consist of rock types which also can be found in the stratigraphically upper part of the Moro Marlstone outcrop. Only a few are identical to the rocks of the Quipar and Jorquera Formations.

Fig. 8. Twisted plate-like olistholith predominantly composed of larger foraminifera. Olisthostrome of the Moro Marlstone Formation, Barranco del Moro.

Fig. 9. Close-up of the conglomeratic marlstone; the book is 16 cm long. Olisthostrome of the Moro Marlstone Formation, Barranco del Moro.



1. Inclusions resembling some of the parallel-laminated limestone interbeds in the stratigraphically upper part of the Moro outcrop: grey fine-grained (50-60μ), well-sorted, bioclastic lime-packstone composed of angular fragments which look like broken foraminifera, and 3 to 5 percent angular siliciclasts (quartz, quartzite, felspar, colourless mica and sporadically glauconite). Fossils generally present are: globigerinids, globorotalids, smaller benthonic foraminifera, and echinid-spines. In some samples also a few *Nummulites, Discocyclina, Operculina,* and calcareous algae were found.
2. Inclusions similar to the coarse basal parts of the limestone interbeds and to the beds entirely composed of larger foraminifera in the stratigraphically upper part of the Moro outcrop: coarse-grained, bioclastic pseudo lime-grainstone almost exclusively composed of *Nummulites* and *Discocyclina.* Less common are: *Asterocyclina, Operculina, Assilina, Amphistegina,* alveolinids, *Rotalia,* and fragments of various kinds of calcareous algae, echinoids and bryozoans. Here and there micrite with globigerinids, globorotalids, and smaller benthonic foraminifera was found between the larger foraminifera. Only 2 percent of the rock consists of sub-rounded quartz and quartzite grains (20-1200μ).
3. Inclusions corresponding to the finer grained upper part of the interbeds with calciruditic basal part in the upper Moro Formation: grey, bioclastic pseudo lime-grainstone with a considerable percentage of angular siliciclastic detritus (20-200μ) of quartz, quartzite, felspar, colourless mica; only a few tourmaline and glauconite grains occur. A few reworked grains made up of bioclastic lime-wackestone with globotruncanids were found. Among the bioclasts were recognized many globigerinids, globorotalids, smaller benthonic foraminifera, and fragments of *Nummulites, Discocyclina, Operculina,* various kinds of calcareous algae and of echinoids.
4. Inclusions comparable with some of the limestone interbeds in the stratigraphically upper part of the Moro outcrop: globigerinal lime-wackestones and packstones with a very small percentage of siliciclastic detritus (0,5%). The packstones may have been recrystallized into pseudo lime-grainstones. Among the bioclasts were recognized globigerinids (abundant), globorotalids, smaller benthonic foraminifera, fragments of calcareous algae, echinoids, bryozoans and molluscs.
5. An earliest Eocene inclusion of soft grey marlstone, in which *Globorotalia aragonensis, Gl. quetra, Gl. multica,* and *Globigerina linaperta* were determined.
6. A few Maastrichtian inclusions of soft, white, calcilutite in which were observed *Globotruncana contusa, Gt. stuarti, Gt. falsostuarti, Gt. gansseri, Gt. mayeroensis, Gt. arca, Gt. aegyptiaca, Gt. havanensis, Gt. intermedia, Globigerinelloides messinae, Gl. benthonensis, Hedbergella petaloidea, Rugoglobigerina rugosa,* and *R. rotundata.*
7. Many isolated larger foraminifera (mainly *Amphistegina, Nummulites and Discocyclina),* echinoid fragments and grains of quartz and quartzite are scattered in the grey marlstone binder. In a wash-residue from the binder the following planktonic foraminifera were determined: *Globigerina linaperta, Globorotalia rex, Gl. wilcoxensis, Gl. aequa,* double-keeled globotruncanids and heterohelicids. This association is characteristic for the *Globorotalia rex* zone of earliest Eocene age or slightly younger.

The spherical inclusions may reach a diameter of 3 m. The plate-like slabs may have a longest apparent diameter of more than 5 m. The plates, which are predominantly composed of closely packed larger foraminifera, are folded and twisted (fig. 8). The harder inclusions which generally consist of fine-grained grey limestone with a few larger foraminifera, may be equidimensional, rectangular, triangular, or lenticular in cross-sections (fig. 9). Also an inverted rectangular fragment of limestone turbidite with a basal part of closely packed larger foraminifera was found. The inclusions of softer, laminated grey limestone are twisted, plicated and balled-up.

The binder consists of grey marlstone. In a few places a faint lamination is preserved, which is found to be deformed into isoclinal, disharmonic folds with curved or undulating axial planes tending to be parallel to the upper and lower boundaries of the conglomeratic marlstone unit (fig. 10). In the zone of maximum curvature the laminae are thicker than in the limbs, indicating a plastic behaviour of the binder during the deformation. The fissility is concentric with the fold structures and curves around the inclusions (fig. 12). At several places in the grey bind-

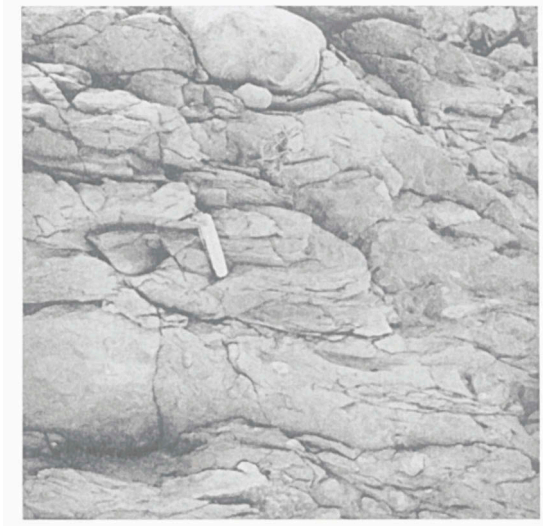


Fig. 10. Recumbent fold in a laminated part of the binder. Olisthostrome of the Moro Marlstone Formation, Barranco del Moro.

Fig. 11. The anastomosing network of serrately fissiled zones enclose lenticular, almost non-fissiled marlstone bodies, which are twisted and drawn out. These bodies are considered as parts of the binder, because they contain small inclusions. The folded, weakly fissiled part right above the hat merges gradually into serrately fissiled marlstones to the right, whereas it shows a sharp discontinuity with the fissiled marlstones to the left. Olisthostrome of the Moro Marlstone Formation, upper course of the Barranco de los Barranquillos.

er an anastomosing network of zones was found. These zones are characterized by parallel, serrate fissility enclosing lenticular and oblong bodies of marlstone of a slightly lighter shade in which the fissility is markedly less (fig. 11). Some of them display curved and twisted shapes with tapering ends; some others may even pass gradually into the enclosing marlstone at one side, whereas a sharp discontinuity with the binder occurs at the other side. These bodies do not, or only faintly, show internal lamination revealing the internal structures. Moreover, they commonly contain small inclusions. These bodies should therefore be considered as parts of the binder instead of as soft inclusions. At one site the fissility pattern was found to form chevron folds (fig. 13). This may be due to subsequent tectonization of the conglomeratic marlstones, because this pattern was found close to the bordering thrust-contact (page 86).

The conglomeratic marlstone is a rock unit that has suffered extensive small-scale deformation. The absence of slickensides and fracture cleavage (except in a narrow zone adjoining the thrust fault that forms the southern boundary of the Moro-outcrops in the westernmost part of the Sierra de la Puerta) indicates that the deformation has been brought about by very low stresses. The undeformed state of the delicate structures of the many foraminifera in the conglomeratic marlstone seems to point against a tectonic (autoclastic) origin of the conglomeratic texture. As the stratigraphically younger beds of the Moro Formation are not deformed except for an inverted attitude (strike N 65 E, dip 60° to the south), it can be concluded that the visco-plastic deformation had taken place before the conformable deposition of these beds. The deformation is likely to be due to sliding and a similar strike, dip and inverted attitude ought to be accepted for the conglomeratic marlstone. This view is favoured by the fact that the average direction of the bedding fissility roughly parallels the inferred attitude of the conglomeratic marlstone.

The oligomictic and almost monogenic nature of the conglomeratic marlstone indicates that mainly the rocks of only one bedded formation became involved in the sliding. Deposition then continued on top of the slide. The conglomeratic marlstone is therefore interpreted as the result of incoherent sliding of a bedded part of the Moro Marlstone Formation itself when it was still in a visco-plastic state. The inclusions are interpreted as the disrupted portions of the harder beds of the Formation, which became entirely wrapped up in the softer marls. The random distribution of these inclusions indicates that the binder had a consistence which did not permit the inclusions to sink. The resulting deposit may therefore be designated as 'olisthostrome', the inclusions as 'olistholiths'.

The axes of the few small antiform structures found in alternating soft grey marlstone and harder grey marly limestone beds (perhaps large inclusions) are parallel to the average strike of the inverted sequence (N 65 E) indicating a stress directed N 25 W. Since the deformations caused by subsequent tectonization are only confined to a narrow zone along the thrust-contact, we may consider these structures to be due to the sliding. The larger tabular olistholiths in the olisthostrome are dipping towards the south. Their dip, however, is notably less than the inferred dip of the olisthostrome unit (fig. 14). This implies that if the sequence were in a horizontal normal attitude, which should be accepted to be the original one, the plates would have a northward dip. If we assume that a slow eddying motion took place in the binder during the sliding, the consistent northward dip of the plate-like olistholiths would indicate a more or less northern provenance. This is in accordance with the stress direction inferred from the antiform structures.

Fig. 12. Close up of the conglomeratic marlstone. Note the fissility curving around the olistholiths. The two white, drawn-out, fissiled olistholiths in the lower part of the photograph consist of rocks of the Quipar Formation. Olisthostrome of the Moro Marlstone Formation, Barranco del Moro.

Fig. 13. The chevron folds displayed by the fissility pattern probably are due to the subsequent tectonization of the olisthostrome of the Moro Marlstone Formation, Barranco del Moro.

The steeply dipping stratification of the Moro Marlstone Formation is truncated by the "couches rouges" of the Quipar Formation. Locally the structural contact, which is slightly dipping to the south, shows polished and striated surfaces. Towards the south the dip increases until the contact plunges steeply away, forming the southern boundary of the Moro outcrops. In several places the "couches rouges" are intimately fused with the conglomeratic marlstone; they even have exactly the same direction of fissility. At several places directly below the low-angle structural contact large wedge-shaped bodies of "couches rouges" were found, between which tapering tongues of conglomeratic marlstone are intercalated. These "couches rouges" masses are interpreted as dislodged slices pressed into the olisthostrome. They should not be considered as olistholiths. Directly above the structural contact the alternately vivid-red an white Santonian calcilutites are strongly crumpled and locally show small chevron folds. The bedding fissility exactly follows the contortions. A similar relation has been found by Paquet (1969a) south of the Casa de la Puerta about 5 km to the east. But if we trace the rocks of the Moro Marlstone Formation and the "couches rouges" further to the east it appears that both rock units become part of one and the same stratigraphic sequence of steeply dipping inverted attitude and that the structural contact is replaced by Campanian to Paleocene rock units, which are lacking towards the west. Hence in the westernmost extension of the South-Prebetic zone the "couches rouges" are pinched between two parallel faults (see page 85).

Thickness - Along the Barranco de la Puerta the thickness of the Moro Marlstone Formation is about 100 m.

Age - Planktonic foraminifera are indicative for a Paleocene to lowermost Eocene age.

Rasa Sandstone Formation

In the Barranco de la Puerta this Formation consists of yellowish to white, mainly friable sandstones, locally alternating with green marlstones (Paquet, 1969, p. 108, rock units 2-6 of his "Coupe de Barranco de la Puerta").

Derivatio nominis - In the Cañada Rasa directly to the north of the aquaduct (Canal de Taibilla) over the Barranco del Moro (encl. 4) the upper part of the Formation crops out, whereas the best exposures of the lower part can be found in the Barranco del Moro directly west of the aquaduct (type-section).

Lithology - The thick middle part of the Formation is well-bedded and consists of very coarse to medium-grained, friable, yellowish quartz sandstone. The beds may reach a thickness of several metres. They show a normally graded division at the base, followed by a lower parallel-laminated division, a current-bedded division with convolutions and, rarely, a lutitic division consisting of dark green shale. An upper parallel-laminated division was not found. Almost all beds are amalgamated. In some beds the laminae show fold structures with axial planes that tend to be parallel to the stratification, whereas the top of the bed remains flat and undeformed. These folds are taken to be due to internal sliding during the deposition of the sand beds. The base of most of the beds is flat, but some of the thicker beds show downward bulges (loadcasts?). Irregular basal planes truncating the flat and undeformed laminae of the underlying bed were also found. These irregularities are apparently due to scouring during the deposition of the sand bed (fig. 15). On account of these observations the Rasa Sandstone Formation is interpreted as a sequence of proximal tubidites.

Fig. 14. The larger tabular olistholiths in the olisthostrome of the Moro Marlstone Formation are dipping notably less than the dip (dashed line) of the whole Formation. Photo taken almost parallel to the strike. Barranco del Moro.

Fig. 15. The succession of Rasa Sandstone is inverted, for the irregular base of the conglomeratic lower part of a thick proximal turbidite (below) is truncating the parallel-laminated, fine-grained upper part of the underlying bed (above). The basal irregularity is probably due to scouring. 700 m west of the Cortijo de la Monja.

The Rasa Sandstone is predominantly composed of angular grains of quartz and quartzite. Less common are felspars among which microcline and plagioclase. Some colourless mica, amphibole, tourmaline, zircon, and epidote were found. The sandstone is badly sorted; the grains grade in size from very fine to very coarse sand. The larger ones are well-rounded. The interstices are filled partly with a muddy matrix, partly with a cement of calcite crystals. Among the few organic remains fragments of echinoids and some smaller foraminifera were recognized; in one sample a fragment of a bryozoan and one of a mollusc were found.

The lowermost 5 to 7 m of the Rasa Formation is more calcareous and exposed directly south of the aquaduct. It forms a low, oblong topographic ridge. Higher in the sequence, but still in the lower part of the Formation, three intervals of several metres thick occur consisting of a regular alternation of yellowish, parallel-laminated, very fine-grained (20-200μ), graded sandstone beds of 10 to 20 cm thick and green shale beds of the same thickness. In one of these intervals a graded bed with many larger foraminifera at its base was found. *Globorotalia rex,* and *Globorotalia aequa* were determined in the green shales, indicating the *Globorotalia rex* zone. At the base of some of these thin sandstone beds excellent flute marks were observed (fig. 16). At one place very coarse isolated quartzite grains were found in the "deepest" part of the flutes. This is taken to be the evidence for the existence of a very thin bed-load at the base of the turbidity current responsible for this bed. The average direction of the flutes of three different beds was measured. By reconstructing the original attitude of the inverted beds these average directions are in stratigraphic order N27E, N40E, N18E. Thus the turbulent flow that made these flutes must have come from the northeast, which is in harmony with the N25W provenance of the stratigraphically underlying Moro olisthostrome. Also small prod-marks, bounce-marks and structures interpreted as the result of bioturbation were found.

In some of the thin, fine-grained turbidite beds with flute marks, a peculiar kind of fracturing was observed (fig. 16), which brought forth a regular alternation of strikingly equidistant, parallel and narrow (1-3 cm) "fault ridges" separated by broad (15 cm) "fault throughs". The flute marks are cut by the fractures, so that the fracturing was apparently subsequent to the formation of the flutes. The fractures are at an acute angle to the flutes and almost parallel to the strike of the beds. They are very conspicuous at the basal side of the beds. The top side, however, is either flat without any trace of the fracturing, or shows a much smaller shift along the fracture planes. Hence the fractures probably were formed when the bed was not yet lithified. A similar fracturing was found in some of the thick amalgamated sandstone beds (fig. 17). Here the fractures can be observed only in the fine-grained, laminated upper part of the bed, whereas the thick coarse-grained lower part is structureless. In our opinion the fractured parts of the sandstone beds reacted in a dilatant way to a tensional stress which may have been caused by a stretching of the incompetent marlstone or liquified coarse sandstone layer adjacent to the base of the fractured parts.

The strinkingly regular spacing of the fractures and their peculiar paralellism to the strike of the beds suggest that the fracturing was brought about by the bending of the whole rock sequence of the South-Prebetic zone when it acquired its overturned attitude. In a later chapter arguments will be added in favour of a still non-lithified state of the Rasa Sandstone during its deformation and reworking in the Middle Miocene deposits of the Prebetic zone (page 144, 147).

In the upper part of the Rasa Sandstone Formation the coarse basal parts of the sandstone beds contain many well-rounded, smooth quartzite pebbles which

Fig. 16. Sole of a thin turbidite with flute marks showing a regular pattern of parallel, narrow "fault ridges" and broad "fault troughs" which is probably caused by the fracturing (due to the dilatant behaviour) of the still non-lithified bed, when the whole sequence of the Sierra de la Puerta became inverted. Rasa Sandstone Formation, Barranco del Moro. Photo taken by J. J. Hermes.

tend to be ellipsoidal in shape. Their shape and smoothness may indicate a prolonged stay in a marine surf-zone before they were deposited in the present position. The longest axis of the largest pebble is approximately 6 cm. At the top of the Formation the rock can even be designated as a sandy conglomerate composed of rounded quartzite pebbles cemented together by limestone in which *Nummulites* occur.

Of particular geologic interest is the presence of an intrusive body of diabase found in the Rasa Sandstone 100 m north of the Cortijo de la Puerta (encl. 5).

Thickness - Along the Barranco de la Puerta the Formation is 130 m thick (Paquet, 1969). At the aquaduct of the Canal de Taibilla it reaches a total thickness of 200 m.

Age - On account of the planktonic foraminifera in the lower part of this Formation and those at the base of the stratigraphically overlying Formation an Early Eocene to early Middle Eocene age is attributed. In view of its lithology and fossils the Rasa Sandstone Formation can be correlated with the lowermost 16 m (member A) of the Subbetic Gredero Formation (van Veen, 1969), in age it can be correlated with both the A and B members.

Miñano Limestone Formation

Derivatio nominis - This Formation builds the summital parts of the Sierra de la Puerta. The highest summit is called Miñano (encl. 4). The Formation is defined by Paquet's rock units 7 up to and including 11 of his "Coupe du Barranco de la Puerta" (1969, p. 108, fig. 48), which serves as the type section.

Lithology - In the Barranco de la Puerta it consists predominantly of limestones (containing *Nummulites,* discocyclinids, *Assilina,* alveolinids, *Orbitolites, Operculina,* and *Actinocyclina)* with several interbeds of grey and white marlstones, and yellowish sandstones.

From a marlstone intercalation at the base of the Formation were determined: *Globigerina linaperta, G. yeguaensis, Globigerapsis kugleri, G. index, Globorotalia spinulosa, Gl. spinuloinflata, Gl. lehneri, Gl. frontosa, Truncorotalia topilensis, Hantkenina dumblei, H. mexicana.* This association is characteristic for the *Globigerapsis kugleri* zone to *Porticulisphaera mexicana* zone.

In the southwesternmost part of the South-Prebetic zone it consists of bioclastic lime-packstone and bioclastic pseudo lime-grainstone without or with a small percentage (up to 10%) of sand-sized siliciclastic detritus. The fossils recognized are: *Discocyclina, Nummulites,* alveolinids, *Operculina,* various kinds of calcareous algae, echinoid fragments, smaller benthonic foraminifera (among which miliolids), and globigerinids. The siliciclastic detritus is composed of quartz, quartzite, felspar, colourless mica and tourmaline.

Thickness - 40 m, gradually thinning in southwestern diretion to 2 m.

Age - The associations of planktonic foraminifera and larger foraminifera are indicative for a Middle Eocene age. The presence of Upper Eocene rocks (Paquet, 1969) could not be established with certainty. The Miñano Formation corresponds in age with the member C of the Gredero Formation (van Veen, 1969).

Roble Limestone Formation

Derivatio nominis – see page 113.

Lithology - In the Barranco de la Puerta this rock unit is thin and separated from the Miñano Limestone by a 10 to 20 cm thin interval of reddish rock. The Formation consists of white bioclastic limestone in which many *Lithothamnium* oncoliths,

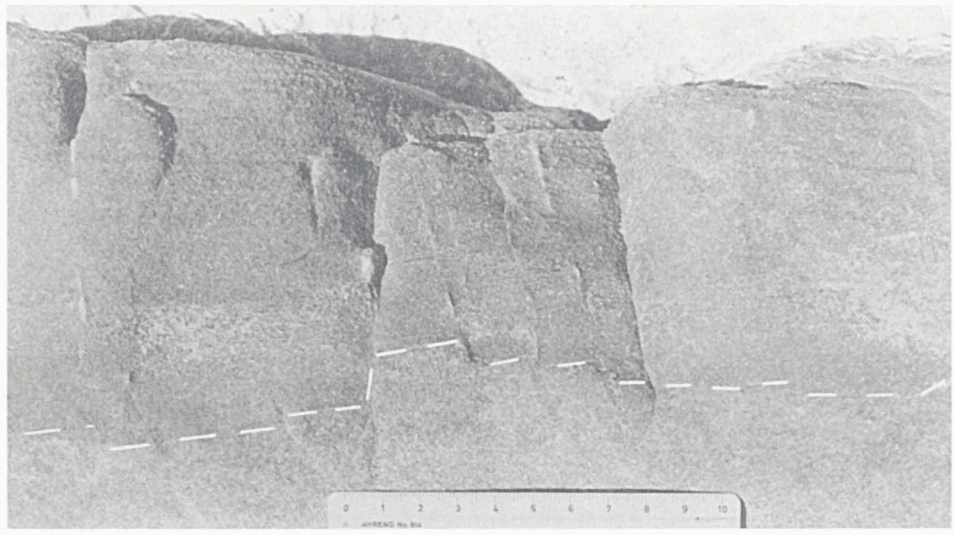


Fig. 17. Close-up of a strongly oblique cut through a "fault ridge" developed in the parallel-laminated, fine-grained top part of a thick proximal turbidite. Rasa Sandstone Formation, Barranco del Moro. Photo taken by J. J. Hermes.

Fig. 18. The underside of the Asares Conglomerate, which has an upright attitude. The large tabular blocks of the rock-fall deposit are parts of the micaceous sandstone beds of the Subbetic Argos Formation. 100 m south of the Cortijo de los Asares.

Amphistegina and *Heterostegina* were found. It has exactly the same lithology and fossil association typical for the Roble Limestone Formation in the Prebetic zone northwest of Moratalla (which will be described on page 113) and in the Sierra de los Falcones (encl. 4; Paquet, 1969, p. 153, Fig. 47, rockunit m 1). In the westernmost part of the Sierra de la Puerta this rock unit is absent.
Thickness - 3 m along the Barranco de la Puerta.
Age - In the Prebetic zone a Middle Miocene age was attributed to the Roble Limestone Formation (page 114). This age implies the existence of an important hiatus between the Miñano Limestone and the Roble Limestone in the Sierra de la Puerta. In the Subbetic zone the greatest part of the Gredero Formation (members D, E, and F of van Veen, 1969) has been deposited in the time interval represented by this hiatus. In the Prebetic zone a sequence of about 200 m of mainly red-coloured rocks separates the Roble Limestone from the equivalents of the Miñano Limestone. The interval of reddish rock that separates the Roble Limestone from the Miñano Limestone in the Sierra de la Puerta probably represents a considerable lapse of time. Northwest of Moratalla the Roble Limestone is directly overlain by the Moratalla Marlstone Member of the Trieta Formation, whereas along the Barranco de la Puerta Moratalla Marlstone is separated from the Roble Limestone by the Puerta Limestone Member.

Cuervo Sandstone Formation

Derivatio nominis - In the large slabs of South-Prebetic rocks reworked in the Middle Miocene deposits of the Prebetic zone, a rock-unit of different lithology was found in the same stratigraphic position as the Roble Limestone Formation. This rock unit is not developed in the Sierra de la Puerta, but only in the reworked rock slabs; it is well exposed in the Cerro del Cuervo where the type section is situated (encl. 1 and 2).
Lithology - The rock unit consists of calcareous sandstone. In thin sections it generally cannot be distinguished from the Rasa Sandstone, except perhaps by the roundness of its grains. In the field, however, the difference is more distinct. The greyish-brown Cuervo Sandstone is very hard owing to a calcareous sparitic cement, whereas the light yellowish-brown Rasa Sandstone is friable. In the Cuervo Sandstone, particularly in the upper part, sandy limestone intervals occur which cannot be distinguished from the Roble Limestone. At several locations clypeasterids and other echinoids were found.

Thin sections made of the Cuervo Sandstone reveal that the non-calcareous detritus predominantly consists of subrounded to rounded quartz and quartzite grains up to a diameter of about 2 mm. Also much felspar (among which plagioclase and microcline) and subordinately tourmaline, amphibole, and glauconite were found. The material between the non-calcareous grains consists of bioclastic lime-packstone and bioclastic lime-grainstone in which were recognized: *Heterostegina, Amphistegina,* globigerinids, rotalids, smaller benthonic foraminifera, calcareous algae (*Lithothamnium),* and fragments of echinoids, bryozoans and molluscs. The reworked limestone clasts present in the upper part of the Formation are precursory symptoms of the stratigraphically overlying Asares Conglomerate. They consist of bioclastic lime-wackestone with *Pithonella* (Represa Formation), and clasts encrusted by calcareous algae and consisting of bioclastic packstone with *Heterostegina,* rotalids, and fragments of echinoids and bryozoans.

The boundary with the stratigraphically overlying Asares Conglomorate is either sharp, undulating and locally lined with an iron crust (hiatus?), or represented by a thin transition zone which at its base is composed of Cuervo Sandstone with a few pebbles of globigerinal lime-packstone (of which the stratigraphically overlying

Asares Conglomerate is composed) grading upward into a mixture of Cuervo Sandstone pebbles and pebbles of the Asares Conglomerate. The amount of the former pebbles rapidly diminishes upwards in favour of the latter. The Cuervo Sandstone is in direct stratigraphic contact with the Rasa Sandstone with which it has the siliciclastic detritus in common. No trace of the Miñano Limestone was found in between! The lower boundary, therefore, probably represents a hiatus.
Thickness - The greatest thickness, 20 m, was measured in the Cerro del Cuervo.
Age - On account of its stratigraphic position directly below the Puerta Limestone Member, and because of the limestone intervals identical in composition to the Roble Limestone, the Cuervo Sandstone is interpreted as the lateral equivalent of the Roble Limestone and is presumably also of Middle Miocene age.

Trieta Formation

Derivatio nominis – see page 114. In this Formation 4 Members (see page 113) are distinguished of which only the lowest two, i.c. the Puerta Limestone Member and the Moratalla Marlstone Member, are developed in the South-Prebetic zone.

Puerta Limestone Member

Derivatio nominis – This Member is best exposed in the pass (= puerta) where the Barranco de la Puerta breaks through the highest ridge of the Sierra de la Puerta (type section, encl. 4). The Member is defined by rock unit g of Paquet's "Coupe du Barranco de la Puerta" (1969, p. 133, fig. 48).
Lithology - Beige, thin-bedded, fine-grained limestone in which fracture cleavage gives rise to numerous parting planes at an angle of 40° to 45° to the bedding planes; bedding planes and fracture cleavage produce a characteristic pattern of rhombs. As we shall see many large slabs of this Limestone have been reworked in the Middle Miocene deposits of the Prebetic zone.

About 50 thin sections were made from these slabs. From these it appeared that the rocks of the middle and upper part of the Member consist of globigerinal lime-packstone almost entirely composed of closely packed globigerinal tests. The small interstices between the tests are filled with micrite in which numerous silt-sized fragments of globigerinal tests occur. In almost every thin section small *Lithothamnium* fragments, some echinoid fragments, and some smaller benthonic foraminifera were found. In many thin sections also mollusc fragments, *Amphistegina, Rotalia,* radiolarids and sponge-spicules were recognized. In fact these thin sections can hardly be distinguished from those taken from the limestone interbeds in the stratigraphically overlying Moratalla Marlstone Member. But in the latter the percentage of micrite is generally greater. The few thin green marlstone interbeds contain: *Orbulina universa, Globigerinoides trilobus,* and *Globorotalia* cf. *Gl. scitula.* In almost all thin sections the non-calcareous detritus (up to 5%), of medium sand size at most, is composed of glauconite, quartz, quartzite, felspar, tourmaline, colourless mica, and sporadically brown mica.

Towards the base of the Member the amount of globigerinids diminishes in favour of the amount of other fossil fragments and of the siliciclastic detritus. Nearly always *Heterostegina* and fragments of bryozoans were found. In addition the whole or a part of the following fossil association is present: many echinoid fragments, many *Lithothamnium* fragments, many smaller benthonic foraminifera, *Amphistegina,* rotalids, globigerinids and mollusc fragments, but here no radiolarians and sponge-spicules were found. This lower part is characterised in particular by the presence of reworked fossils and exotic rock-fragments. Reworked fossils are: *Discocyclina* (many), several Nummulites and *Operculina* (Eocene), very few *Globotruncana* (Upper Cretaceous), and *Orbitolina* (Lower Cretaceous). Among the exotic rock fragments were recognized lime-wackstones with *Pithonella* and *Stomiosphaera* (Represa Formation).

In the reworked slabs a gradual transition occurs towards the overlying Moratalla

Marlstone Member; here marlstone beds of the Moratalla type are intercalated which grow in importance towards the top.

In the westernmost extension of the South-Prebetic rock sequence this lower part of the Puerta Limestone Member was found in direct stratigraphic contact with the Miñano Limestone. Along the Barranco de la Puerta it is separated from the latter by the Roble Limestone.

Thickness - Along the Barranco de la Puerta it is some 50 m; in the reworked rock slabs about 6 m.

Age - The planktonic foraminifera indicate a Middle Miocene age for the upper part of the Member. The age of the lower part may also be Middle Miocene, as the stratigraphically lower Roble Limestone is supposed to be of Middle Miocene age. Paquet (1969) attributes an Oligocene age to the Puerta Limestone on account of the presence of *Lepidocyclina* (not found during the present investigation).

Asares Conglomerate Bed

Derivatio nominis – This Conglomerate only occurs in the large slabs of Puerta Limestone reworked in the Middle Miocene deposits of the Prebetic zone, and forms the basal part of this Member. A conspicuous exposure of this Conglomerate is situated in the limestone mass upon which the Cortijo de los Asares has been built (stratotype, encl. 1 and 2).

Lithology - In this exposure (fig. 18) the most conspicuous components of the Conglomerate are the tabular blocks of laminated brown calcareous sandstone of the Argos Formation. These blocks are in touch with each other and many of them are more than 1 m long. They are apparently the disrupted parts of beds of the Argos Formation.

The rock-material between these blocks consists of bioclastic lime-packstone with about 10% of quartz, quartzite, and felspar grains (up to 2 mm diameter) and with many small exotic limestone clasts of bioclastic lime-wackestone with *Pithonella* and *Stomiosphaera* (Represa Formation). In the bioclastic packstones pectinids are common and in thin sections were frequently recognized: *Amphistegina,* globigerinids, *Heterostegina,* rotalids, smaller benthonic foraminifera, fragments of echinoids, bivalves, bryozoans, and calcareous algae.

The coarseness of the components and the presence of many large, angular, platy parts of the Argos Formation indicate that this Conglomerate was deposited close to its source area and probably represents a rock-fall deposit.

This development of the Asares Conglomerate, however, differs in many aspects from the usual development found in the reworked rock slabs where it consists of intraformational conglomerates.

These conglomerates are composed of subangular to subrounded granules, pebbles and cobbles of globigerinal lime-packstone with only a few percent of angular, silt to medium sand-sized siliciclastic detritus. Within the pebbles and cobbles the closely packed globigerinal tests are generally unbroken. The pebbles and cobbles are in mutual contact. The material between the pebbles and cobbles is markedly richer in micrite than the pebbles themselves and the globigerinal tests are commonly broken.

A fragment of sandstone of the Argos Formation was found only in one other outcrop. This indicates that the two kinds of conglomerate, both occurring at the base of the Puerta Limestone, most likely represent two different developments of one and the same Conglomerate.

Thickness - 1 to 15 m.

Moratalla Marlstone Member

Derivatio nominis – called after the village of Moratalla (page 114).
Lithology - Along the Barranco de la Puerta it consists of light grey marlstones with many medium to very thick interbeds (up to 3.5 m) of generally laminated, graded, calcareous sandstones. Flute marks were found at the base of a 30 cm thick sandstone bed, from which a current-direction of N75E towards the west was measured. Three thick beds show internal contortions. The grain-size reaches a maximum of very coarse sand. The succession is interpreted as a rather proximal turbidite sequence.
Thickness - Along the Barranco de la Puerta the sequence is exposed over a thickness of 290 m.
Age - According to planktonic foraminifera a Middle Miocene age can be attributed to these rocks (Hermes, appendix in MacGillavry et al., 1963).

In the westernmost part of the Sierra de la Puerta inverted Puerta Limestone was found to lie directly upon Moratalla Marlstone of normal attitude. Hence also the stratigraphic contact between the Puerta Limestone and the Moratalla Marlstone along the Barranco de la Puerta develops westward into a fault contact. Thus the number of thrust faults in the western part of the South-Prebetic zone amounts to three (page 85, and 92). They are more or less parallel to the strike of the beds.

General aspects of the stratigraphy of the South-Prebetic zone

The Represa, Quipar, Jorquera and Moro Marlstone Formations of the South-Prebetic zone show close lithologic affinities to their equivalents in the Subbetic zone in the south. The Rasa Sandstone, Miñano Limestone, Roble Limestone Formations, and Moratalla Marlstone Member of the South-Prebetic zone have close lithologic affinities to their Prebetic equivalents in the north. The Puerta Limestone is typical for the Sierra de la Puerta. Only the Canara Formation is lithologically intermediate between the Argos and Represa Formations of the Subbetic zone and the Tosca Limestone Formation of the Prebetic zone. Therefore the name "Unité intermédiair" should be rejected in favour of "South-Prebetic zone".

Outline of the lithostratigraphic succession of the Prebetic zone

From old to young the following Groups, Formations and Members were distinguished (encl. 5):

1. Bañador Limestone Formation (oldest)
2. Muela Group
   - Bolos Sandstone Formation
   - Tosca Limestone Formation
   - Ubacas Dolostone Formation
3. Alárabe Group
   - Mojeyes Dolarenite Formation
   - Somogil Marlstone Formation
   - Almanillas Limestone Formation

Aguacerinida Limestone Member
Fresne Limestone Member
Murtas Formation
Grero Marlstone Member
Celada Sandstone Member

4. Ulea Group
Roble Limestone Formation
Trieta Formation
Puerta Limestone Member (only in South-Prebetic zone)
Moratalla Marlstone Member
Turmal Conglomerate Member
Benamor Geröllton Member with Bejar Beds (youngest)

The rocks of the Prebetic zone are folded into a boxfold, the Sierra de la Muela, which is situated northwest of a broad, basin-shaped syncline, the Moratalla basin. The Tertiary rocks of the southeastern limb of the Muela anticline have slipped down from its Mesozoic substratum forming gravity collapse structures (encl. 1, 3 and 4).

BAÑADOR LIMESTONE FORMATION

Derivatio nominis – These Limestones are mainly exposed in the southeastern slope of the valley of the Barranco del Bañador (encl. 1). They are described by Jerez Mir (1971), and Fourcade et al. (1972), by the latter with special emphasis on the algae and foraminifera. The Formation is defined by the rock units 1 up to and including 4 of section d of Fourcade et al. (1972, p. 217, 218, fig. 2 and 3).
Lithology – Irregular alternation of light brownish-grey to white, bioclastic lime-wackestones and packstones (which may be more or less recrystallized into bioclastic pseudo lime-grainstones) and pseudo-oolitic lime-grainstones (in which the rounded bioclasts generally have a micritic coating). The bedding planes are uneven and in some beds the limestones are slightly nodular. The thickness of the beds varies considerably, though they generally are about 1 m thick.

In the uppermost 12 m (unit 4 of Fourcade et al.), however, the beds are more constantly about 30 cm thick and made up of pseudo-oolitic pseudo lime-grainstone composed of abundant rounded echinoderm fragments. In this upper part two marlstone beds are intercalated and in a few beds burrows have been observed. The topmost bed consists of hard bioclastic packstone containing many iron-ooliths. The irregular upper surface of this 20 cm thick bed is iron-encrusted. The percentage of silt to sand-sized siliciclastic detritus in the Bañador Limestone is low, but in some beds it may reach 15 percent.

Siliciclastic detritus: common constituents are: quartz, quartzite, felspar (of which many grains have autigenous borders); tourmaline is an accessory constituent; micas are virtually absent. *Fossils: Clypeina jurassica* Favre, other calcareous algae, trocholinids, miliolids, *Pseudocyclammina,* other smaller benthonic foraminifera, fragments of echinoderms, molluscs, and bryozoans. Towards the top an increasing number of coral fragments were found except in the upper 12 m where *Alectryonia,* other ostreids, and gastropods were observed. The topmost bed is rich in fossils: belemnitids, rhynchonellid brachiopods, ammonites, bivalves, and again coral fragments.

Thickness – The lower boundary of the Bañador Limestone Formation is not visible, only the upper 160 m are exposed. The 40 m measured by Fourcade et al.

(1972) with respect to their rock unit 4 is too great because of repetition by faulting.

Age – On account of the ammonite association (page 104) collected 3 m above the top, the Bañador Limestone would be Tithonian-Hauterivian. According to Alloiteau (1960) the association of corals in rocks of similar lithology and stratigraphic position in the Sierra de Mariola (Prov. Alicante) is comparable with that of the Portlandian rocks of Stramberg (Carpathians). This may also be the case in the Bañador Limestone, which would imply that the rocks of the upper 12 m of the Bañador Limestone had been deposited mainly in Valanginian and/or Hauterivian times and can be correlated with the lower 700 m of the Subbetic Miravetes Formation. The remainder can be correlated with the upper part of the Subbetic Tollo Formation.

MUELA GROUP

The Sierra de la Muela is predominantly constituted of rocks of this group. A complete and representative section is exposed along the Barranco de la Muela. Jerez Mir (1971) and Fourcade et al. (1972) gave an excellent description of this section. All 25 units distinguished by them can be well matched with my own observations. Three Formations have been taken together in this Group, namely:

1. the Bolos Sandstone Formation (oldest) (Unit 5 of Fourcade et al.),
2. the Tosca Limestone Formation (Units 6-24 of Fourcade et al.),
3. the Ubacas Dolostone Formation (youngest) (Unit 25 of Fourcade et al.).

The thickness of the Group is about 900 m when measured along the barranco. At the crest of the Muela anticline the thickness only amounts to some 350 to 400 m. As the Bolos Sandstone and the Ubacas Dolostone are rather constant in thickness, the greatest lateral difference in thickness should be allotted to the Tosca Limestone, during the deposition of which a structural hinge in the basement must have come into existence with a throw of some 500 m at the place where now the Sierra de la Muela is situated (encl. 5).

Bolos Sandstone Formation

Derivatio nominis – Bolos is the name of one of the summits of the Sierra de la Muela in the northwestern flank of which the Formation crops out (encl. 1). The type section is situated along the Barranco de la Muela. The Bañador and Bolos Formations are shortly described together by Jerez Mir (1971, p. 121, 122, fig. 1, rock units 8 and 9) under the heading "Kimmeridgiense-Portlandés-Neocomiense".

Lithology – Fine, dark-grey, friable, polymictic, mica-bearing sandstone. In the uppermost 6 m exposed, the mica flakes are clearly discernable. The lowermost 5 m contain iron-ooliths which decrease upwards in abundance. Several yellowish calcareous interbeds occur in the Bolos Sandstone which consist of bioclastic packstone with a variable percentage of siliciclastic detritus of very fine sand-size.

Siliciclastic detritus: quartz, quartzite, felspar (plagioclase), and a conspicuous amount of colourless mica.

Fossils: in thin sections from the calcareous interbeds were recognized: smaller benthonic foraminifera, and fragments of calcareous algae, echinoderms, and molluscs. Macrofossils were found only in the lower 45 m of the Formation. In a 15 cm thick calcareous

bed 3 m above the base of the Formation numerous cephalopods were found: microconchs and macroconchs of *Endemoceras (?) catulloi* (Parona) occur in abundance; far less numerous are: *Crioceratites (Sornayites) simionescui* (Sarkar), *Anahamulina* sp., *Desmoceras* cf. *D. tenuicinctum* (Sar. & Schön.), *Cymatoceras radiatum* (Sow.) (nautilid). In the following 42 m of the Bolos Sandstone were found: *Barremites difficilis* (D'Orb.), *Raspailiceras* cf. *R. cassidoides* (Sar. & Schön.), *Hamulina haueri* Uhlig, many indeterminable ammonites, many large *Exogyra,* rhynchonellid and terebratulid brachiopods, belemnites, and both regular and irregular echinoids.

Thickness – The upper boundary was chosen at the base of the first pseudo-oolitic limestone bed of the Tosca Limestone Formation. The thickness when measured along the Barranco de la Muela amounts to 80 m. Fourcade et al. report a thickness of 40 m near the Cortijo de la Muela.

Age – The ammonite association indicates an Early Barremian age. The lower part of the Bolos Sandstone Formation can be correlated with the upper part of the Miravetes Formation of the Subbetic zone. On account of the large amount of mica the upper part of the Formation is tentatively correlated with the lower part of the Subbetic Argos Formation.

Tosca Limestone Formation

Derivatio nominis – This Formation is called after the Cortijo de la Tosca, a farm just below the summit of the Muela de Moratalla, which is called Puntal de la Tosca by the local population (encl. 1). Its type section is situated along the Barranco de la Muela (encl. 1). The Formation is shortly described by Jerez Mir (1971) under the heading: "Cretácico inferior del sector meridional, b) Facies central o interior del surco sedimentario" (p. 122-124, fig. 1, rock units 10-12).

Lithology – Very heterogeneous, though predominantly composed of intervals of medium to thick-bedded, grey to white pseudo-oolitic (bioclasts generally rounded with a micritic coating) and bioclastic lime-packstones (bioclasts generally without micritic coating). The lime-packstones may have been recrystallized into pseudo lime-grainstones. Also bioclastic and oolitic lime-grainstones were found. Several intervals consist of bioclastic lime-wackestones and mudstones. In the limestones the amount of siliciclastic detritus does not exceed a few percent. The grain-size is that of medium sand at the most. Extensive reworking has taken place during the deposition. In many samples, for instance, both ooliths and pseudo-ooliths may be present, cemented together by micrite. Also a few true ooliths may be present in badly sorted bioclastic lime-packstone in which the bioclasts are not worn off and do not have a micritic coating. In these cases the true ooliths are supposed to have been reworked. In a few samples reworked limestone clasts were found which are of slightly different lithology than that of the surrounding limestone. Also some coral fragments carry a veneer of rock of somewhat different lithology and are therefore considered as reworked. In the upper 250 m of the Tosca Formation a variable percentage of euhedral dolomite crystals are present in the limestones. They have grown independently from the textures of the limestones and most of them show a cloudy centre and a clear rim; they are therefore of secondary origin. Several beds even have been entirely dolomitized, a feature, which is a precursory symptom of the overlying Ubacas Dolostone Formation.

Between the limestones intervals of softer rocks occur, which are only locally exposed and appear to consist of friable, grey and ochre-coloured, mica-bearing sandstones, in one interval with plant remains and iron concretions, in another with numerous small limestone concretions and a large amount of

glauconite. Also beds of green shales were found.

Siliciclastic detritus: quartz, quartzite, quartz-felspar rock fragments, mica-schist fragments, felspar (among which plagioclase, perthite, and microcline), colourless and brown mica.
Fossils: in thin sections were recognized: many kinds of calcareous algae, orbitolinids (particularly abundant about 100 m below the top of the Formation), miliolids, many other smaller benthonic foraminifera, ostracodes, aptychi, fragments of echinoderms and molluscs. A green shale bed 214 m above the base of the Formation (unit 13 of Fourcade et al., 1969) yielded *Atopochara trivolvis* Peck.

In the field were sampled: *Acteonella*-like gastropods, large *Natica*-like gastropods, nerineids, brachiopods, ostreids (concentrated in abundance in a few beds), and rudists (for instance *Monopleura)* which were also found in growth position and which are particularly abundant about 250 m below the top (top of unit 17 of Fourcade et al., 1972), where many *Pseudotoucasia santanderensis* and *Polyconites* cf. *P. verneuilli* (determination van Dommelen) were collected. Also one large, rounded, quartzite pebble was found which was extremely polished even in the shallow depressions of the surface. In view of the exceptional size of this unique find (the largest size of non-calcareous detritus ever found in the Muela Group is 600 μ) and on account of the particularly striking polish, the pebble was interpreted as a gastrolith (stomach-stone).

In the Barranco de Hondares about 6 km southwest of the Barranco de la Muela the few topmost metres of the Formation yielded the following fauna: *Mariella bergeri crassituberculata* (Spath), *Ostlingoceras puzosianum* (D'Orb), *Anisoceras picteti* Spath, *Mortoniceras rostratum* (Sow.), *Mantelliceras martimpreyi* (Coq.), *Stoliczkaia dispar* (D'Orb.), *Puzosia* cf. *P. planulata,* (Bayle), *Protocardia* cf. *P. hillana* (Sow.), *Janira quinquecostata* D'Orb., gastropods, and irregular echinoids. Upstream along the same barranco several sandstone intervals of the Tosca Formation contain large amounts of amber which was exploited in historical times.

Thickness – The upper boundary was chosen at the top of the uppermost limestone bed. The thickness of the Tosca Limestone Formation measured along the Barranco de la Muela is 627 m. The thickness is considerably less on the crest of the Muela anticline. There it is estimated to be about 150 m.
Age – *Atopochara trivolvis* has hitherto been found only in rocks to which an Aptian age has been attributed. Therefore probably rocks of Aptian age occur in the lower two thirds of the Formation. Wiedmann (1965) claims, in view of the ammonite faunas, that in Spain the association of *Pseudotoucasia santanderensis* and *Polyconites verneuilli* occurs in the Clansayesian and probably Lower Albian. The ammonite association in the topmost beds of the Formation indicates a latest Vraconian age. Therefore an Albian age has been attributed to the rocks of the upper 250 m of the Formation. The Aptian-Albian boundary should be drawn within unit 18 of Fourcade et al. (1972).

The lower two thirds of the Formation may be correlated in age with a part of the Subbetic Argos Formation. The reworked ooliths in the latter may have been derived from the Tosca Formation. At least a part of the upper third of the Tosca Formation can be correlated in age and to a lesser degree in lithologic respect with the Canara Formation (South-Prebetic) and with the Represa Formation (Subbetic).

Ubacas Dolostone Formation

Derivatio nominis – This Formation is shortly described by Jerez Mir (1971, p. 125, fig. 1, rock unit 13) under the heading "Complejo dolomítico del Cenomanense-Turonense". It is named after the Barranco de Ubacas which transects the whole Formation (reference section). Its type section is situated along the Barranco de Hondares directly upstream of the Baños de Mojeyes (encl. 1).
Lithology – White and yellowish sugary dolostones devoid of siliciclastic detritus.

The rock is generally friable. The colour of wheathered surfaces is generally light grey.
Fossils – In some thin sections, even from close to the top of the Formation, orbitolinids have unmistakably been recognized.
Thickness – The upper boundary was chosen at the top of the uppermost dolostone bed. The thickness of the Ubacas Dolostone Formation measured along the Barranco de la Muela is 193 m.
Age – On account of the presence of orbitolinids and of the latest Vraconian age of the rocks at the very top of the Tosca Limestone, we may attribute a Cenomanian age to the Ubacas Dolostones. But the orbitolinids may have been reworked, so that a younger age may still be possible.

Rocks of Cenomanian age are thin or perhaps locally even absent in the stratigraphic successions of the Buitre unit and of the Sierra de la Puerta. Perhaps an unconformity may account for this, though a faunal hiatus has never been found as yet. The existence of conglomeratic marlstone of Turonian age in de Sierra de la Puerta, in which inclusions of "cenomanien inférieur" (Paquet, 1969, including the Vraconian) have been found, indicates that rocks of older age than Turonian were eroded during that time. Paquet (1969, p. 96) reports the presence of reworked fragments of orbitolinids in Turonian micro-conglomerates (Subbetic zone). The otherwise inexplainable occurrence of large rudist fragments near the base of the Quipar Formation (van Veen, 1969, p. 73) may be explained by assuming a slight angular unconformity between Cenomanian and Turonian rocks along which the erosion products of the truncated underlying strata have been reworked (encl. 5). In fact Paquet (1969, fig. 42 bis A) draws an angular unconformity between the Cenomanian and Turonian strata of the Loma de Solana Unit.

Depositional environment of the Muela Group

The lithologic and faunistic characteristics indicate a deposition mainly in a marine sublittoral environment above wave base. The local abundance of amber even may indicate temporary emergence above sea level. The section of the Sierra de la Muela may compete in importance with the classical one of the Sierra Mariola (Prov. Alicante, Busnardo & Durand Delga, 1960).

ALÁRABE GROUP

This Group is named after the Río Alárabe, the upper course of the Río de Moratalla (erroneously called Río Benamor on the topographic map sheet number 889: Moratalla). The rock sequence has already been described by Fallot (1945) and by Jerez Mir (1971) under the heading "Nummulitico S. lat." (p. 125-127). The following Formations and Members are grouped together in the Alárabe Group:

1. Mojeyes Dolarenite Formation (oldest)
2. Somogil Marlstone Formation
3. Almanillas Limestone Formation
   Aguacerinida Limestone Member
   Fresne Limestone Member

4. Murtas Formation
   Grero Marlstone Member
   Celada Sandstone Member (youngest)

The thickness of the Group is about 658 m when measured along the Río Alárabe. Near the Cortijo de los Campaneros the thickness is about 350 m and considerably diminishes towards the northeast. Just beyond the northwestern corner of the investigated area (northwestern limb of the Muela anticline) the Alárabe Group is absent and rocks of the Roble Limestone Formation are directly overlying those of the Muela Group. In the area west of the Sierra de la Muela the Group is very thin (internal report by van Heerden, Geological Institute of the University of Amsterdam). This implies that also during the deposition of the Alárabe Group a structural hinge existed in the basement, at the place where now the Sierra de la Muela is situated. This is probably the same hinge that caused the difference in thickness of the Tosca Formation. As the topographic expression of this hinge has been blotted out by the Ubacas Dolostone Formation, a renewed growth of the hinge must be assumed after the deposition of the Ubacas Dolostones, possibly concurrent with a primordial upwarping of the Muela anticline. The additional 600 m makes the total throw of the hinge more than 1 km (encl. 5).

Mojeyes Dolarenite Formation

Derivatio nominis – This Formation is best exposed along the Barranco de Hondares near the Baños de Mojeyes (type section), a spring that wells up from one of the oblique peri-anticlinal faults of the Muela anticline (encl. 1).
Lithology – Alternately more and less resistant beds of arenite composed of dolomite crystals, presumably a more or less cemented detritus derived from the directly underlying Ubacas Dolostone Formation. This reworking can be inferred from the fact that these dolomite crystals are associated with many well preserved non-dolomitized calcareous bioclasts. The faces of the euhedral to subhedral dolomite crystals are not crossing the boundaries of these bioclasts, though some agglutinant foraminifera have used some dolomite crystals to build their tests. Some internal cavities of the foraminifera are filled with a cement of anhedral dolomite crystals interfingering like a jigsaw puzzle. The percentage and the average size of the non-calcareous detritus diminishes towards the top (from 15% to 2%, and from fine sand to silt size).

Siliciclastic detritus: quartz, quartzite, felspar (among which perthite and plagioclase), colourless mica, and tourmaline were recognized.
Fossils: in thin sections globorotalids, globigerinids, smaller benthonic foraminifera and rotalids were observed. In the lower part of the Formation only a few fragments of echinoids and bivalves were found. In the field burrows were seen.

Thickness – The range of the detrital dolomite crystals determines the boundaries of the Formation. The thickness measured along the Barranco de Hondares is 9 m.
Stratigraphic relation – When tracing the upper boundary of the Ubacas Dolostone in northeastern direction the overlying beds continue to consist mainly of detrital dolomite crystals, but gradually show more characteristics of the younger rock units of the Alárabe Group. In the valley of the Arroyo de las Murtas calcareous algae, alveolinids, and occasionally siliciclasts of granule size were found in the Mojeyes Formation; this combination is typical for the upper part of the Somogil Formation and for the Almanillas Formation. In fact in the valley of the

Murtas the former unit is in direct contact with the Mojeyes Formation; the lower part of the Somogil Formation tapers out. From these data it can be inferred (1) that the Mojeyes Formation represents the transgressive facies of both Somogil and Almanillas Formations, (2) that the relation between these Formations and the Ubacas Formation is an overstep relation, and (3) that the relation between the various units within the Somogil and Almanillas Formations can be designated as an onlap relation (see enclosure 5).

Age – The age of the Mojeyes Formation varies at the different outcrop localities. According to the ages of the directly overlying rocks in the investigated area, the age of the Mojeyes Formation ascends from latest Ypresian to Lutetian.

Somogil Marlstone Formation

Derivatio nominis – This Formation is best exposed near the Cortijos de Somogil (type area), in the widening of the Alárabe valley (encl. 1). The type section is situated along the Barranco de Hondares and the Río Alárabe.

Lithology – This Formation is rather heterogeneous. A large part consists of green marlstones with a small amount of siliciclastic detritus. Several intervals with beds of yellowish calcareous sandstone or bioclastic limestone occur in these marlstones. These intervals become more important towards the top of the Formation.

In the lower 191 m of the Formation the green marlstones predominate. The sandstone interbeds are generally graded, laminated, and may show flute-marks at their base and ripples at their top. The coarse basal parts of the lowest sandstone interbeds are almost exclusively composed of larger foraminifera. Many of these sandstone beds are considered to have been deposited by turbidity currents. The three intervals of yellowish bioclastic lime-packstone beds (lime-grainstones are subordinate) in the lower 191 m generally contain a very small amount of angular, sandsized siliciclasts.

The lithology of the upper 141 m of the Somogil Marlstone Formation is transitional between that of the lower part of the Formation and of the overlying Almanillas Limestone Formation. The green marlstones are subordinate but were found up to the top of the Formation. The limestone and sandstone intervals predominate and become more important towards the top of the Formation. The yellowish calcareous sandstones are coarse-grained, generally conglomeratic, and characterised by the presence of abundant very smooth, ovoid quartzite granules and pebbles, which are absent in the lower 190 m of the Formation. The bioclastic lime-packstones, in many beds recrystallized into bioclastic pseudo lime-grainstones, are in contrast to the lower 190 m of the Formation white, thick-bedded and also commonly contain many quartzite granules and pebbles.

Siliciclastic detritus: mainly quartz and quartzite grains, invariably with a few felspars (among which plagioclase, microcline, and perthite); in several samples, quartz-felspar rock fragments, colourless mica, tourmaline and zircon were found; in a few samples garnet and rhyolite fragments were recognized.

Fossils: the marlstones of the lowest 133 m of the Formation yielded the following dwarfed planktonic foraminifera; *Globorotalia aragonensis, Gl. soldadoensis, Gl. wilcoxensis, Gl. broedermanni, Globigerina linaperta, G. primitiva, Pseudohastigerina premicra* and *P. micra.* This association is characteristic for the *Globorotalia formosa* zone. The limestone intervals in the lowest 133 m contain many globigerinids and globorotalids, but also *Discocyclina, Asterocyclina, Amphistegina, Operculina,* fragments of echinoids and smaller benthonic foraminifera. In the coarse-grained basal parts of the sandstone interbeds in the lowest 133 m in addition *Nummulites, Assilina,* and alveolinids were recognized. In the limestone intervals above the lowest 133 m of the Formation almost no globigerinids and globorotalids were

found. On the other hand many fragments of calcareous algae were observed among which *Distichoplax biserialis.* Also *Nummulites,* (the macrosphaeral protoconchs exceeding 500 μ in diameter) *Assilina, Asterocyclina, Discocyclina, Operculina, Amphistegina,* alveolinids, rotalids, miliolids, *Halkyardia, Lockhartia,* other smaller benthonic foraminifera, and both regular and irregular echinoids were recognized. The uppermost 60 m of the Formation, which is composed of conglomeratic calcareous sandstone with sandy limestone interbeds, contain a comparable fauna; notable, however, is the absence of *Amphistegina* and *Operculina.*

Thickness – 332 m, when measured along the Barranco de Hondares and the Río Alárabe. Near the Cortijo del Salto the lower 191 m are absent.

Age – The planktonic foraminifera from the lower 133 m of the Formation are indicative for the *Globorotalia formosa* zone, which would, according to Lutenbacher (1964), be of latest Ypresian age. The large diameter of the protoconchs of the macrospheric *Nummulites* from the upper 199 m of the Formation already indicates a Lutetian age, probably Early Lutetian, because the overlying Almanillas Formation is of Late Lutetian age.

The Somogil Marlstone Formation has many lithologic affinities with the South-Prebetic Rasa Sandstone Formation. In age it can be correlated with the members A and B of the Subbetic Gredero Formation (van Veen, 1969).

Remarks – One thick laminated sandstone bed in the middle of the Formation has been folded into a recumbent isoclinal fold. The zone of maximal curvature does not show any fracturing and the upper limb of the folded bed is undulating and tapering. It is therefore assumed that the deformation had taken place before the sandstone bed became indurated. The fissility of the green marlstones surrounding this sandstone bed also shows strongly folded structures, which are truncated by the subhorizontal, non-deformed overlying sandstone bed. The structures of this deformed interval are interpreted to be due to sliding. This would imply the existence of a sloping surface at that time. The direction of this sliding, as inferred from the strike of the fold axis, would have been N 140 E towards the southeast, which is about perpendicular to the strike of the Muela anticline, where the structural hinge in the basement is located.

Almanillas Limestone Formation

Derivatio nominis – The Cerro de las Almanillas (encl. 1) consists for the greater part of rocks of this Formation. Along the Río Alárabe, where the type section is located, the total thickness is 154 m.

Aguacerinida Limestone Member

Derivatio nominis – The rocks of this Member form a large cuesta enclosing a wide stretch of the valley of the Río Alárabe. At both ends of the bowl-shaped depression the river breaks through this limestone sequence. In the eastern breakthrough the type section is located. The cuesta is known as the Aguacerinida (encl. 1).

Lithology – Generally thick-bedded bioclastic lime-grainstones and pseudo lime-grainstones of which several samples appeared to be bioclastic lime-packstones in origin. The amount of siliciclastic detritus is generally less than 5%, but in several horizons the percentage may amount to 30% and the grain-size to that of granules.

Siliciclastic detritus: generally composed of quartz and quartzite, commonly accompanied by felspar and tourmaline. In a few samples colourless and brown mica was recognized enclosed in the quartzite fragments.

Fossils: the impoverishment and specialization of the fossil association that already started in the top part of the Somogil Formation, is continued in the Aguacerinida Member. This change can be followed in 5 stages (or assemblage zones) of which four are incorporated in the Aguacerinida Limestone. (1) Whereas the topmost 60 m of the Somogil Formation is characterized by the disappearence of *Amphistegina* and *Operculina, (2)* the lowermost 10 m of the Aguacerinida Member is distinguished by the disappearence of discocyclinids concomitant with an increase in the number of calcareous algae; the presence of *Assilina* becomes obvious. (3) The next 45 m are characterized by the increasing number of calcareous algae, of large *Nummulites,* of large flosculinized alveolinids, and by the disappearance of *Assilina.* (4) The next 57 m are characterized by the appearance of *Orbitolites* and the abundant presence of complete calcareous algae (among which *Distichoplax)* with fructifications (frequently, meandering and horseshoe cross-sections were observed), miliolids, large flosculinized alveolinids and many echinids *(Eupatagus).* Large *Nummulites* are restricted to a few beds (which generally contain a relatively large amount of siliciclastic detritus). (5) In the uppermost 10 m of the Aguacerinida Limestone only calcareous algae, miliolids and a few other smaller benthonic foraminifera are present. No alveolinids and *Nummulites* were found anymore. Among the *Nummulites* in the Aguacerinida Member were determined (by J. Werner) *N. perforatus, N. brongniarti* and *N. aturicus.*

Thickness – The lower boundary was chosen at the base of the limestone unit that forms the cuesta. Along the Río Alárabe the thickness of the Aguacerinida Limestone Member amounts to 123 m. At the northern boundary of the investigated area it is only a few metres.

Age – According to the *Nummulites* a Late Lutetian age could be attributed to the Aguacerinida Limestone Member. The Aguacerinida Limestone can be correlated with respect to its lithology and age with the Miñano Limestone Formation (South-Prebetic zone). In age it can be correlated approximately with member C of the Subbetic Gredero Formation (van Veen, 1969).

Stratigraphic relation – In the valley of the Arroyo de las Murtas the colour of the few topmost metres of the Aguacerinida Limestone gradually changes from white to purple-red concomitant with a gradual increase in the amount of quartzite pebbles. This change culminates in a brownish-red sandy conglomerate composed of quartzite pebbles with sandstone in the interstices. Near the Cortijo de los Campaneros the conglomerate is 1 m thick and directly overlain by the Fresne Limestone Member. At the highest point reached by the road from Moratalla to las Murtas (about 2 km to the southwest) the conglomerate is only 20 cm thick and overlain by a 3 m thick sequence with exactly the same lithology as that found in the upper 10 m of the Aguacerinida Limestone along the Río Alárabe (another 3 km to the south). In the latter locality the conglomerate is reduced to a 30 cm thick bed of white bioclastic lime-packstone with a sparse sprinkling of quartzite pebbles. Here this bed is situated approximately at the base of assemblage zone 5, thus within the Aguacerinida Member itself.

The relation of the various rock units is interpreted in the same way as for the Mojeyes Formation. The uppermost part of the Aguacerinida Member together with the Fresne Member are in overstep relation with the remainder of the Aguacerinida Member, whereas an onlap relation exists within the overstepping rock units. The quartzite conglomerate represents a regressive phase during the deposition of the Aguacerinida Limestone itself.

Fresne Limestone Member

Derivatio nominis – Extensively exposed near the Cortijo del Fresne (encl. 1). The type section is located along the Río Alárabe.

Lithology – At first sight the Fresne Limestone seems to be homogeneous lutitic

limestone. But this Member appears to consist of white intraformational limestone conglomerate composed of granules to boulders of lime-mudstone with in the interstices lime-mudstone that may contain a variable amount of siliciclastic detritus, and that may be slightly more translucent than the lime-mudstone of the granules and pebbles (probably due to slight recrystallization). Locally the limestone consists of closely packed, rounded lime-mudstone clasts of 100 to 400 μ. Notwithstanding the irregular shapes of the clasts they fit well into each other; the few spaces between them are filled with sparite. In a few samples this texture was also discovered between the large pebbles of the limestone conglomerates.

Siliciclastic detritus: sand-sized quartz, quartzite and felspar.
Fossils: a few *Natica*-like gastropods were found. In thin sections bioclasts are virtually absent.

Thickness – In the valley of the Arroyo de las Murtas the lower boundary was chosen at the top of the red quartzite conglomerate. But along the Río Alárabe the lower boundary was chosen at the top of the upper sandy algal limestone bed of the Aguacerinida Member. Here the thickness amounts to 31 m; 3 km to the northeast it is only 14 m thick.
Age – Unknown, but the apparently uninterrupted sequence from Aguacerinida Limestone to Fresne Limestone suggests a latest Lutetian or perhaps a Priabonian age. The latter age has been attributed to it by Fallot (1945).

Murtas Formation

Derivatio nominis – The southeastern slope of the valley of the Arroyo de las Murtas is constituted for the greater part of rocks of this Formation with accessible exposures along the road from Moratalla to las Murtas (encl. 1). The section (type section) described was taken from the highest point of this road and along it in southern direction. The total thickness of the Murtas Formation measured along this road is 190 m. It is thickest along the Río Alárabe and thins towards the northeast. It has been shortly described by Jerez Mir (1971).

Grero Marlstone Member

Derivatio nominis – The Barranco del Grero has eroded a deep valley in the soft rocks of this Member (encl. 1). The type section is the same as that of the Murtas Formation.
Lithology – Heterogeneous, but consisting predominantly of grey, greyish-green and reddish-brown marlstones, which at several places contain a notable amount of sand-sized siliciclastic detritus. Between these marlstones several white generally thin to medium thick beds occur, made up of soft, nodular lime-mudstone. A few limestone beds, however, are thicker, harder, and consist of bioclastic lime-packstone or wackestone and one limestone bed is composed of pseudo-oolitic lime-grainstone. Along the Río Alárabe even gypsiferous intercalations were found. Near the base of the Member there is a 13 m thick interval of coarse, brownish-red quartz sandstone with a few quartzite granules and pebbles. This interval is overlain by a few metres of fine grey sandstone.

Siliciclastic detritus: quartz, quartzite, chert-fragments and tourmaline.
Fossils: most wash-residues from the marlstones appear to be devoid of fossils. A few samples, however, yielded: smooth-shelled ostracodes, oogonians of *Chara, Microcodium* also in accumulations; reef?), skeletal parts of holothuroids and ophiuroids, miliolids, pene-

roplids and agglutinating benthonic foraminifera. Thin sections made from the bioclastic limestone interbeds predominantly contain miliolids and peneroplids, a few globigerinids, *Amphistegina,* other smaller benthonic foraminifera *(Nonion)* skeletal parts of ophiuroids, cross-sections through *Chara*-stems, *Microcodium* and other calcareous algae. One of the lime-mudstone interbeds contains many *Cerithium.*

Thickness – The upper boundary is unknown, but must be situated within the 77 m thick interval that is not exposed along the road. The exposed thickness of the Grero Marlstone Member measured along the road is 105 m.

Age – Durand Delga & Magné (1958) have collected some fossils from a few levels about 50 m from the base and attributed a Chattian and (or) Aquitanian age to the sampled levels. During that time the top part of the Subbetic Gredero Formation was deposited (van Veen, 1969).

Celada Sandstone Member

Derivatio nominis – Well-exposed along the Barranco de la Celada (reference section, encl. 1); but its type section is the same as that of the Murtas Formation.

Lithology – Along the road from Moratalla to Las Murtas, it consists of coarse, friable sandstone. The first 5 m exposed are constituted of an alternation of brown and grey beds on top of which a 2 m thick purple sandstone bed occurs. This is overlain by a 4 m thick sandstone sequence in which from bottom to top the colour changes from dark red-brown into pale-brown. The dark-brown sandstone fills erosion gullies in the underlying purple sandstone. In these gullies rounded quartzite pebbles, plant remains and reworked fragments of the purple sandstone are concentrated. The top is formed by an 1 m thick bed transitional to the overlying Roble Limestone Formation and composed of calcareous conglomeratic sandstone.

Thickness – The lower boundary is unknown, but must be situated within the 77 m thick interval that is not exposed along the road. The level directly below the first appearance of *Lithothamnium*-oncoliths halfway the conglomeratic sandstone was chosen as the upper boundary. The exposed thickness is 11 m.

Age – Unknown, but, on account of the ages of the underlying rocks, probably Miocene.

Depositional environment of the Alárabe Group

The succession of the rocks of the Alárabe Group gives the impression that it was deposited in a marine environment which gradually became shallow, and that finally even temporary exundations occurred.

The rocks of the lower third part of the Somogil Marlstone Formation were probably deposited largely below wave base with occasional influxes of turbidity currents. The transgression initiated in the Mojeyes Formation was succeeded by rapid deepening of the sea: planktonic foraminifera are common. In the younger units of the Alárabe Group, however, the planktonic foraminifera were only sporadically found and they are absent in the Grero Member (except in a few samples of the bioclastic limestones). The Somogil and Almanillas Formations become more calcareous towards the top, concomitant with an impoverishment and specialisation of the fossil association. The occurrence of complete calcareous algae with fructifications in the upper part of the Aguacerinida Limestone indicates that the associated fauna is autochthonous. Perhaps the lime-mudstones of the Fresne

Member may represent the climax of this directed change in facies. The deposition of the upper part of the Somogil Formation and the Almanillas Limestone is thought to have taken place in a very shallow marine environment. The intensive red colour of the conglomerate bed near the top of the Aguacerinida Limestone probably indicates a partial exundation of the sea bottom. The peculiar fossil association of the Grero Marlstone Member is characteristic for a lagoonal environment (Jerez Mir, 1971). This is confirmed by the presence of gypsum beds which indicates restricted water circulation. The existence of a barrier between the open sea and the lagoon can be inferred from the stratigraphic columns given on enclosure 5. The lagoonal environment may already have come into existence during the deposition of the Aguacerinida Limestones and the Fresne Limestones, which would explain the specialisation of the fossil community. The purple sandstone interval in the Celada Member with at its top erosion gullies containing plant remains represents another period of emergence. It is quite possible that the Celada Sandstones had been deposited in a littoral environment.

ULEA GROUP

This Group is named after the large estate in the middle of the Moratalla basin where the road from Moratalla to Calasparra crosses the Rambla de Ulea (map sheet 890, Calasparra, and encl. 4). The following rock-units are taken together in this group:

Prebetic zone:
Roble Limestone Formation (oldest)
Trieta Formation
Moratalla Marlstone Member
Turmal Conglomerate Member
Benamor Geröllton Member
(with Bejar Beds) (youngest)

South-Prebetic zone:
Cuervo Sandstone Formation
Trieta Formation
Puerta Limestone Member
(with Asares Conglomerate at its base)
Moratalla Marlstone Member

Near Moratalla the thickness of the Group was estimated at 350 m. Large quantities of rocks from the South-Prebetic and Subbetic zones were redeposited in the rocks of the Ulea Group and this reworking may even be considered as characteristic for the Trieta Formation.

Roble Limestone Formation

Derivatio nominis – The rocks of this Formation constitute the major part of the Sierra del Roble (encl. 1).
Lithology – Along the road from Moratalla to Las Murtas (type section) the calcareous conglomeratic sandstone of the basal bed (1 m thick) is transitional to that of the Celada Sandstone, the amount of calcite increases towards the top where the lithology changes into bioclastic lime-grainstones with 35% sand-sized siliciclasts and numerous quartzite granules and pebbles. This bed is overlain by 3.50 m bioclastic lime-grainstone with a lower amount of sand-sized siliciclasts (15% at the most), which disappears completely in the next 10.50 m of bioclastic lime-packstone. Along the Río Alárabe, however, the entire sequence contains a considerable percentage of siliciclastic detritus (reference section).

Siliciclastic detritus: predominantly quartz and quartzite but also brown mica and tourmaline.
Fossils: in the basal bed many *Lithothamnium*-oncoliths and *Chlamys* were sampled.

In the following 3.50 m two levels with abundant *Amphiope* were found. In the upper 10.50 m numerous large *Lithothamnium*-oncoliths occur. In thin sections many *Lithothamnium, Amphistegina* and *Heterostegina* were recognized among which *Heterostegina complanata* Menighini and *H. complanata spiralis* Papp & Küpper has been determined (J. P. Krijnen). Also globigerinids, smaller benthonic foraminifera and fragments of echinoids, bryozoans and molluscs are present.

Thickness – The lower boundary was chosen at the first appearance of *Lithothamnium* above the Celada Sandstone, the upper boundary at the top of the algal limestone. The thickness measured at the outcrop transected by the road from Moratalla to Las Murtas is 14 m. Along the Río Alárabe it amounts to 40 m.
Age – The concurrent range of *Heterostegina complanata* and *H. complanata spiralis* would indicate a "Late Miocene or Pliocene" age. In the mediterranean region *Amphiope* exclusively occurs in "Helvetian" rocks. As the base of the overlying Moratalla Marlstone contains *Orbulina universa* and *Globorotalia fohsi peripheroronda,* indicating a late Langhian age, it seems justified to attribute a Middle Miocene age to the Roble Limestone.

Trieta Formation

Derivatio nominis – Trieta is the old roman name for Moratalla.

Moratalla Marlstone Member

Derivatio nominis – This Member is named after the village of Moratalla which is built upon the rocks of this Member (encl. 1). The type area is the investigated part of the valley of Moratalla.
Lithology, fossils and non-calcareous detritus – The lithology slightly differs north and south of the Río de Moratalla. North of the Río de Moratalla it mainly consists of a thick, rather homogeneous sequence of soft, white, laminated bioclastic lime-wackestones with bedding fissility. They contain globigerinids, globorotalids, radiolarids and sponge-spicules. Only at the base a few harder, light-grey beds of medium thickness are intercalated, which consist of globigerinal lime-packstone with 5 to 10% sand-sized angular siliciclastic detritus.

In these beds abundant globigerinids, radiolarians, sponge-spicules, and fragments of *Amphistegina, Heterostegina,* rotalids, bryozoans, molluscs and *Lithothamnium* were observed. Also reworked orbitolinids and dolomite crystals were found, presumably derived from the Ubacas Dolostone and the Tosca Limestone which at that time apparently were exposed to erosion.

South of the Río de Moratalla the lower part of the Member, of which only the upper 25 m are exposed, consists of grey marlstones with bedding fissility. At specific levels 1 to 2 m thick intervals with grey, somewhat nodular limestone beds of medium thickness occur composed of globigerinal lime-packstones with a variable percentage of sand-sized siliciclastic detritus. The upper 50 m are constituted of grey marlstones with a low percentage (up to 2%) of fine silt-sized siliciclasts. In these marlstones numerous laminated, calcareous sandstone beds are intercalated, which vary in thickness from 1 mm to 67 cm, though mainly thin-bedded (fig. 19). The percentage of angular, siliciclastic detritus in these sandstone beds is about 50%. Apart from a few exceptions the grain size does not exceed 500 μ, the largest volume is occupied by grains between 100 and 200 μ.

The siliciclastic detritus of these sandstone beds is mainly composed of quartz and quartzite grains, but also many felspar grains (among which plagioclase and microcline) were found.

Fig. 19. Turbidites in the Moratalla Marlstone Member along the Río Benamor.

Fig. 20. Close-up of a turbidite showing a parallel-laminated, cross-laminated, and a convolute-laminated division. Moratalla Marlstone Member, near the mouth of the Barranco de Azafran.

Colourless mica, glauconite, tourmaline and zircon are present in minor quantities. The material between the siliciclasts is composed of bioclasts and micrite, which is commonly recrystallized into sparite. The bioclasts are mainly globigerinids. Also sponge-spicules, radiolarians, *globorotalids,* smaller benthonic foraminifera, and fragments of *Heterostegina,* rotalids, *Lithothamnium,* echinoids and molluscs occur.

Next to the calcareous sandstone beds also light-grey laminated interbeds of globigerinal lime-packstones were found in the marlstones.

They contain abundant globigerinids, sponge-spicules, radiolarians, smaller benthonic foraminifera, and fragments of calcareous algae and echinoids.

The sandstone beds generally are only slightly graded and show parallel lamination (fig. 20). The upper and lower boundaries of the sandstone beds are sharp. The thickest beds however show a marked grading and the following sequence of division characteristic for turbidites: a coarse grained basal part, a parallel laminated middle part which is always thickest, an upper part with distinct current-ripple lamination and convolute lamination, and in one bed a thin fine-grained top part. In other beds the structure of the latter division is obliterated by intensive bioturbation. The grey marlstone between the sandstone beds may represent wholly or partly the upper lutitic division; they contain reworked Cretaceous and Lower Tertiary planktonic foraminifera. The thinnest sandstone beds do not show internal structures. The current directions inferred from the few reliable measurements on suitable current-laminated outcrops are from the east and southeast (N 140 E, N 120 E and N 90 E.). The base of most sandstone beds is flat; only in a few sandstone beds were load marks observed at the base; only in one bed were flute marks found, from which also a current direction from the southeast can be inferred (N 140 E to N 120 E). The upper surface of a few sandstone beds shows current ripples indicating a current direction from the east and southeast (N 147 E, N 122 E, N 117 E, N 100 E, N 90 E and N 85 E). A peculiar structure found is that several thin sandstone beds show parallel long-stretched undulations affecting the whole bed (fig. 21). In the crestal parts of the undulations the bed is thinnest, whereas in the depressions it is thickest. The maximum thickness, however, is found on the side and always on the same side of the central part of the depressions. The marlstone bed immediately below shows flame-like structures pointing towards the crests of the undulations. Tentatively these undulations are interpreted as ripples that subsequently became loaded into the visco-plastic marl below, which produced flame structures. The current directions inferred from the strike of these undulations also are from the southeast (N 135 E, and N 140 E). The measurements on the current directions suggest that the latest currents came from the east to southeast (N 85 E to N 147 E). The internal structures and the reworked fossils found in the calcareous sandstone beds strongly suggest deposition by turbidity currents.

In one exposure (in the northern branch of the Barranco de Agüica) three successive beds of non-graded calcareous sandstone were found, which contain many randomly distributed rounded quartzite pebbles (fig. 22). Only in the lower bed are the pebbles concentrated in the basal part. These beds are thought to have been deposited by sand-avalanching. Since the only Formation known in the southeast containing similar pebbles is the Eocene Rasa Sandstone Formation, presumably a great part of the sand grains of the turbidites of the Moratalla Marlstone is derived from the Rasa Formation, which at that time would have been subject to erosion.

Also the more calcareous interbeds in the marlstones, which mainly differ

Fig. 21. Turbidite sequence showing load marks and peculiar parallel undulations in which several beds are involved. These undulations are probably due to loading of ripples. Moratalla Marlstone Member, near the mouth of the Barranco de Azafran.

from the marlstones in the amount of globigerinids and from the calcareous sandstones only in the amount of siliciclastic detritus, may be deposited by turbidity currents, though apparently lacking the association of structural features characteristic for turbidites.

Characteristic for the grey marlstones is the consistent reworking of Cretaceous and Lower Tertiary planktonic foraminifera. This indicates that during the deposition of this part of the Moratalla Marlstone Cretaceous and Tertiary rocks, mainly with pelagic foraminifera, which are only found in the Subbetic and South-Prebetic zones, were eroded. Also the grey colour of the marlstones in the southern part of the Moratalla basin may therefore be explained by the intensive reworking of the grey marlstones of the Lower Cretaceous Represa formation.

In the bed of the Río Benamor about 150 m west of the bridge where the road from Caravaca to Moratalla crosses the river, the Moratalla Marlstone appears to be divided into segments (which may be more than 20 m in length) by gently curving planes which truncate the stratification at low angles (fig. 23). Locally the strike and dip of the beds in each segment differs markedly from those of the adjacent segments. The beds within the segments are curved. This is a striking deviation from the normally flat-lying Moratalla Marlstone beds, which are only intersected by numerous steep fractures. These steep fractures are invariably lined with calcite, which may be slickensided. Commonly a relative shift of the strata at both sides of these fractures could be observed. The low-angle planes, however, are not lined with calcite and the calcite-filled fractures cross these planes unmodified. Polished surfaces or striations were not detected along these planes. It can therefore be concluded that the formation of the steep fractures is of a later date than that of the low-angle planes. Directly below the basal plane of a lenticular segment (fig. 24a and b) the laminated marlstone shows many small isoclinal, recumbent folds, which are overturned in northward direction (fig. 25). The strike of the axes of 5 of them was measured: N 92 E, N 92 E, N 97 E, N 107 E, N 87 E. This implies an average stress-direction towards N 6 E. Also along other low-angle planes the marlstones may be corrugated. A thin section of a sample with corrugated laminae also shows microscopic corrugations. As the continuity of the stratification of the Moratalla Marlstone has apparently been disturbed by movements along these planes, as no slickensides or whatsoever were found along these planes, and as the disturbed interval is overlain by a non-disturbed horizontal sequence of Moratalla Marlstone, it can be inferred that the marlstone segments are the intersections of slump-packets bordered by curved, low-angle slip-planes. The Moratalla Marlstones have apparently suffered rotational sliding along spoon-shaped slip-planes, while the slipped rock masses remain internally undisturbed except for some minor visco-plastic deformation near the slip-planes. The direction of overturning of the fold structures indicates that the slump-packets have slid towards the north. The stratigraphic position of this slumped interval is about 50 m below the base of the overlying Benamor Geröllton Member (which is 150 m thick) and may therefore be correlated with the base of the Turmal Conglomerate (page 124). Similar structures were observed in the Moratalla Marlstone in the southeastern slope of the Cerro de San Jorge.

A strongly folded interval of approximately 5 m thick occurs below the lowest unit of the Benamor Geröllton Member (fig. 27), which overlies the Moratalla Marlstone Member. The interval is generally separated from the Geröllton unit by a few undisturbed, or slightly folded beds. The many tight, recumbent

Fig. 22. Close-up of the sand avalanche deposit intercalated in the Moratalla Marlstone Member exposed along the northern tributary of the Arroyo de Agüica. Note the absence of grading and the random distribution of the rounded quartzite pebbles. Many pebbles were dislodged and fell down, which accounts for the holes in the sand bed.

Fig. 23. Slip plane separating two slump packets in the Moratalla Marlstone. The left-hand packet has a finely corrugated zone along the slip plane. The dip of the beds in the right-hand packet reaches 30°. Northern bank of the Río Benamor, 150 m west of the bridge (road from Caravaca to Moratalla).

folds shown by the turbidite beds in this interval are piled upon each other and are invariably overturned towards the north (fig. 28). They do not show any fractures in the zones of maximum curvature, so that a visco-plastic behaviour of the material during the deformation can be assumed. Some beds are fallen apart into isolated slabs that are folded and balled up. Locally the Moratalla Marlstone is even converted into a monomictic bouldery marlstone whose boulders are partly balled-up fragments of incompetent beds, partly non-deformed broken-off pieces of harder beds. In one outcrop of this interval 10 strikes of anticlinal axes were measured. They vary considerably; especially those of the smaller folds may deviate notably from those of the folds with long axes. The average strike is N 77 E, corresponding to a stress direction of N 13 W towards the north. The intensity of deformation in this "slump sheet" increases in upward direction.

A good exposure of penecontemporaneous deformation in the turbidite sequence of the Moratalla Marlstone was found in the bed of the Barranco de Azafran about 300 m northwest of the bridge over which the road from Moratalla to Sabinar crosses the barranco (fig. 26). Here two directly superposed subhorizontal turbidite beds diverge towards the south making room for marlstones with turbidites laid into a large isoclinal recumbent fold. The fractures in the turbidite beds in the zone of maximum curvature suggest that they were somewhat indurated before the folding. The bedding fissility in the marlstones remains parallel to the folded stratification. The average strike of the fold axis is N 85 E indicating a stress direction of N 5 W towards the north.

Near the top of the southernmost spur of the Cerro de Moratalla la Vieja also slide structures were observed below the northernmost exposure of the Benamor Geröllton Member. Here laminated calcareous sandstone beds are balled up, twisted and broken; the broken pieces form winding strings in the white marlstones. In this outcrop, 4 m below the base of the conglomeratic marlstone, a nearly straight sandstone dike of 1 m length was found cutting obliquely through the horizontal stratification. It begins and ends abruptly without showing its source. Its thickness is irregular and varies between 2 and 5 cm.

The direction of the dip of the palaeoslope that can be inferred from these slide-structures should have been due north. Hence the Moratalla basin is no exception to the common feature that the direction of turbidity currents differs from that of the slides.

Thickness – As a result of the construction of the cross-sections the thickness of the Moratalla Marlstone Member near Moratalla would be about 200 m. Near the Barranco de los Barranquillos it would be more than 300 m.

Age – The associations of planktonic foraminifera (determined by J. Manuputti) from the Sierra del Roble, Cerro de Moratalla la Vieja, and from the southeastern corner of the investigated area (first three columns of table 2) are characteristic for the *Globorotalia fohsi peripheroronda* zone. The association from the upper 60 m of the Moratalla Marlstone Member exposed along the Río Benamor and Barranco de Azafran (last column of table 2) is characteristic for the *Globorotalia fohsi fohsi* zone. On account of the planktonic foraminifera the Moratalla Marlstone Member is assumed to be of latest Langhian and Early Serravallian age (Middle Miocene).

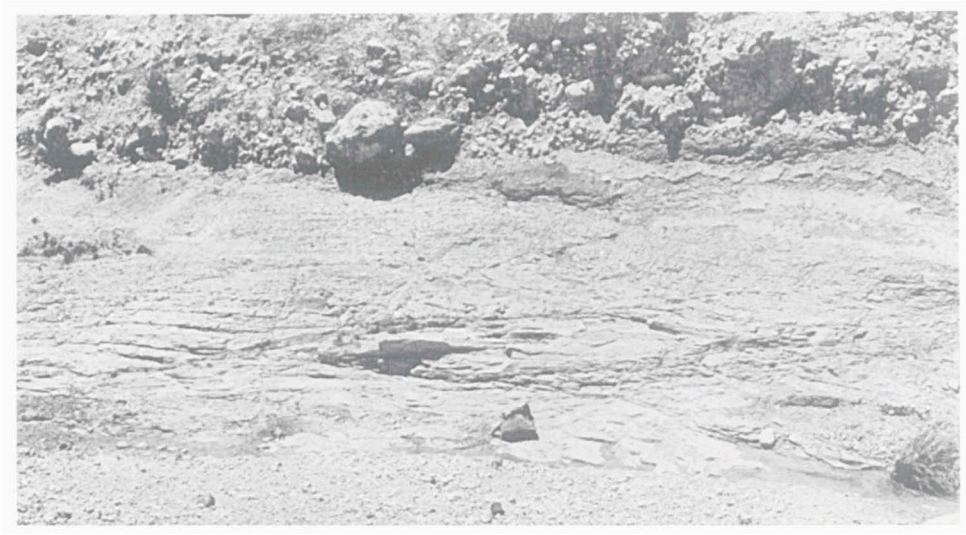


Fig. 24a. Cross-section of a lenticular slump packet, which contains a drawn-out calcareous sand bed. The packet is bordered by four other slump packets. Moratalla Marlstone Member, southern bank of the Río Benamor, 150 m west of the bridge (road from Caravaca to Moratalla).

Fig. 24b. Schematic drawing of Fig. 24a.

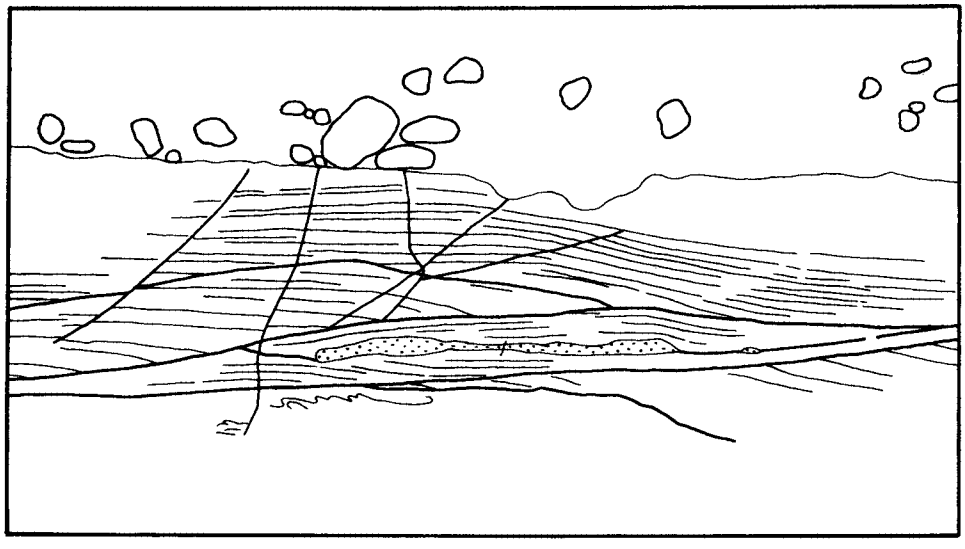


Table 2. Planktonic foraminifera of the Moratalla Marlstone Member.

	Basal 40 m in the Sierra del Roble	Interval in southeastern-most corner of investigated area	Sequence exposed in the slopes of the Cerro de Mortalla la Vieja	Directly below and within turbidite sequence along Río Benamor and Barranco de Azafran
Catapsydrax unicavus			x	
Globigerina angustiumbilicata	x	x		x
Globigerina cf. G. bulloides	x	x		x
Globigerina cf. G. nepenthes	x	x		
Globigerina praebulloides		x		x
Globigerinita naparimaensis	x			
Globigerinita naparimaensis naparimaensis		x		
Globigerinoides aperturus			x	
Globigerinoides ruber		x	x	
Globigerinoides sacculifer		x		x
Globigerinoides trilobus	x		x	x
Globigerinoides ex gr. G. trilobus		x		
Globorotalia archeomenardii		x		
Globorotalia fohsi fohsi				x
Globorotalia fohsi peripheroacuta				x
Globorotalia fohsi peripheroronda	x	x	x	x
Globorotalia cf. G. fohsi peripheroronda	x	x		x
Globorotalia mayeri	x	x		x
Globorotalia obesa	x		x	x
Globorotalia ex gr. G. opima				x
Globorotalia opima continousa	x			
Globorotalia opima nana		x		
Globorotalia praemenardii	x			
Globorotalia praescitula	x	x		x
Globorotalia cf. G. scitula	x	x	x	
Globorotaloides suteri			x	
Globoquadrina ex gr. G. altispira	x	x		x
Globoquadrina dehiscens	x	x		x
Globoquadrina venezuelana		x		x
Orbulina bilobata	x			x
Orbulina suturalis	x	x	x	x
Orbulina universa	x	x	x	x
Porticulisphaera glomerosa			x	x
Porticulisphaera glomerosa circularis		x		
Porticulisphaera glomerosa glomerosa		x		

Turmal Conglomerate Member

Derivatio nominis – The rocks of this Member are well exposed along the road from Moratalla to Las Murtas directly north of the bridge over the Río de Moratalla. This exposure and those along the Río de Moratalla east and west of this bridge constitute the stratotype of the Member. The nearest farm is called Turmal (encl. 1).

Lithology – It consists of coarse, polymictic conglomerate of which the components are made up of rocks, which, as far as recognized, were found only in the Prebetic rock-sequence of the Sierra de la Muela directly to the northwest of the

Fig. 25. The isoclinal recumbent folds in the laminated Moratalla marlstone are probably caused by drag exerted by the directly overlying lenticular slump packet of fig. 24a. Southern bank of the Río Benamor, 150 m west of the bridge (road from Caravaca to Moratalla).

Fig. 26. Large slump fold in the turbidite sequence of the Moratalla Marlstone Member. Note the fractures in the curved turbidite beds. Direction of photograph parallel to fold axis. Barranco de Azafran, 300 m upstream from the bridge (road from Moratalla to Sabinar).

Turmal deposits; Subbetic and South-Prebetic rocks were not found. A very common component of the Turmal Conglomerate is the Roble Limestone. Also limestones of the Somogil, Almanillas and Tosca Formations were recognized. The pockpitted appearance of many boulders (fig. 29) appears to be due to numerous burrows indicating their prolonged stay in a shallow nearshore marine environment. Of particular importance is the presence of a very large (1.5 m) rectangular block of soft green shales, probably derived from the Grero Marlstone. It indicates that this block cannot have travelled over a large distance before it became trapped into its position with respect to the surrounding boulders. Probably this is also the case with the other large blocks in the conglomerate on account of their presumed derivation from the immediate hinterland in the northwest. The Turmal Conglomerates may therefore be interpreted as rock-fall deposits. Also lenticular slabs of Moratalla Marlstone were found which are twisted and drawn out. In the right bank of the Río de Moratalla some 150 m west of the bridge, also medium-bedded fine-grained rocks are interfingering with and overlying the conglomerates. The conglomerate beds, which may reach a thickness of more than 10 m, do not show any internal layering. All grain sizes are present up to an apparent longest diameter of 1.5 m. In some exposures the model size of the clasts is markedly smaller than in other exposures. The commonly rounded, irregularly shaped pebbles, cobbles, and boulders, are very tightly packed (fig. 30); the percentage of sand-sized fraction is conspicuously small. The smallest interstices are filled with calcarenite with a variable percentage of siliciclastic detritus.

The Turmal Conglomerate has two interesting aspects:

1. It marks an angular unconformity within the Moratalla Marlstone Member (encl. 3 and 5).
2. Large slabs of it have slipped down along the unconformity plane into the Moratalla basin.

The first aspect is the most important one. Near the highest top of the Sierra del Roble (reference section) the slightly dipping Turmal Conglomerate truncates steeply southward dipping Roble Limestone and Moratalla Marlstone Near the Cortijo del Collado del Roble, approximately 1 km to the east and only 50 m lower in altitude, however, the Conglomerate lies already upon a very thick sequence of Moratalla Marlstone that obviously belongs to the part below the unconformity. In the Collado del Roble the conglomerate is directly overlain by the Cretaceous rocks of the Nevazo Unit. But immediately south of the Río de Moratalla it is separated from the Nevazo Unit by an approximately 200 m thick sequence of Moratalla Marlstone that obviously belongs to the part above the unconformity. As the basal plane of the Nevazo Unit is horizontal, this implies that in the Río Benamor, where no angular unconformity was found, the level corresponding with the unconformity in the north would be situated some 50 m below the base of the Benamor Geröllton Member (which is 150 m thick and directly overlain by the Nevazo Unit). As we have seen, at this level an interval with intensive slumping occurs in the Moratalla Marlstone (page 118, 120). This interval is therefore tentatively correlated with the unconformity plane at the base of the Turmal Conglomerate.

The presence of the tongue of Turmal Conglomerate unconformably intercalated in the Moratalla Marlstone has important consequences for the geological history of the area. It has to be assumed that in the immediate hinterland intensive erosion must have taken place (down to the Albian strata of the Tosca

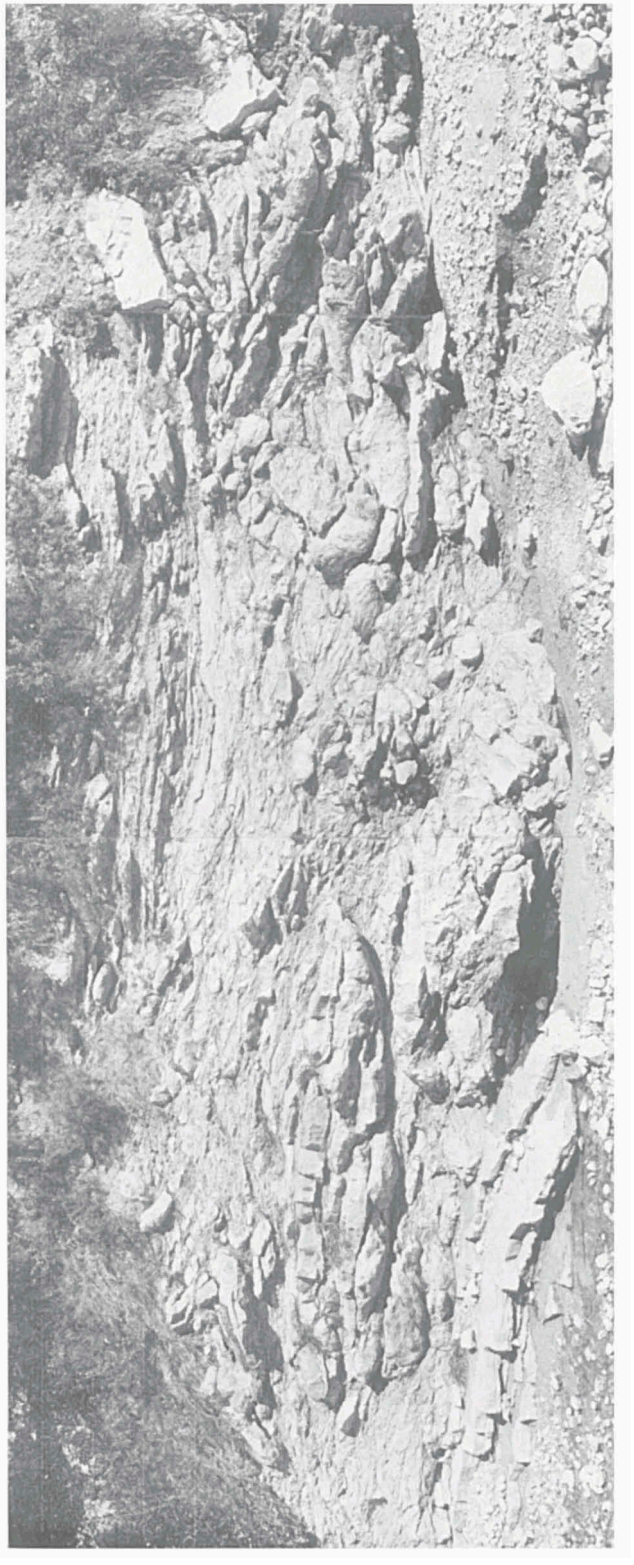


Fig. 27. The "slump sheet" in the Moratalla Marlstone Member. The turbidite beds are deformed into a chaotic heap of piled-up recumbent folds. In the middle of the exposure the beds have lost their coherence and have fallen apart into folded and balled-up portions. Photograph taken in the direction of provenance. Southern bank of the Río Benamor.

Limestone Formation), and that the Sierra de la Muela probably was an island or a promontory in the Middle Miocene sea. The angle of about 50° found between the Turmal Conglomerate and the Roble Limestone - Moratalla Marlstone sequence at the top of the Sierra del Roble can be explained by assuming that the gravity collapse structures in the southeastern limb of the Muela anticline, in which the Roble Limestone and the lower part of the Moratalla Marlstone have taken part, had already come into existence. The Turmal Conglomerate has not been involved in this deformation.

The second aspect of the Turmal Conglomerate is the most striking one. Large tabular blocks of the Turmal Conglomerate have slipped down in indurated state into the Moratalla basin. This can be seen in an exposure in the Río de Moratalla just east of the bridge (encl. 6). Plates of Turmal Conglomerate with its proper matrix were broken up during their sliding over the marlstone substratum, became imbricated, or plunged like wedges between the beds of the Moratalla Marlstone, which were tilted up. The marlstones were broken into slump-packets truncating the adjacent beds, were pushed into flame-like structures, became intruded between the imbricated plates, or were kneaded into a mush in which the harder beds were balled up or broken into lozenge-shaped chips which float in the softer matrix. Often cobble and boulder-sized components of the conglomerate became dislodged and were kneaded into the Moratalla marlstone or were pushed between the marlstone beds. The deformation of the Moratalla marlstone suggests that the marls were still plastic, but had already gained some stiffness before the tabular blocks slid down from the northwest (fig. 29). Also the large mass of Turmal Conglomerate exposed in the right bank of the Río de Moratalla 150 m west of the bridge appears to be a slip-block. The fine-grained beds in the mass do not show any sign of disturbance, whereas the Moratalla Marlstone directly below the sharp lower contact is highly contorted.

Above this level of slipped-down blocks of Turmal Conglomerate four other, thinner conglomerate beds are intercalated within the Moratalla Marlstone (fig. 30). In these beds, however, the interstices between the roundstones are filled with a Moratalla marlstone matrix. In the lower two of them the percentage of marlstone is very small, but in the upper bed this percentage is estimated at 40%. Upward in this sequence the beds become thinner and the apparent diameter of the large components decreases. In the upper bed also twisted blocks of Moratalla Marlstone are incorporated in the conglomerate, while the Moratalla Marlstone directly below shows curved slabs with tapering ends floating within the softer surrounding rock. The sliding conglomerate apparently tore the marlstone beds apart. In the western most exposure directly south of the Río de Moratalla the percentage of grey Moratalla marlstone matrix is far greater than that of the pebbles, cobbles and boulders of Prebetic rocks floating in it. This conglomeratic marlstone is 2.5 m thick. Many rock types were recognized which are also present in the Prebetic zone, but no components of the Subbetic or South-Prebetic zone were observed. Also balled-up lamellar portions of the Moratalla Marlstone were found included in the non-stratified mass of the grey marlstone binder. Thread-like structures can be seen draped around the inclusions. Irregular shaped patches can be discerned in which numerous coarse sand-sized grains of quartz and quartzite are concentrated in the binder. These patches may also be drawn out into threads. Since the inclusions have not been subject to any differentiation, the rock can be considered as an olisthostrome. This olisthostrome is separated from the Roble Limestone Formation by a few metres thick alternation of Moratalla Marlstone

Fig. 28. Close-up of the folds of Fig. 27. The direction of sliding was from the right. Note the absence of fractures in the zones of maximum curvature of the beds. Southern bank of the Río Benamor. Photo taken by J. J. Hermes.

Fig. 29. The large, pock-pitted boulder of the Turmal Conglomerate has pushed the Moratalla Marlstone beds into a curve. The push came from the northwest. Photographed obliquely downwards. Río de Moratalla, 150 m east of the bridge (road from Moratalla to Las Murtas).

beds, globigerinal lime-packstones beds, polymictic conglomerate beds with components of Prebetic rocks, and coarse bioclastic lime-grainstone beds with fragments of *Lithothamnium, Amphistegina, Heterostegina,* echinoids, bryozoans, molluscs, smaller benthonic foraminifera, and 5% to 30% siliciclastic detritus. In these lime-grainstone beds reworked orbitolinids and dolomite crystals were found. These beds are therefore correlated with the basal part of the Moratalla Marlstone north of the Río de Moratalla. The conglomeratic marlstone is overlain by a calcareous sandstone bed with a flat top and a strongly undulating base (loading?), causing the thickness of this bed to vary from 0 to 30 cm.

The Turmal Conglomerate is interpreted as the near shore facies (rock-fall deposits) of the Moratalla Marlstone Member. It advanced basinward after the upheaved part of the tilted Moratalla Marlstone was attacked by erosion. Ultimately, large pieces of this tongue came sliding down. The period was apparently one of general tilting and upheaval in both flanks of the Moratalla basin, followed by a considerable intensification of redeposition in the basin.

Thickness – does not exceed 15 m.

Age – The time of the emplacement of the slabs of Turmal Conglomerate into their present position is approximately the same as that represented by the boundary between the biozones of *Globorotalia fohsi peripheroronda,* found in the Moratalla Marlstone below the conglomerate, and *Gl. fohsi peripheroacuta – Gl. fohsi fohsi,* found in the Moratalla Marlstone along the Río Benamor in and above the interval that presumably can be correlated with the angular unconformity at the base of the Turmal Conglomerate.

Benamor Geröllton Member

The principle aim of the project was the investigation of the Benamor Geröllton Member. This Member is largely composed of olisthostromes, mud flow deposits, and gravel avalanche deposits in which large volumes of rocks from the Subbetic and South-Prebetic zones have been reworked. Since good descriptions of olisthostromes are rare in the literature, the rocks will be described in detail.

DERIVATIO NOMINIS AND GENERAL DESCRIPTION

In the steep northern slope of the Río Benamor valley very large isolated rock masses are protruding over a length of two kilometres, well visible from the southern valley slope of the Río Benamor (encl. 8a and 8b). At several levels conspicuous horizontal rows of these harder rock masses, the largest of which is more than 100 m in length, can be followed over a considerable distance. The rock masses are separated from each other by softer rocks. The major part of this softer material appears to consist of 1) isolated rock masses of lesser consistence, also of considerable dimensions, in several horizontal levels, 2) grey, non-stratified, pebbly to bouldery marlstones, and 3) non-stratified, grey marlstones with only locally a sparse scatter of boulders. As all size grades are found to exist between the large isolated rock masses and the pebbles enclosed by the grey marlstones, the large rock masses should be considered as the largest boulders of the conglo-

Fig. 30. The second of the four beds of Turmal Conglomerate overlying the interval of imbricated plates. It rests upon non-disturbed, bedded Moratalla Marlstone. Almost no marlstone was found between the closely packed clasts. 200 m east of the bridge in the Río de Moratalla.

meratic marlstones. The lesser part of the softer material is found to be restricted to several horizontal intercalations at different levels and to consist of a thin to medium-bedded alternation of grey marlstones and sandy limestones (or calcareous sandstones) with interbeds of medium to thin-bedded, polymictic limestone conglomerates and coarse calcarenites. Some of these intercalations can also be followed over considerable distances. The polymictic conglomerate and calcarenite beds are called Bejar Beds after the Campo de Bejar where they are thick and well exposed.

The whole rock sequence has a vaguely layered appearance due to the horizontal rows of large rock masses and the horizontal intercalations of bedded rocks. The latter can properly be assumed to be rock-stratigraphic units intercalated conformably with the sheet-like associations of non-stratified conglomeratic marlstones and the non-stratified marlstones containing only locally a few boulders. These sheet-like associations should therefore also be considered as rock-stratigraphic units.

Most of the larger rock masses and many pebbles, cobbles, and boulders floating with no apparent order within the non-stratified marlstones appear to be exotic. Many of them are composed of Cretaceous or Tertiary rocks similar to those found in the Subbetic and (or) South-Prebetic zones several kilometres to the south. Evidently they must have travelled northward along a considerable distance over the Moratalla Marlstones. As exotic rock masses of all size grades are randomly distributed in the non-stratified marlstones which entirely surrounds them, it may be assumed that they travelled together with these marlstones. As the underlying Moratalla Marlstones are disturbed only to a negligible degree, apparently large stresses were not involved during the northward transport of the huge marlstone masses with inclusions, so that visco-plastic flowing is the only possible mode of transportation. The whole association of grey non-stratified marlstones with varying percentages of inclusions should therefore be considered

as olisthostromes and mud flow deposits, the inclusions of the former as olistholiths.

The whole 150 m thick rock sequence, from the first conglomeratic marlstone unit at the top of the Moratalla Marlstones up to the structural contact with the overlying Nevazo Unit (composed of Cretaceous rocks) is considered to constitute the Benamor Geröllton Member, henceforward abbreviated as B.G.M. The petrographic term 'geröllton' is preferred to the less specific term 'diamictite' and the lengthy term 'conglomeratic marlstone'.

The B.G.M. can be subdivided into two parts. The lower part is characterized by the presence of many inclusions larger than 5 m in diameter (meso, macro and megolistholiths) and by the absence of thick conglomerate deposits. The upper part is characterized by the presence of many thick conglomerate deposits and by the virtual absence of components larger than 60 cm in diameter. The upper part is mainly exposed in the Campo de Benamor and the Campo de Bejar more to the west (encl. 1). Only locally thick conglomerate deposits were found east of the Campo de Benamor.

Because of the softness of the main constituents of the conglomeratic marlstones, the rocks of the B.G.M. rapidly become subject to wheathering, so that large exposures are rare. The largest and cleanest exposure of conglomeratic marlstone is 200 m by 12 m and situated in the right vertical bank of the Río Benamor 850 m SE of the Cortijo de Santa Anna (encl. 1 and 2). This exposure may be designated as typical for the lower part of the B.G.M. (encl. 7).

The lower part of the B.G.M. has a composite stratotype. In stratigraphic order it is composed of the Benamor succession (p. 148), the Cerro de las Arcas succession (p. 149) and the Collado de San Jorge succession (p. 151). The upper part can be defined in two type areas, viz. the Campo de Benamor for the lower half and the Campo de Bejar for the upper half.

LOWER PART OF THE BENAMOR GERÖLLTON MEMBER

Lithology and features

The various features of the B.G.M. are grouped into categories. For each category the features of the typical exposure are described first; after that, the variants found in other localities are described.

Upper boundaries of depositional units: typical exposure – The upper boundary of the olisthostrome cropping out in the typical exposure is horizontal, sharp and nearly flat. It is either overlain by a thin-bedded sandy marlstone unit (0-1 m) followed by a medium-bedded sandy calcarenite unit (40-80 cm), or directly by the latter (encl. 7). Apparently before or during the deposition of the sandy calcarenite beds the undulating upper surface of both the olisthostrome and the sandy marlstone on top of it have been truncated. The calcarenite is again overlain by conglomeratic marlstones.

Sandy calcarenite beds also occur between some other conglomeratic marlstone units in the lower part of the B.G.M.

They are composed of exotic limestone clasts, a varying percentage of siliciclasts, and of bioclasts. The exotic limestone clasts consist of bioclastic packstone with *Pithonella* and *Stomiosphaera* (Represa Formation). In many samples pieces of Argos Formation and Quipar

Formation were recognized. The 10 to 25% siliciclastic detritus is predominantly composed of quartz and quartzite, always with a minor percentage of felspar and glauconite. In most samples mica and a few tourmaline grains were found. Characteristic are the greenish to yellowish brown egg-shaped, sand-sized grains composed of flaky material. In thin sections these grains show opaque spots and frequently regional to zonal extinction. They are interpreted as coprolites. The bioclast-association of the different samples is nearly the same, viz. *Heterostegina, Amphistegina,* globigerinids, smaller benthonic foraminifera, and fragments of calcareous algae, echinoids, bryozoans, and molluscs. In some samples also sponge-spicules and radiolarians were found, which are typical for the Moratalla Marlstone. Reworked *Discocyclina* fragments were recognized in almost all samples, generally in combination with *Nummulites.* In a few samples reworked orbitolinids were observed.

Also in other exposures the structure of the conglomeratic marlstone may be truncated by the commonly sharp upper boundaries.

Upper boundaries of depositional units: variants in other exposures

1. The base of the horizontal calcareous sandstone bed overlying the lowermost conglomeratic marlstone unit of the B.G.M. (right bank of the Río Benamor 200 m upstream of its confluence with the Barranco de Azafran) shows downward bulges, presumably caused by loading.

2. In another conglomeratic marlstone outcrop (in a barranco 500 m east of the Cortijo de Santa Ana) small irregular cloudy patches, in which numerous sand to granule-sized grains are concentrated, were found distributed in the marlstone of the upper 10 cm just below the sharp boundary with the overlying coarse polymictic calcarenite (which is the basal part of a reversely graded conglomeratic marlstone unit described on page 136). Marlstone patches occur in the few basal centimetres of the calcarenite. These features were interpreted as irregular protuberances of calcarenite protruding down from the boundary which is the evidence of the incipient mingling of waterladen rocks at both sides of the boundary.

3. In two outcrops a normal and continuous upward gradation was found from conglomeratic marlstone via polymictic limestone conglomerate into coarse polymictic calcarenite (fig. 32). An instructive example can be found in the southeastern slope of the Cerro de San Jorge. In the upper part of the conglomeratic marlstone, in which the size of the constituent elements ranges from clay to boulder, the diameter of the largest clasts diminishes rapidly down to small cobbles in upward direction; concomitantly the clasts come more and more in contact with each other and the mud content decreases. From the level in which all clasts are just in mutual contact the ratio mud to clasts remains approximately constant over 35 cm. The more or less equidimensional, facetted limestone pebbles of this badly sorted conglomerate commonly show rounded edges and are of the same lithologic types as the inclusions of the conglomeratic marlstone below, only the percentages of the various rock types differ somewhat. The conglomerate grades into coarse calcarenite (20 cm) in which the mud percentage gradually increases again. This calcarenite shows a layering brought about by three horizons in which coarser limeclasts are concentrated. Whether this is due to alternately normal and reverse grading or to recurrent grading is doubtful. The top layer grades normally via globigerinal pseudo lime-grainstone into globigerinal lime-packstones and wackestones.

A thin section of this upper layer yielded reworked fragments of *Discocyclina* and globorotalids of Eocene age. In the lower part of the sample exotic clasts from the Represa

Formation, Quipar Formation, and Moratalla Marlstone were found. Globigerinids, sponge-spicules, *Lithothamnium,* smaller benthonic foraminifera, and echinoid fragments were recognized. The siliciclastic detritus is composed of quartz, quartzite, felspar, colourless and brown mica (of which the brown colour had been for the greater part converted into green), glauconite, and nests of pyrite crystals.

Since the conglomeratic marlstone layer gradually grades into the conglomerate-calcarenite layer, both are to be considered as belonging to one and the same depositional unit, that is, deposited during one depositional event. Such a normally graded conglomeratic top can readily be explained by assuming that the mud of the upper part of the downflowing rock mixture of diamictonic composition goes into suspension as a result of increased inmixing of water in the turbulent flow. This would cause a gradual decrease in the density of the mud in upward direction. The larger clasts (suspended in the mud and the mud-laden water) would, when the flow slackens, sink to certain density levels within the thixotropic mud, whereas the smaller ones would drop out from the mud suspension.

The differentiation of the downflowing rock mixture into a lower conglomeratic marlstone and an upper conglomerate gives rise to internal layering. Since olisthostromes should not show internal layering by definition, this deposit should be designated as a differentiated mud flow deposit.

4. If the upward decrease in density in a flowing diamictonic mass was not gradual, but discontinuous, also the grading of the resulting conglomerate would be discontinuous. Subsequent differential loading of the gravel at the top into the soft gravelly marl below would result in a strong undulating base of the gravel. Both discontinuous grading and strongly undulating base can be observed in the conglomerate top of a conglomeratic marlstone unit exposed in the northern side of the road leading to Casa Nueva 100 m before it crosses the Acéquia de Benamor (fig. 31).

Above this bed an interval of 1 m thickness crops out consisting of polymictic, badly sorted calcirudite and calcarenite beds with thin intercalations of marlstone with bedding fissility. These Bejar beds rapidly wedge out in eastern or western direction, thinning from one or more decimetres to zero over a distance of only three metres. The calcarenites and calcirudites contain many quartzite grains of sand to granule size and show symmetrical grading, recurrent grading, and parallel lamination. The coarse levels and the laminae within these beds are at an angle of a few degrees to the base of these beds.

A sample of a coarse calcarenitic interval shows a micro-breccia of limestone clasts with calcareous mud in the interstices and with approximately 10% of angular to subrounded siliciclastic detritus (quartz, quartzite, felspar, glauconite, and tourmaline) varying in size from 20 to 1400 μ. The subrounded to angular limeclasts are exotic and consist of globigerinal lime-packstone (Moratalla Marlstone or Puerta Limestone), and of bioclastic mudstones and wackestones with single and (or) double-keeled globotruncanids, *Heterohelix, Pithonella,* and *Stomiosphaera* (Represa and Quipar Formations). Among the bioclasts were recognized: fragments of *Heterostegina, Lithothamnium,* bryozoans, and reworked *Discocyclina.*

The way in which these beds were deposited can tentatively be explained by assuming currents of heavily clay-laden water carrying their load of non-cohesive grains partly in suspension and partly as bed load. The movement of the latter may also have been partially induced by gravity. The mud-laden currents are thought to be slow with respect to the velocities which are usually supposed to be reached by true turbidity currents, because the beds rapidly wedge out, and the internal lamination as well as the coarse-grained levels are at a small though

Fig. 31. The bed below the book (16 cm long) is the normally graded top part of the mud flow deposit cropping out at the base of the exposure. The downward bulges of the base of the bed are caused by loading. The calcarenite and calcirudite beds above the book rapidly wedge out to the right and to the left. They were interpreted as the basinward extension of the normally graded top parts of other mud flow deposits. Benamor Gerölllton Member. Southern slope of the Cerro de las Arcas.

conspicuous angle to the base of the various beds. The currents of clay-laden water are assumed to represent the liquified and suspended upper parts of descending mud flows. After the latter came to a halt these slurries moved further onward giving rise to the deposits just described.

5. In the southern bank of the Río Benamor 25 m upstream of its confluence with the Barranco de Escaramujo a Bejar bed was observed between two conglomeratic marlstone units forming a gully scoured into the underlying conglomeratic marlstone (fig. 51).

Lower boundaries of depositional units: typical exposure – The lower boundary of the conglomeratic marlstone unit of the typical exposure cannot be observed 300 m downstream, however, large exposures of horizontally bedded Moratalla Marlstone Member were found. If a horizontal continuation of this bedded sequence is assumed, the lower boundary of the conglomeratic marlstone unit should not be more than a few metres below the bed of the river, so that its total thickness can be inferred to be only a little over 12 m.

Lower boundaries of depositional units: variants in other exposures – In other exposures the lower boundaries of the conglomeratic marlstone units are generally sharp.

1. In two out of some twenty cases a gradual transition was found from non-bedded marlstone with a parallel bedding fissility (indicating an undisturbed stratification), via non-bedded marlstone with a crumpled fissility pattern, into conglomeratic marlstone. The transition between the latter two is shown by the appearance and the gradual increase in number of inclusions. Localities: south-eastern slope of the Cerro de San Jorge and in the barranco running southeast from the Cortijo de St. Ana just downstream of the bridge.

2. The bedded Moratalla Marlstone directly underlying the lowermost conglomeratic marlstone unit of the B.G.M. (southern bank of the Río Benamor 250 and 700 m upstream of its confluence with the Barranco de Azafran) is strongly folded over an interval of approximately 5 m (fig. 27). As the intensity of deformation in this slump sheet increases in upward direction, the folding may be partially due to the movement of the overlying olisthostrome, which can therefore be assumed to have moved in the same direction (N 13 W) over the Moratalla Marlstone which was pushed forward and squeezed out from under the overburden suddenly imposed upon it (a similar relation has been reported by Lucchi & d'Onofrio (1967) in delapsional intervals in Tortonian deposits along the Savio river near Sarsina, Italy). The sharp lower boundary of this conglomeratic marlstone unit is highly irregular. The conglomeratic marlstone is separated from the Moratalla Marlstone by a thin bed of sandy polymictic calcirudite with a coarse-grained basal part in which were found: quartzite pebbles (Rasa Sandstone Formation) and many exotic limestone clasts of the same rock types as those in the conglomeratic marlstone directly above.

3. The basal parts of some conglomeratic marlstone units consist of polymictic conglomerates of limestone clasts which are in mutual contact. Such basal conglomerates seem to be confined to the thinner units (few metres) of the B.G.M., in

Fig. 32. Normally graded top part of a mud flow deposit in the southern slope of the Cerro de San Jorge. Benamor Geröllton Member.

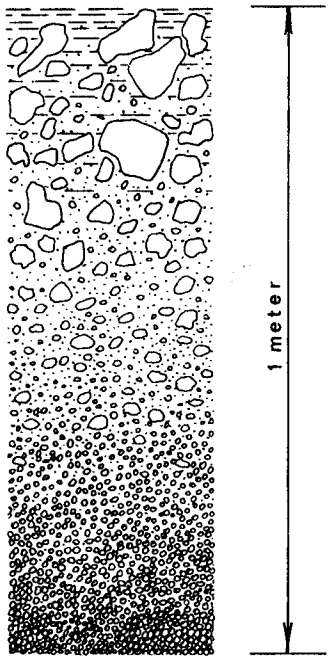


Fig. 33. Reversely graded basal conglomerate of a mud flow deposit in the B.G.M., northern slope of Benamor valley.

which only a few clasts exceed the diameter of 30 cm. The basal conglomerates may grade reversely into the conglomeratic marlstone on top it (fig. 33). A striking example of a reversely graded basal conglomerate is exposed 500 m east of the Cortijo de Santa Anna in a barranco 100 m upstream from the bridge of the road from Moratalla to Sabinar.

Near the sharp base of the lower part of this 1 m thick conglomerate the smallest clasts (limestone fragments, echinoid fragments, calcareous algae and quartzite grains) are of sand size, the largest granule-sized, and the distribution seems to approach a phi-normal one. There is hardly any mud in the interstices. Upward the amount of ever larger clasts becomes greater as does the mud content. A faint layering is brought about by two lamellae of granules.

The upper part contains many cobble-sized limestone clasts. The mud content is very large; nevertheless all grain sizes between clay and boulder are present. At the top the mud content is preponderant and the limestone clasts gradually lose their mutual contacts. The clasts are generally isodiametric. The two largest ones are 25 by 20 by 15 cm; one consists of hard mica-bearing calcareous sandstone of the Argos Formation, the other of grey marlstone of the Represa Formation. The remaining clasts predominantly consist of Moratalla Marlstone. The latter clasts are generally well-rounded to spherical, whereas those of Argos sandstone are commonly angular and prismatic, and those of Represa marlstone rounded and facetted.

At the base of the overlying conglomeratic marlstone the volumetric percentage of the inclusions is great, but it decreases gradually upward, while that of the marlstone binder increases.

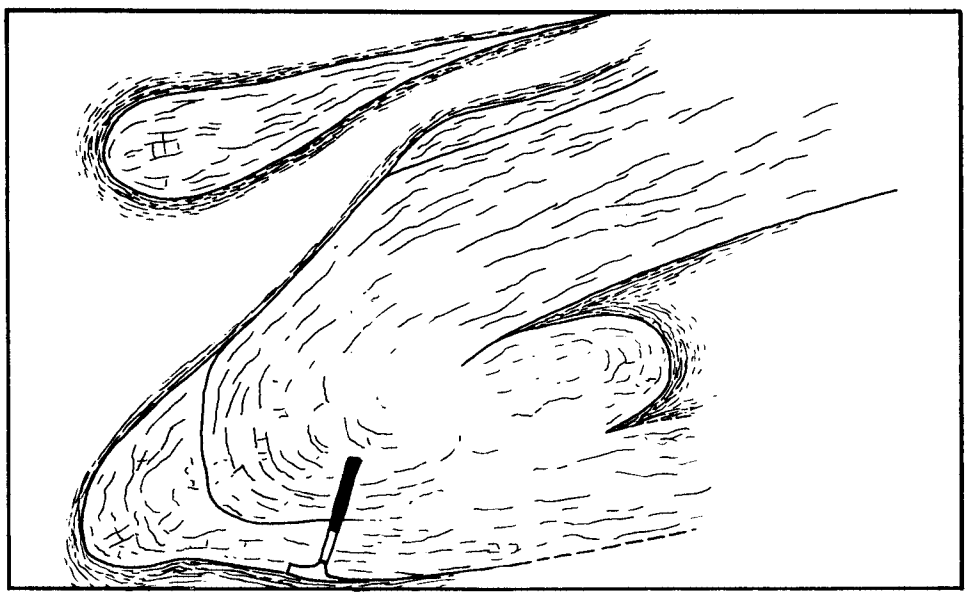
Recapitulating one may say that the peculiar reverse grading of the basal conglomerate is in fact an upward gradation from monomodality or moderate bimodality to extreme bimodality concomitant with a divergence of the constituent modes. If we consider the conglomerate to be built up of "matrix" and "clasts", we may describe the grading by saying that the modal diameter of the "clasts" becomes progressively larger, whereas that of the particles in the "matrix" becomes progressively smaller from base to top.

Lateral transitions of depositional units: typical exposure – Fissile marlstones of the same lithology and fossil content as the Moratalla Marlstones crop out over some 20 m at the eastern extremity of the typical exposure. In bedded or laminated Moratalla Marlstones the fissility is invariably parallel to the stratification (bedding fissility), however disturbed this may be. In the non-bedded, apparently homogeneous Moratalla marlstones, therefore, the fissility pattern is assumed to be the external evidence of otherwise invisible internal structures. These homogeneous marlstones were designated as horizontally stratified only on account of the parallelism of the fissility. But the fissility may also cause some parts of homogeneous marlstones to fall apart into angular, approximately isodiametric blocks, and other parts to fall apart into flat, lenticular chips which may be curved. The former parts are less quickly eroded than the serrately fissiled ones and are bulging outward (fig. 35).

The fissility pattern of the ill-bedded marlstones at the eastern extremity of the typical exposure gives the strong impression of folding and crumpling of the marlstone (evidenced by the folded structure of an intercalated thin calcareous sandstone bed). For example, recumbent fold structures are found in which long, sinuous zones of flat, lentil-shaped chips occur along the outer margin of the

Fig. 34a. Fissility pattern in the marlstone disclosing the structure of a recumbent fold. Eastern extremity of the typical exposure of the B.G.M. Southern bank of the Río Benamor.

Fig. 34b. Schematic drawing of Fig. 34a.



folds (fig. 34a and b), whereas isodiametric blocks occur in their nuclei. Also lenticular to subspherical marlstone bodies were found in which the parting planes tend to be perpendicular to each other, and which are surrounded by serrately fissiled zones in which the marlstone falls apart into curved, lenticular chips (fig. 35). Similar spherical bodies in other exposures may show faint laminations which are wholly or partly concentric with the outer margin. These bodies were therefore interpreted as balled-up marlstone portions.

Proceeding in western direction along the typical exposure the complexity of the fissility pattern increases considerably: bodies of less serrate fissility are lenticular, drawn out, or strongly curved. Exotic inclusions appear and gradually increase in number towards the west. Over the first few metres they consist of greyish-beige limestone composed of globigerinids in closest packing, of coarse sandy limestone, and of grey calcilutitic limestone with globigerinids (all belonging to the Trieta Formation). Apparently also in lateral direction marlstones with a not or relatively little disturbed stratification may pass gradually into conglomeratic marlstone. The fold structures may be explained in an analogous way as those occurring below the conglomeratic marlstone units, viz. by a lateral push of the passing olisthostrome.

Binder of conglomeratic marlstones: typical exposure – The fissility pattern reaches the climax of complexity in the binder of the olisthostrome in the typical exposure. The marlstone binder falls apart into lenticular chips, equidimensional blocks, and spheres. The parting planes are generally curved. At several places subparallel parting planes were seen to curve around the olistholiths. The fissility is especially marked at the surface of those exposed parts that are somewhat influenced by wheathering. More internally, as can be seen in fresh exposures, the moist binder does not show any fissility and makes a homogeneous impression. Also under-water exposures of the binder do not show any fissility and the marl appears kneadable. The parting is therefore thought to be due to shrinkage as a result of desiccation of the marls, which presumably contain a high amount of montmorillonite derived from the Represa Formation.

The grey marlstones of the binder contain a sparse sprinkling of sand-sized quartzite grains. Next to globigerinids of Miocene age also many globotruncanids of Cretaceous age were found.

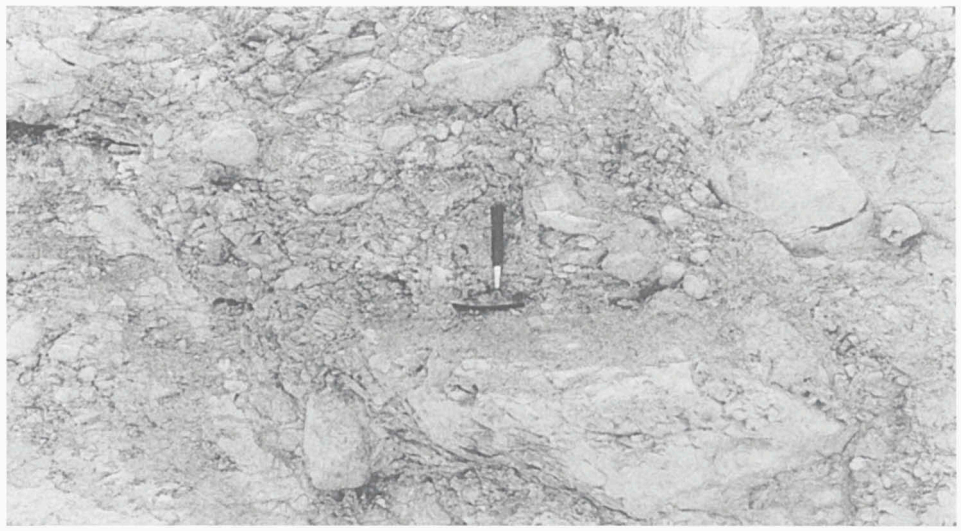
The following Miocene globigerinids were determined: *Globigerinoides trilobus, G. bisphericus, Globorotalia mayeri, Gl. scitula, Globigerina tripartita, Globoquadrina dehiscens, Catapsydrax* cf. *C. dissimilis.*

This fauna is comparable with that of the Moratalla Marlstone Member. Also the petrography of the binder of the conglomeratic marlstone is similar to that of the Moratalla Marlstone, except for the colour of the former which is a slightly darker shade of grey. The binder of the conglomeratic marlstones of the B.G.M. is therefore taken to be in reality non-stratified Moratalla marlstone containing exotics.

The fissility pattern delineates two types of spherical marlstone bodies, one with widely spaced parting planes (which tend to be perpendicular to each other) surrounded by serrately fissiled marlstone, the other with a stronger fissility than the surrounding marlstone. These bodies can be differentiated from the surrounding marlstone by discontinuities in the fissility pattern; their fissility shows a different spacing and configuration and (or) a different preferential direction than that of the surrounding marlstones. Therefore, they should not be regarded as

Fig. 35. Balled-up marlstone portion bordered by serrate fissility concentric with the outer margin of the ball. The parting planes within the ball show the tendency to be perpendicular to each other. Eastern extremity of the typical exposure of the B.G.M., southern bank of the Río Benamor.

Fig. 36. Part of the conglomeratic marlstone of the typical exposure of the B.G.M. The olistholiths are randomly distributed; the larger ones have a horizontal attitude. Southern bank of the Río Benamor.



parts of the binder, but as olistholiths, although a clear difference in lithology cannot be observed. These olistholiths are interpreted as fragments of the Moratalla Marlstone that became balled up instead of assimilated in the binder. The figures 36 and 37 show the conglomeratic marlstone of the typical exposure.

Binder of conglomeratic marlstones: variants in other exposures

1. In the marlstone binder of the lowest depositional unit of the B.G.M. (in the Río Benamor 250 m upstream of its confluence with the Barranco de Azafran) many faint slickensides and meandering threads of sandy marlstone were observed.

2. In another exposure (in a barranco 500 m east of the Cortijo de Santa Ana) strings of small inclusions (0.5 to 3 cm) were found wrapped around larger inclusions (0.5 m). In these strings the longest apparent axes of the small inclusions and the fissility between them are aligned in the curvature of the string. Among these small inclusions curved, lenticular marlstone parts of light grey and dark grey colour were found. The light grey lentils are generally drawn out at the extremities or pulled apart. The boundaries between the light and dark grey lentils are generally sharp, but colour transitions and mutual wrapping up of the small marlstone parts were also seen. These strings clearly suggest an eddying flow of the binder marl like a thick paste around the larger inclusions. It seems probable that the larger inclusions repeatedly turned over. This is in contrast to the idea of Görler & Reutter (1968) that inverted inclusions would not have turned over more than once. Figure 38 shows the conglomeratic marlstone exposed in the barranco that borders the northern slope of the Cerro de las Arcas.

Microlistholiths: typical exposure

Quantitative analysis: a time consuming quantitative analysis was done on the accessible lower parts of the vertical bank in the middle section of the typical exposure, with the double purpose to establish the volumetric percentages of the different kinds of inclusions and the percentages of the different size classes of the inclusions (Wentworth scale) for each depositional unit. This was done because there were indications that in successively higher depositional units increasingly older inclusions appear in increasing numbers.

It has been argued that there is no simple relation between the grain size of the rock and the point distance chosen for its modal analysis (Chayes, 1956). Van der Plas & Tobi (1965), however, came to the conclusion that the point distance chosen should be larger than the largest grain fraction that is to be included in the analysis. But from a practical point of view the point distance should, when a point-counting is done in the field, be chosen in relation to the total accessible clean rock surface available. Since the apparent diameter of the largest inclusion in the typical exposure is approximately 30 m, a modal analysis in the way of van der Plas & Tobi is technically impossible. But they pointed out that a modal analysis can also be carried out in two or more steps at different scales.

In our quantitive study it turns out that a modal analysis carried out only for the smaller fraction of the olisthostrome is not representative for the depositional unit as a whole, because some rock types occur only as large rock slabs, whereas they are sparse among the fragments of smaller size, and the reverse. The results nevertheless show the strong bimodality of the conglomeratic marlstone (though all size grades are present) and a modal diameter for the olistholiths

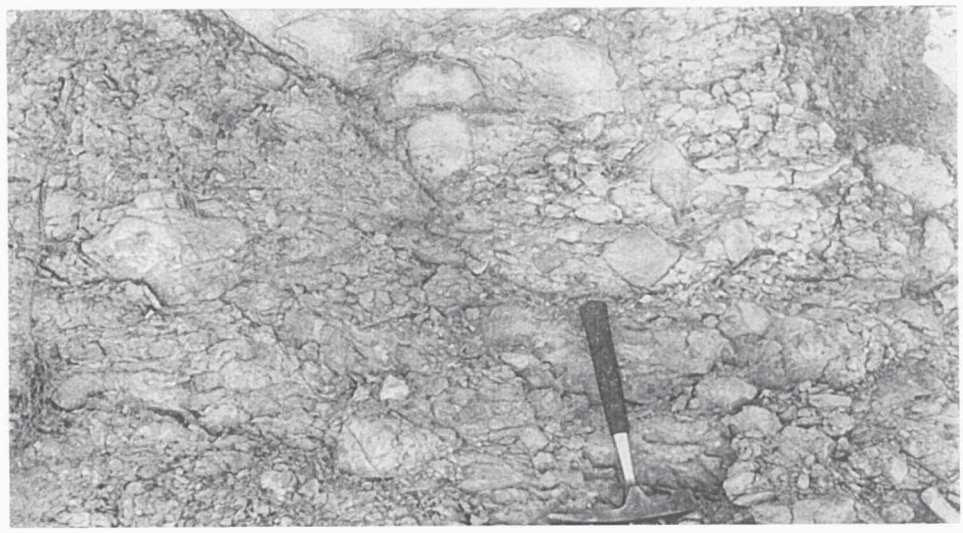


Fig. 37. Close-up of the conglomeratic marlstone in the typical exposure of the B.G.M. Southern bank of the Río Benamor.

Fig. 38. Clean part of the conglomeratic marlstone exposure in the barranco that borders the northern slope of the Cerro de Las Arcas. Benamor Geröllton Member.

lying between 16 and 32 cm. The volumetric percentage of the inclusions appears to be about 25%. It could also be concluded that the majority of the smaller olistholiths (0 to 32 cm) is composed of Cretaceous rocks, whereas the majority of the larger ones (4 to 128 cm) is composed of Moratalla Marlstone. It should be noted that the percentage of Moratalla Marlstone is considerable (about 80-90%), because 30 to 50 percent of the inclusions are derived from this Member and because the matrix is Moratalla marlstone.

Shapes and internal structures: Many of the larger microlistholiths show internal stratification which is apparently the original one. The flattened, plate-like microlistholiths are obviously pieces of beds; they are generally folded and drawn out. The angular and short prismatic ones are apparently fragments of competent beds. Twisted and drawn-out microlistholiths were observed which consist invariably of parallel-laminated white Moratalla Marlstone (fig. 39). The boundary between these olistholiths and the binder is smooth and sharp and at least for the greater part parallel to the internal lamination. Where the surface truncates the internal lamination it is hummocky and irregular, apparently representing broken-off bonds. These olistholiths are therefore interpreted as the fragments of originally larger olistholiths which, during the emplacement of the olisthostrome have been deformed in a visco-plastic way and subsequently broken. A large, irregularly bossed microlistholith of grey, layered Moratalla Marlstone appeared to be internally broken with a horizontal shift of the layers of only a few centimetres. It is not clear whether this breaking took place prior to or during the down-sliding motion of the olisthostrome. The globular or egg-shaped inclusions consist of soft material. For example subspherical bodies of grey marlstone were found with little or no fissility and of a slightly lighter shade than the grey binder, from which they can hardly be distinguished; no internal lamination was found. The boundary between such an inclusion and the binder is sharp and relatively smooth. Here and there small limestone fragments were found to adhere to the outer surface of such inclusions (fig. 40), whereas no limestone fragments were found within them. This indicates that they should not be considered as parts of the binder, but as olistholiths. The small limestone fragments apparently were picked up by the swirling inclusion when still in a visco-plastic state.

Spatial distribution and orientation: The spatial distribution of the microlistholiths according to their size is entirely random; there is no sign of any concentration of the larger ones in the lower half of the depositional unit. The flat inclusions with a longest apparent diameter of 1 m or more are commonly oriented parallel to the horizontal upper boundary of the olisthostrome.

From the spatial distribution, the orientation, and the contorted shapes of the olistholiths it may be concluded that the consistence of the down-flowing mud paste was rather stiff, but not so stiff as to prevent some kind of slow eddying motion.

Rock types: the microlistholiths are generally composed of grey foraminiferal lime-wackestones with globigerinids, radiolarians, and sponge-spicules (Moratalla Marlstone Member). Many microlistholiths also consist of grey bioclastic lime-mudstones or wackestones with non- or single-keeled globotruncanids, *Pithonella,* and *Stomiosphaera* (Represa Formation). Other ones appear to be composed of globigerinal lime-packstones in which the globigerinid tests are in closest packing with micrite in the interstices (Puerta Limestone Member). Also several grey bioclastic lime-packstones were sampled with globigerinids, globorotalids, and *Nummulites* (Moro Marlstone Formation). Less frequent are microlistholiths of

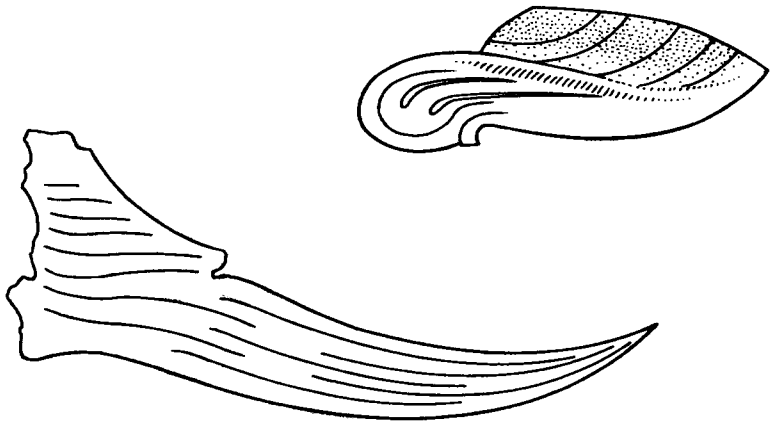


Fig. 39. Twisted inclusions of parallel-laminated Moratalla Marlstone in the typical exposure of the B.G.M., southern bank of the Río Benamor.

Fig. 40. Small limestone inclusions adhering to the outer surface of a spherical olistholith of light-grey marlstone. Typical exposure of the B.G.M., southern bank of the Río Benamor.

white bioclastic lime-wackestones with double-keeled globotruncanids and *Heterohelix* (Quipar Formation); calcareous sandstones of which in one sample globigerinids were found (presumably Moratalla Marlstone Member); and bioclastic pseudo lime-grainstone which is for the greater part composed of tests of larger foraminifera (Moro Marlstone Formation). Moreover one very rounded quartzite pebble was sampled which was assumed to be derived from the Rasa Sandstone Formation.

Microlistholiths and small inclusions: variants in other exposures – In other exposures also smaller inclusions of the Argos Formation, Miñano Limestone Formation, and sandstones of the Rasa Formation were found. Moreover a loose fossil (gastropod) was sampled as inclusion in the binder. And finally black chert nodules (Quipar Formation) and fully limonitized pyrite crystals (Represa Formation) were found as inclusions.

Many Rasa Sandstone microlistholiths (up to 2 m apparent longest axis) resemble rumpled-up wads of paper (fig. 41). At the surface an intricate pattern of small steps and ridges can be seen, whereas fresh cuts and thin sections do not reveal any internal structure. These Rasa inclusions consist invariably of fine sandstone and have a more or less twisted shape. These structures show some analogies with the trough-and-ridge structure found in the fine sandstone beds of the Rasa Sandstone in the Sierra de la Puerta (page 94); the latter are however very regular. The irregular step-and-ridge structure at the surface of the inclusions is also interpreted as due to a dilatational behaviour of the still non-lithified fine sandstone when a sudden stress was exerted upon it by the churning marlstone during the movement of the olisthostrome.

Olistholiths larger than 5 m: typical exposure – Near the eastern end of the conglomeratic marlstones of the typical exposure a mesolistholith is protruding (encl. 2). It consists of a bedded, white, intraformational breccia of bioclastic lime-mudstone with planktonic foraminifera of Cretaceous age (Quipar Formation of the South-Prebetic zone); here and there very rounded pebbles and boulders were found between the subangular fragments. The rock mass shows signs of deformation and is intersected by many slickensides and calcite-filled fractures. At the surface of the olistholith loose breccia fragments form lenticular breccia bodies sticking to the parent rock mass. A halo of breccia fragments, diffused in the binder, occurs around the olistholith. Presumably the surrounding marl had moved with respect to the brecciated rock mass, through which the breccia fragments became dislodged and were taken up by the mobile marl. As it seems most unlikely that the soft marls were able to transmit sufficiently large stresses to cause the plications and the slickensides in the large competent limestone olistholith, the deformation of the latter must have taken place prior to its downsliding.

Directly west of the Cretaceous rock mass a large mesolistholith of conglomeratic marlstone of the South-Prebetic Moro Formation is included in the conglomeratic marlstone of the typical exposure (encl. 2). Superficially there is no difference between the two conglomeratic marlstones. The one of the Moro olistholith, however, has a binder that consists of grey marlstone in which many angular sand-sized quartzite grains and larger foraminifera (among which *Amphistegina* and small *Nummulites)* are scattered. The inclusions of the Moro conglomeratic marlstone consist of grey fine-grained limestone with larger foraminifera; most of them are pieces of beds of medium thickness which are bent, folded, and

Fig. 41. Rasa Sandstone olistholith resembling a rumpled-up wad of paper. It shows an intricate pattern of small steps and ridges at the surface, which is explained by a dilatant behaviour of the sand when sudden stresses were exerted upon it during the olisthostromization. Southeastern slope of the Cerro de San Jorge.

Fig. 42. Rasa Sandstone olistholith with limestone clasts of Cretaceous rock adhering to the outer surface. The adhesion is explained by assuming that the olistholith had sucked up water from the enclosing marl owing to the dilatational behaviour of the still non-lithified sandstone. Southeastern slope of the Cerro de San Jorge.

contorted; one of them is more than 5 m long. It was impossible to detect the exact boundary between the conglomeratic marlstone of the Moro olistholith and that of the B.G.M. in which it is included. Nevertheless the Moro rocks are found to be continually exposed over a distance of about 30 m (encl. 7).

At the western extremity of the typical exposure a mesolistholith was found, consisting of bedded, light-grey bioclastic lime-mudstone of Cretaceous age. The upper boundary of the olistholith parallels the horizontal apparent longest axis but truncates the folded stratification. Also the lower boundary is horizontal except for its eastern end where it curves upward to become parallel with the stratification, which at that spot dips about 45° west. Apparently this rock mass is a slice broken off from a folded Cretaceous rock sequence and slumped down to become incorporated in the olisthostromization.

Olistholiths larger than 5 m: other exposures – The large olistholiths are generally the only outcropping components of the B.G.M. (encl. 2). They may attain a considerable diameter. Apparent longest diameters of more than 100 m are not exceptional at all. The largest olistholith has an apparent longest axis of more than 500 m.

All rock types found as inclusions in the B.G.M. appear to be identical to those found in the South-Prebetic zone and in the northern part of the Subbetic zone. The great majority of the large exotic olistholiths are parts of the Ulea Group as it is developed in the South-Prebetic zone, and parts of the Rasa Sandstone Formation. The Ulea-blocks consist of Cuervo Sandstone, Asares Conglomerate, Puerta Limestone, and Moratalla Marlstone. The first two rock units occur exclusively in the exotic olistholiths. As in most of these olistholiths several rock units can be found together in normal stratigraphic order, a fairly complete picture of the parent rock sequence can be obtained and the stratigraphic position of the Cuervo Sandstone and the Asares Conglomerate can be established. The Moratalla Marlstone found in some of the exotic rock masses differs slightly from that underlying the B.G.M. and from that of the Sierra de la Puerta. It consists of a regular alternation of soft, fissiled, white limestone beds and harder white limestone beds largely composed of globigerinids; in the soft beds the globigerinids are less numerous; sandstone beds were not found. These occurrences are the only evidence of a different development of the Moratalla Marlstone Member in the source area of the olisthostromes. Less common are large olistholiths composed of Cretaceous rocks of the Represa and Quipar Formations. Rare are large olistholiths of Moro Marlstone, Miñano Limestone, and Gredero Formation (white fissiled limestones with interbeds consisting of Eocene larger foraminifera). Very rare are the inclusions of the Jorquera Formation. Rocks of the Argos Formation only occur in fragments smaller than 2 m. Only two large masses of Moratalla Marlstone comparable with that underlying the B.G.M. were found.

The large tabular olistholiths composed of rocks from the South-Prebetic Ulea group and of Rasa Sandstone are characterized by two important features.
1. The internal stratification of these olistholiths is always parallel to the basal plane of the rock mass.
2. Their attitude is always normal, except for the two cases in which these plate-like olistholiths were found to be recumbently folded. These rock slices apparently became separated from a rock sequence of normal attitude along the planes of stratification and slid down while

retaining this attitude. The olistholiths of internally folded Cretaceous limestones and Moro Marlstone, however, do not show such marked parallelism between basal plane and internal stratification, neither could their attitude be determined. The shape of the large olistholiths is predominantly flattened with tapering, rounded, or rectangular margins. Large isodiametric inclusions are rare: some of the Rasa Sandstone inclusions are subspherical. Some of the large tabular olistholiths of South-Prebetic Ulea rocks and Rasa Sandstone are folded in such a way that the whole thickness of the plate participates in the concentric folds. These folds vary from faint undulations to northward recumbent folds.

1. One tabular olistholith of Cuervo Sandstone (50 m west of the bridge over the barranco that borders the northern slope of the Cerro de las Arcas) displays a northward recumbent fold. The absence of strong fracturing indicates that the Sandstone was not yet indurated during the folding. The strike of the fold axis is N 120 E corresponding to a stress from the south directed N 30 E. This is in accordance with the striation (N 20 E) found on a calcite fill along a subhorizontal fault plane within the Upper Cretaceous micritic limestone in the core of this fold. This limestone possibly belongs to an olistholith of the underlying depositional unit of the B.G.M.
2. The megolistholith of South-Prebetic Ulea rocks over which the Barranco de Azafran makes a high waterfall, shows an overturned anticline and a syncline. The measured strikes and dips correspond to an average stress from the south directed N 26 E. The white micritic limestone in the core of the anticline is of unknown affinity.
3. The strike and dip of the most inclined part of a bent, tabular olistholith of Ulea rocks in the westernmost cluster of large exotic olistholiths exposed in the Río Benamor is N 68 E/32 S which corresponds to a stress directed N 22 W.
4. The strike of the axis of a recumbent fold in a conglomeratic intercalation in the lower part of the B.G.M. (southern slope of the Cerro de las Arcas) appeared to be N 90 E corresponding to a stress directed due north.

The folds were probably formed by stresses exerted by the surrounding conglomeratic marlstones upon the rock plates in the direction of movement during their joint downward travel. The stress directions found are about the same as those of the folds formed by the soft-rock deformation in the underlying Moratalla Marlstone Member (page 118, 120), and as those in the slid-down blanket of Cretaceous rocks (Nevazo Unit) overlying the B.G.M. (page 172). The consistency of the strike measurements, which only show relatively small deviations from the E-W direction, indicates that a possible horizontal rotation had been relatively small.

Many large mesolistholiths and macrolistholiths of coarse-grained Rasa Sandstone show strong internal contortions which are small in relation to the size of the olistholiths. In a macrolistholith composed of a medium-bedded alternation of coarse sandstone and green clay, for instance, the folded, curved and contorted beds do not show any trace of fracturing. The beds are irregularly thinned, thickened, and tapered; they are pulled apart into strongly plicated and rolled-up fragments commonly with drawn-out ends; some parts are even balled up into large sandstone balls. Obviously the Eocene Rasa Sandstone was not lithified when the deformation took place. But deformations of similar style and amplitude were not found in the parent Rasa Sandstone outcrops in the Sierra de la Puerta. The opinion is therefore favoured that this deformation is not a syndepositional one, but that it came about when the folding of the South-Prebetic

rock sequence took place, thus during the Middle Miocene.

Also in other exposures, for instance in the southern flanks of the Cerro de las Arcas and the Cerro de San Jorge, haloes of dispersed breccia fragments were found in the marlstone binder surrounding the olistholiths of white intraformational limestone breccia of the Quipar Formation. The radii of the haloes measure a few metres to some tens of metres. Some of these breccia fragments even adhere to the surface of neighbouring inclusions of relatively fine-grained Rasa Sandstone and are somewhat pushed inward (fig. 42). Perhaps the sucking up of water from the surrounding gravelly marl as a result of the dilatational behaviour of the sandstone may be responsible for the surficial adherence of the white breccia elements, for otherwise sandstone has no adhesive properties. Apparently the Rasa Sandstone was not yet lithified even when it became involved in the olisthostromization.

Zonal succession in the lower part of the B.G.M.

The faintly layered appearance of the B.G.M. can be assumed to be the result of several successive olisthostrome influxes and mud flows superposed upon each other. An attempt was made to delimit the successive depositional units and to find a correlation between the successions found at different localities. The degree of exposure, however, is too poor to trace directly the boundaries of the various units. In fact almost everywhere only the larger olistholiths crop out, whereas the finer-grained parts of the units are covered with a veneer of rubble. Hence these larger olistholiths are the only means by which the correlation can be carried out.

It soon appears then that, at least in a restricted area, almost every row of large olistholiths shows a remarkable preponderance of olistholiths of one rock type. Also it appeared that at different localities more or less similar successions of rows could be distinguished each carrying its distinctive preponderant type of olistholith. Since it may be doubted whether each row actually corresponds to one depositional unit, the informal rockstratigraphic term 'zone' is used for each row of olistholiths that occur at the same level and show comparable characteristics

The zones are delimited according to the following principles:

1. Every row or rows of olistholiths that can be distinguished from the rows directly above and below comprise one zone.
2. Every interval that can be assumed to be one depositional unit comprises one zone.
3. Every horizon that can be assumed to be the boundary between two successive depositional units is also the boundary between two successive zones.

The successions of the Río Benamor, the Cerro de las Arcas, and the Collado de San Jorge constitute the composite stratotype of the lower part of the B.G.M.

Río Benamor succession – Exposed in the right bank of the Río Benamor 250 m upstream from its confluence with the Barranco de Azafran (encl. 2).

Zone 1: this conglomeratic marlstone unit is the stratigraphically lowest unit of the B.G.M. found. It is directly underlain by horizontally bedded Moratalla Marlstone Member of which the upper 5 m are intricately folded (page 118).

At the base of the unit three small mesolistholiths of about equal size were found imbricated upon each other. Two of them consist of brick-red limestone ("couches rouges", Quipar Formation, Upper Cretaceous), the other of Moro Marlstone (Paleocene-lowest Eocene). Also folded fragments of beds of the Moro Marlstone Formation with abundant foraminifera were found. The upper part consists of

conglomeratic marlstone with smaller inclusions (from granule size to 80 by 60 by 30 cm) mainly of Moratalla Marlstone (Middle Miocene). The depositional unit is directly overlain by an undisturbed horizontal sequence of alternating sandstone and marlstone beds similar to those of the Moratalla Marlstone Member below. It was proved by a time-consuming bed-to-bed correlation of various exposures of the Moratalla Marlstone Member that in the Barranco de Azafran, 350 m north of the exposure in the Río Benamor, the conglomeratic marlstone of this zone is replaced by horizontally bedded Moratalla Marlstone with turbidites.

Cerro de las Arcas succession – Exposed in the southern slope of the Cerro de las Arcas (encl. 2, 8a, 8b, and 8c).
Zone 2: in the Cerro de las Arcas this depositional unit comprises the lowest horizontal row of mesolistholiths of Middle Miocene Puerta Limestone or Asares Conglomerate (relatively thick). The base of the unit is not exposed, but in the level directly below a horizontally bedded sequence of the Moratalla Marlstone Member is exposed. The largest and westernmost Puerta Limestone mass of this zone was found to be associated with conglomeratic marlstones. This rock mass probably is also the westernmost limit of the depositional unit, because directly west of it a horizontally bedded sequence with turbidites of the Moratalla Marlstone Member is continuously exposed over a thickness of about 20 m. This undisturbed sequence is overlain by a B.G.M. unit tentatively correlated with zone 6. The mud flow deposits and Bejar Beds described on page 132 are included in this zone, just as the Bejar Beds cropping out under the bridge over the barranco that borders the northern slope of the Cerro de las Arcas. Zone 2 is overlain by a 2 to 3 m thick, horizontally bedded sequence which is similar in lithology to the Moratalla Marlstone.
Zone 3: two macrolistholiths were found which are composed of the following successive rock units: at the base Rasa Sandstone (Lower and Middle Eocene), in the middle Cuervo Sandstone (Middle Miocene), and at the top Asares Conglomerate (Middle Miocene) which is about 10 to 15 m thick. In the westernmost olistholith, the Asares Conglomerate is absent. Also a mesolistholith exclusively composed of Rasa Sandstone was found in this zone. Only in this zone were olistholiths seen in which the Cuervo Sandstone unambiguously is in conformable contact with the Rasa Sandstone. Directly west of the western-most macrolistholith this zone pinches out between the bedded sequence overlying zone 2 and the rocks of zone 4 by way of a tapering end of conglomeratic marlstone. It is doubtful whether the easternmost mesolistholith also marks the other end of the tranverse section of this zone, because exposures are lacking directly east of it.
Zone 4. the next horizontal level is characterized by a row of sheet-like olistholiths, among which megolistholiths and macrolistholiths, consisting of white micritic limestone from which a rudist fragment was collected (Quipar Formation, Upper Cretaceous). Locally these olistholiths are in contact with those of the zones directly below or above. The strong brecciation makes that the rock falls apart into angular, isodiametric fragments (many are nearly cubical) of about 1 cm^3. Neither can this zone be traced further westward, because it is replaced there by a continuous sequence of horizontally bedded Moratalla Marlstone Member.
Zone 5: over, and generally in contact with, the sheets of Cretaceous limestone of zone 4 lies a row of sheet-like olistholiths (among which a megolistholith) composed of Puerta Limestone and Asares Conglomerate (Middle Miocene). In one

macrolistholith the conglomerate is underlain by Cuervo Sandstone (Middle Miocene). At one place along the contacts with the olistholiths of zone 4 an intimate mixture was observed of the sandy Asares Conglomerate of the upper olistholith with fragments of the brecciated Cretaceous limestone of the lower one. This may be explained by assuming that the Asares Conglomerate was still in a non-cemented state when the upper olistholith slithered over the lower one, while the Cretaceous limestone was already brecciated and lithified, because the breccia fragments remained angular. The recumbently folded mesolistholith of Cuervo Sandstone described on page 147 is correlated with this zone, because it occurs at the same altitude and contains brecciated Cretaceous limestone in the core of the fold. Also this zone is replaced to the west by horizontal beds of the Moratalla Marlstone Member.

Zone 6: this zone fills up the space between the rows of olistholiths of zone 5 and 7. It consists of conglomeratic marlstone and marlstones with only a few inclusions. In this zone small mesolistholiths of Puerta Limestone (Middle Miocene), Rasa Sandstone (Lower and Middle Eocene), and Moro Marlstone (Paleocene and lowest Eocene) are included. The typical exposure of the B.G.M. is tentatively correlated with this zone, because the zone overlying the typical exposure has the characteristics of zone 7. But the conglomeratic marlstone unit of the typical exposure probably represents a separate depositional unit at the same level.

Zone 7: the olistholiths of this row are composed of Puerta Limestone (Middle Miocene) with only a few metres thick basal part of Asares Conglomerate which in some olistholiths is underlain by Cuervo Sandstone (Middle Miocene). This string of olistholiths characteristically contains many megolistholiths and macrolistholiths which render the zone very conspicuous. The associated conglomeratic marlstones also contain contorted Rasa Sandstone microlistholiths (Lower and Middle Eocene), contorted portions of sandstone interbeds of the Moratalla Marlstone Member, white micritic limestone inclusions of Cretaceous age, and white hard limestone blocks of the Moratalla Marlstone Member. Also a large mesolistholith of Represa Formation with Vraconian ammonites was found in this zone. This occurrence is rather exceptional, for all other large olistholiths of Represa Formation are restricted to zone 14.

Zone 8: the conglomeratic marlstones of this zone are separated from those of the underlying zone by a thin sequence of grey marlstones with several thin to medium thick interbeds of polymictic calcirudite (Bejar Beds). This zone can be distinguished only in the southern slope of the Cerro de las Arcas. The conglomeratic marlstones contain mesolistholiths of Puerta Limestone (Middle Miocene), facetted boulders of white micritic limestone of Cretaceous age, rounded sandstone cobbles of the Moratalla Marlstone Member (Middle Miocene), and rhomboedric cobbles of Puerta Limestone. The thin horizontal sequence with Bejar interbeds was found to cut halfway the vertical height of two large olistholiths of zone 7. This relation is interpreted as follows: when the olisthostrome of zone 7 came to rest, these large olistholiths stood out above the surface of the associated conglomeratic marlstones giving the bottom topography a hummocky appearance. The calcirudite then became deposited only in the depressions between the hummocks.

Zone 9: This zone is characterized by extensive sheets (of 300 and 150 metres across) of white and brick-red micritic limestones and bedded intraformational limestone conglomerates in which a fragment of *Eupachydiscus* (ammonite) and double-keeled globotruncanids were sampled. These sheets of Upper Cretaceous

Quipar Formation, in a facies that was found only in the South-Prebetic zone, are either lying upon zone 8 or directly upon the large olistholiths of zone 7. The sheet in the Cerro de las Arcas succession is the only olistholith of inverted attitude found, for the brick-red beds are located below the intraformational conglomerates.

Zone 10 and higher: directly upon these sheets of "couches rouges" lies a flat, small mesolistholith of Rasa Sandstone (Lower and Middle Eocene). It is followed by a very thick sequence of homogeneous, non-stratified grey marlstones, lithologically and palaeontologically similar to the Moratalla marlstones, showing a heavily contorted fissility pattern which is interpreted as the outward expression of internal contortions that are otherwise obscure. In this sequence only a few widely scattered mesolistholiths to megolistholiths occur. The virtual absence of microlistholiths is striking. The morphology of the outcropping rocks is of the bad-land type. In the southern slope of the Cerro de las Arcas this sequence cannot be subdivided into zones, but in the Collado de San Jorge and in the northern slope of the Cerro de las Arcas it can be subdivided into six zones.

Collado de San Jorge succession – Exposed in the east bank of the Barranco de Azafran directly north of the road from Moratalla to Sabinar (encl. 2, 9a, and 9b).

Zone 7: the megolistholiths of the lowest row in this exposure are composed of the following succession of rock units: Cuervo Sandstone (Middle Miocene), Asares Conglomerate of a few metres thick (Middle Miocene), Puerta Limestone (Middle Miocene), and a thin bedded sequence of Moratalla Marlstone as it presumably was developed in the source area of the olisthostromes (Middle Miocene). This level of olistholiths is correlated with zone 7 of the Cerro de las Arcas succession,

(1) because of the large size of the olistholiths and the composing rock units,

(2) because it is immediately overlain by sheet-like olistholiths composed of "couches rouges" (Quipar Formation, Upper Cretaceous),

(3) because its vertical distance above the level within the underlying bedded Moratalla Marlstone Member that corresponds exactly with the depositional unit of zone 1 is approximately the same as in the Cerro de las Arcas,

(4) because its vertical distance below the base of the Nevazo unit is the same as in the Cerro de las Arcas,

(5) and because the material between the large olistholiths of the higher zones in this exposure consists of homogeneous, non-stratified marlstones without microlistholiths like in the Cerro de las Arcas. Directly east of the megolistholith of zone 7 in the northern slope of the Cerro de San Jorge a continuous sequence of horizontally bedded Moratalla Marlstone Member is exposed, so that this olistholith can be interpreted as the easternmost limit of the depositional unit of zone 7.

Zone 8: was not observed.

Zone 9: the sheet-like megolistholiths of white and brick-red micritic limestones (Quipar Formation, Upper Cretaceous) were placed in this zone. Also the mesolistholith of Upper Cretaceous bedded intraformational limestone conglomerate (Quipar Formation) in the southeastern slope of the Cerro de San Jorge is correlated with this zone. In the northern slope of this hill, however, the zone could not be traced further eastward, because there it is replaced by bedded Moratalla Marlstone Member.

Zone 10: the mesolistholith of Rasa Sandstone (Lower and Middle Eocene) above zone 9 is thought to represent a separate zone. In this zone also mesolistholiths

of Puerta Limestone (Middle Miocene) are included. Like zones 7 and 9 also this zone is bounded to the east by a megolistholith and replaced by bedded Moratalla Marlstone Member in the northern slope of the Cerro de San Jorge.

Zone 11: two megolistholiths of brecciated and contorted white micritic limestone, locally associated with Upper Cretaceous brick-red limestones (Quipar Formation), in this level are grouped into zone 11. East of the megolistholith in the northern slope of the Cerro de San Jorge this zone is replaced by horizontally bedded Moratalla Marlstone Member.

Zone 12: This zone comprises a conspicuous row of flat olistholiths in which the following stratigraphic succession was wholly or partly found: Rasa Sandstone (Lower and Middle Eocene), Cuervo Sandstone (Middle Miocene), Asares Conglomerate of only a few metres thick (Middle Miocene), Puerta Limestone with interbeds of light green marlstones (Middle Miocene) passing upward into Moratalla Marlstone as it was presumably developed in the source area of the olisthostromes (Middle Miocene). Several megolistholiths are included in this zone, for instance (1) in the southwestern slope of the Cerro de San Jorge (where small Upper Eocene Miñano Limestone masses were found between the Cuervo and the Rasa Sandstones), (2) the huge folded one over which the Barranco de Azafran is compelled to cascade, and (3) the one in the west bank of this barranco.

Zone 13: mesolistholiths of white micritic limestones of Cretaceous age are directly lying upon the olistholiths of zone 12. Between this level of Cretaceous olistholiths and those of zone 14 is an interval of grey homogeneous, non-stratified marlstones without inclusions, which is about 10 m thick. The Cretaceous olistholiths together with the marlstone interval are incorporated in zone 13.

Zone 14: in this zone are incorporated the megolistholiths of bedded, light-grey micritic limestone with thin grey marlstone interbeds. In the olistholith at the Collado de San Jorge, which is exposed over a length of more than 500 m, many ammonites were collected indicative for an Upper Vraconian age *(Stoliczkaia dispar* zone).

The following ammonites were determined: *Mariella (Mariella) bergeri bergeri* (Brongn.), *M. (M.) bergeri crassituberculata* Spath, *Mortoniceras (Durnovarites) subquadratum* Spath, *Discohoplites* cf. *D. subfalcatus* (Sem.), *Anisoceras perarmatum* (Pict. & Camp.), *Stoliczkaia (Stoliczkaia) dispar* (d'Orb.), *Lechites moreti* Breistr., *L. gaudini* (Pict. & Camp.), *Hamites (Stomohamites) subvirgulatus* Spath, *Desmoceras latidorsatum* (Mich.), *Puzosia mayoriana* (d'Orb.), *Hemiptychoceras* sp., *Phylloceras* sp. The microscope revealed *Pithonella, Stomiosphaera,* and single-keeled globotruncanids.

The megolistholiths of Cretaceous limestone in the southern slopes of the Cerro de San Jorge and the Cerro de las Higueras, and in the northern slope of the Cerro de las Arcas are correlated with this zone. The Higueras olistholith lies directly upon a plate-like megolistholith of Puerta Limestone of zone 7; the zones 8 to 13 are lacking there.

Zone 15: the uppermost tabular megolistholith is composed of Cuervo Sandstone (Middle Miocene), Asares Conglomerate (Middle Miocene), and Puerta Limestone (Middle Miocene), and is separated from the megolistholiths of zone 14 by a thin intercalation of grey marlstone. The highest row of mesolistholiths of the same composition in the southern slope of the Cerro de San Jorge is correlated with this zone. And so are the megolistholiths of similar composition directly above the large Cretaceous olistholith near the top of the northern slope of the Cerro de las Arcas and at the top of the Cerro de las Higueras.

In the ridge from the Collado de San Jorge to the Cerro del Cuervo the next higher level consists of coarse polymictic limestone conglomerates (a few metres thick) belonging to the Bejar conglomerates of the upper part of the B.G.M. This is, in turn, overlain by the Nevazo Unit.

The total thickness of all 15 zones of the Cerro de las Arcas and the Collado de San Jorge successions together, thus of the lower part of the B.G.M., is approximately 150 m. The average thickness of the zones must therefore be about 10 m. Their thicknesses, however, range from 0 to 25 m.

Succession of the Cerro Hilario and the Cerro de las Piedras Llanas. – The zonal succession of the B.G.M. in this region can only tentatively be correlated with those north of the Río Benamor (encl. 10). In the Cerro de las Piedras Llanas (encl. 2) a level could be indicated with megolistholiths of brick-red and white micritic limestones of Late Cretaceous age in which many *Inoceramus* and *Ananchites* were found. This level is correlated with zone 9 of the exposures north of the Río Benamor, (1) because it is the only level in which large quantities of "couches rouges" were found, and (2) because this level is at the same altitude as zone 9 in the northern exposures. The two rows of olistholiths below zone 9 in the Cerro Hilario, therefore, may tentatively be correlated with zone 7 (megolistholiths of Rasa Sandstone in the eastern slope of the Cerro Hilario) and zone 8 (mesolistholiths of Asares Conglomerate and Puerta Limestone; also a few mesolistholiths of micritic limestone of Cretaceous age were found in this level). Zone 10 comprises megolistholiths of Rasa Sandstone, one of which forms the top of the Cerro Hilario. Small slices of Miñano Limestone are stratigraphically associated with the Rasa Sandstone. Zone 11 contains megolistholiths of white micritic limestone of Cretaceous age and also some "couches rouges". The row of olistholiths that includes the megolistholith on top of the Cerro de las Piedras Llanas, which is composed of Cuervo Sandstone, Asares Conglomerate, and Puerta Limestone, is tentatively correlated with zone 12 of the successions north of the Río Benamor. The altitude of this zone is however some 50 m higher than that in the Collado de San Jorge. The white micritic limestone megolistholith of the next higher level may be correlated with zone 13.

It was not possible to correlate the olistholiths found at higher levels in the region southwest of the Cerro de las Piedras Llanas.

A kind of cyclic development can be detected in the lower part of the B.G.M. Each cycle comprises 2 or 3 successive zones with increasingly older olistholiths of large size. The successive cycles comprise zones 2-4, 5-7, 8 and 9, 10 and 11, 12-14 respectively. Each cycle starts with a zone characterized by large Middle Miocene Ulea olistholiths, the last zone of a cycle is characterized by large Cretaceous Quipar and Represa olistholiths. The zone in the middle, if present, generally contains large Eocene Rasa and Moro olistholiths. Towards the top of the lower part of the B.G.M. the first zone of a cycle may already contain large Eocene olistholiths and the middle zone already Upper Cretaceous olistholiths, whereas the upper zone contains Lower Cretaceous ones. The cyclicity may be explained by assuming a strip by strip recession of the source area by delapsion. From each strip first the upper and younger rocks were shed and subsequently the lower and older ones, then it starts all over again for a more inward strip and so on. In the course of this process increasingly older rocks got involved in the

delapsion, probably due to a gradual upwarping of the source area.

Stratigraphic relations, shape and macro-texture of the olisthostromes

The B.G.M. is only found to the south of the Río de Moratalla with the exception of the top of the Cerro de Moratalla la Vieja; the most easterly occurrence known is situated on top of a hill 500 m east of the eastern boundary of the investigated area at the same latitude as the Cerro de Rótova; the two hills are crowned by several large blocks of Roble limestone. The B.G.M. does not extend more than about 1 km west of the western boundary of the area.

1. It was shown (page 149) that the olisthostrome of zone 1 of the B.G.M. is towards the North replaced by horizontally bedded Moratalla Marlstone Member with turbidites in the Barranco de Azafran 100 m southeast of the bridge with the road from Moratalla to Sabinar. But olistholiths of zone 5 were indeed found at this latitude. Some 150 m further north along the Barranco de Azafran the lowest level of olistholiths is correlated with zone 7, as is the lowest row of olistholiths in the Cerro de las Higueras again 1 km further north. At a yet more northern latitude, at the altitude of 800 m in the Barranco de la Celada, the only olisthostrome found is directly overlain by the Nevazo Unit, and underlain by a 200 m thick continuous sequence of bedded Moratalla Marlstone Member. It may be concluded that the olisthostromes of the B.G.M. are replaced by the Moratalla Marlstone Member in the north, and that the higher units extend further northward than the lower ones.
2. In the western part of the northern slope of the Cerro de San Jorge olisthostromes occur which can be correlated with zones 7 to 13, whereas in the middle and eastern part of this slope a continuous sequence of Moratalla marlstone can be traced almost to the top of the hill. The lower part of this sequence is well bedded up to the level of zone 11. The upper part is non-stratified. In the southern slope this upper part contains large olistholiths correlated with zones 11 to 15, so that this upper part apparently belongs to the B.G.M. Hence in the northern slope the olisthostromes correlatable with zones 7 to 11 have a limited eastward extension and are replaced to the east by bedded Moratalla Marlstone Member. In the Cerro Hilario the lowermost row of olistholiths was correlated with zone 7. Below this row continuous sequence of horizontally bedded Moratalla Marlstone occurs. This implies that the olisthostromes of zones 1 to 6 exposed in the Cerro de las Arcas also have a limited eastward extension and also are replaced to the east by bedded Moratalla Marlstone.
3. In the southern slope of the Cerro de las Arcas the western ends of zones 2 to 5 abut against a continuous sequence of horizontally bedded Moratalla Marlstones with turbidites. Thus these olisthostromes have also a limited westward extension and are about 1 km broad.

These three points may indicate that the successive olisthostromes of the lower part of the B.G.M. have the shape of flat tongues (about 10 m thick and 1 km broad) which are either amalgamated or separated from each other by thin intercalations of rocks similar to the Moratalla Marlstone Member, or by Bejar Beds. The higher tongues extend further northward than the lower ones. To the east, the north, and the west the olisthostrome tongues are replaced by the Moratalla Marlstone Member which also occurs below the B.G.M. Apparently the lower

part of the B.G.M. is interfingering with the upper part of the Moratalla Marlstone Member.

On account of this interfingering relation it can be inferred that at least the greater part of the B.G.M. has the same age as the upper part of the Moratalla Marlstone Member, i.c. Serravallian.

The fact that the western boundaries of the zones 2, 4, and 5 are coinciding, and that this also is the case with the eastern boundaries of the zones 7, 9, 10, and 11, cannot be regarded as accidental. It seems that the spreading of the olisthostromes has been obstructed in that direction. Presumably at both localities these coinciding boundaries represent the steep sides of broad, shallow channels in the Moratalla Marlstone worn by several successive olisthostrome influxes which have flowed along the same course.

Whenever the eastern and (or) western extremities of the transverse sections through the olisthostrome tongues are irrefutably recognized, it appears that they are marked by a large olistholith. This is, as we have seen, the case for the western extremities of zones 2, 3, and 4 in the southern slope of the Cerro de las Arcas and for the eastern extremities of zones 7, 9, 10, and 11 in the northern slope of the Cerro de San Jorge. Also the eastern extremity of the typical exposure is marked by a large olistholith which even has an upright attitude. These observations suggest that the lateral margins of the olisthostrome tongues tend to be lined with large olistholiths.

It also appears that the largest olistholiths, with an apparent longest axis of more than 300 m, are situated in the northern (= distal) part of the B.G.M., north of the Barranco de Azafran. For instance:

1. The Ulea olistholith (zone 7) in the east bank of the Barranco de Azafran: 300 m.
2. The Ulea olistholith (zone 7) in the Cerro de las Higueras: 300 m.
3. The Quipar olistholith (zone 11) in the northwestern spur of the Cerro de San Jorge: 350 m.
4. The Rasa olistholith (zone 12) in the southwestern slope of the Cerro de San Jorge: 350 m.
5. The Ulea olistholith (zone 12) in the Barranco de Azafran: 350 m.
6. The Rasa (+Ulea) olistholith in the northwestern spur of the Cerro de San Jorge: 400 m.
7. The Represa olistholith (zone 14) in the Collado de San Jorge: more than 500 m.
8. The Represa olistholith (zone 14) in the Cerro de las Higueras: more than 500 m.

The only olistholiths of comparable size found south of the Barranco de Azafran are: (1) the "couches rouges" olistholith (zone 9) in the southern slope of the Cerro de las Areas (300 m) and (2) the Quipar olistholith (zone 13) 700 m southwest of the Cerro de las Piedras Llanas (400 m). In the northwestern spur of the Cerro de San Jorge tabular olistholiths of six successive zones (7 to 12) are imbricated upon each other without matrix between them. Also in the Cerro de las Higueras and in the spur between the two main tributaries that unite into the Barranco del Commendador the large tabular olistholiths are imbricated upon each other without matrix between them. The tapering distal parts of the olisthostromes of zone 7 to 15 are probably situated in these localities. This distribution, though there are statistically insufficient data, suggests that the largest olistholiths have the tendency to become concentrated near or at the distal ends of the

olisthostrome tongues, hence also at a marginal position.

With respect to the vertical distribution of the olistholiths according to their apparent diameter, the occurrence of the largest ones in the highest zones suggests that the largest inclusions tend to occur in those olisthostrome tongues that extended farthest from the source area.

A similar shape and macro-texture can be observed for the olisthostromes of the Haymond Formation (Texas, U.S.A.) described by King (1938) and McBride (1966, 1970). From the map of King can be concluded that

(1) the Haymond olisthostromes are thickest near their distal (= western) ends,
(2) the maximum thickness of the successive olisthostromes increases from the lower to the middle ones and decreases from the middle to the highest units,
(3) the thicker the olisthostrome, the larger the inclusions,
(4) the larger inclusions are concentrated where the olisthostrome is thickest,
(5) the concentration of larger inclusions diminishes towards the proximal (= eastern) end,
(6) the largest inclusion found is situated in the tapering distal end of the thickest olisthostrome.

Some general aspects of the olisthostromes

Since the olisthostromes consist for about 80 to 90% of Moratalla marlstones, which have flowed like a though paste in which a kind of slow eddying motion has taken place, they can be designated as endolisthostromes. As most exotic olistholiths are apparently rock slices that fell, slumped and slipped down from an uplifted area (built up of a Subbetic and South-Prebetic rock sequence of normal attitude) to become incorporated in the Moratalla marlstones, they can be designated as exolistholiths. The mainly delapsive denudation of the source area probably proceeded strip by strip; probably each marginal strip of rock had completely delapsed away (first the upper and younger parts, then the lower and older ones) before the next strip became subject to delapsion. The intensity of denudation suggests that the source area was emerged above sea level. The delapsed rock masses became embedded in the Moratalla marlstone accumulations (for a large part composed of detritus of the same source rocks) bordering the emerged source area probably in the same way as Stanley & Unrug (1972) describe from the submarine slope off the French rivièra. These soft accumulations repeatedly flowed down together with the inclusions, which resulted in a progressive succession of at least 15 olisthostromes and mud flow deposits in the deeper parts of the Moratalla basin. It is possible that in some cases the downflowing of these soft accumulations was triggered by the delapsion of a very large rock slice upon them. The appearance of increasingly older inclusions in growing numbers in the successively higher olisthostromes (inversion), indicates that successively deeper parts of the rock sequence in the source area were exposed to denudation.

The downflowing diamictonic rock mixtures scoured out broad, shallow channels in the sloping substratum, along which also several subsequent flows found their course. Locally the rocks of the substratum and channel walls were squeezed and pushed into wrinkles by the passing flow. The Moratalla Marlstone masses dislodged from the channel walls and the substratum, also became incorporated in the olisthostromes and may be designated as hypolistholiths. The harder beds of the original Moratalla marlstone accumulations were broken,

disrupted, folded, and balled up, and may be designated as endolistholiths. As a result of increased inmixing of water, the upper part of the downflowing rock mixtures sometimes went into suspension. The suspended parts were subsequently deposited on top of the conglomeratic marlstone units and farther basinward.

The resulting olisthostromes and mud flow deposits are flat tongue-shaped. The thickest part of the tongue would be near the distal end. Probably also the width of the tongue would be greatest in this part. The greatest volume of inclusions would tend to concentrate in the thickest and widest part, whereas the larger ones would tend to get a marginal position. The largest inclusions would tend to reach the tapering tip of the tongue. The larger olistholiths may stick out above the undulating surface of the olisthostrome, which results in a hummocky surface, a typical feature of olisthostromes.

Marine currents acting on the just emplaced olisthostromes and mud flow deposits may have winnowed away the mud of the top part and displaced the smaller inclusions, which finally resulted in Bejar intercalations between the successive olisthostromes. The deposition of Moratalla marlstone was resumed after the emplacement of each of the first olisthostromes. Between the later olisthostromes, however, only sporadically Moratalla marlstones were found in the depressions of the undulating surface of the olisthostrome. Presumably the olisthostrome influxes more rapidly succeeded one another, so that the possible thin deposits were eroded by the next olisthostrome influx. The increasing intensity of delapsion, suggested by the ever more rapid succession of influxes of olisthostromes and olistholiths of increasingly greater volume, indicates the growing intensity of the diastrophism in the region

It should be emphasized that the chaotic stratification can be recognized by the fissility pattern, which helps to identify flowed-down mudstones, especially when inclusions are scarce or absent.

Although the olisthostromes in the Moratalla area are rather incompletely exposed, the results obtained are a good base for further investigations.

UPPER PART OF THE BENAMOR GERÖLLTON MEMBER

Bejar Beds

Derivatio nominis – All those beds and bedded tongues which are intercalated between the thick conglomeratic marlstone units of the B.G.M. and even constitute the final deposits of this Member, and which consist of coarse to fine-grained polymictic limestone conglomerates (composed of exotics) with some minor interlayers of conglomeratic marlstone, and of detrital calcarenites and sandstones (both of similar exotic composition), are called Bejar Beds. In fact also the basal and topmost conglomeratic parts of the mud flow deposits described in the preceding pages are of exactly the same lithology and are therefore called Bejar conglomerates. The Bejar intercalations predominantly occur in the upper part of the B.G.M., which is almost exclusively exposed in the Campo de Benamor and Campo de Bejar (from which the Beds derive their name). As the Bejar Beds are closely related to the conglomeratic marlstones of the B.G.M. with respect to their composition as well as to their mechanism of deposition they cannot be considered as a separate member.

Lithology of the Bejar conglomerates – Campo de Benamor – The upper tributaries of the Río Benamor, viz. the Barranco de Escaramujo, the Barranco del Sas, the Barranco de los Álamos, and the Barranco del Morrón (encl. 11), form cascades over a thick (up to 40 m), subhorizontal sequence of bedded, coarse, polymictic conglomerates with some interlayers of conglomeratic marlstones. These units of Bejar conglomerates are interfingering with the upper part of the B.G.M. The conglomerates are predominantly composed of limestone clasts of pebble, cobble, and boulder size; they are rounded or facetted with rounded-off ribs. The interstices are filled with small limestone clasts, bioclasts, siliciclasts of granule and sand size, and mud (figs. 43 and 44). The mud content generally amounts to a few percent, but locally about 50% of the conglomerates consists of mud (fig. 45). In the Barranco de Escaramujo a point counting showed that the modal diameter of the constituents is somewhat smaller than 1 cm (fig. 43). Only in the interstices between the smallest clasts approximately 5% mud was found (100 points were counted; a point distance of 16 cm was used).

Thin sections from the fine-grained material between the pebbles and cobbles of the conglomerate exposures along the Barranco del Sas show that most of the smaller limestone clasts consist of bioclastic lime-wackestone with *Pithonella* and *Stomiosphaera* (Represa Formation). Other ones are constituted of Moro Marlstone, Rasa Sandstone, Puerta Limestone, Moratalla Marlstone, and a few dolostones of doubtful affinity. Among the bioclasts, which are chiefly fragmental, were recognized: *Amphistegina* and *Heterostegina* (both kinds generally with chalcedony-filled body chambers), well preserved rotalids, echinoid spines and plates, bryozoans, smaller benthonic foraminifera, molluscs, and calcareous algae. Reworked *Discocyclina* and orbitolinids were found. Only once a thin valve of an oyster was found. About 10% of the interstitial material consists of angular siliciclastic detritus (40-700 μ), predominantly quartz, quartzite, felspar, and glauconite.

The matrix of the muddy conglomerates of the exposure farthest upstream in the Barranco de los Álamos revealed: sponge-spicules, *Globigerinoides trilobus, Globorotalia mayeri, Globigerina bulloides,* and reworked *Hedbergella trocoidea* (Lower Cretaceous) and *Pseudohastigerina micra* (Eocene).

A. The sampled clasts predominantly consist of Cretaceous rocks as developed in the Subbetic and South-Prebetic zones:

1. Argos Formation - Canara Formation: grey, calcareous, parallel-laminated sandstones with brown colour on wheathered surfaces. On the bedding planes the mica flakes are conspicuous. The sparitic cement consists of calcite crystals that surround poecilitically the angular grains of quartz, quartzite, chert, plagioclase, other felspars, brown and colourless mica. In the fine-grained samples the matrix may be micritic. Many samples of these calcareous sandstones contain some ooliths and orbitolinids. In fact all types were found between those containing only a few ooliths to oolitic lime-grainstones with only a small percentage of siliciclastic detritus. Colourless mica, however, is always present. In a few samples the matrix is entirely replaced by haematite. These are the only vivid-red clasts found in the Bejar conglomerates. Components of Argos and Canara Formations were observed only in the upper zones of the lower part and in the upper part of the B.G.M.
2. Represa Formation: grey bioclastic lime-wackestones or mudstones with *Pithonella* and *Stomiosphaera,* in some samples also with some globigerinids and (or) single-keeled globotruncanids. One *Hemiptychoceras* (Albian ammonite) was collected.
3. Quipar Formation: white, soft, fissiled bioclastic lime-wackestones or packstones with *Heterohelix* and double-keeled globotruncanids, in some samples also with single-keeled globotruncanids and many small bivalves. One sample contained a colony of *Durania* (rudist, Det. MacGillavry). Also a few black chert nodules were collected.

B. In subordinate percentages clasts of Tertiary rocks were sampled:

4. Moro Marlstone Formation: grey marlstones containing *Globorotalia velascoensis, Gl.* ex gr. *Gl. formosa, Gl. occlusa, Gl. mckannai, Globigerina linaperta;* this association is indicative of the *pseudo-menardii*-zone of Late Paleocene age. Also a sample of hard, fine bioclastic lime-packstone with 5% angular non-calcareous detritus and abundant tiny fragments of

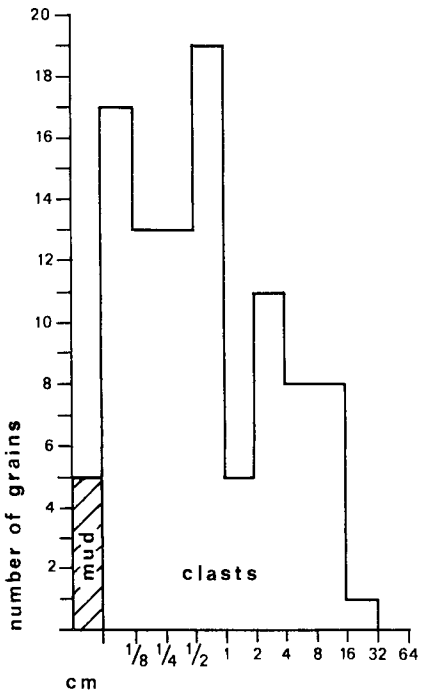
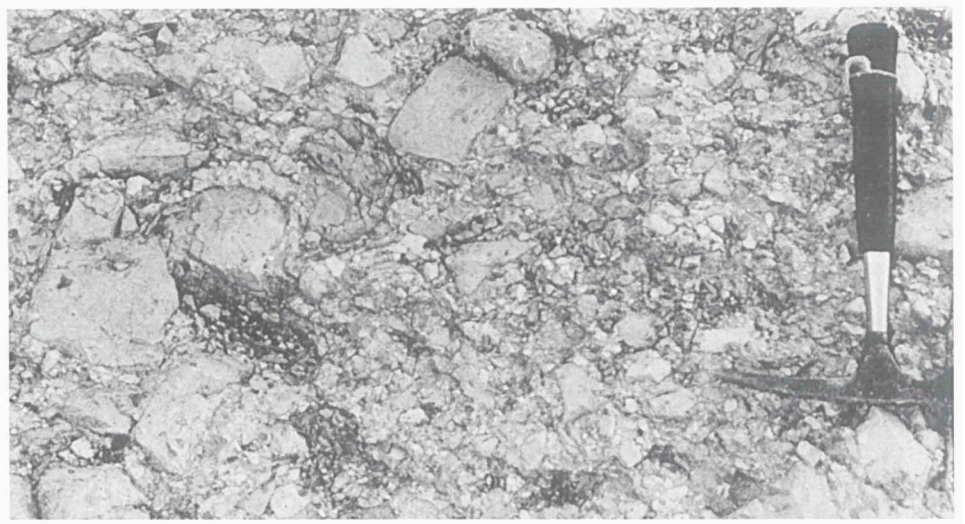


Fig. 43. Frequency distribution of the size classes in the Bejar conglomerates along the Barranco de Escaramujo.

Fig. 44. Close-up of the Bejar conglomerate in the eastern bank of the Barranco de Escaramujo. Note that the interstices between the larger clasts are filled with smaller ones, so that the mud content is small.



foraminiferal tests. *Nummulites, Operculina,* globigerinids, globorotalids, smaller benthonic foraminifera, and fragments of bryozoans and bivalves were recognized.
5. Rasa Sandstone Formation: quartzite pebbles.
6. Minaño Limestone Formation: coarse bioclastic lime-packstones which may be recrystallized into pseudo lime-grainstones, in some samples with authigenic quartz. Recognized fossils: *Discocyclina, Asterocyclina, Operculina, Nummulites, Amphistegina,* alveolinids, rotalids, globigerinids, smaller benthonic foraminifera, and fragments of calcareous algae, echinoids, and bryozoans. Siliciclastic detritus is virtually absent.
7. Puerta Limestone Member: bioclastic lime-packstones with various percentages of siliciclastic detritus of quartz, quartzite, felspar, glauconite, and colourless mica. Recognized fossils (all fragmental): *Heterostegina, Amphistegina,* globigerinids, rotalids, *Lithothamnium* (also in oncoliths), smaller benthonic foraminifera, echinoids, molluscs, and bryozoans. In these bioclastic lime-packstones grains have been reworked that consist of bioclastic lime-wackestones with *Pithonella* and *Stomiosphaera* (Represa Formation), and fragments of *Discocyclina.*
8. Puerta Limestone Member or limestone interbeds of the Moratalla Marlstone Member: globigerinal lime-packstone with a variable percentage of siliciclastic detritus.
9. Moratalla Marlstone Member: bioclastic lime-wackestone with a small percentage (up to 10%) of silt-sized siliciclastic detritus of quartz, quartzite, colourless mica, brown mica, and glauconite. The fossil association: many globigerinids, sponge-spicules, radiolarians, and sparse globorotalids.

The inclusions in the conglomeratic marlstones are of the same rock types as the components of the conglomerates between which they are intercalated. Along the Barranco del Sas, however, a thin part of conglomerate was found curling up into the overlying conglomeratic marlstone (slump overfold, fig. 46) and ending in a curved string of Bejar inclusions. Apparently the Bejar bed was on its way to be consumed by the overflowing rock mixture when the latter came to a stand and stiffened.
10. Along the Barranco de Escaramujo an isolated inclusion of a somewhat finer grained variety of the Bejar conglomerate was found in a conglomeratic marlstone intercalation (fig. 47).

C. The affinity of the following clasts sampled in the Bejar conglomerates is unknown or doubtful:
11. Only in a few cases dolostone fragments of unknown derivation were sampled. Dolostones occur in the Gavilán Formation (Subbetic zone) and in the Ubacas Formation (Prebetic zone). Derivation from the Gavilán Formation seems most plausible.
12. Two of the collected samples consist of echinodermal lime-grainstone with some siliciclastic detritus. The echinodermal fragments have a coating of micrite. Other bioclasts recognized are: fragments of calcareous algae, bryozoans, molluscs, and of a punctate brachiopod. Derivation from the uppermost part of the Gavilán Formation seems most likely.

Lithology of the Bejar conglomerates – Campo de Bejar – In addition to the exotic clasts already described from the Campo de Benamor the following clasts were found only in the Campo de Bejar and in the westernmost outcrop along the Barranco de las Álamos (Campo de Benamor), thus in the southwestern corner of the investigated area.

13. Gredero Formation: coarse bioclastic lime-packstone characterized by the presence of *Lepidocyclina, Nephrolepidina,* and *Eulepidina.* Also *Heterostegina, Amphistegina, Lithothamnium,* other calcareous algae, globigerinids, globorotalids, rotalids, miliolids, other smaller benthonic foraminifera, and fragments of echinoids, bryozoans, and bivalves were recognized. This is the youngest Subbetic component found in the B.G.M.
14. Miravetes Formation: grey micritic limestones with the ammonite *Crioceratites (Crioceratites) sarasini sarasini* (Sarkar) indicating a latest Hauterivian age. This is the oldest Subbetic component found in the B.G.M.

Of doubtful affinity are the following clasts:
15. In the Campo de Bejar and in the Barranco de la Celada dark-green shale chips are very

Fig. 45. Close-up of the Bejar conglomerates in the most upstream exposure in the Barranco de los Álamos. The mud content of this conglomerate is large.

Fig. 46. Bejar bed that was on its way to be consumed by the rock mixture flowing over it when the latter came to a stand and stiffened. Benamor Geröllton Member in the southern bank of the Barranco del Sas.

conspicuous components of the conglomerates. They might be derived from the shale interbeds in the Rasa Sandstone Formation.
16. In the Barranco de los Álamos three limestone clasts were sampled composed of micritic balls and eggs and other rounded micritic grains of approximately 100 μ diameter, which are separated from each other by sparry cement. The lithologic structure is an alternation of lamellae with more micritic balls, and lamellae with more sparry cement. A few smaller benthonic foraminifera were found. In one sample also dolomite rhomboedra were noted in the micritic balls, in the sparry cement, and on the boundaries between the two. The Gavilán Formation (Subbetic zone) and the Fresne Limestone Member (Prebetic zone) are the only rock units of which the lithology remotely resembles that of these clasts.

Lithology of the Bejar conglomerates – Barranco de Hollarionda – In the Bejar conglomerates exposed along the upper course of the Barranco de Hollarionda, in the southeastern guadrangle of the investigated area, clasts composed of red and white bioclastic lime-wackestones and packstones of the Quipar Formation are present in abundance. Far less common are the components of Moro Marlstone, Rasa Sandstone, and Miñano Limestone. A wash-residue of the marlstone matrix contained many reworked Cretaceous planktonic foraminifera, but also reworked *Globorotalia spinulosa* of Eocene, and *Globigerinoides trilobus* of Miocene age.

From the conspicuous difference in composition of the thick Bejar conglomerates in the westernmost, easternmost and middle part of the total area of deposition, it may be concluded that they originated from several different parts of the source area.

Reverse grading in the Bejar conglomerates – The bedded nature of these limestone conglomerate deposits is shown by the alternation of layers with large clasts and layers with smaller ones and by the layers of conglomeratic marlstone between them. Often an unmistakable reverse grading was detected (fig. 48, 49, 50). In these cases the layer with fine clasts and the layer with coarse clasts on top of it form one bed. The beds vary in thickness between 0.5 and 2 m. In some of them the grading may pass laterally into non-grading, in which the various sizes are randomly distributed. The reverse grading is particularly well developed in the thinner beds (fig. 49), whereas the thicker beds are predominantly non-graded.

The type of reverse grading is similar to that already described from the basal parts of the differentiated mud flow deposits in the lower part of the B.G.M. It is an upward gradation from moderate to extreme bimodality, concomitant with a divergence of the constituent modes. At the base of the beds in which the reverse grading is well developed only smaller clasts (up to pebble size) were found almost without interstitial mud. Towards the top the size of the clasts progressively increases to boulder size (the greatest boulder found has an apparent longest axis of 60 cm), while the percentage of mud in the wider interstices increases conspicuously. Erosion selectively attacked these conglomerates; from base to top the surface of exposure of a single bed is progressively more protruding; at the contact of two beds a narrow platform is carved out (fig. 48). In some beds the limestone conglomerate grades upward into conglomeratic marlstone in which large clasts are scattered in a muddy matrix. The conglomeratic marlstone layers may be very thin (even 10 to 20 cm). In case a conglomerate layer is non-graded it is difficult to decide whether the conglomeratic marlstone on top of it belongs to the same depositional unit or to a separate one.

In many conglomerate beds the reverse grading is inconspicuous. This may

Fig. 47. Inclusion of Bejar conglomerate consumed by the conglomeratic marlstone. Western bank of the Barranco de Escaramujo.

Fig. 48. Reverse grading of the Bejar conglomerates in the eastern bank of the Barranco del Morrón. Note the selective erosion according to the mode of the larger fraction, and note the scarcity of smaller pebbles below the lower boundary in contrast with the abundance of them directly above it.

be the case if the beds are too thin with respect to the diameter of the largest clasts, so that also at the base large clasts can be found (in some beds the diameter of the largest clasts even equals the thickness of the beds), and (or) if the upper part with coarse clasts is sharply bounded from the lower part with fine clasts, and (or) if the small basal clasts of the overlying bed have penetrated into the interstices between the large clasts at the top of the underlying bed. This may give the impression of an ostensible normal grading. A good aid in recognizing the reverse nature of the grading in these cases is the modal diameter of the clasts directly below the division with the coarsest clasts, for it is generally coarser than that of the clasts above that coarsest division. Moreover, in most beds a sharp and distinct separation exists between the coarse upper part of the underlying bed and the fine lower part of the overlying bed.

A fine-grained variety of the limestone conglomerates can be found in the upper part of the left bank near the mouth of the Barranco del Sas. The clasts are of the same rock types as those in the coarse conglomerates. Their dimensions vary from sand size to 15 cm. Virtually no mud was observed in the interstices. The beds are generally 20 to 30 cm thick, though varying from 7 to 100 cm. Most of these beds also show a distinct reverse grading, though many thicker interbeds are non-graded. The few topmost centimetres of several of these reversely graded beds are found to be grading normally into a very fine grain-size.

Amidst these bedded fine-grained conglomerates a very large isolated block occurs. The rectangular cross-section is 1.5 by 9 m, the longest axis is parallel to the stratification in the conglomerates. The conglomerate beds abut against the vertical sides of the block. The conglomerate bed directly below the block does not show any sign of deformation. The only way in which this block could have reached its present position is by slivering down over a surface of rolling grains. This block is a typical example of a 'klippe sédimentaire' (or olisthothrymma).

Mechanism of deposition of the Bejar conglomerates – Some facts relevant to the explanation of the mechanism of deposition are summarized first:

1. The constituent clasts are exotic and can only be derived from the South-Prebetic and (or) Subbetic zone immediately to the south.
2. The medium to thick-bedded conglomerates contain many rounded cobbles and boulders that are too large to have been transported by traction of flowing water, though not too large to have been transported by flowing mud or by avalanching. Also the drag phenomena locally seen near the lower boundaries of the conglomerate units suggest ultimate emplacement by delapsion.
3. Many medium to thick-bedded conglomeratic marlstone beds are intercalated between the conglomerates.
4. Many conglomerate beds are reversely graded in the described peculiar way; some of the finer conglomerate beds show a fine-grained normally graded topmost part.
5. The conglomerates are badly sorted; the size of the constituents ranges from clay to boulder; the mud content may amount to 50%.

In some beds the limestone conglomerate grades upward into conglomeratic marlstone. These deposits can be designated as differentiated mud flow deposits. Their depositional mechanism is supposed to be the same as that already described on page 136.

The depositional mechanism of the muddy conglomerate beds that do not have a top-part of conglomeratic marlstone is interpreted to be intermediate

Fig. 49. Reverse grading of the Bejar conglomerates in the northern bank of the Barranco del Sas. Note the normal grading in the matrix and the difference in size between the grains in the basal parts of the different beds.

Fig. 50. Reverse grading of the Bejar conglomerates in the eastern bank of the Barranco del Morrón. Note the relatively better sorted, fine-grained basal part and the badly sorted upper part which contains clasts of boulder size; note the sharp lower boundary.

between those of mud flows and gravel avalanches without mud. The dynamical conditions in the moving coarse-grained system did not pass the critical threshold beyond which the inter-particle bonds of the mud fraction break, as a result of which the mud would lose its cohesion and would go into suspension. But there is no mud enough to build a top layer of conglomeratic marlstone. The cohesive grains are always in contact with each other. This condition is required for frequent collisions to take place, which are assumed to have caused the reverse grading. Owing to frequent collisions the larger grains work their way up to the top leaving behind the smaller grains. But the presence of mud will set a lower dimensional limit to the grain sizes contributing to the formation of reverse grading. The grains of dimensions below this limit will be unable to move with respect to the surrounding mud and will be transported passively in the mud. This mixture serves as a medium filling the interstices between the grains that are larger than the limit size. Occasionally the mud content may be just large enough to create a thin top layer of conglomeratic marlstone.

The depositional mechanism of the fine-grained reversely graded conglomerates with normally graded top is interpreted as the result of avalanching of muddy gravels during which the interstitial mud did go into suspension together with the finest non-cohesive components. The latter subsequently settled down in normally graded fashion. The dynamical conditions in the moving fine-grained system are such that the critical threshold can be reached much earlier. The very thinness of the normally graded top layer is an indication of the slight suspensive capacity and consequently of the low density of the mud suspension above the gravel avalanche. Presumably the amount of mud was only small. If the amount of mud had been greater the normally graded part would probably have been considerably thicker.

The deposition of the conglomerates in which the amount of mud is only a few percent and which have not a normally graded top part can be explained by avalanches of gravel without interstitial mud.

It is important to note that all transitions exist from olisthostromes via mud flow deposits and muddy gravel-avalanche deposits to gravel-avalanche deposits without interstitial mud.

Lithology of the detrital calcarenites – The thick deposits of Bejar conglomerates are interfingering with and laterally replaced by bedded limestones composed of sand-sized organogenic and inorganogenic detritus. This interfingering relation is seen by the wedge of detrital calcarenite intercalated between the fine-grained conglomerate beds in the exposure near the mouth of the Barranco del Sas. The wedge thickens from 0 to 60 cm over a distance of 5 m. The interfingering relation can also be observed in the exposures along the Barranco de los Álamos. Here can be seen that the conglomerate beds are tapering off over quite short distances. Also the detrital calcarenite beds have tapering ends.

The calcarenite beds have the same lithology as those separating the conglomeratic marlstone units in the lower part of the B.G.M. (page 130). The few washable beds intercalated in the detrital limestones along the Barranco Seco contained *Orbulina bilobata, Globigerinoides trilobus, Globorotalia mayeri, Globorotaloides* cf. *G. suteri,* and reworked Cretaceous planktonic foraminifera. If the Miocene association is not reworked the Bejar Beds would be slightly older than Tortonian.

The Bejar beds that are exposed along the Barranco Seco in the Campo de Bejar constitute the final deposits of the B.G.M. Here the 5 m thick unit of detrital

Fig. 51. Gully scoured by the Bejar conglomerates and calcarenites into the underlying conglomeratic marlstone unit. Southern bank of the Río Benamor, 25 m upstream from its confluence with the Barranco de Escaramujo.

Fig. 52. Reverse grading of the Bejar conglomerates in the eastern bank of the Barranco Seco.

calcarenites show foreset-bedded intervals. Locally isolated pebbles, cobbles, and boulders of various rock types were found within these calcarenites. This calcarenite unit lies upon a 5 to 10 m thick conglomeratic marlstone unit and is laterally interfingering with and replaced by thin to medium-bedded, polymictic limestone conglomerates, in which also reverse grading was found (fig. 52). Near the Cortijo de la Pava the detrital calcarenite unit is overlain by a foreset-bedded sandstone unit of about 2 m thick, which in turn is covered by the Cretaceous rocks of the Nevazo Unit.

At several places in the Campo de Bejar the foreset-dips were measured: from the sandstone unit near the Cortijo de la Pava: N 60 E, and N 55 E; from the intervals in the detrital calcarenites: N 60 E, N 100 E, N 30 E, N 85 E, N 65 E, N 80 E, N 90 E, N 40 E. The average current direction apparently was from the southwest (N 66 E).

In the Bejar exposures southeast of the Rambla de las Buitreras the lateral substitution of non-bedded conglomeratic marlstone (= an olisthostrome of about 8 m thick) by well-bedded, fine, polymictic conglomerates with detrital calcarenite interbeds was observed. Near the contact some of these conglomerate and calcarenite beds are curved upward and folded, whereas other beds are sharply truncated. At the border, the non-bedded part is constituted of very fine, relatively well-sorted, polymictic conglomerate with mud in the interstices. This body seems to wedge itself into the bedded conglomerates. Towards the olisthostrome rounded cobble- and boulder-sized inclusions (almost no pebble-sized ones) enter into these fine conglomerates, while the mud content gradually increases until the genuine lithologic composition of a conglomeratic marlstone is reached. The interrelation of the various rocks in this exposure suggests that the well-bedded rock-sequence existed before the deposition of the olisthostrome. This would imply that the olisthostrome has flowed along a channel. The discrepancy in grain size and composition between (1) the non-bedded, fine well-sorted polymictic conglomerate along the border of the olisthostrome and (2) the olisthostrome itself may be explained by assuming that these non-bedded conglomerates represent the erosion products of the bedded, polymictic conglomerates which had been detached from the channel wall by the moving olisthostrome.

In the area round the Cortijo de Chillarion it was observed that in the west the horizontal beds of the uppermost Bejar unit (developed as coarse conglomerates) lie directly upon the red sandstones of the Murtas Formation and upon a 2.5 m thick unit of Roble Limestone dipping 5° to the east. Eastward, in the Barranco Seco, the same Bejar unit (but developed as detrital calcarenites) lies directly upon a several metres thick conglomeratic marlstone unit. Hence these uppermost Bejar Beds are in overstepping relation with the Murtas Formation and the Roble Limestones; the contact is interpreted as an angular unconformity.

The horizontal, 6 m thick rock unit that caps the Cerro de Moratalla la Vieja consists of calcarenites and calcirudites. It lies with a slight angular unconformity upon a thick, slightly southward dipping, bedded sequence of soft, white limestones of the Moratalla Marlstone Member. Turbidites occur in the upper part of the latter sequence. Only at the southern tip of the southern spur of this hill directly below the unconformity plane was a wedge of conglomeratic marlstone found which is incorporated in the B.G.M. It contains many inclusions of bioclastic lime-wackestones and mudstones with *Pithonella, Stomiosphaera,* non-keeled and single-keeled globotruncanids. The calcarenites and calcirudites are

Fig. 53. Panoramic view of the Bejar conglomerate unit along the Barranco del Sas. The unit tapers out towards the west, which suggests that the conglomerates probably are fan deposits and not deltaic deposits. Note the gully scoured out in the underlying conglomeratic marlstones. Photo taken in northwestern direction.

included in the Bejar Beds and tentatively correlated with the uppermost Bejar Beds in the Campo de Bejar.

The rock is composed of an association of exotic limestone clasts, bioclasts, siliciclasts, sparite, and micrite locally recrystallized into pseudo-sparite. Among the exotic limestone clasts were recognized: Moratalla Marlstone, Roble Limestone, Quipar Formation, and Represa Formation. In the material between the exotic limestone clasts the following bioclasts were recognized: *Heterostegina, Amphistegina,* globigerinids, rotalids, miliolids, other smaller benthonic foraminifera, and fragments of echinoids, bryozoans, molluscs, *Lithothamnium,* and other calcareous algae. The percentage of siliciclastic detritus varies between 10% and 30%. The angular to rounded siliciclasts are composed of quartz, quartzite, felspar (among which plagioclase), and glauconite. The grains are generally not larger than 1 mm, but may reach a diameter of 2.6 mm.

Depositional history of the Ulea Group

The Middle Miocene transgression is materialized in the Roble Limestone, which is assumed to have been deposited on a shallow marine platform within the photic zone above wave base. Along the southern border of the Moratalla basin the Cuervo Sandstone, which is the lateral equivalent of the Roble Limestone, was formed by reworked material from the Rasa Sandstone. The transgression is succeeded by a rapid deepening of the Moratalla basin, for it is assumed that the Moratalla Marlstone, which contains mainly pelagic fossils, was deposited in a rather deep marine environment below wave base. The reconstructed amplitude of the folds, however, does not permit a depth of more than a few hundreds of metres.

Directly from the beginning of the deposition of the Trieta Formation a reworking of elements from the borders of the basin took place: dolomite rhomboedra and orbitolinids from the north, Cretaceous rock-elements of Subbetic facies (Asares Conglomerate) and planktonic foraminifera from the south.

A sudden increase in the upheaval of the borders induced intraformational slumping within the Moratalla Marlstones. In the north the Tertiary rock sequence slid down from the southeastern limb of the Muela anticline to form gravity collapse structures. This gave rise to the formation of a local angular unconformity in the Moratalla Marlstone sequence, enabled the erosion to attack severely the newly uncovered rocks, and caused the deposition of the denudation products (Turmal Conglomerates) upon the unconformity plane. After the lithification of the Turmal Conglomerate large parts of it slipped deeper into the basin presumably due to further tilting.

Gradually the reworking from the south intensified resulting in turbidity currents (with current directions from the east to southeast) and finally in the olisthostromes and differentiated mud flow deposits of the Benamor Geröllton Member (with flow directions from the south). This points to the existence of a palaeoslope due north and to a gradual intensification of the denudation of the upheaved southern border.

The thick tongues of coarse conglomerate in the upper part of the Benamor Geröllton Member are regarded as submarine fan-deposits (fig. 53). The detrital calcarenites interfingering with these conglomerates are interpreted as the lateral fine-grained equivalents of these fan-conglomerates, because they consist of the same components, and because they contain isolated pebbles, cobbles, and boul-

Fig. 54. Surprisingly undisturbed part of the lower Nevazo slip sheet composed of Subbetic Represa Formation. This part appears to be the upper limb of a large recumbent fold. Southern main tributary of the Barranco de Azafran.

ders. The fans probably were fed from different parts of the source area because of the conspicuous differences in composition.

The presence of mega-cross-stratified intervals in the detrital calcarenites in the Campo de Bejar points to strong marine currents from the southwest. Apparently the top of the B.G.M. was deposited in a rather shallow marine environment.

The turbidites, olisthostromes, differentiated mud flow deposits, and fan-conglomerates probably filled the Moratalla basin almost up to the rim. The sequence of the Trieta Formation may in some way be compared with the classic flysch-wildflysch-molasse succession, though of miniature size. The succession is closed by the Nevazo Unit, which consists of two extensive slip sheets of Cretaceous rocks mainly in Subbetic facies.

Nevazo Unit and Azafran slice

NEVAZO UNIT

Derivatio nominis – The thick and almost undisturbed Middle Miocene deposits

of the investigated (= western) part of the Moratalla basin are covered by a thick blanket of soft Cretaceous rocks in which erosion has carved large openings through which the underlying Miocene rocks are peeping out. Van Veen (1969) has called this rock sheet the Nevazo Unit, because of the good exposures along the Barranco del Nevazo (encl. 4). Owing to their softness the rocks of the Nevazo Unit lend themselves to agricultural purposes. The non-cultivated parts are covered by a thick veneer of coarse rubble shed from the Subbetic Gavilán outliers (pag. 80) situated upon this sheet of Cretaceous rocks. Good exposures are therefore restricted to the steep banks of barrancos. The largest exposure was found in the headwater region of the Barranco de Azafran.

Lithologic succession and structures in the Azafran valley – From Casa de Cristo through the gorge of the southern main tributary of the Azafran in northeastern direction down to where it unites with the northern main tributary to form the main course of the Barranco de Azafran, the following features were observed:

1. Dark-grey marlstones with strong fissility, in many places associated with sandstone beds of the Argos Formation. The non-stratified marlstones are intersected by numerous curved, polished, and striated surfaces which are anastomosing or crossing each other. These slickensides enclose bodies of various sizes and of lenticular or lozenge-shaped cross-sections which also may be curved or have tapering ends as if they had been drawn out. The thickness of this dark-grey, non-stratified marlstone unit is about 200 m.

One ammonite, *Oxytropidoceras* sp., was found in this unit, indicative of a Middle Albian age. Fossil planktonic foraminifera recognized in the marlstones of this unit are: *Globigerinoides benthonensis G. breggiensis, Rotalipora ticinensis, Hedbergella infracretacea, H. amabilis.* This association suggests a Late Albian age.

This non-bedded marlstone unit has a sharp, though undulating lower boundary, which truncates the beds of the underlying, regularly stratified unit. This boundary marks a sharp change in the intensity of deformation.

2. This underlying unit consists of a sequence of alternating grey marlstone beds and grey marly limestone beds (both 10 to 30 cm thick). Close to the boundary most beds are turned over towards the north. The marly limestone beds are here thickened in the zones of maximal curvature and do not show any feature that might be designated as fractures. The curvature of the beds near the undulating boundary is explained by drag exerted by the moving sheet of non-bedded marlstones overlying this well-bedded sequence. As a matter of fact a large part of the latter is astonishingly undisturbed (fig. 54). But when descending further along the barranco, it appears that the thick, slightly inclined, undisturbed sequence forms the upper limb of a large recumbent fold of which the axis strikes N 90 E (encl. 12a en b). Further downstream the front of the recumbent fold displays excellent cascade folding. Here about five disharmonic recumbent folds are found piled upon each other. According to the E-W strike of the fold-axes the direction of sliding would have been from the south. The folds of the cascade are box-shaped, chevron-shaped or isoclinal. Still further downstream the isoclinally folded sequence merges gradually into a chaotic jumble of folded and broken beds (fig. 55). Near the base of the sequence the bedded nature cannot even be discerned anymore and the rock becomes a non-stratified, chaotic accumulation of lozenge-shaped and lenticular bodies, which are commonly bounded by slickensides. Lentils

Fig. 55. Chaotic jumble of disrupted beds in the lower part of the lower Nevazo slip sheet. Southern main tributary of the Barranco de Azafran.

and large irregular masses of harder limestone are scattered in a "matrix" of smaller lentils of softer rock. Occasionally some beds have been preserved which are tightly folded. Finally, large parts can even be designated as monomictic conglomeratic marlstones and as such can be distinguished from the conglomeratic marlstones of the Benamor Geröllton Member (fig. 56, 57). Subangular inclusions of some 5 to 10 cm diameter consisting of grey marly limestone are rather widely dispersed in a grey marlstone binder. This bedded unit consists entirely of Subbetic Represa Formation; no Argos sandstone was found.

In thin sections the marly limestones can be designated as bioclastic lime-wackestone with *Pithonella, Stomiosphaera,* single-keeled globotruncanids, globigerinids, and some echinoid spines. In this unit Upper Vraconian and uppermost Albian ammonites were sampled: *Anisoceras perarmatum* (Pict. & Camp.), *Stoliczkaia dispar* (D'Orb.), *Mariella bergeri* (Brongn.), *Mortoniceras rostratum* (Sow.), *Mortoniceras (Durnovarites) vraconiense* Renz, *Hamites* sp. (Upper Vraconian association) and *Mortoniceras inflatum orbignyi* (Spath) (uppermost Albian).

The association of planktonic foraminifera is also indicative of latest Albian and Vraconian age: *Globigerinoides breggiensis, G. benthonensis, Rotalipora ticinensis, Ticinella primula, T. roberti,* (uppermost Albian association); *Globigerinoides benthonensis, Rotalipora appenninica, Planomalina buxtorfi, Hedbergella delrioensis* (Lower Vraconian association).

The lower boundary of this unit is sharp and undulating. Astonishingly, below this boundary the rocks of the Benamor Geröllton Member appear to be undis-

turbed. In the barranco which runs from the Cortijo de St. Ana towards the southwest about 100 m upstream of the road from Moratalla to Sabinar where the lower boundary is well exposed, only in a thin zone (1 m thick) below the basal contact are the underlying grey homogeneous marlstones slightly deformed and show numerous slickensides. In the barranco which discharges from the south into the Río Benamor, the deformed zone is 2 m thick. Here the deformation does not even alter the shape of the intercalated Bejar Beds, but is confined to the marlstone inclusions which are stretched out into parallel flat-lenticular shapes.

3. Only locally, for instance where the two main tributaries unite to form the main course of the Barranco de Azafran, large lentil-shaped masses of white and red calcilutites of Late Cretaceous age (Quipar and Jorquera Formation) were found under the lower boundary. The rocks in these lentils are generally intricately crumpled.

The Nevazo Unit in other exposures – The 100 m thick lower unit has a limited westward extension. Directly to the east of the line formed by the Barranco del Morrón and the Barranco de Escaramujo it ends abruptly and is replaced by the upper part of the Benamor Geröllton Member, which is predominantly constituted of conglomeratic fan-deposits and in which inclusions larger than 60 cm are virtually absent. East of the Cortijo del Buitre the only moderately exposed unit probably has a normal attitude, because rocks of increasingly younger ages occur at successively higher levels. The sequence starts with heavily deformed rocks of the Represa Formation. At higher levels the rocks of this Formation become less deformed. Also the overlying Quipar Formation appears to be only moderately deformed. The zone of black chert nodules and the brick-red beds (in the Collado del Buitre) were found in normal stratigraphic order. The few intraformational conglomerate beds in the Quipar Formation appear to be normally graded.

The thicker (about 200 m) upper unit, however, can be traced much further westward. The westernmost outcrops of the rocks of this unit were found in the Campo de Bejar. Along the Barranco Seco, about 150 m north of the road from Moratalla to Sabinar, rocks of the Subbetic Miravetes Formation were found forming part of this sequence. In these light-grey, bedded, lutitic limestones, which are strongly folded, were collected: *Barremites difficilis* (D'Orb.), *Partschiceras infundibulum* (D'Orb.), *Partschiceras baborense* (Coq.), *Holcodiscus vandenheckei* (D'Orb.), indicating a Barremian age.

Also the Subbetic dolostone massifs (page 80, 171) of the Serrata, Cerro de Gamonár, Sierra Pelada, and the Sierra del Cerezo appear to be situated, at least partly, upon a sheet of Cretaceous rocks. However, in the Collado del Roble the rocks of this sheet differ markedly from those described from the Azafran valley. In an outcrop in the barranco just north of the Cortijo del Collado the rock consists of grey, slightly nodular marlstone with numerous brownish, 1 to 2 mm thick, sandy lamellae at regular distances of 1 or 2 cm. In another exposure in the same barranco, but northeast of the Cortijo, it consists of a regular alternation of grey shale beds (40 cm thick) with a notable percentage of siliciclastic detritus and light grey, slightly nodular marly limestone beds (20 cm thick).

The following fossil foraminifera were determined by van Helden and Manuputti from washresidues: *Globigerinelloides benthonensis, Biticinella breggiensis, Rotalipora ticinensis, Ticinella primula, Hedbergella amabilis, H. planispira, H. trocoidea, H. washitaensis, Spiro-*

Fig. 56. Transitional stage of disruption between that of Fig. 55 and Fig. 57.

Fig. 57. Monogenic conglomeratic marlstone, which can only be distinguished from the conglomeratic marlstones of the B.G.M. by the absence of exotic inclusions. Basal part of the lower Nevazo sheet as it is exposed in the southern main tributary of the Barranco de Azafran.

plectinata annectens, S. complanata, S. bettenstaedti, Epistomina spinulifera polypoides, Dentalina nana, Saracenaria sp., *Gavellinella* sp., many ostracods. This fossil association is indicative for a Late Albian age.

Rocks of the same age, with the same fossil association and of similar lithology have been found only in the Canara Formation in the Sierra de la Puerta. (Paquet, 1969, p. 82, summital part of rock unit c; and p. 83, rock unit g; own observations). Such rocks have never been found in the Subbetic zone.

General aspects of the unit – Summarizing, the Nevazo Unit appears to be composed of two sheets of strongly deformed, internally imbricated, incompetent Cretaceous rocks overlying each other en echelon, both having an extremely deformed lower part and a less deformed upper part. The lower sheet probably has a normal attitude (with the exception, of course, of the rocks in the overturned limbs of the recumbent folds in this sheet) and shows cascade structures which are apparently due to sliding from the south. The surprising fact that directly below the lower contact of the Nevazo Unit undisturbed horizontal deposits of the Benamor Geröllton Member were found, strongly suggests that it was emplaced by sliding. The presumed normal attitude of the rocks of the Unit would in turn imply that it is composed of at least two slip sheets, a lower one composed of Upper Albian to Upper Cretaceous rocks in a facies similar to that found in the Subbetic zone, and an upper one composed of Barremian to Upper Albian rocks also in Subbetic facies. In the extreme east the Upper Albian rocks of the Unit are in a facies similar to that found in the Sierra de la Puerta of the South-Prebetic zone. The lentils of Upper Cretaceous rocks locally occurring along the basal contact of the Unit should be interpreted as dislodged slices dragged along with the lower slip sheet.

An important aspect of the Nevazo slip sheets (or diverticules) is that some parts of the incompetent bedded sequences were so severely broken up and mixed together that the results can be designated as monogenic conglomeratic marlstones. If the beds of the whole sheet had been disrupted, for instance if the distance of sliding had been far greater and (or) if the rocks had been more incompetent, the result would have been an allolisthostrome (Elter & Raggi, 1965). The Nevazo Unit is therefore an example of one of the transitional olisthon-types between diverticules and olisthostromes. As the Nevazo Unit has the essential character of a deposit, i.e. deposited on a surface of deposition, it must be regarded, like all olisthons, as a rockstratigraphic unit instead of as a tectonic unit, for the latter are separated from each other primarily by faults that truncate the rockstragraphic units. The Nevazo Unit consists of rocks of Cretaceous age that are redeposited by delapsion in the Middle Miocene. The age of this rockstratigraphic unit is therefore Middle Miocene. (For further discussion about the Nevazo Unit see page 190).

AZAFRAN SLICE

In the headwater region of the Barranco de Azafran six huge rock slices occur entirely surrounded by rocks of the Nevazo Unit (encl. 13). The Cerro del Cuervo is built of one of these slices and the Cortijo de los Asares is situated on top of an other one. All these slices are composed of the same Middle Miocene rock units which are from old to young:

1. Cuervo Sandstone

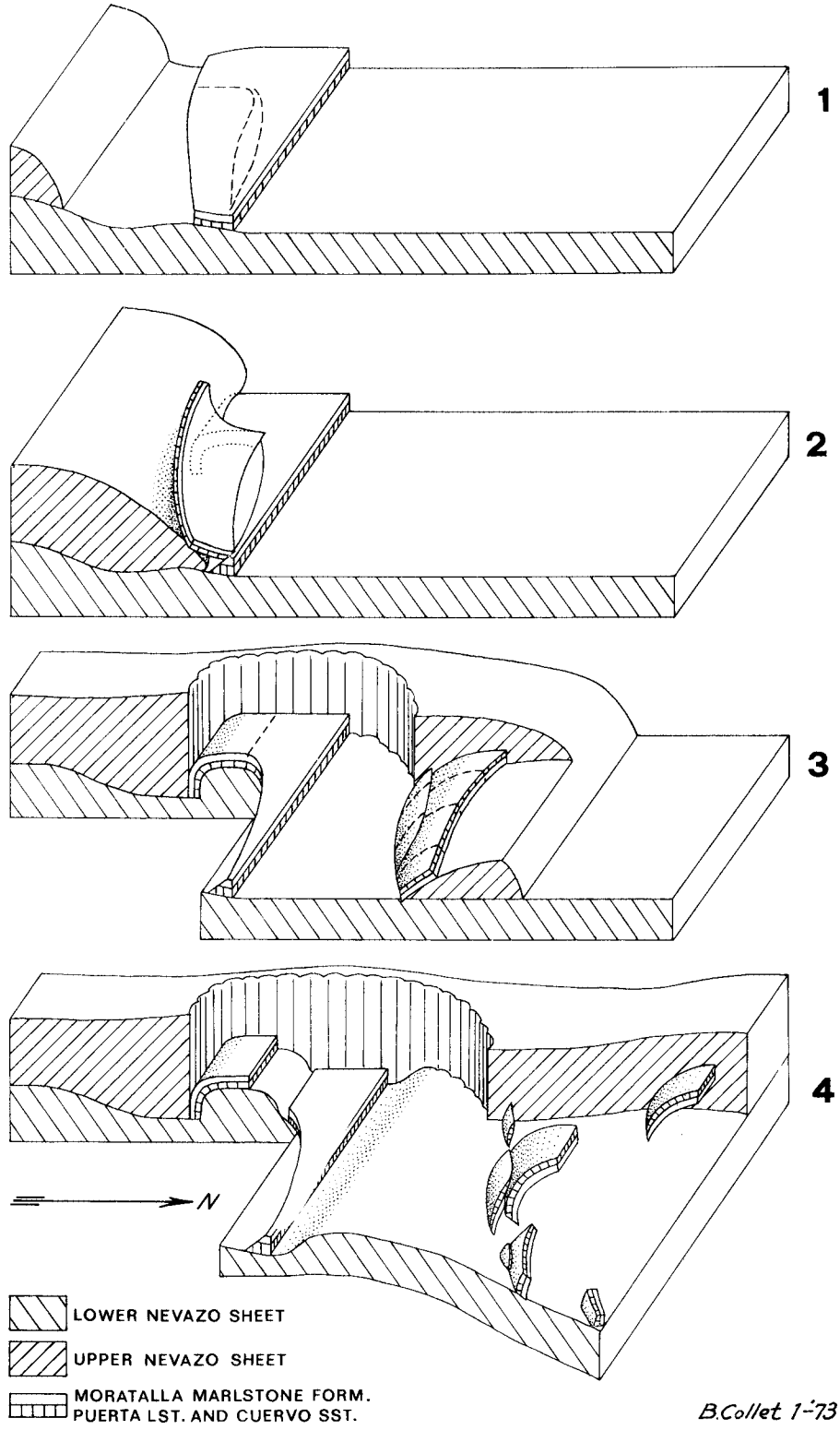


Fig. 58. Diagrammatic representation of the development of the Azafran slice.

2. Asares Conglomerate
3. Puerta Limestone
4. Moratalla Marlstone as it presumably was developed in the source area of the slices.

Two of the six rock slices have a normal attitude and are situated at the boundary separating the lower from the upper Nevazo sheet. The other four have a vertical to overturned attitude and occur amidst the marlstones of the lower part of the upper sheet. The strikes of the latter four are nearly parallel and three of the vertical slices are roughly in a line with each other. The mutual closeness and the distribution of these rock slices suggest that, shortly before reaching their present position, they had probably formed one large slice of normal attitude and about 1.25 km long and some 300 m broad. This slice is named the Azafran slice.

The provenance of this Azafran slice is still obscure. One possibility is that it has been a gigantolistholith of zone 12 of the Benamor Geröllton Member, which has been picked up by the down-sliding lower Nevazo sheet. In fact in the headwater region of the Barranco de Azafran the lower boundary of this sheet reaches the abnormal low level of zone 12. Moreover zone 12 contains several large megolistholiths composed of the same rock units. The other possibility is that it represents a remnant of the original Upper Tertiary cover of the Nevazo Unit itself, which slipped from its Lower Tertiary substratum, but did not reach the Moratalla basin. Subsequently it would have moved together with, and on top of, the sliding lower Nevazo sheet down to its present position. The first interpretation is given on the geologic map, though the second interpretation is equally plausible.

The middle and eastern portion of the Azafran slice may subsequently have been pushed up by the approaching upper Nevazo sheet into an upright attitude and then carried a little further northward. During the last movement the upright slab would have fallen apart into four separate portions lying in a row in upright to overturned position (fig. 58).

Stratigraphic succession in the source area

From the spatial distribution of the various kinds of exotics in the Benamor Geröllton Member (table 3) we can derive some clues about the stratigraphic succession of the source area (encl. 5):

1. We must conclude to the presence of Asares Conglomerate and Cuervo Sandstone exclusively in the source area. The latter directly underlies the Asares Conglomerate and overlies the Rasa Sandstone.
2. The Roble Limestone is replaced to the west by the Cuervo Sandstone.
3. In the South-Prebetic zone the Miñano Limestone is rapidly thinning in western direction and the absence of it in the Benamor Geröllton Member (except sporadically in the easternmost outcrops) suggests that it was absent in the middle and western part of the source area.
4. Large volumes of Lower Cretaceous rocks can be found mainly in the higher olisthostrome units and in the Nevazo Unit, viz. Represa, Argos, and Miravetes Formations. This indicates that only during the emplacement of these units the source area was sufficiently denuded to supply these rocks.

Tabel 3 Distribution of the various inclusions in the B.G.M. and of the rocks in the Nevazo Unit.

	Western portion Campo de Bejar, Campo de Benamor. (Upper B.G.M.)	*Middle portion* Rio Benamor valley, Cerro de las Arcas, Collado de San Jorge. (Lower B.G.M.)	*Eastern portion* Cerro de San Jorge, Cerro Hilario, Cerro de las Piedras Llanas, Cerro de Rótova. (Lower B.G.M.)
Middle Miocene	Moratalla Marlstone	Moratalla Marlstone; and Moratalla Marlstone of northernmost Subbetic derivation in megolistholiths of zones 7, 12, and in Azafran slice.	Moratalla Marlstone
	Puerta Limestone (abundant)	Puerta Limestone (abundant)	Puerta Limestone (abundant)
	? (small samples cannot be distinguished from Puerta Limestone).	Asares Conglomerate (frequent), thick in lower zones, thin in upper zones, with reworked Argos Formation in the Azafran Slice.	Asares Conglomerate (frequent)
	———	Roble Limestone (as thin intercalations in upper part of Cuervo Sandstone) in upper zones (one occasion) and in Azafran slice.	Roble Limestone (thick) only in Cerro de Rótova
	? (small samples indistinguishable from Rasa Sandstone)	Cuervo Sandstone (frequent)	Cuervo Sandstone (less frequent)
Lower Miocene	Upper Gredero Formation (frequent)	———	———
Middle Eocene	———	———	Miñano Limestone (thin, rare)
Middle + Lower Eocene	Lower Gredero Formation	Lower Gredero Formation (one occasion)	Lower Gredero Formation (two occasions)
Lower Eocene	Rasa Sandstone	Rasa Sandstone (abundant)	Rasa Sandstone (abundant)
Paleocene + Lower Eocene	Moro Marlstone (rare)	Moro Marlstone (rare)	Moro Marlstone (rare)
Upper Cretaceous	Jorquera Formation	Jorquera Formation	Jorquera Formation
	Quipar Formation	Quipar Formation (frequent), thick intraformational conglomerates	Quipar Formation (frequent), many Inoceramus
Upper Albian + Vraconian	Represa Formation (frequent)	Represa Formation (frequent), very large olistholiths in the upper zones from zone 7	Represa Formation (frequent), very large olistholiths only in zone 14
Upper Barremian to Upper Albian	Argos Formation (frequent) + Canara Formation	Argos Formation (frequent) + Canara Formation, only in the westernmost part (Benamor valley) and in uppermost zones (zone 10 and higher)	Argos Formation (once)
Upper Hauterivian	Miravetes Formation, only in the westernmost part (Campo de Bejar)	———	———
Lower Nevazo Unit	———	Upper Albian to Upper Cretaceous rocks in Subbetic facies	Upper Albian rocks in South-Prebetic facies (Collado del Roble)
Upper Nevazo Unit	Barremian to Upper Albian rocks in Subbetic facies	Aptian (?) + Upper Albian rocks in Subbetic facies	———

5. The older the rocks, the more west they can be found in large quantities. This indicates that in the western part of the source area the denudation had proceeded far more rapidly than in the eastern part (the Bejar Beds in the eastern part consist almost exclusively of components of the Quipar Formation). This may indicate that the source area had a greater elevation in the west than in the east.
6. The Asares Conglomerate is less thick in the olistholiths of the higher zones than of the lower ones. Presumably the Asares Conglomerate in the olistholiths of zone 12 was deposited nearer to the shore than that in the olistholiths of the lower zones and is, therefore, probably derived from a more southern part of the source area than that occurring in the lower zones.

The source area of the olisthostromes and diverticules should be an uplifted, probably emerged area to the south of the present extension of these olisthons. Most likely the stratigraphic sequence of the source area contained Miravetes and Represa Formations in a facies similar to that found in the Buitre Unit of the Subbetic zone along the Río Argos west of Caravaca. The source area must also have contained rocks of Middle Albian age (with *Oxytropidoceras)* which are absent in the rock sequence along the Río Argos. The facies of the Argos Formation obviously had Subbetic (micaceous sandstones) as well as South-Prebetic affinities (oolitic limestones). The facies of the Quipar and Jorquera Formations were partly similar to that of the Sierra de la Puerta (thick intraformational conglomerates and many *Inoceramus);* the intraformational conglomerates in the Nevazo Unit near the Cortijo del Buitre are reduced to a few thin intervals in the white soft limestones of the Quipar Formation. Hence the transition of the Quipar Formation in South-Prebetic facies to Subbetic facies probably was situated within the source area. The Tertiary rock units are predominantly developed in a facies equal to that of the Sierra de la Puerta (Moro Marlstone Formation, Rasa Sandstone Formation, Miñano Limestone Formation, and Puerta Limestone Member). The remaining part of the Tertiary rocks, however, is in Subbetic facies (Gredero Formation). The transition from Tertiary rocks in South-Prebetic facies into the Tertiary Gredero Formation of the Subbetic zone evidently was situated in the source area.

The fact that virtually all the larger olistholiths of the B.G.M. have a normal attitude indicates that the rock sequence of the source area would also have had a normal attitude. The presence of rocks of South-Prebetic facies as well as of the facies developed in the Buitre Unit implies that the source area was situated to the north of the Subbetic rock sequence that is exposed along the Río Argos.

Recent and subrecent deposits

The front of the Subbetic thrust sheet and almost all the Subbetic delapsional outlier (pages 80, 171, 174, 182), which are composed of dolostones of the Gavilán Formation, are situated upon a very thick sequence of highly incompetent rocks which are subject to rapid denudation. The borders of the front and outliers are repeatedly undercut and when losing their support they gave rise to huge landslides of the slump-type. Large rock masses have slumped, leaving behind spoon-shaped scars on the parent rock body. Most of the slumped rock masses

were broken into pieces. The bedding of these rock masses is dipping towards the parent rock body.

The slump blocks may also be designated as toreva blocks. Numerous examples can be found along the northwestern slope of the Sierra del Buitre, the northeastern slope of the Sierra del Enmedio (for instance the Peñon de la Molata), and the southern slopes of the Sierra Pelada and the Sierra del Cerezo (for instance the Copa del Cerro). The Peñon del Morrón is a good example of a toreva block slumped from the southern slope of the Sierra de los Álamos. Most of the depressions between the parent rock body and the toreva blocks are filled with rock-waste. In fact, the huge rock mass that forms the highest summits of the Sierra de los Álamos is subject to incipient slumping, since the spoon-shaped fractures have already come into existence, along which they have already slumped a few metres. The large toreva blocks along the northern slope of the Sierra de los Álamos (for instance the Beñales) have been wrongly interpreted by Fallot (1945) as remnants of a frontal wedge of the Subbetic overthrust nappe. Locally the Jurassic rocks of the Tollo Formation slumped down from the front of the Buitre thrust sheet. Along the Barranco de las Tablas these slump blocks are crumpled into several tight folds. The slip horizon probably is the basal, serrately fissiled lime-mudstone of the Tollo Formation.

The dolostone massifs of the Subbetic Gavilán Formation became subject to intensive erosion. This results in talus slopes (scree), especially along the northern slopes, and in land-slides of the earth-flow type (for instance north of the Cerezo and west of the Cerro de las Higueras). The dolostone massifs are therefore surrounded by a broad fringe of rock debris, which covers the slopes with a thick veneer of coarse rubble entirely composed of dolostone fragments of the Gavilán Formation.

Isolated relicts of this cover of rock debris can be found capping the hills preserved between the barrancos, for instance on top of the Cerro de San Jorge, the Cerro de las Arcas, the two hill-tops between the Cerro de las Arcas and the Barranco de Azafran, and on top of de Cerro de las Almanillas. The bottom of the valleys bordering these hills are locally more than 200 m below the level of the sheet of rubble. As it is likely that the fringe of rock debris was originally a continuous one, it can be inferred that the valleys of the Río Benamor, the Barranco de Azafran and its confluents, the Barranco del Commendador and its tributaries, the Río Alárabe, and the Barranco del Grero were deepened some 200 m after the rubble was deposited. This rubble and the associated toreva blocks may therefore have been deposited in the Pleistocene.

One debris avalanche was found (southeast of the Cerro de las Piedras Llanas) of which the dolostone components are cemented together into a very hard mass. The shape of this cemented debris avalanche, however (narrow proximal part and broad, flat distal part with tongue-shaped endings), suggests that it flowed down as a watery mass. The extensive rubble-fan at the lower end of the valley of the Barranco de los Barranquillos is supposed to have been deposited by a subaerial mud flowing.

In various places the rivers and barrancos have formed terraces, for instance, along the Río Alárabe near the Cortijo de Somogil and at its confluence with the Barranco del Grero, at various localities along the Arroyo de las Murtas, and along the Río Benamor southwest of Moratalla (the latter occurrence is not marked upon the geological map).

Tectonics

GENERAL

Summarizing, the tectonic structure of the investigated area can be described as follows (encl. 3 and 4): the deep, basin-shaped syncline of Moratalla is flanked in the northwest by the anticline of the Sierra de la Muela, which can be described as a box-fold striking northeast-southwest. The Tertiary rocks of the southeastern limb of this box-fold have slipped down from their Mesozoic substratum giving rise to gravity collapse structures. The rocks of both the Muela anticline and Moratalla syncline are autochthonous and belong to the Prebetic zone.

In the south the Moratalla syncline is bounded by a zone of complicated structures of which the also northeast-southwest striking Buitre Unit is the most prominent one. The latter can be designated as a thrust sheet situated on top of the rocks of the Moratalla syncline. This Buitre Unit belongs to the Subbetic zone. Large outliers of this Subbetic Unit are situated on top of the Prebetic rocks lying in a curved row against the northwestern limb of the Moratalla syncline. These are the Nogueras, the Álamos, and the Cerezo-outliers.

Between the Prębetic rocks of the Moratalla syncline and the Subbetic rocks of the Buitre Unit a rock complex can be found, which is separated from both the Prebetic and the Subbetic by structural contacts. This complex is named the Puerta-Nevazo complex, and is grouped into the South-Prebetic zone.

TECTONIC STRUCTURE OF THE PREBETIC ZONE

Muela anticline

The tectonic structure of the Sierra de la Muela can be described as a box-fold with steep limbs and a broad, subhorizontal crest (encl. 3). The northwestern limb is on the average somewhat steeper than the southeastern one. This box-fold can be traced over a distance of some 18 km. The fold-axis plunges towards the northeast as well as the southwest. The limbs are steepest (90°) at both sides of that part in which the axis is culminating, this is between the Muela de Moratalla and the Bolos. They become less steep in those parts where the axis is plunging down.

In the flat crestal part at least five longitudinal crestal faults were observed. Along these faults the longitudinal middle strip of the crest has subsided a little with respect to the side strips of the crestal part. Moreover a few cross-faults were found. By this more or less perpendicular fault-system the broad crestal part of the box-fold has been divided into blocks which have subsided or were uplifted with respect to each other. An example of a relatively deeply subsided block can be found 1.5 km ENE of the Cortijo de la Muela. In the southeastern limb of the Muela anticline west of the Cerro de las Almanillas at least five east-west directed oblique peri-anticlinal faults were noted, whose northern sides are upthrown.

A reconnaissance trip around the Sierra de la Muela revealed that in the north, northwest, and south the box-fold is bounded by faults. The block of the Sierra de la Muela enclosed by these faults has been uplifted with respect to its

surroundings. The form and fault-pattern of the box-fold suggest that next to lateral compression also a diapiric action of deeply buried rocks (probably gypsiferous Triassic rocks) has played a part in shaping the Muela anticline. The fact that the Muela anticline originated at the place where, inferred from the considerable difference in thickness of the stratigraphic rock-units, an important hinge existed in the basement, suggests that this hinge favoured the accumulation of the Triassic rocks.

The larger part of the Tertiary rocks of the southern limb of the Muela anticline has slipped down towards the southeast. This can be inferred from the structures displayed by the 150 m thick Almanillas Limestone. These structures resemble the gravity collapse structures of Falcon & Harrison (1936). The following structures were found:

1. In the valley of the Arroyo de las Murtas about 500 m SW of the Cortijo de los Campaneros these Limestones display a slip-sheet structure. Steeply overturned NW-dipping Almanillas Limestone is situated with a structural contact upon steeply SE-dipping Grero Marlstone. The net slip along the slightly SE-dipping slip plane is about 50 m towards the southeast.
2. In the Cerro de las Almanillas the Almanillas Limestone has a roof-and-wall structure of which the wall-part is slightly overturned towards the southeast. The subhorizontal roof-part, which has a very shallow synclinal structure, is however faulted by steeply SE-dipping reverse faults. Hence the imbricated blocks slipped underneath each other from the north.
3. South of the Río Alárabe the roof-and-wall structure reduces to a knee-fold (fig. 59).
4. The Almanillas Limestone in the area between the Cerro de las Almanillas and the valley of the Arroyo de las Murtas shows a collapse structure. The junction between roof and wall has broken down and the roof has caved in according to a parallelogram-shaped fault-pattern. The wall even stands out above the collapsed roof.

It is thought that the gravity collapse structures came into existence during the formation of the Muela anticline. The Murtas Formation, Roble Limestone Formation, and even the lower part of the Moratalla Marlstone Member have taken part in the gravity collapse structures. The Turmal Conglomerate, however, truncates the structures, so that the formation of the collapse structures probably has taken place at the end of the time-interval represented by the *Globorotalia fohsi peripheroronda* biozone (Middle Miocene).

Moratalla syncline

The structure of the Moratalla syncline is broad and basin-shaped. It is filled with Moratalla Marlstone and Benamor Geröllton. The stratification is horizontal and subhorizontal over large distances. Numerous small faults intersect the horizontal strata of the Moratalla Marlstone; these fault planes are invariably lined with calcite. The displacement along these faults is generally less than 1 m. Only locally the throw amounts to several metres. In the southeastern part of the basin a broad, feebly developed fold striking SW-NE was found, of which the crest rises about 300 m above the flanks and the southern limb plunges below the overturned rock sequence of the Sierra de la Puerta. Since in the eastern part of the Sierra de la Puerta the overturned succession ends with a 300 m thick sequence of the same Moratalla Marlstone, we must conclude to a tight syncline with an

overturned southeastern limb marking the southern border of the Moratalla basin.

TECTONIC STRUCTURE OF THE SUBBETIC ZONE

The Triassic rocks: arguments favouring a parautochthonous position

The current opinion is that the Mesozoic and Tertiary rocks of the Subbetic zone are allochthonous and have moved a considerable distance over the equivalent rocks of the autochthonous Prebetic zone (Fallot, 1945; Durand Delga, 1966; Paquet, 1969). Paquet (1969, fig. 106) draws diagrammatic cross-sections in which Mesozoic rocks of the Prebetic are underlying the northern part of his Ponce Unit. This would imply a horizontal displacement of the Subbetic over the Prebetic at the NW-SE line Cehegin-Totana of some 20 km at least.

In the rocks of the Subbetic zone horizontal deformative stresses could have been transmitted only through the thick competent Jurassic carbonate rock sequence (Gavilán and Tollo Formations). This is bounded at its base by incompetent, marly, gypsiferous Triassic rocks (Peñuelas Formation) and at the upper side by incompetent, marly Cretaceo-Tertiary rocks (Miravetes, Argos Represa, Quipar, Jorquera, and Gredero Formations). These latter would have adapted in a passive way to the structures of the Jurassic rocks. When stressed, such a constellation fulfills all requirements to become deformed in a similar way as the Jura mountain system in Switzerland and the High Atlas Mountains in North Africa, that is concentric epidermal folding by way of box-folds, which in turn may develop either into diapiric folds (de Sitter, 1965) or, when asymmetric, into fold overthrusts. Thus the relatively undisturbed condition of the thick sequence of marly Cretaceous rocks along the Río Argos (west of Caravaca), and of that of the Loma de Solana (south of Caravaca) can be explained only by assuming that they are directly underlain by a rigid substratum of Jurassic carbonate rocks. Also the vertical to overturned attitude of the Cretaceo-Tertiary rocks of the Sierra de la Puerta can only be explained by assuming that they have a rigid backing of Jurassic carbonate rocks.

According to the current opinion the major disharmony, a subhorizontal thrust plane, is thought to be situated below the incompetent Triassic rocks of the Subbetic zone. Hence these, too, would have been dragged along over some 20 km towards the north in squeezed position between a stable substratum of Mesozoic and Tertiary rocks of the Prebetic zone and a rock-sequence of about 3 km thick on top of them. However, the large masses of Triassic rocks in the area between Caravaca, Bullas, and Calasparra (encl. 4), which are thought to have been thrust over the Prebetic (Paquet, 1969, fig. 76, pag. 178), have retained a remarkably well preserved and locally complete stratigraphic order (Paquet, 1969, pag. 28, and own observations) and are astonishingly little disturbed over large distances. This cannot be reconciled with the idea that these rather incompetent Triassic rock masses are the basal part of a 3 km thick thrust sheet (Paquet, 1969, pag. 29 and 30, fig. 10 and 11). Only along the borders and in some restricted areas the deformation is severe. Moreover all observed structural contacts between the Triassic rocks of this extensive outcrop and the rocks of the Prebetic and of the northern part of the Rameles Unit are steep (about 45°) or even vertical (Paquet, 1969, fig. 11, 40, 69, 70, 72, 75, 93; van Veen, 1969; own observations).

These observations and the extraordinary massiveness of the accumulations

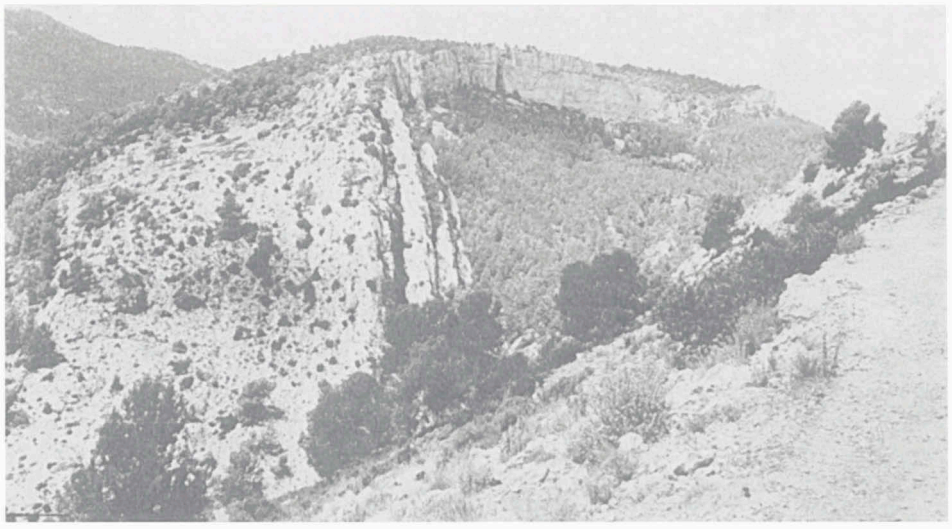


Fig. 59. Kneefold of the Almanillas Limestone Formation south of the Río Alárabe. Photo taken towards the southwest.

Fig. 60. Plate of lower Tertiary rocks lying in inverted attitude upon subhorizontal Moratalla Marlstone of normal attitude as exposed along the service road of the Canal de Taibilla. The hill in the background is the Cañada Rasa.

of these Triassic rocks are incompatible with the hypothesis advocated by Paquet (1969, fig. 76, p. 178) that these accumulations form the basal part of the Subbetic nappe. At all those places, however, where Triassic rocks can be unmistakably shown to be in squeezed position between the Jurassic carbonate rock sequence, which has moved horizontally, and a relatively stable substratum, the thrust plane below the Triassic rocks appears to be subhorizontal; in all these cases the rocks are smeared out into thin sheets, or may be drawn out into thin lenses and are constituted of a chaotic jumble of rock-fragments of various sizes, shapes, and lithologies.

The most reasonable conclusion, therefore, is to assume that the main body of these massive Triassic rock accumulations has not been thrust over a distance of some 20 km, but are autochthonous to parautochthonous, and that these rocks represent the cores of diapiric folds that were broken through upwards. This would account for the steep abnormal contacts, the relatively slight disturbance, and the remarkable stratigraphic order of the Triassic rocks. It also explains the consistent absence of tectonic inliers of Prebetic rocks south of the extended strike-line of the Sierra de la Puerta, the large extension and massiveness of the Triassic accumulations, and the merely slight sinuosity of the outcrop of the structural contact between these Triassic massifs and the Prebetic rocks (the contact would be highly frilled if it had been a subhorizontal thrust plane). Finally it explains the consistent absence of rocks older than Triassic within the Subbetic zone; these are supposed not to have taken part in the diapirism. The hypothesis of a subhorizontal thrust plane can hardly explain the steep plunge of the Jurassic carbonate rock sequence south of the Sierra de Gavilán down to a depth of at least 1.5 km (van Veen, 1969, enclosure 2, cross-section A), unless it is assumed that a considerable part of these competent rocks is scraped off either by the underlying highly incompetent Triassic rocks or by the incompetent Cretaceo-Tertiary rocks of the overthrust Prebetic. The extreme difference in competence of the rocks involved makes this unlikely.

If an autochthonous to parautochthonous position for the greater part of the Triassic rocks in the Subbetic realm is assumed, we also must assume the major disharmony to be situated at or near the lower boundary of the Subbetic Jurassic carbonate rock sequence, since there is evidence for a thrusting, of at least several kilometres, of the Jurassic rock sequence of the Buitre Unit over Prebetic rocks (van Veen, 1969, p. 110). This inferred position of the major disharmony is in accordance with the assumption that the deformative stresses could be transmitted only through these competent Jurassic carbonates. It is unlikely that the Jurassic carbonate rock sequence with its Cretaceo-Tertiary cover represents a distinct nappe thrust from the south over a lare distance on top of the parautochthonous Triassic rocks, as in the Subbetic zone no cover proper to the Triassic rocks other than the Jurassic carbonate rock sequence has ever been found. It is therefore more likely that this rock sequence was deposited more or less in situ upon the Triassic rocks. Also from a stratigraphical point of view there is no need to suppose a very large distance between the original site of deposition of the South-Prebetic rocks of the Sierra de la Puerta and that of the Subbetic Buitre Unit, for the Albian rocks in the southwesternmost part of the Sierra de la Puerta cannot be distinguished from those of the Buitre Unit. Moreover, a rapid facies change was demonstrated within the Sierra de la Puerta in the direction of the strike over a distance of merely 10 km from Upper Albian rocks (Canara Formation) of a character intermediate between those of the Prebetic

Sierra de la Muela and those of the Subbetic zone to Upper Albian rocks of full Subbetic character (Kuhry, in preparation, and own observation). This implies that the tectonic structures cut obliquely the various facies boundaries and it indicates the existence of rapid changes in facies in a direction that makes a large angle with the tectonic structures. Hence there is no need to presume a distance of more than a few kilometres between the original sites of deposition of the Cretaceo-Tertiary rocks of the Sierra de la Puerta and those of the Subbetic.

The proximity of the completely different facies of the Cretaceous rocks of the Prebetic Molino anticline and those of the Subbetic Rameles unit is adduced by Paquet (1969) as an argument in favour of a thrusting of the Subbetic rocks over a considerable distance. This juxta-position can however be explained by a shift of some 40 km along the postulated E-W directed, dextral transcurrent fault, the Calasparra fault zone, which forms the southern boundary of the Molino anticline (encl. 4). The northwestern extention of this major tectonic feature has been traced by Jerez Mir (1971) as far as Peñarrubia. It separates significantly differing developments in the stratigraphy of the Prebetic zone. This fault is analogous to the dextral transcurrent fault of Garobera (Paquet, 1969, 1972), which is responsible for the marked difference in facies of the Subbetic rocks directly to the north and south of this fault zone (encl. 4).

So (1) we have adduced arguments in support of an autochthonous to parautochthonous position of the large accumulation of slightly disturbed Triassic rocks in the area between Caravaca, Bullas, and Calasparra. (2) We also have advanced arguments in favour of a displacement of far less than 20 km for the Jurassic carbonate rock sequence of the Subbetic Buitre Unit. (3) Finally we have favoured the idea of a "décollement" process leading to concentric folding of the Subbetic rocks above the detachment plane into box-folds that have evolved either into diapiric folds or, when asymmetrical, into fold-overthrusts. Due to these three aspects and the obvious analogy with the Sierra de la Muela, the Buitre Unit west of Caravaca should be considered as a fold-overthrust evolved from a broad asymmetric box-fold with a core of Triassic rocks. East of Caravaca this box-fold has evolved into a diapiric fold of which the Triassic rocks of the core has broken through in upward direction.

The characteristic circular depressions around the diapirs can be indicated (encl. 4): for the Caravaca area the Cerro Gordo, the Cerro del Trompetero, the Peña del Gato, and the Montadores are interpreted as being located in this depression (van Veen, 1969, enclosure 1); for the Cehegin area the subsided blocks along the northeastern and eastern margins of the Peña Rubia (Paquet, 1969, p. 180, 181) are interpreted in the same way.

Consequently it can be assumed that the rocks in this part of the Subbetic zone are rather autochthonous than allochthonous. Many unsolved questions and puzzling phenomena can readily be explained when this assumption is accepted. For instance the "rétrochevauchement" assumed by Paquet (1969) to explain the structures of his Cabras-Burete Unit can easily be explained by strong diapiric folding.

Basement hinges

The tentative scheme set up to illustrate the relation between the stratigraphic rock units in the Caravaca-Moratalla area (encl. 5) clearly shows the existence of two important hinges, which probably came into existence by block-faulting of the Palaeozoic basement presumably at the beginning of the Albian. The southern

block would then have been uplifted about 900 m in relation to the block in between, the northern one about 500 m. In the foundered area between these hinges the thick pre-Albian deposits were therefore preserved against denudation and are virtually complete. In this area, the rocks were deposited that on account of their nature would ultimately be assigned to the Subbetic, South-Prebetic as well as Prebetic zones. But the Prebetic rock sequence north of the northern hinge and the Subbetic rock sequence south of the southern hinge also differ considerably from the Prebetic and Subbetic rock sequences deposited between these hinges. Perhaps the zones in which the Betic cordilleras are subdivided had better be redefined in connection with these and possibly other important basement hinges.

At the downthrown sides of these hinges the primordial box-fold of the Buitre Unit and the Muela box-fold came into being, probably partly as a consequence of the diapiric action of the Triassic rocks (because of the great thickness of pre-Albian deposits in the Charco Unit (Paquet, 1969) the diapiric fold of the Cabras-Burete Unit (encl. 4) possibly marks the downthrown side of the hinge that forms the southern boundary of the southern block). On account of the relations between the Tertiary rock units the shaping of these box-folds would already have started approximately at the beginning of the Lutetian. Then also the northern block would have been uplifted another 500 m, whereas the southern block sank some 400 m. During the Eocene, Oligocene, and Early Miocene the growing Buitre box-fold gradually separated the basin between the two box-folds (the primordial Moratalla syncline) from the basin south of the Buitre box-fold (the primordial syncline of the Loma de Solana Unit). In the former basin lagoonal circumstances gradually became prevalent, whereas in the latter the Gredero Formation was deposited in open marine environments. In the Middle Miocene started the most important phase in the development of the region, during which the two box-folds gained several hundred metres in amplitude and during which the structures, now observed, came into shape. Then the Moratalla syncline again stood in open connection with the sea and became filled with the sediments and olisthons of the Trieta Formation.

Buitre Unit

The northern limb of the theoretical primordial asymmetric Buitre box-fold would have broken, after which the crestal part was thrust over the northern limb, while retaining its normal position and dragging along a few slices of Triassic rocks at its base. The average strike direction in this thrust sheet is about N 75 E (van Veen, 1969, enclosure 1), which is taken to be about perpendicular to the stress direction. The most prominent fault direction has an average SE-NW strike of about N 120 E (van Veen, 1969, enclosure 1), so at an angle of some 45° to the supposed stress direction (encl. 4). Hence the direction of these faults is one that could be expected of wrench faults. This is in accordance with the prominent horizontal grooves observed over a distance of several metres on a vertical fault-plane in Jurassic carbonate rocks at the Puerta de las Tablas. In fact this pass is situated on the conspicuous rectilinear SE-NW fault that forms the northeastern boundary of the Sierra de Benamor and the Sierra de Tejero and the southwestern boundary of the Sierra de Enmedio (encl. 1 and 4). Another equally conspicuous rectilinear SE-NW fault forms the northeastern boundary of the Sierra de las Muelas. South of the Sierra del Buitre and the Sierra de Tejero these two faults

are accompanied by fault-troughs in which the successive down-thrown blocks are separated from each other by faults parallel to the wrench-fault direction.

The Subbetic outliers of the Sierra del Enmedio, the Sierra del Frontón, and the western part of the Sierra de los Álamos fit like a jig-saw puzzle into positions along the front of the Buitre thrust sheet when shifting them southeastward along these SE-NW trending faults. They fit so surprisingly well that it can hardly be a coincidence.

It may be concluded from these observations that the SE-NW trending faults are indeed wrench faults cutting the Buitre thrust sheet into oblong strips. The otherwise unexplainable origin of the Prebetic tectonic inliers visible through openings in the Buitre thrust sheet (van Veen, 1969) can readily be explained by assuming that the northwestern parts of some of these strips got detached from their southeastern parts and moved northwestward leaving a gap between the two parts. The Prebetic tectonic inliers are situated in a row which is the prolongation of the broad NE-SW striking fold in the southern part of the Moratalla basin (encl. 4). The northwestward movement of the northwestern parts of the strips is therefore assumed to be largely gravitative. The frontal parts of the thrust sheet became detached from the main body of the sheet, moved on their own along these wrench faults in northwestern direction, and collided against the northern limb of the Moratalla syncline. They should be designated as delapsional outliers. The tension that caused the Escaramujo and Pajarejo fault-troughs may have been brought about by the incipient moving away of the Buitre and Tejero massifs respectively.

The question may be raised where the rocks can be found that originally covered the diapirically risen anticlinal core of Triassic rocks between Caravaca and Valentin (encl. 4). It appears that the string of delapsional outliers of Subbetic Jurassic carbonate rocks west and north of Moratalla and west of Calasparra exactly fills up this gap of 15 km. The outliers derived from this gap are: the Sierra de los Álamos east of the Puerta de las Canalejas, the hills of the Collado del Agua, the Peñon de Serrata, the Cerro de Gamonár, the Sierra Pelada, the Sierra del Cerezo (encl. 1), and the Sierra de San Miguel (encl. 4). The latter is not the westward continuation of the Prebetic Upper Cretaceous dolostones of the Sierra del Molino, as is advocated by Paquet (1969, p. 80), but a delapsional outlier composed of Subbetic Jurassic dolostones of the Gavilán Formation.

TECTONIC STRUCTURE OF THE PUERTA-NEVAZO COMPLEX

In the investigated area the Puerta-Nevazo complex is composed of two parts of which the upper one lies with a subhorizontal contact upon the lower one. In the Sierra de la Puerta the lower part consists of an undisturbed rock succession which lies in overturned position over the entire length of the mountain range. Rocks of Albian to Miocene age have been observed of which many, in particular the Albian rocks, occupy, with respect to their facies, an intermediary position between their Subbetic and Prebetic equivalents (Paquet, 1969). In the upper part, along the Barranco del Nevazo, rocks have been found of Barremian to Early Tertiary age in exactly the same facies as their Subbetic equivalents. In general the rocks of this upper part are strongly deformed; there are, however, arguments suggesting a generally normal position (see p. 174). Originally this marked difference in facies had been the strongest argument for considering both parts as

belonging to two independent structural units of which the lower part had been named Puerta Unit (encl. 4) and the upper part Nevazo Unit (encl. 4; van Veen, 1969). Notwithstanding this marked difference in facies it can be shown that both parts form one entity, because, when the Albian rocks belonging to the lower part are traced towards the west, they appear to pass uninterruptedly into those belonging to the upper part of the complex, changing gradually in facies. This is the reason for the combined name: Puerta-Nevazo complex.

Following the Sierra de la Puerta from east to west, actually a gradual transition can be observed from a complete overturned succession in intermediary facies (Puerta Unit) into an incomplete succession, in which at three stratigraphic levels rock-units are lacking due to tectonic dislocations – faulting – approximately parallel to the stratification. These three dislocations occur (1) within the Miocene rock sequence (between Moratalla Marlstone and Puerta Limestone), (2) between the Upper Cretaceous and Lower Tertiary rock sequence (Paquet, 1969, p. 109, and own observations) and, (3) between the Lower Cretaceous and Upper Cretaceous rock sequence. Further west the Upper Cretaceous rock sequence becomes progressively thinner and finally wedges out, so that the two latter dislocations fuse together; simultaneously the dislocations evolve into low angle thrust faults. The resulting configuration can be observed directly east of the road from Caravaca to Moratalla. Here overturned Cretaceous rocks, which are already in Subbetic facies, lie with a subhorizontal structural contact upon steeply dipping overturned Lower Tertiary rocks, which are still in intermediary facies and which are truncated by the contact; these Lower Tertiary rocks, which have fallen apart into imbricated wedges, in turn lie with a moderately southward dipping abnormal contact upon subhorizontal, undisturbed Miocene rocks of the Moratalla syncline (Moratalla Marlstone; fig. 60). Directly west of the road from Caravaca to Moratalla the two remaining structural contacts fuse to one subhorizontal contact. Above this contact lies a sheet of predominantly Lower Cretaceous rocks which are conspicuously deformed (van Veen, 1969, pl. X b); below the contact one finds subhorizontally stratified undisturbed Moratalla Marlstones, and along the contact there are lentil-shaped remnants of Upper Cretaceous and Lower Tertiary rocks, the latter still in intermediary facies.

In the area where this structural transition from the undisturbed inverted stratigraphic sequence (Puerta Unit) into the "thrust sheet" (Nevazo Unit) occurs, SE-NW striking dextral wrench faults are superposed upon these structures, which are progressively shifted by these wrench faults, so that it can be concluded that in the Puerta-Nevazo complex these wrench faults have been formed in a later stage than the dislocations of thrust-fault type.

STRUCTURAL EVOLUTION OF THE PUERTA-NEVAZO COMPLEX AND THE BUITRE UNIT

1. From a mechanical point of view it is impossible to explain the vertical to overturned attitude of the undisturbed rock sequence by assuming that a rigid horizontal slab of rock of more than 12 km long and 1 km broad became detached over its entire length from its substratum and turned over without rigid backing into its present inverted attitude, acquiring at a certain intermediate moment a vertical attitude of the entire slab (see Hsü, 1967). The anticline, of which the sequence of the Sierra de la Puerta is the steep northern limb, cannot

be a hypothetical Prebetic anticline that is completely covered by the Subbetic nappe (hypothesis favoured by Paquet, 1969), because the steeply plunging Jurassico-Cretaceous rock sequence of the Buitre Unit along the Río Argos (van Veen, 1969) does not leave enough room to permit the existence of such an anticline, and because it cannot explain the moderate disturbance of the Triassic rocks south of the Sierra de la Puerta, for these rocks should then have scraped off the crestal part of this anticline.

The only remaining possibility is to consider the rock sequence of the Sierra de la Puerta as the Cretaceo-Tertiary part of the overturned remnant of the northern limb (flap structure) of the inferred primordial Buitre box-fold. Hence the structure of the Sierra de la Puerta is a strong argument in favour of the inference that a primordial Buitre box-fold once existed. The rock sequence of the Sierra de la Puerta is situated rather in front of than below the Triassic rocks directly south of it.

2. There must exist a normal limb connected with the overturned sequence of the Sierra de la Puerta, or at least remnants of it. The only place to look for this normal limb is in the upper part of the Puerta-Nevazo complex, the Nevazo Unit. The observations suggesting a generally normal attitude of the rocks of this upper part in the region of the Cortijo del Buitre are in favour of this view.

If this Nevazo sheet does not represent the Cretaceo-Tertiary part of this normal limb, then it must be assumed that it was the cause of the squeezing out of the absent normal limb, but this is very unlikely because of the incompetence of the rocks in this sheet.

If we accept the Subbetic-nappe hypothesis of Paquet the explanation of the existence of the Nevazo Unit would be to assume that it represents a Cretaceo-Tertiary rock sheet stripped off from the crest of the hypothetical Prebetic anticline and pushed northward retaining its normal position without being entirely crushed. But it has already been pointed out that the existence of such a Prebetic anticline is improbable (page 190).

Also, it is evident that this upper part of the Puerta-Nevazo complex had slid down instead of having been thrust over the rocks of the Moratalla basin, for large parts of the Nevazo Unit show a totally undisturbed subhorizontal stratification which merges into a pile of cascade folds (upper course of the Barranco de Azafran). The totally undisturbed condition of the rocks of the Moratalla basin directly below the basal plane of the Nevazo Unit is however the most convincing argument in favour of sliding. For this reason we may conclude that the upper part of the Puerta-Nevazo complex, the Nevazo Unit, represents the original Cretaceo-Tertiary cover of the crestal part of the inferred primordial Buitre box-fold, which subsequently slid down into the Moratalla basin.

The fact that the Nevazo Unit slid down is a third and strong argument in favour of a generally normal attitude of the rocks of this sheet. The succession in the valley of the Barranco de Azafran, however, seems to suggest an inverted attitude: slivers of "couches rouges" along the basal contact, a middle portion of Upper Albian and Vraconian rocks (highly crumpled and brecciated lower part and little disturbed upper part), and an upper portion of strongly disturbed marlstones containing remnants of Argos Formation. However, in view of its emplacement by sliding it is more likely that the Nevazo Unit is in reality composed of two sheets of which the upper sheet contains rocks older (Barremian to Upper Albian) than those of the lower sheet (Upper Albian to Lower Tertiary). Both

sheets then have a strongly crumpled lower part and the rocks of both sheets would have a normal attitude. The slivers of Upper Cretaceous rocks along the base of the lower sheet are interpreted as remnants of the lower part of the Puerta-Nevazo complex (Puerta Unit) dragged along in northern direction.

The extent of the lower Nevazo sheet is restricted to the eastern part of the investigated area. The basal plane of the sheet has an altitude of about 900 m. The upper Nevazo sheet is restricted to the western part of the area with a basal plane at an altitude of about 1000 m. The sheets overlap in the middle part.

3. The Subbetic Jurassic carbonate rock masses of the Buitre Unit are lying in normal attitude only on top of this upper part of the Puerta-Nevazo complex. In the southernmost parts of the investigated area the subhorizontal basal planes of these Jurassic rock masses are lined with Triassic rocks; further to the north only locally thin slivers of Triassic rocks were found along this basal plane.

4. The question remains where the upper part of the Puerta-Nevazo complex that once was connected to the overturned lower part exposed in the eastern side of the Sierra de la Puerta has gone to. This upper part was actually found as thin sheets of Cretaceous rocks, one directly below the delapsional outlier of the Cerezo, and the other cropping out in normal attitude (in the Llanos de Carrasca, encl. 4) west and north of (and probably below) the delapsional outlier of the San Miguel (Paquet, 1969, p. 171 and fig. 72A). It has already been pointed out (p. 189) that these delapsional outliers should have originated from the region between Caravaca and Valentin. The Cretaceous rocks in these sheets are not in Subbetic facies like the western portion of the Nevazo Unit, but in intermediary facies (Paquet, 1969, p. 85, and p. 174 of this paper). This could be expected, since the tectonic structures are crossing obliquely the facies boundaries.

In consequence of these considerations the following interpretative explanation for the structure of the Puerta-Nevazo complex can be given:
Stage 1 – During the growth of the Buitre box-fold, of which the northern limb would have been steeper than the southern limb, the incompetent Cretaceo-Tertiary cover would have behaved disharmonically (décollement) with respect to its competent Jurassic carbonate substratum. The disharmony would have given rise to an overturned kneefold structure or roof-and-wall structure in which the roof (the normal limb) represents the future upper part of the complex (the Nevazo Unit), and the wall (the overturned limb) the future lower part (the Puerta Unit). The box-fold probably emerged above the level of the sea in which the Moratalla Marlstones were being deposited. During this stage the structure would have been attacked by erosion, and the denudation products that accumulated along the coast would have given rise to the turbidites, olisthostromes, mud flow deposits and fan-conglomerates that rapidly filled the Moratalla basin almost up to the rim. The normal attitude of virtually all the large tabular olistholiths can readily be explained by assuming that they are derived from the normal limb of the Puerta-Nevazo complex. Most of them would represent parts of the Tertiary cover of the Nevazo Unit slid down into the Moratalla basin (this is a fourth, though indirect argument in favour of a general normal attitude of the Nevazo Unit). They may therefore be designated as "klippes sédimentaires". But as these successive slabs, which might have triggered the successive incoherent slidings that gave rise to the olisthostromes, became incorporated in the olistho-

stromes, they should be called olistholiths. After the support of the roof-and-wall structure had been eroded away and the denudation products were deposited in the Moratalla basin by way of incoherent sliding, mud flowing, and gravel avalanching, the remaining Cretaceous cover of the crestal part of the Buitre box-fold (that is the roof) slid down in two steps as slip sheets, or diverticules, into the Moratalla basin over the olisthostromes. Probably they were emplaced subaqueously like the Ultrahelvetic diverticules in Switzerland (Badoux, 1973). The olisthostromes contain the elements that originally were situated upon the Cretaceous rocks that later slid down as Nevazo Unit over the olisthostromes.
Stage 2 – In this stage the steep northern limb of the Buitre box-fold broke, and the Jurassic carbonate rock sequence was thrust northward over its former Cretaceous cover. Hence the Nevazo Unit represents the original cover of the Buitre thrust sheet. Only are the lowermost Cretaceous rocks (up to Barremian) preserved in the fault throughs. At this stage the wrench faults would have come into existence, dividing the thrust sheet into oblong strips of which the southwestern ones would have moved over a larger distance (at least 9 km, van Veen, 1969) than the northeastern ones (about 5 km). The front of the Buitre thrust sheet was probably situated on the northern limb of the broad anticline in the southeastern part of the Moratalla basin. In this stage the frontal parts of the thrust sheet would have broken away and moved independently northward to lay themselves in a curved row against the northern margin of the Moratalla basin. The Cerezo outlier presumably dragged along its substratum, that is a part of the Nevazo slip sheet. These moving masses presumably still had their basal part in a shallow sea, since it is unlikely that they could move over a dry substratum.
Stage 3 – In this ultimate stage the fault-troughs had come into existence and the tectonic inliers of the Barquilla, Nevazo, and the Vicario (van Veen, 1969) were formed.

There are several indications suggesting that the primordial Buitre box-fold had an axial plunge towards the northeast:
1. The undisturbed condition of the inverted succession of the Sierra de la Puerta in the northeast is in contrast with the disturbed, thrust succession in the southwest.
2. The SE-NW trending wrench faults, which divide the thrust sheets into oblong strips, are all dextral, except where the frontal broken-away masses had followed these faults during their independent movement; the southwestern strips have successively been thrust over a greater distance than their northeastern neighbours.

From the observations 1 and 2 the conclusion can be drawn that the horizontal stress had been far greater in the west than in the east.
3. The erosion products shed into the Moratalla basin reveal that in the west the erosion of the crest of the box-fold had reached a far deeper level than in the east. In the west (in the Campo de Bejar) Hauterivian, Barremian, Aptian, and Albian to Miocene components were found, whereas in the east the oldest components are of Late Albian age.
4. The upper western sheet of the Nevazo Unit is composed of Barremian to Upper Albian rocks, whereas the lower eastern sheet contains Upper Albian to Lower Tertiary rocks.

From these observations it may be concluded that the western part of the Buitre box-fold had been considerably higher than the eastern part. The fact that, at present, in the eastern part the Triassic rocks of the core of the anticline are

cropping out, whereas in the western part the Jurassic and Lower Cretaceous cover is still present can be explained by diapiric rising of the Triassic rocks in the east in a late stage.

Summary and conclusions

In this part the olisthons occurring in the Moratalla area were described within the framework of the regional geology. Special emphasis was laid on the olisthostromes of the Benamor Geröllton Member, which are particularly suited to thorough investigation,

(1) because they are entirely undisturbed and relatively well exposed, as they are deeply transected by the valleys of the Río Benamor and its confluents,

(2) because their extension is restricted owing to the smallness of the receiving basin, and

(3) because the geology of the source area is well known.

The olisthostromes form part of a progressive succession of olisthons that delapsed from the primordial Buitre box-fold. This fold was formed along the down-thrown (= northern) side of an important basement hinge that came into being at the beginning of the Albian and again became active in the earliest Eocene. The rocks of the southern limb of the anticline are incorporated in the Subbetic zone, those of the northern limb in the South-Prebetic zone (the two lithostratigraphic successions are briefly discussed). The shaping of the Buitre box-fold presumably started at the beginning of the Eocene, but the fold, which probably had an axial plunge towards the east, did not rise above sea level until the Middle Miocene. Then it formed the southern border of the subsiding Moratalla basin and became the source area of the olisthons that delapsed from it towards the north.

The succession of olisthons, which filled the basin up to the rim, shows an increase in the intensity of delapsion corresponding to the increasing intensity of the diastrophism in the region. The whole succession is comprised in the Middle Miocene Ulea Group, and is in a way comparable with the classical flysch-wildflysch-diverticule-nappe sequence in the Alps, though of miniature size.

The succession started with the *redeposition* of South-Prebetic Rasa Sandstone elements (Eocene) in the Cuervo Sandstone Formation (20 m thick) (page 98), which was deposited along the southern border of the basin. It was followed by the coarse *rock-fall deposits* (1-2 m thick) of the Asares Conglomerate (page 100) probably along a cliff where already the Subbetic Argos Formation (Lower Cretaceous) was being eroded. In the centre of the gradually deepening Moratalla basin the Roble Limestone, Puerta Limestone, and the basal part of the Moratalla Marlstone were deposited yet without any sign of reworking.

But the delapsion gradually intensified, for along the southern border of the Moratalla basin the overlying Moratalla Marlstone Member consists of a 300 m thick sequence of *turbidites of proximal character* (page 101), while more basinward the marlstones (which contain reworked Cretaceous pelagic foraminifera) became subject to northward slumping and sliding which gave rise to *slump packets* (page 118) associated with *slump folds* (page 120) and *sandstone dykes* (page 120). The turbidite facies gradually progressed towards the centre of the

basin where only the upper 50 m of the Member contains *turbidites of distal character* (page 116) with flow directions from the east to southeast. On account of the rounded pebbles found in an interval of *sand avalanche deposits* (page 116), the siliciclasts of the turbidites are assumed to be derived mainly from the South-Prebetic Rasa Sandstone Formation, which is an Eocene sequence of *proximal urbidites* of northwestern provenance (page 92).

The progressive turbidite sequence of the Moratalla Marlstone Member was followed by the progressive succession of *olisthostromes* and *differentiated mud flow deposits* of the Benamor Geröllton Member (150 m thick), evidencing the general increase of the intensity of delapsion. The olisthostromes, which are interfingering with the Moratalla Member, are composed for about 80-90% of Moratalla marlstones *(endolisthostromes)* and for 10-20% of exotic inclusions *(exolistholiths)* which consist of Subbetic and South-Prebetic rocks. It is presumed that the exotic rock masses, which are derived, partially by delapsion, from the Buitre anticline, became embedded in the Moratalla Marlstone accumulations (for a large part composed of detritus from the same area) bordering the emerged anticline. These soft accumulations repeatedly flowed down towards the north together with the inclusions.

Among the rock types found as exolistholiths were observed: (1) parts of the Lower Eocene *endolisthostrome* of the South-Prebetic Moro Marlstone Formation (page 90), and (2) parts of the Turonian intraformational conglomerates (possibly *grain-supported slide conglomerates)* and conglomeratic marlstones of the South-Prebetic Quipar Formation (page 85). The latter can be regarded as the calcilutitic variety of olisthostromes of the *argille brecciate* type. The South-Prebetic deposits of the Moro and Quipar olisthostromes are, like the Rasa turbidites, presumably of northern provenance. The presence of exolistholiths derived from the Ulea Group as it is developed in the South-Prebetic zone is the evidence that the marginal deposits of the basin itself are 'cannibalized'.

At least 15 olisthostromes were distinguished above each other which successively extend further basinward (progression). The higher the olisthostrome, the older the exotic elements it contains, and the greater the volume of these older elements (inversion). The olisthostrome bodies appear to be tongue-shaped (about 10 m thick and 1 km wide on the average). The larger the inclusions (the largest is more than 500 m long), the more marginal and distal the position they seem to occupy within the tongue. The greatest volume of olistholiths presumably is concentrated where the tongues are widest and thickest.

The downward flowing rock mixtures probably scoured out broad, shallow channels of which the wall and bottom rocks became wrinkled by the stresses exerted by the passing flows. Also wall and bottom segments became dislodged and were dragged along with the flowing rock mixture. After the emplacement of each of at least the first few olisthostromes the deposition of Moratalla marlstone was resumed, but the increasingly rapid succession of olisthostrome influxes prevented the deposition of Moratalla marlstones above the later olisthostromes. This acceleration in the successive flowing in of olisthostromes and olistholiths of increasingly greater volume indicates the growing intensity of the delapsion and of the diastrophism in the region during the deposition of the olisthostromes.

Again the intensity of delapsion increased, for what still remained of the incompetent Cretaceo-Tertiary cover of the Buitre box-fold collapsed northward into a roof-and-wall structure. The wall subsequently developed into the *flap structure* of the Puerta Unit (page 190), which is composed of South-Prebetic

rocks, and the roof into the two *slip sheets (or diverticules)* of the Nevazo Unit (each 8 by 8 km) (page 176, 192), of which the lower sheet is composed of Upper Albian and Upper Cretaceous rocks of Subbetic and partly South-Prebetic character. In this 100 m thick lower sheet excellent *cascade structures* were developed (page 172). Locally the basal part of the sheet was severely distorted, so that the result can be designated as monogenic conglomeratic marlstone. If the whole sheet would have been transformed into conglomeratic marlstones the result could then be designated as *allolisthostrome*. At the base of this sheet several slivers of the Puerta Unit were dragged along.

West of this lower slip sheet the deposition of the Benamor Geröllton Member continued mainly in the form of extensive fans of *gravel-avalanche deposits* (page 166), which might be slightly older than Tortonian (Bejar Beds). The regional variation in composition indicates that the fans were fed from different parts of the Buitre anticline. During the deposition of the detrital calcarenites that are interfingering with these gravel avalanche deposits the current direction was from the southwest. A large *klippe sédimentaire* (page 164) is embedded within the fan-conglomerates.

These fan deposits and the western part of the lower Nevazo diverticule subsequently became covered by the 200 m thick upper Nevazo *diverticule* (page 176), which consists of Barremian to Upper Albian rocks of Subbetic character, thus slightly older than the rocks of the lower unit.

The climax in the intensity of the diastrophism was reached when the steep northern limb of the Buitre box-fold broke and the remaining 350 m thick Jurassic carbonate rock sequence (mainly Gavilán Formation) was thrust over its former Cretaceous cover dragging along a few slices of the Triassic Peñuelas Formation at its base. Large frontal parts of this Subbetic thrust sheet (Buitre Unit) broke away, and moved independently towards the north to lay themselves as *delapsional outliers* in a curved row against the northern margin of the Moratalla basin (page 189).

By that, the diastrophism came to an end, except for a regional uplift and a diapiric rising of the incompetent Triassic core of the Buitre anticline in the east, by which its axial plunge became west. The front of the Buitre Unit and the Gavilán massifs of the delapsional outliers became, probably during the Pleistocene, subject to intensive erosion which gave rise to *screes (rock-fall deposits),* to coarse rubble deposits discordantly covering the rocks around the massifs, to large landslides of the earth-flow type, to *debris flows,* and to *subaerial mud flows* (page 181). Associated with the rubble deposits large *slump blocks (toreva blocks)* of the Gavilán Formation came down (page 180). Large parts of the Álamos and Cerezo outliers show incipient slumping. The two slump blocks along the Barranco de las Tablas show excellent cascades.

The delapsion from the northern border of the Moratalla basin, which is formed by the Prebetic Muela box-fold, was less intensive and incidental, but certainly not less spectacular. Also the Muela box-fold was formed along the downthrown (= southern) side of a prominent basement hinge that came into being at the beginning of the Albian and was rejuvenated at the beginning of the Eocene. At the base of the Moratalla Marlstone Member a *reworking* of dolomite crystals and orbitolinids took place (page 114), apparently derived from the Ubacas Dolostones and Tosca Limestones. But when the lower half of the Moratalla Marlstone had been deposited, the Tertiary cover of the southeastern limb of the Muela anticline, when it rose above sea level as a promontory or island, slid down

towards the southeast to form *gravity collapse structures,* such as a *knee fold* (page 183), *roof-and-wall structure* (of which a part subsequently caved in (page 183) and even a small *slip sheet* (page 183). This gave rise to the unconformable deposition of the erosion products as *rock fall deposits* and as an *endolisthostrome* (Turmal Conglomerate Member) within the Moratalla Marlstone Member (page 124). The exotic components of these olisthons consist of rocks of the Muela anticline, of which the lithostratigraphic succession is given. Apparently the uplift was not yet completed, for the Turmal Conglomerates became tilted and large parts of it slipped down as tabular *slip blocks* further basinward (page 126).

In fact all types of olisthons that were distinguished in part 1 of this publication can be found in the Moratalla area except real allolisthostromes and olisthothrymmata. Perhaps the isolated blocks of Roble Limestone on top of the Cerro de Rótova could be designated as such (page 154).

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GEOLOGICAL MAP OF THE MORATALLA AREA - SE SPAIN - by Ph.J. HOEDEMAEKER

surveyed in 1963 1964 1967 1968

LEGEND

- Scree
- Fine-grained rockwaste, subaerial mudflow deposit, river terraces
- Coarse Rubble and landslides

ROCK-UNITS WITH PREBETIC AFFINITIES

- Trieta formation: Benamor gerölliton member / Bejar conglomeratic fan-deposits and detrital limestones — Upper Langhian + Serravallian
- Trieta formation: Turmal conglomerate member
- Trieta formation: Moratalla marlstone member
- Trieta formation: Puerta limestone member
- Roble limestone formation — Middle Miocene
- Murtas formation { Celada sandstone member / Grero marlstone member } — Oligo-Miocene
- Almanillas limestone form. { Fresno limestone member / Aguaderinida limestone member } / Miñano limestone form. — upper Middle Eocene
- Somogil formation, Rasa sandstone formation / Mojeyes dolarenite formation — upper Lower and lower Middle Eocene
- Ubacas dolostone formation — Cenomanian
- Tosca limestone formation / Bolos sandstone formation — Barremian-Albian
- Bañador limestone formation — Jurassic - Hauterivian(?)

ROCK-UNITS WITH SUBBETIC AFFINITIES

- Moro marlstone formation. — Paleocene and Lowermost Eocene
- Jorquera Formation / Quipar Formation — Upper Cretaceous
- Represa Formation / Argos Formation — Aptian and Albian
- Miravetes Formation — Barremian - Barremian
- Tollo Formation / Gavilán Formation — Jurassic
- Peñuelas Formation — Triassic

TOPOGRAPHIC FEATURES

- Barranco, Arroyo, River
- Metalled road
- Non-metalled road
- Foot path
- 1416 Elevation point
- 567 Triangulation point
- Well
- ACEQUIA Irrigation canal
- Isohypse
- Quarry
- Isolated buildings
- Village
- Moorish tower

STRUCTURAL FEATURES

- Fault, barbs at downthrown side
- Strike-slip fault
- Fault, of doubtful nature
- Fault, concealed
- Fault, further trace unknown
- Line of outcrop
- Idem, but concealed
- 12 Strike and dip
- Horizontal
- Vertical
- Inverted
- Slumpscar
- Low-angle structural contacts
- Slid-down block

Olisthostromes and other delapsional deposits, and their occurrence in the region of Moratalla (Prov. Murcia, southeastern Spain)

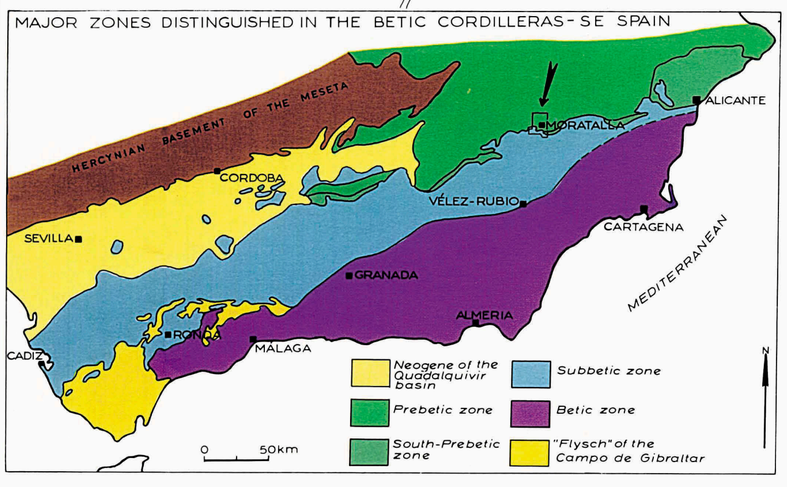
Fuentes de información de la topografía: hojas 889 (Moratalla) y 910 (Caravaca), escala 1:50.000, y aerofotos.

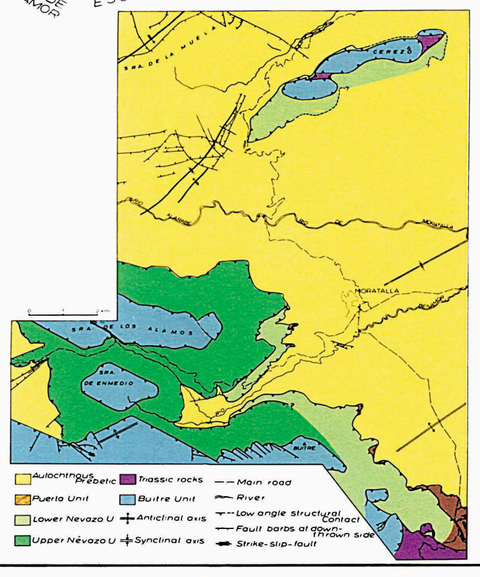
1ª edición autorizada por el Consejo Superior Geográfico en 6-4-1973.

0 1000 1500 m
SCALE 1 : 25.000

N

MAJOR ZONES DISTINGUISHED IN THE BETIC CORDILLERAS - SE SPAIN





DRAWN BY F.H. KIEVITS, GEOLOGICAL INSTITUTE UNIV. AMSTERDAM 1972

Enclosure 2

L E G E N D

The meso-, macro-, and megaolistholiths of the olistho-stromes of the Benamor gerölllton member are furnished with the letter O, the large slump-blocks of landslide origin with the letter V.
The lithostratigraphic units have the same colour as on the main map: the only additional notations are :

- Benamor gerölllton member: binder with micro-olistholiths and undifferentiated.
- Moratalla marlstone member as developed in the source area of the olisthostromes
- Subaerial mudflow deposit.

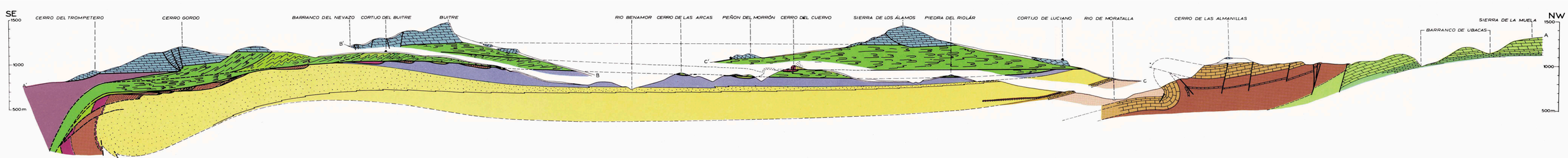
0 1000 m
SCALE 1:10.000

Olisthostromes and other delapsional deposits, and their occurrence in the region of Moratalla (Prov. Murcia, southeastern Spain)

Fuentes de información de la topografía: hojas 889 (Moratalla) y 910 (Caravaca), escala 1:50 000, y aerofotos.

1ª edición autorizada por el Consejo Superior Geográfico en 6-4-1973.

Fuente de las Higueras
Lagena
Parales
Sareon
HIGUERAS
761
744
694
REDONDO
651
Commencaper
600
MORATALLA
681
630
Asares
944
967
CUERVO
Collado de San Jorge
847
SAN JORGE
765
775
Bco
700
900
1000
886
882
32
20
30
BENAMOR
Azafran
Hilario
Tira
CANAL DE TAIBILLA
856
862
ARCAS
Cristo
1041
1036
734
712
HILARIO
802
PIEDRAS LLANAS
799
Santa Ana
Casa Nueva
RIO
800
Hollarionda
Bco de Agrico
40
1100
1200
1300
BUITRE
1428
1274
1305
1258
1264
SIERRA DEL BUITRE
ESCARAMUJO
37
20
15
10
13
48
Buitre
Umbria
Barrancallos


SE
NW
1500
1000
500m
CERRO DEL TROMPETERO
CERRO GORDO
BARRANCO DEL NEVAZO
CORTIJO DEL BUITRE
BUITRE
RIO BENAMOR
CERRO DE LAS ARCAS
PEÑON DEL MORRÓN
CERRO DEL CUERVO
SIERRA DE LOS ÁLAMOS
PIEDRA DEL RIOLÁR
CORTIJO DE LUCIANO
RIO DE MORATALLA
CERRO DE LAS ALMANILLAS
BARRANCO DE UBACAS
SIERRA DE LA MUELA
A
B
C
B'
C'
Peñuelas Formation (Triassic)
Gavilán and Tollo Formations (Jurassic)
Argos and Represa Formations (Lower Cretaceous)
Tosca limestone formation (Lower Cretaceous)
Ubacas dolostone formation (Cenomanian)
Quipar and Jorquera Formations (Upper Cretaceous)
Moro marlstone formation (Paleocene)
Rasa Sandstones Formation, Somogil Marlstone Formation (Ypresian and Lower Lutetian)
Almanillas limestone formation, Miñano limestone formation (Upper Lutetian)
Murtas formation (Oligo-Miocene)
Roble limestone formation (Middle Miocene)
Turmal conglomerate member (Middle Miocene)
Moratalla marlstone member with and without sandstone interbeds (Middle Miocene)
Benamor gerölliton member (Middle Miocene)
Alluvium
Cortijo (farmhous)
Faults and other structural contacts
Cuervo sandstone member and Puerta limestone member (Middle Miocene)
SCALE 1:25000

ENCLOSURE 3

STRUCTURAL UNITS OF THE MORATALLA-CARAVACA AREA AND ITS SURROUNDINGS

(POSTOROGENIC DEPOSITS ARE OMITTED)

1 Bco DE LA PUERTA
2 Bco DE CAÑADA LENGUA
3 TORRE JORQUERA
4 Bco DE GREDERO
5 CUESTA MATA
6 TORRE DE ALCORES
7 Bco DE REPRESA
8 MIRAVETES
9 Bco DE TOLLO
10 LOS AMONTADORES
11 PEÑA DEL GATO
12 CERRO DEL TROMPETERO
13 CERRO GORDO
14 LAS PEÑUELAS
15 SIERRA TEJERO
16 SIERRA DEL FRONTON
17 CASA SILLEN
18 CAÑADA RASA
19 Bco DE LA PUERTA
20 Bco DE LOS BARRANQUILLOS
21 Bco DE HONDARES
22 SIERRA DEL ROBLE
23 Bco DE LA MUELA
24 Cjo DE LA MUELA
25 Bco DEL BAÑADOR

LEGEND

- Parautochthonous Triassic rocks
- South-Subbetic zone
- Charco-Portillo Unit
- Burete-Cabras Unit
- Loma de Solana Unit
- Buitre Unit
- Rameles Unit
- Nevazo Unit
- Puerta Unit
- Prebetic zone
- low angle structural contact
- fault, barbs at downthrown side
- reverse fault
- strike-slip fault, basement not involved
- transcurrent fault, basement involved
- fault, inferred or concealed
- axis of anticline
- axis of anticline with inverted limb
- axis of syncline
- axis of syncline with inverted limb
- fortunite intrusion
- anticlinal or synclinal axis, inferred

SCALE

0 1 2 3 4 5 km

MURTAS
Ao de las Murtas
SRA. DEL PUERTO
SRA. DE LA MUELA
CEREZO
SRA DE LOS FALCONES
Rio Segura
CALASPARRA
SRA. DEL MOLINO
SAN MIGUEL
LLANOS DE LA CARASCA
Rio Argos
investigated area
Rio Alárabe
Rio de Moratalla
ULEA
MORATALLA
SRA. DE LOS ALAMOS
Rio Benamor
Rio
VALENTIN
RA-MELES
ENMEDIO
Rambla de Ulea
MIÑANO
SRA. DE LA PUERTA
LAS MUELAS
PAJAREJO
ESCARA-MUJO
BUITRE
NEVAZO
Bco del Moro
CANARA
CAMBRONES
LAS BAR-QUILLAS
VICARIO
SRA. DE GAVILÁN
CARAVACA
Rio de Quipar
CEHEGIN
PEÑA RUBIA
LOMA DE SOLANA
LA COPA
Rio Argos
BENABLON
LOMA DEL GANSO
Rio Quipar
SRA. QUIPAR
BULLAS
LAS CABRAS
SRA. DE BURETE
Rio de Mula

drawn by B.Collet

S U B B E T I C Z O N E

LOMA DE SOLANA UNIT

south east

BCO. DE LA PUERTA. PAQUET, 1969; KUHRY, 1972

middle

BCO. DE CAÑADA LENGUA, LOMA DEL GANSO. FALLOT ET AL., 1958; PAQUET, 1969; KUHRY, 1972

north west

BCO. DE GREDERO, CUESTA MATA, AND REGION IN BETWEEN. DURAND DELGA & MAGNE, 1958; VAN VEEN, 1969; PAQUET, 1969; KUHRY, 1972

BUITRE UNIT

south

BCO. DE REPRESA, RIO ARGOS, BCO. DE TOLLO, SRA. DE GAVILAN. VAN VEEN, 1969; PAQUET, 1969; KUHRY, 1971

north

CJO. DE PAJAREJO, ESCARAMUJO. KRIJNEN, 1967, INTERNAL REPORT; VAN VEEN, 1969

HINGE ACTIVE IN THE EARLY ALBIAN AND EARLY EOCENE

SOUTH-PREB.Z.

PUERTA UNIT

south west

BCO. DEL MORO, CAÑADA RASA

north east

BCO. DE LA PUERTA. PAQUET, 1969

P R E B E T I C Z O N E

MORATALLA SYNCLINE MUELA ANTICLINE

south

BCO. DE LOS BARRANQUILLOS

middle

RIO BENAMOR

north / south east

BCO. DE HONDARES, RIO ALÁRABE, RIO DE MORATALLA

middle

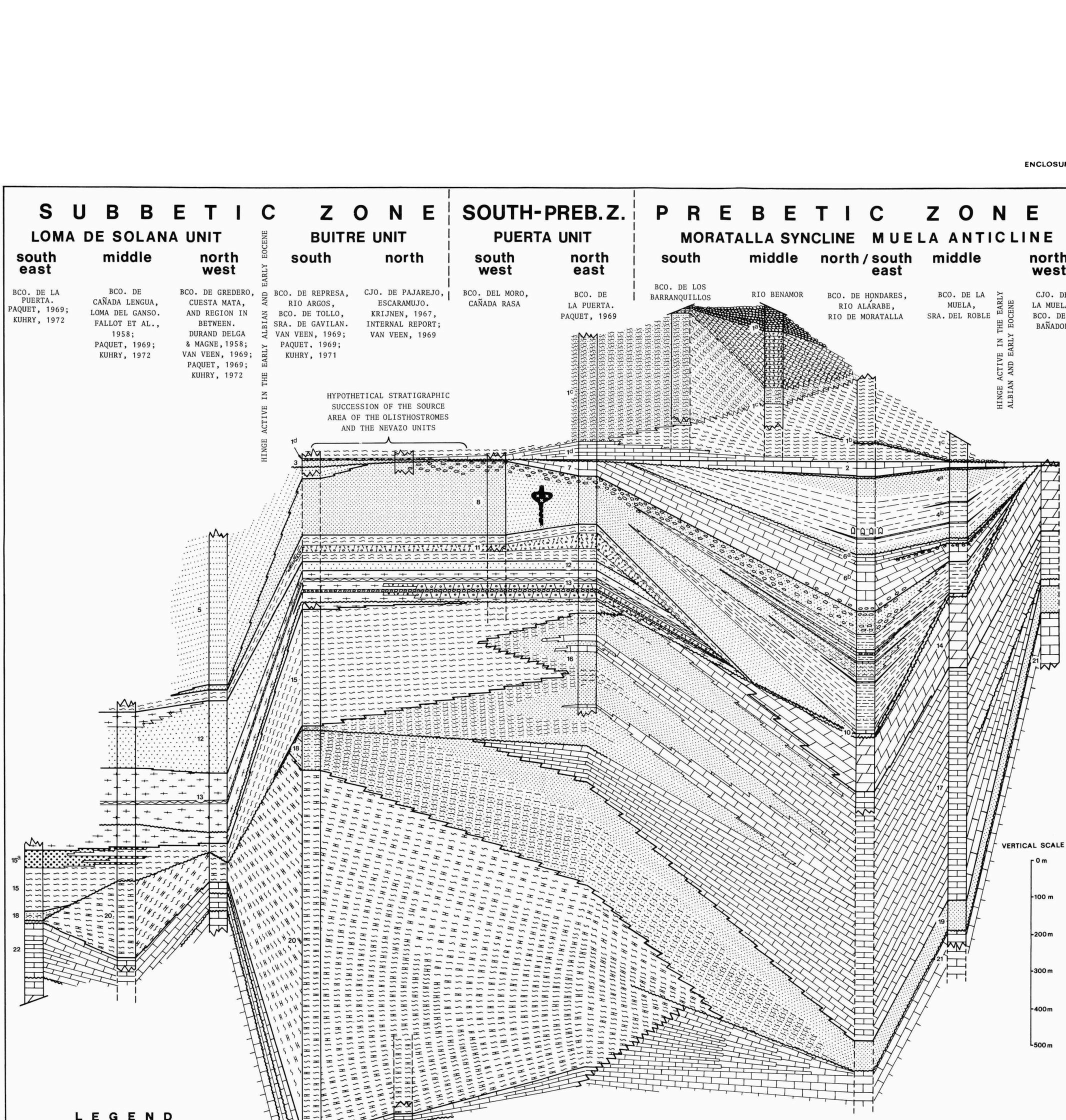
BCO. DE LA MUELA, SRA. DEL ROBLE

north west

CJO. DE LA MUELA, BCO. DEL BAÑADOR

HINGE ACTIVE IN THE EARLY ALBIAN AND EARLY EOCENE

HYPOTHETICAL STRATIGRAPHIC SUCCESSION OF THE SOURCE AREA OF THE OLISTHOSTROMES AND THE NEVAZO UNITS



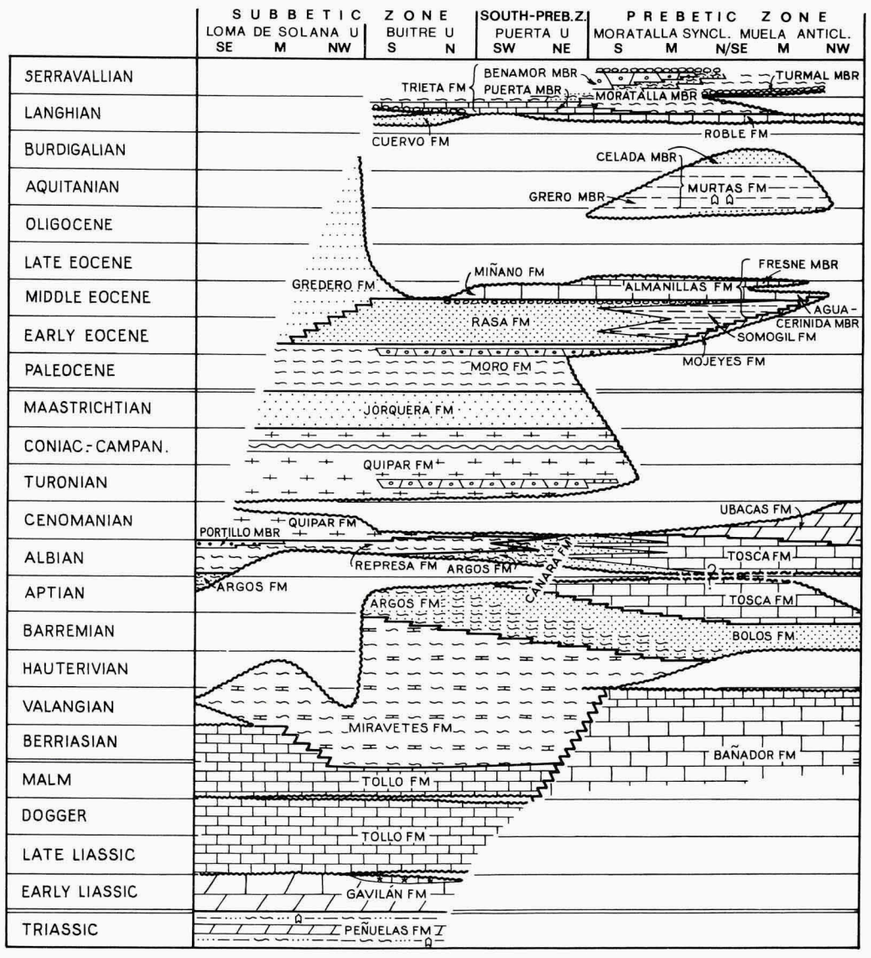
LEGEND

OUTSIDE THE COLUMNS THE THICKNESSES ARE INTERPOLATED AND THE LITHOLOGY-SYMBOLS DRAWN PARALLEL TO THE INFERRED TIME-LINES

- DOLOSTONE
- LIMESTONE
- ALTERNATING MARLY LIMESTONE-MARLSTONE
- ALTERNATING HARD AND SOFT MARLSTONE
- FISSILE CALCILUTITE
- RED BEDS IN QUIPAR FORMATION
- CALCILUTITE WITH INTERCALATED BEDS COMPOSED OF LARGER FORAMINIFERA
- ALTERNATING CALCILUTITE-CALCARENITE
- ALTERNATING MARLSTONE-SANDSTONE
- ALTERNATING SANDSTONE-SHALE
- SHALE
- SANDSTONE; VERY SANDY DEPOSITS
- CONGLOMERATE
- CONGLOMERATIC DEPOSITS
- CONGLOMERATIC MARLSTONE
- PELLET LIME-WACKESTONE
- CRINOIDAL LIMESTONE
- GYPSIFEROUS DEPOSITS
- DIABASE
- UNCONFORMITIES
- TIME-TRANSGRESSIVE BOUNDARIES
- END OF OBSERVED PART OF COLUMN
- SECTION TRUNCATED BY FAULT
- THICKNESS MEASURED
- THICKNESS APPROXIMATE
- THICKNESS INFERRED

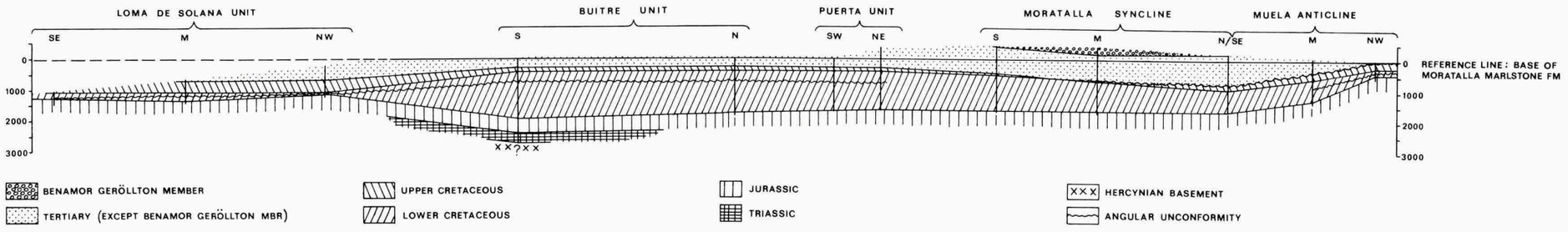
TENTATIVE STRATIGRAPHIC CORRELATION OF THE ROCK UNITS IN THE CARAVACA – MORATALLA AREA (SE SPAIN)

1 TRIETA FM
1a BENAMOR GERÖLLTON MBR
1b TURMAL CONGLOMERATE MBR
1c MORATALLA MARLSTONE MBR
1d PUERTA LIMESTONE MBR
2 ROBLE LIMESTONE FM
3 CUERVO SANDSTONE FM
4 MURTAS FM
4a CELADA SANDSTONE MBR
4b GRERO MARLSTONE MBR
5 GREDERO FM
6 ALMANILLAS LIMESTONE FM
6a FRESNE LIMESTONE MBR
6b AGUACERINIDA LIME-STONE MBR
7 MIÑANO LIMESTONE FM
8 RASA SANDSTONE FM
9 SOMOGIL MARLSTONE FM
10 MOJEYES DOLARENITE FM
11 MORO MARLSTONE FM
12 JORQUERA FM
13 QUIPAR FM
14 UBACAS DOLOSTONE FM
15 REPRESA FM
15a PORTILLO MBR
16 CANARA FM
17 TOSCA LIMESTONE FM
18 ARGOS FM
19 BOLOS SANDSTONE FM
20 MIRAVETES FM
21 BAÑADOR LIMESTONE FM
22 TOLLO FM
23 GAVILAN FM
24 PEÑUELAS FM



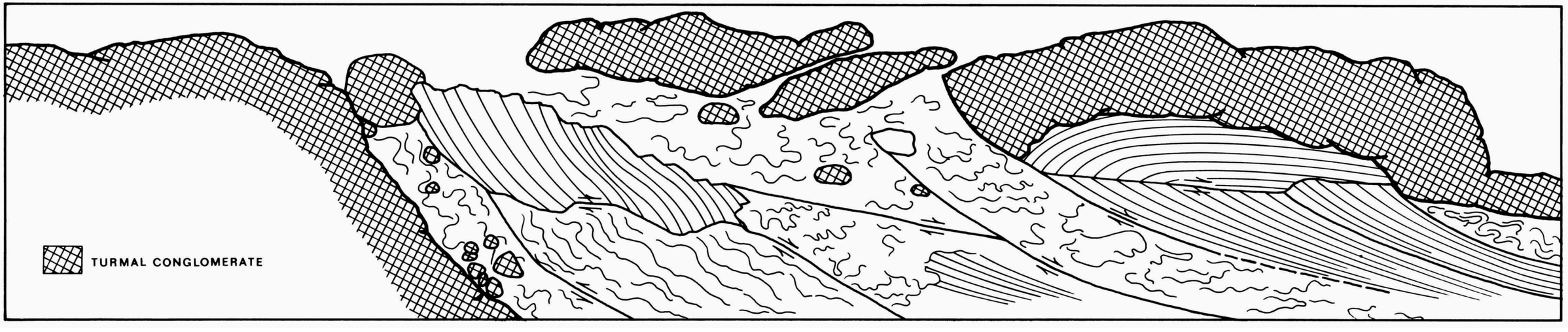
SCHEME IN WHICH THE ORIGINAL HORIZONTAL DISTANCE OF THE COLUMNS IS APPROXIMATED

HORIZONTAL AND VERTICAL SCALE ARE SIMILAR



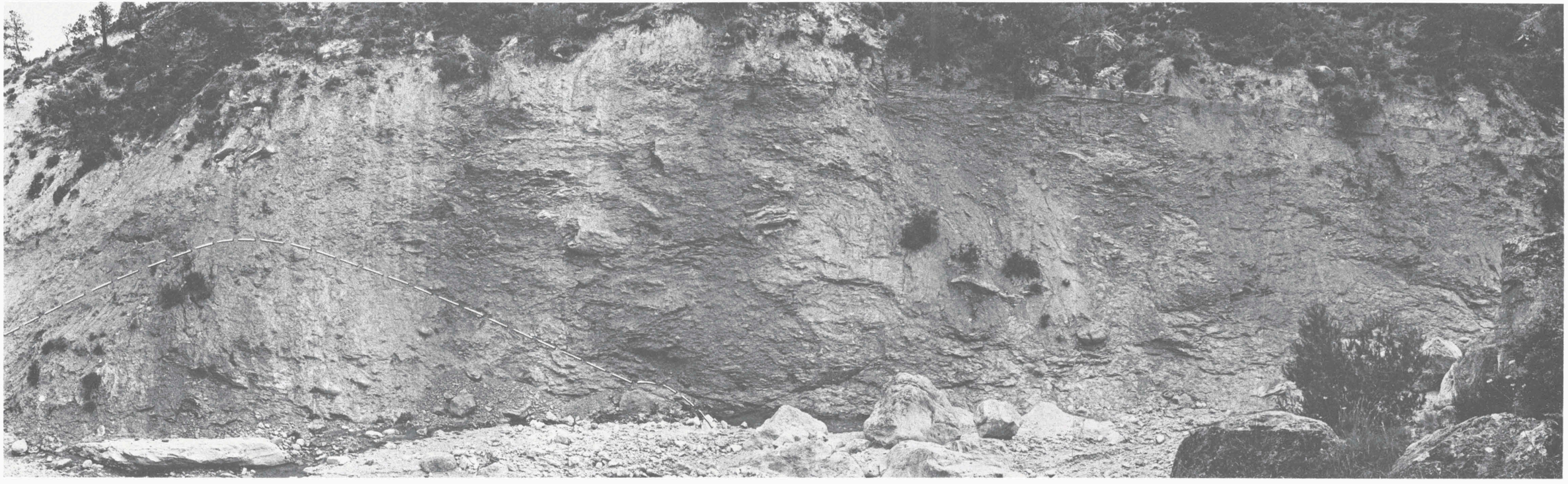
Drawn by B.Collet

6a: Panoramic view of the imbricated slip blocks of Turmal Conglomerate. They slid down in durated state over the still visco-plastic Moratalla Marlstone, which became severely crumpled, or plunged like wedges between the beds of the Moratalla Marlstone, which were tilted up. Río de Moratalla, 150 m east of the bridge (road from Moratalla to Las Murtas).



6b: Schematic drawing of 6a

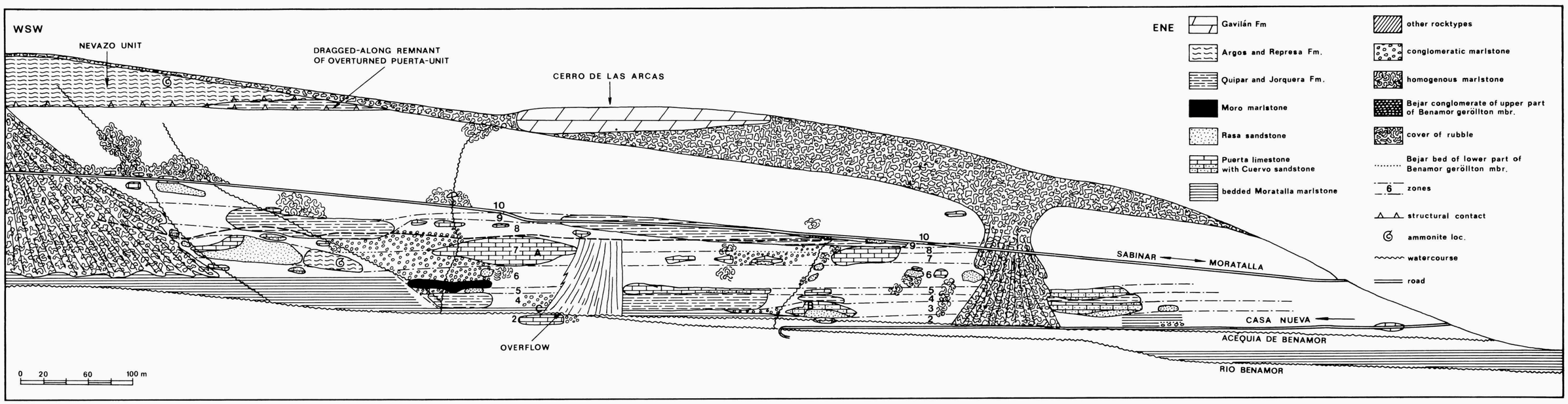
Enclosure 7



Panoramic view of the middle part of the typical exposure of the Benamor Geröllton Member. Note the horizontal Bejar interbed at the top of the unit, and the general horizontal attitude of the larger olistholiths. The dashed line marks the approximate boundery between the conglomeratic marlstone of the B.G.M. and that of the Moro Marlstone olistolith. Southern bank of the Río Benamor, SW of the Cortijo de Santa Ana.

8a: Panoramic view of the southern slope of the Cerro de las Arcas. Photo taken from the southern valley slope of the Río Benamor towards the north.

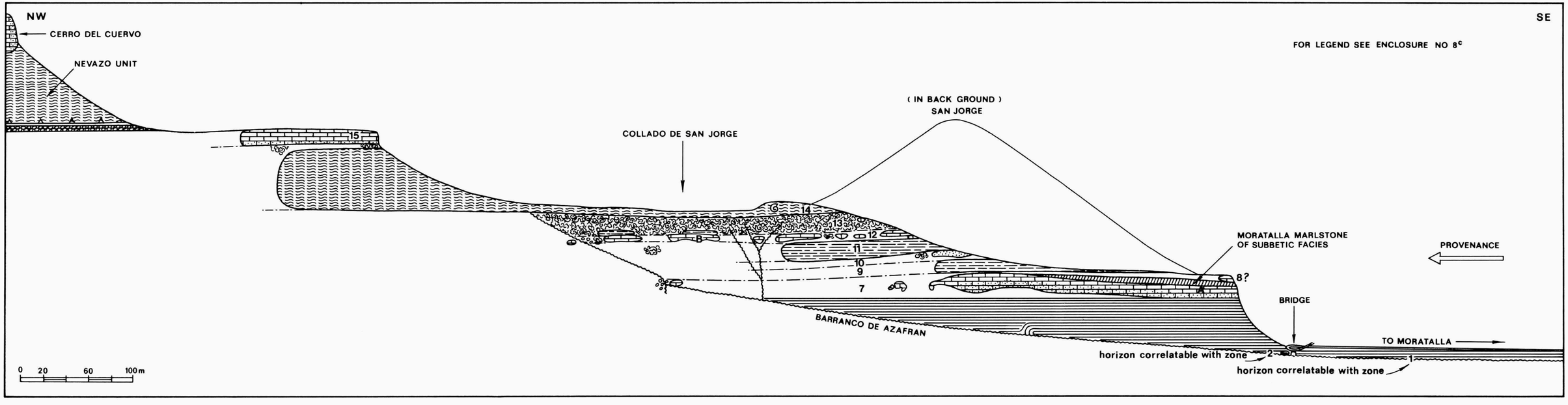
8b: Panoramic view of the southern slope of the Cerro de las Arcas. Photo taken towards the east.

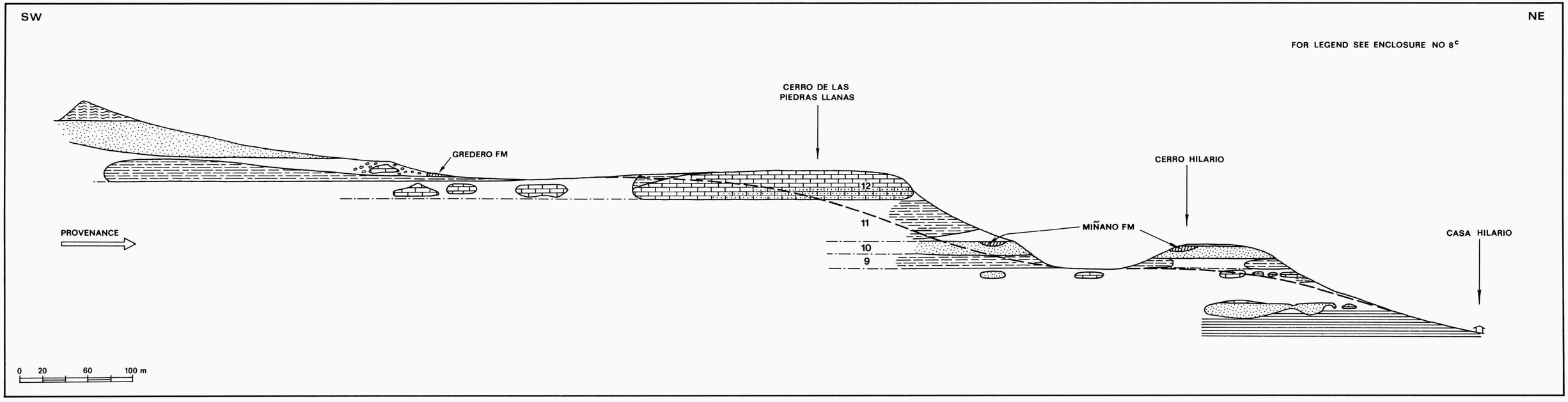


8c: Simplified projection of the southern slope of the Cerro de las Arcas. The letters A, B, and C correspond with those on encl. 8a and 8b.

9a: Panoramic view of the western slope of the Collado de San Jorge. Photo taken towards the east.

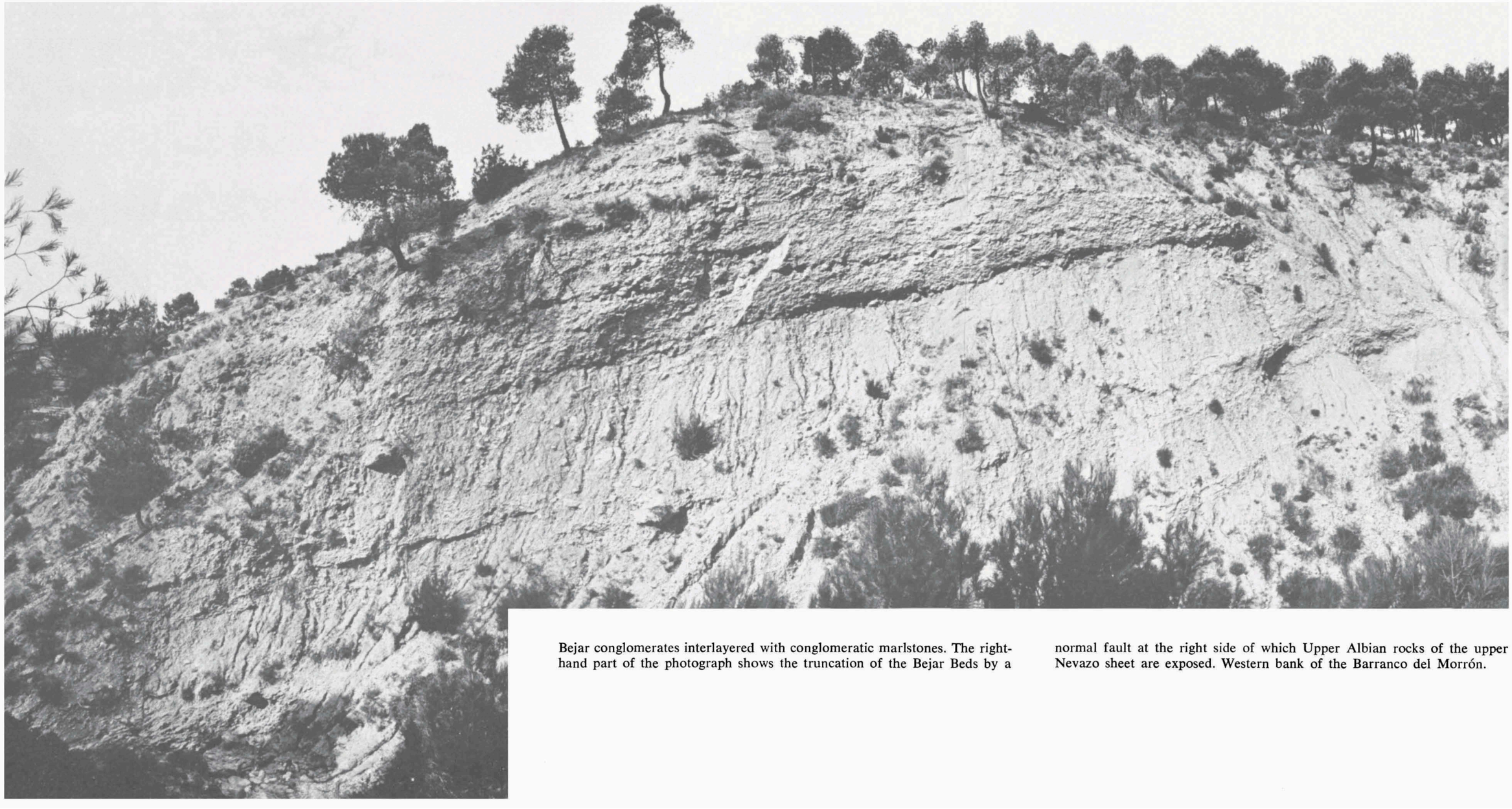
9b: Simplified projection of the western slope of the Collado de San Jorge. The letters A and B correspond with those on encl. 9a.



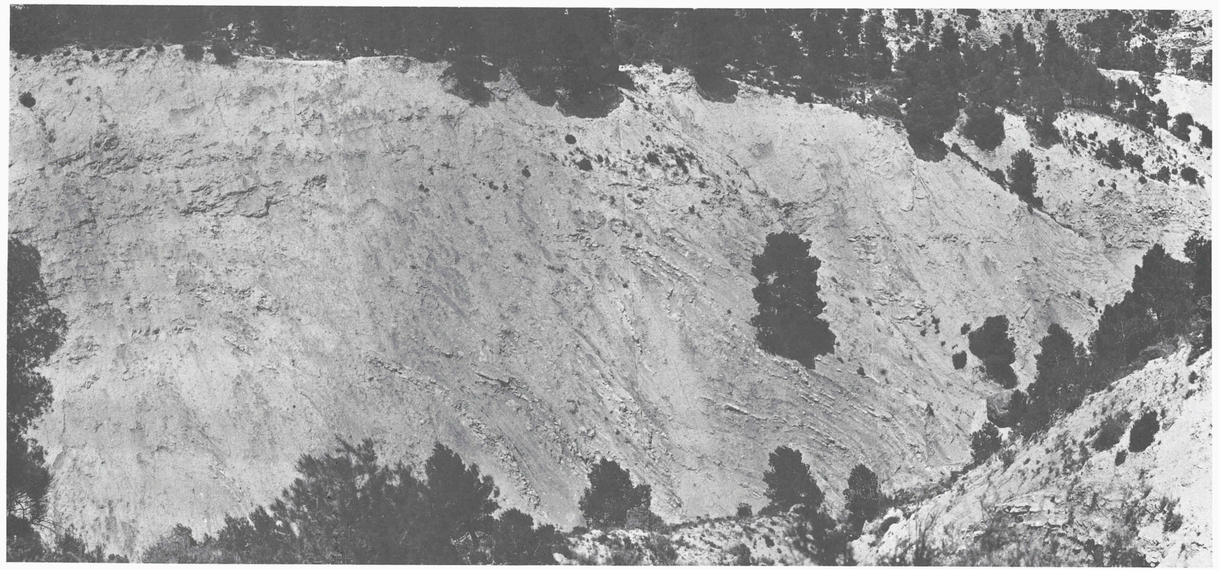


Simplified projection of the eastern slope of the Cerro de las Piedras Llanas and the Cerro Hilario.

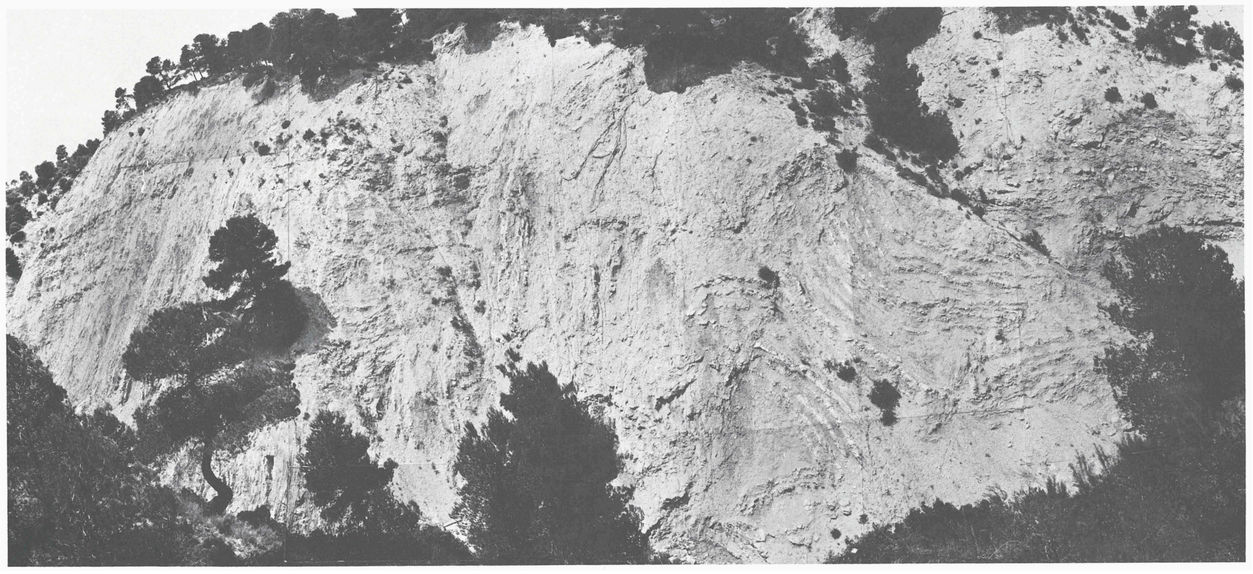
Enclosure 10



Bejar conglomerates interlayered with conglomeratic marlstones. The right-hand part of the photograph shows the truncation of the Bejar Beds by a normal fault at the right side of which Upper Albian rocks of the upper Nevazo sheet are exposed. Western bank of the Barranco del Morrón.

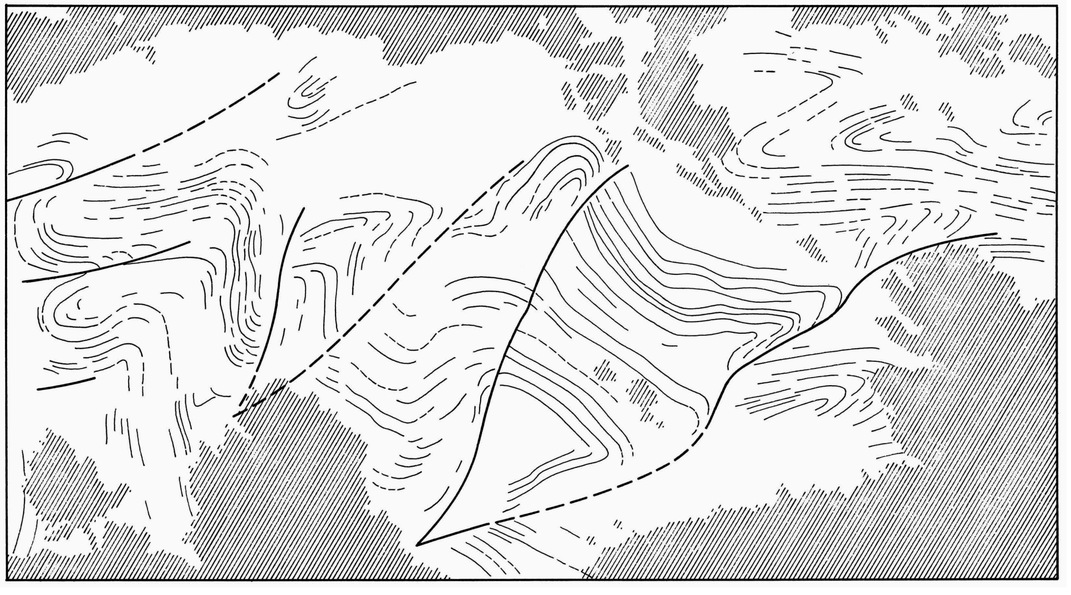


12a: Cascade folding in the frontal part of a large recumbent fold in the lower Nevazo sheet as exposed along the southern main tributary of the Barranco de Azafran.

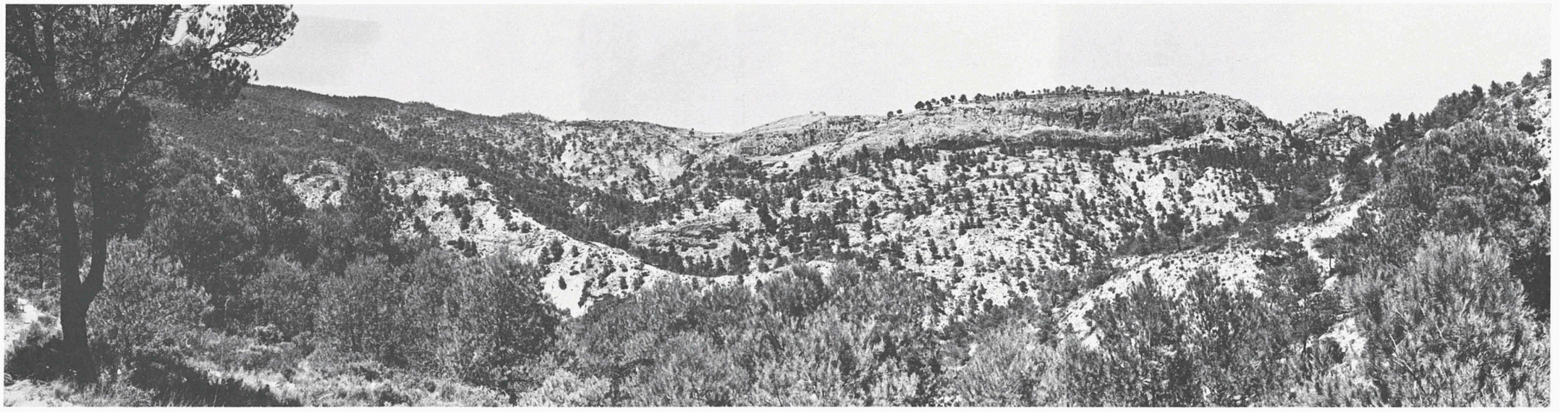


12b: Close-up of the cascades of 12a.

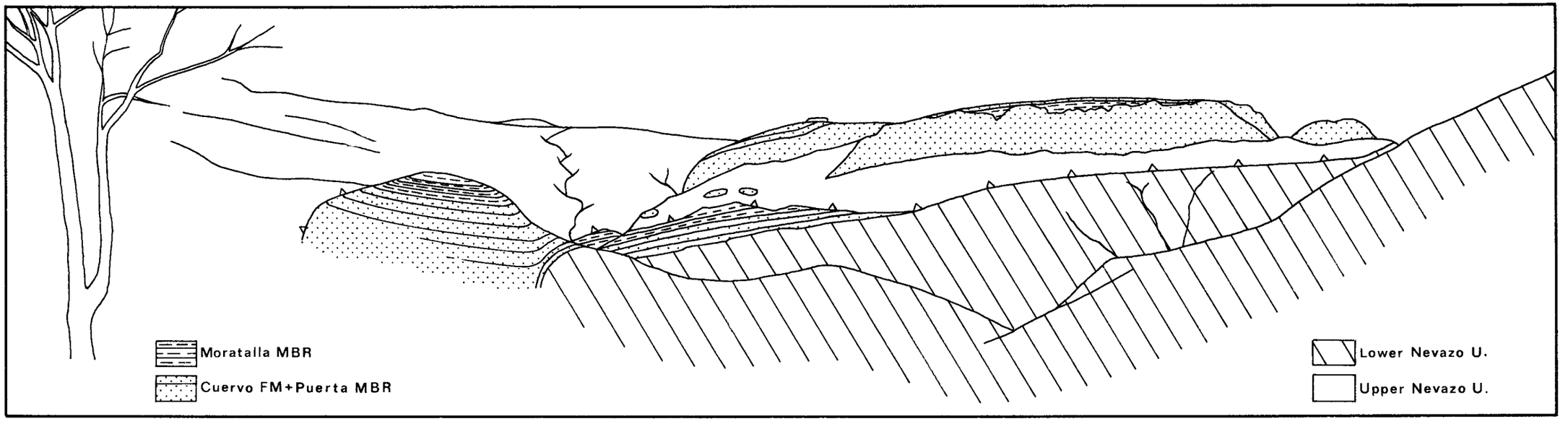
12c: Schematic drawing of 12a.



12d: Schematic drawing of 12b.



13a: Panoramic view of the headwater region of the Barranco de Azafran showing five of the six parts of the Azafran slice.



13b: Schematic drawing of 13a.