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Geosynclinal sequences of the Cordillera de Guaniguanico in western Cuba; their lithostratigraphy, facies development, and paleogeography

ABSTRACT: Jurassic to Paleogene lithostratigraphic units of the Cordillera de Guaniguanico, western Cuba, are formally described. The lithostratigraphic pattern includes Viñales Group subdivided into 5 formations; 9 other formations have not been clustered into any groups. Some formations are further subdivided. The new lithostratigraphic units are correlated with those recognized previously by other authors. The formations are considered within a framework of 4 stratigraphic sequences. One of these sequences occurs in the Sierra de los Organos, the others are in the Sierra del Rosario. The sequences correspond to distinct facies belts of the western part of Greater Antilles geosyncline. In the sequence of Sierra de los Organos, deltaic sedimentation was followed by Late Jurassic shallow-water carbonate sedimentation. The latter sedimentation type influenced also the southern sequence of Sierra del Rosario. The carbonate bank submerged in the Tithonian. Then, carbonate pelagic facies spread throughout the Cordillera de Guaniguanico. During the Early Cretaceous, the facies development in particular stratigraphic sequences became again variable. Deep-water sedimentation started in the Sierra del Rosario; in the northern sequence, it persisted up to the end of the Cretaceous. Pelagic, flysch, and rudaceous sediments were deposited in various facies belts during the Cretaceous. In the sequence of Sierra de los Organos, the Late Paleocene to Early Eocene sedimentation took place under the conditions of an increasing tectonic activity. Relation of the stratigraphic sequences of Cordillera de Guaniguanico to tectonic processes is also considered.

INTRODUCTION

The Cordillera de Guaniguanico occurs in Pinar del Río Province, western Cuba (Fig. 1). It was called either "Pinar del Río intrageanticlinal belt" (Furrazola-Bermúdez & *al.* 1964, Judoley & Furrazola-Bermúdez

1971), "Pinar del Río facies-structural zone" (Khudoley 1967), "meganticlinorium of Pinar del Río" (Pushcharovski & *al.* 1967), or "northern Pinar del Río" (Khudoley & Meyerhoff 1971). The name "Guaniguanico facies-structural zone" has recently been proposed (Pszczółkowski & *al.* 1975) because the structure covers but partly Pinar del Río Province. In this paper, the name "Cordillera de Guaniguanico" is used as synonymous with the latter term.

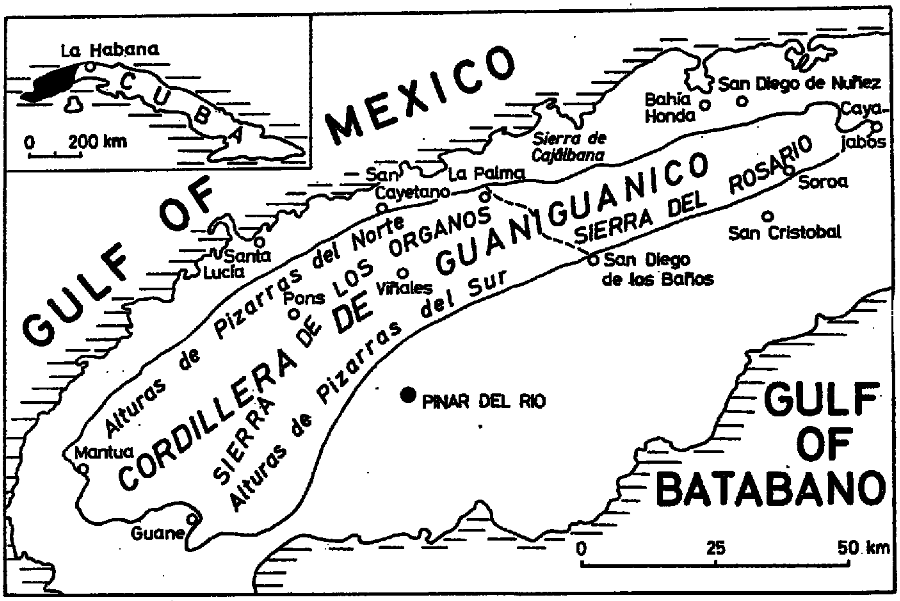


Fig. 1. Geographic setting of the Cordillera de Guaniguanico in Pinar del Río Province (marked black within the inset)

The present author participated since 1970 through 1975 in the geological mapping (scale 1 : 250,000) of Pinar del Río Province, realized by the Polish Academy of Sciences along with the Cuban Academy of Sciences.

The present paper deals with lithostratigraphy and facies development of the geosynclinal sequences (Jurassic to Paleogene) of Cordillera de Guaniguanico, and also with some paleogeographic problems. The author could consider these problems all over the Cordillera owing to the modern studies on biostratigraphy of the Jurassic (Houša & Nuez 1972; Houša 1974b; Myczyński & Pszczółkowski 1975, 1976; Kutek & *al.* 1976; Myczyński 1976; Wierzbowski 1976) and Lower Cretaceous deposits (Myczyński 1977). Modern-style stratigraphy and facies studies of the Mesozoic and Paleogene of the Cordillera de Guaniguanico (mostly in the Sierra de los Organos) have started with Hatten (1957). However, the results of this study have not insofar been published. The work of Hatten (1957, 1967), was commonly taken into account by the other students of

Pinar del Río Province (Gutiérrez-Domech 1968, Judoley & Furrazola--Bermúdez 1968) or the whole Cuba geology (Furrazola-Bermúdez & *al.* 1964, Khudoley & Meyerhoff 1971), even although some authors did not refer to it (Herrera 1961).

The nappe structure of Sierra de los Organos has been demonstrated by Hatten (1957, 1967), Rigassi-Studer (1963), Piotrowska (1972, 1975), and Danilewski (1972). In the case of the tectonic units of Sierra del Rosario, their nappe nature has been recognized by Pszczółkowski (1971a, 1972, 1976b). Some authors regarded the Sierra del Rosario as autochthonous relative to the allochthonous Sierra de los Organos (Rigassi--Studer 1963). It was also considered as a block structure (Pardo 1975). Nevertheless, the nappe structure of Sierra del Rosario (Fig. 2*A*) must be always kept in mind when studying stratigraphy or reconstructing paleogeography of this area.

The Cordillera de Guaniguanico appears very important in the geological structure of Cuba. In the early '70, the problems of Cordillera de Guaniguanico were among the unsolved or controversial points in geology of the Greater Antilles (Khudoley & Meyerhoff 1971). Several paleogeographic, structural, and geodynamic interpretations of the Greater Antilles and even the whole Caribbean referred to the data on Pinar del Río Province, especially on the Sierra de los Organos. Up to a few years ago, the Sierra del Rosario remained almost unknown.

Acknowledgments. The author is grateful to Dr. B. Łącka, Dr. A. Nowakowski, Dr. B. Wierzchołowski, and Dr. A. Skupiński for information on petrography of some clastic and volcanic rocks. Thanks are due to Dr. J. Kaźmierczak for identification of forams occurring in Pan de Azúcar Member, Docent H. Pugaczewska for identification of bivalves of the San Cayetano and Jagua Formations, and Docent A. Radwański for determination of decapod crab fragment. Dr. R. Myczyński, Dr. K. Piotrowska, Dr. G. Haczewski, and Dr. A. Wierzbowski assisted with information and helpful remarks.

STRATIGRAPHIC SEQUENCES OF THE CORDILLERA DE GUANIGUANICO

At present, 4 facies-structural zones are distinguished in Pinar del Río Province (Fig. 2*A*). These are: San Diego de los Baños, Bahía Honda (Khudoley 1967, Judoley & Furrazola-Bermúdez 1971), La Esperanza, and Guaniguanico zones (Pszczółkowski & *al.* 1975, Piotrowska 1976a). Originally, the Cordillera de Guaniguanico was not subdivided into areas different in stratigraphy or facies development. Hatten (1957) noticed some differences in the Jurassic and Cretaceous deposits between eastern and western parts of the Cordillera. Nevertheless, he did not present distinct stratigraphic columns for these two areas. Herrera (1961) recognized distinct lithostratigraphic units in the environs of Soroa, Sierra del Rosario, and in the Sierra de los Organos. Nevertheless, he did not introduce any subdivision of the Cordillera into some regions or strati-

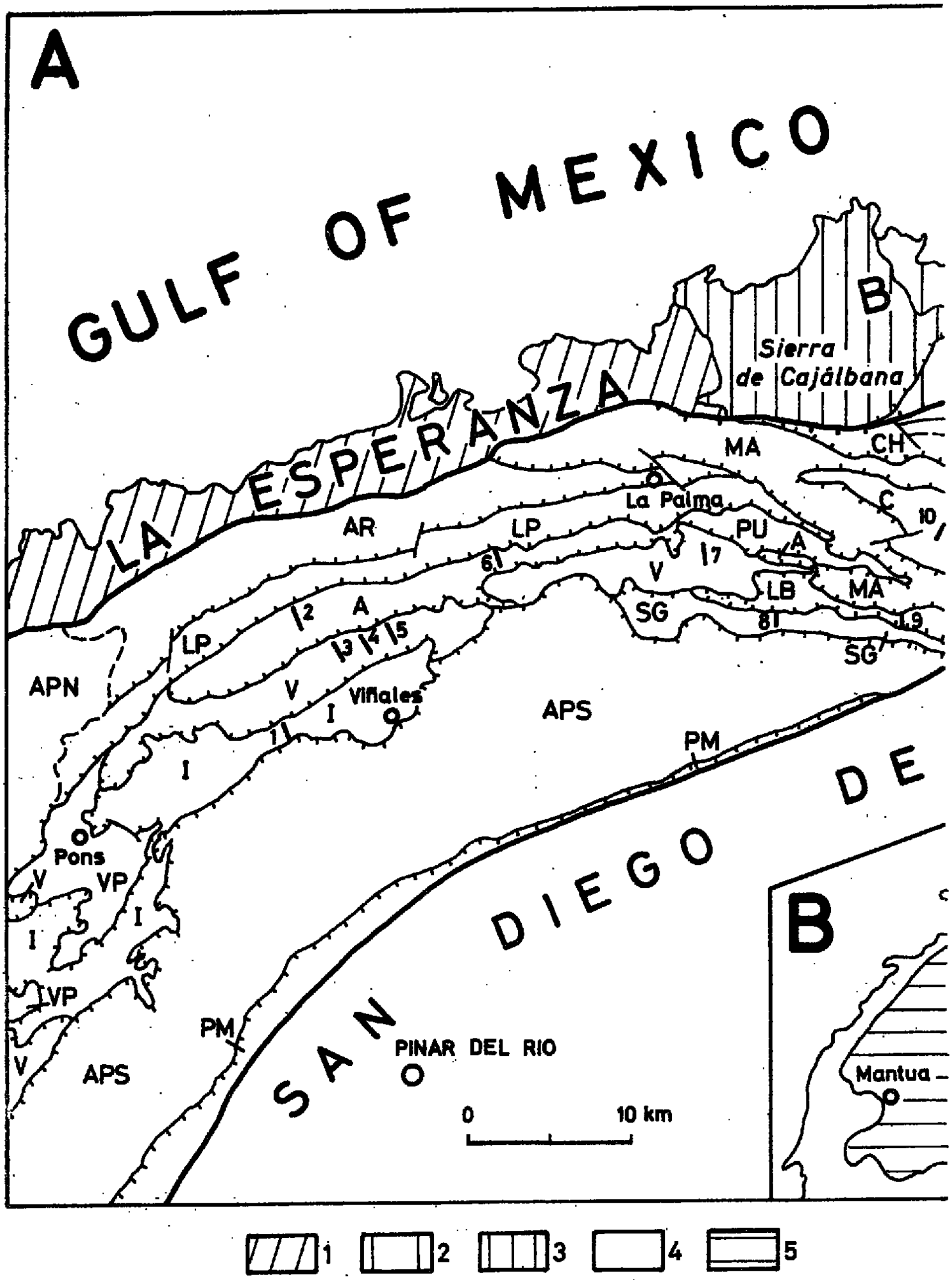
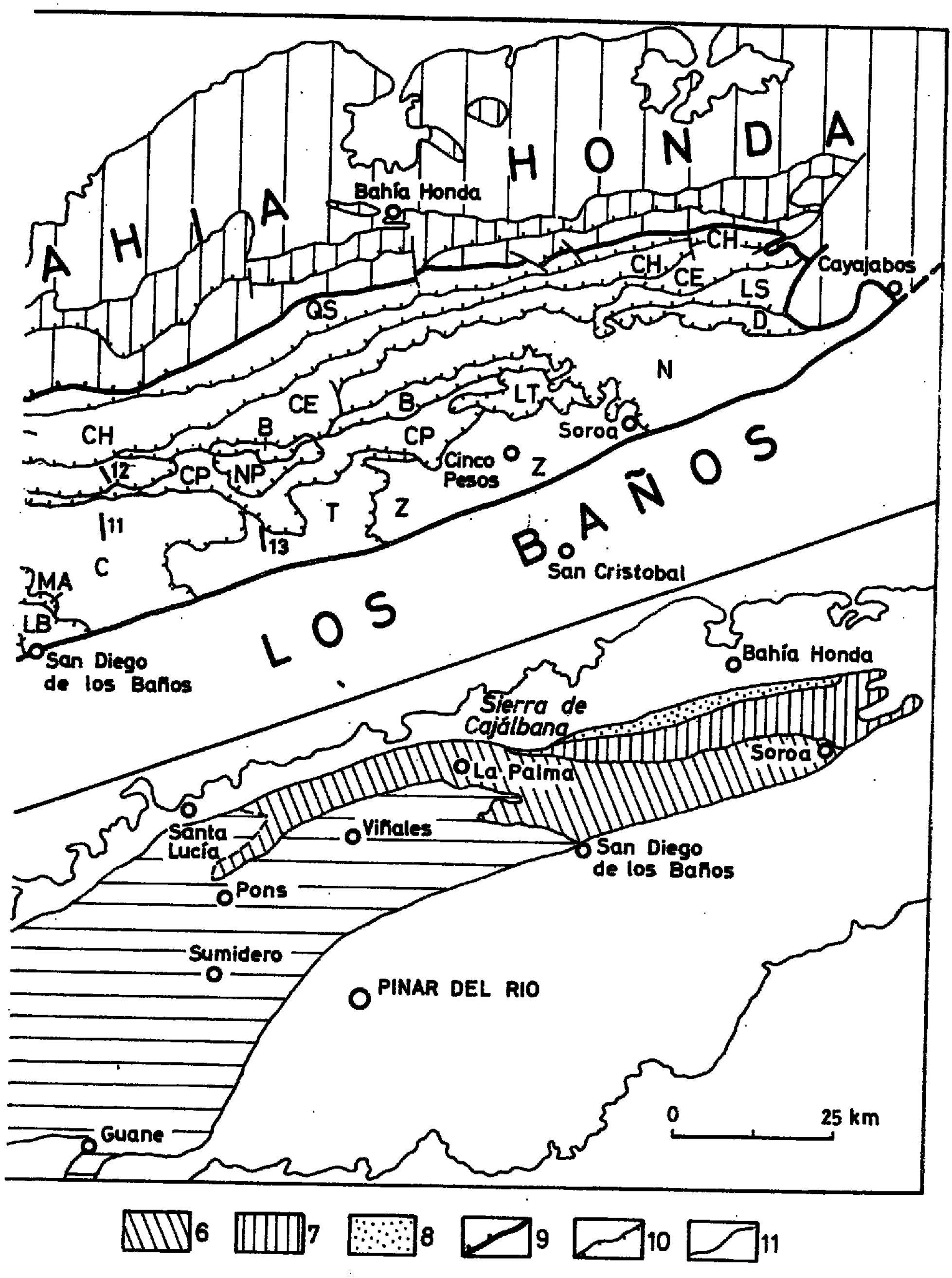


Fig. 2

A — Tectonic sketch of the northeastern Cordillera de Guaniguanico, based on the data gathered by the present author and other members of the Polish expedition to Cuba, *viz.* D. Danilewski, Dr. J. Piotrowski, Dr. K. Piotrowska, and Dr. A. Skupiński

Tectonic units of the Sierra de los Organos: *VP* — Valle de Pons, *I* — Infierno, *V* — Vinales, *A* — Ancón; *SG* — Sierra de la Güira, *APN* — Alturas de Pizarras del Norte, *APS* — Alturas de Pizarras del Sur, *PM* — metamorphosed units of Pino Solo and Mestanza zone; tectonic units of the Sierra del Rosário: *AR* — tectonic units of the northeastern Alturas de Pizarras del Norte, *LP* — La Paloma, *PU* — Loma del Puerto, *LB* — Los Bermejales, *MA* — El Mameyal, *C* — Caimito, *T* — Taco Taco, *Z* — La Zarza, *CP* — Cinco Pesos, *NP* — Niceto Perez, *LT* — Los Tumbos, *B* — Belén Vigoa, *N* — Naranjo, *CE* — Cangre, *D* — Dolores, *LS* — La Serafina, *CH* — Sierra Chiquita, *QS* — Quinones



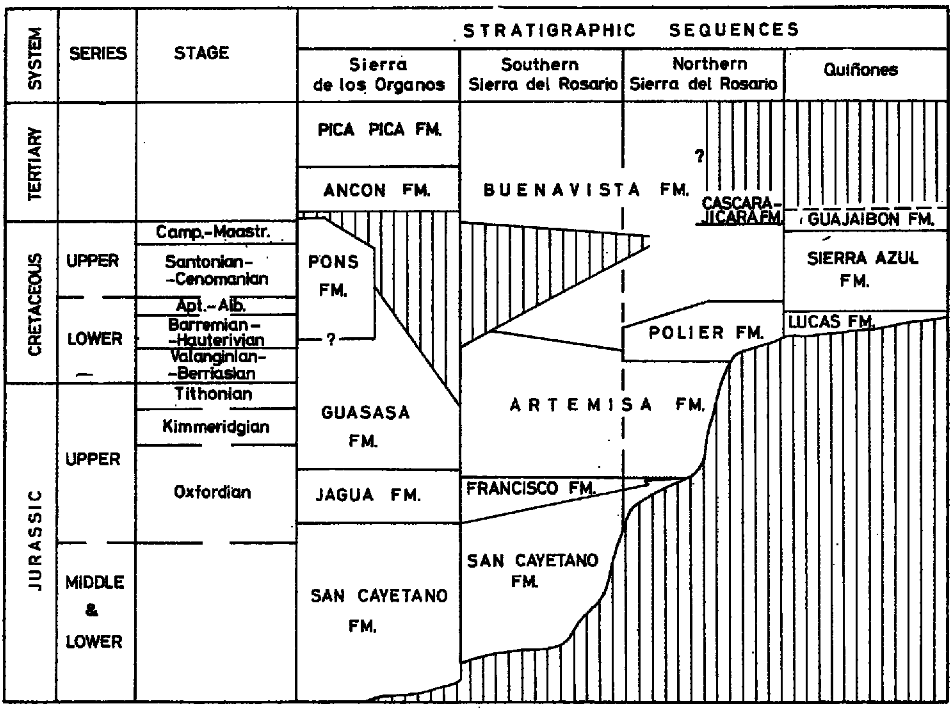
1 facies-structural zone of La Esperanza, *2* sedimentary and volcanic rocks of Bahía Honda facies-structural zone, *3* serpentinites, gabbroids, and ultrabasic rocks of Bahía Honda zone, *4* facies-structural zone of San Diego de los Banos, *9* tectonic boundaries (faults or overthrusts) of the Cordillera de Guaniguanico, *10* overthrusts, *11* faults

The lithologic-microfacies sections of Guasasa and Artemisa Formations *cf.* Text-fig. 5—9 and 11—13) are marked by strokes and consecutively numbered (1—13)

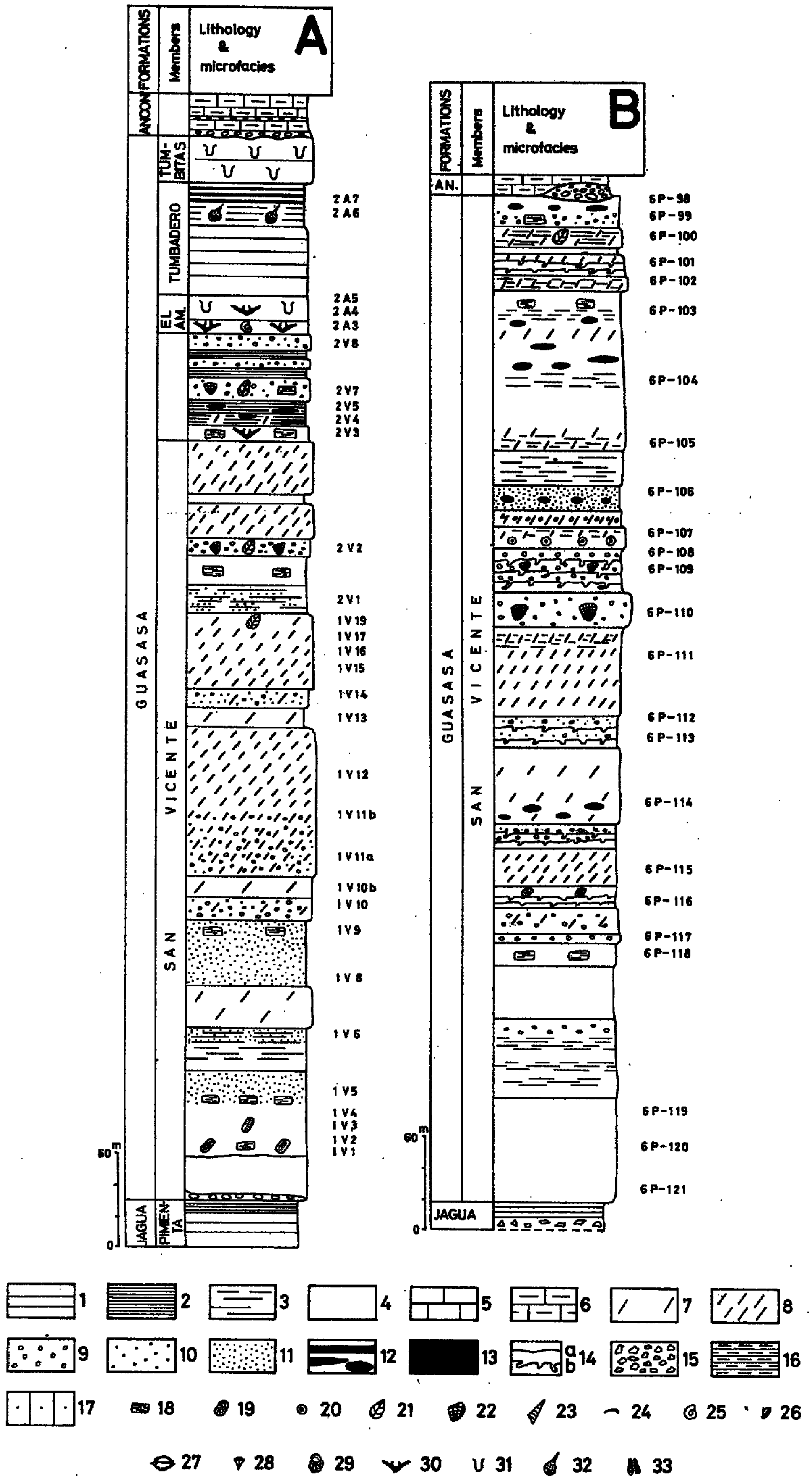
B — Present-day pattern of the stratigraphic sequences of the Cordillera de Guaniguanico

5 sequence of the Sierra de los Organos (and its metamorphic equivalents), *6* southern sequence of the Sierra del Rosario, *7* northern sequence of the Sierra del Rosario, *8* Quinones sequence of the Sierra del Rosario graphic sequences. Rigassi-Studer (1963) was the first to make a clear distinction between the overthrust units of Sierra de los Organos ("lames chevauchantes des Sierras") and "the autochthon of Sierra del Rosario". The differential stratigraphy and facies development of the Sierra de los Organos and Sierra del Rosario have been recognized by the present author (Pszczółkowski 1971a, b, 1972, 1973). A. A. Meyerhoff (*in*: Khudoley & Meyerhoff 1971), too, conceived as reasonable to treat these two areas separately. In contrast, Pardo (1975) subdivided Pinar del Río Province into 4 belts, *viz*. Bahía Honda, Organos, Rosario, and Cacarajícara belts. His interpretation of these belts implied a comparison among the geological units of differential rank.

Table 1
Formations of the Cordillera de Guaniguanico within the framework of stratigraphic sequences



The geosynclinal Jurassic, Cretaceous, and Paleogene deposits of Cordillera de Guaniguanico can be considered within a framework of 4 stratigraphic sequences (Fig. 2*B*, Table 1). In the Sierra del Rosario, three sequences are recognized; namely the northern, southern, and Quiñones sequences (Pszczółkowski 1976a). There is but a single stratigraphic sequence in the Sierra de los Organos. Rigassi-Studer (1963) suggested that the autochthon underlying the overthrust units of Sierra de los Organos occurs in the environs of Pons among others. However, this claim



A — lithologic-microfacies section of Guasasa Formation at the eastern margin of Sierra de Viñales (*cf.* section *4* in Text-fig. 2*A*); EL AM. — El Americano Member

SYMBOLS:

1 well-bedded, micritic limestones, *2* horizontally laminated limestone units, *3* thick-bedded and massive limestones indistinctly laminated, *4* micritic, massive and thick-bedded limestones, *5* well-bedded, micritic limestones (Ancón Formation), *6* marly limestones (Ancón Formation), *7* partly dolomitized limestones, *8* dolomites and significantly dolomitized limestones, *9* intraclasts, *10* ooids, *11* peloids (mostly pellets), *12* cherts and banded or lense-like cherts, *13* banded cherts (Buenavista Formation), *14* erosional surfaces (*a*) and burrow-bearing surfaces (*b*), *15* breccias, *16* shales, *17* sandstones, *18 Favreina*-form coprolites, *19* other coprolites, *20* onkoids, *21* benthic forams, *22* algae, *23* gastropods, *24* bivalves, *25* ammonites, *26* aptychi, *27 Globochaete alpina* Lombard, *28* fish teeth, *29* planktic forams, *30 Saccocoma*, *31* tintinnids, *32* radiolarians, *33 Nannoconus*

To the right of the section, samples are situated

B — Lithologic-microfacies section of Guasasa Formation at Mogote La Mina (*cf.* section *6* in Text-fig. 2*A*); *AN.* — Ancón Formation; for other explanations *see* Text-fig. 5*A* has not been confirmed, as there are in the tectonic windows formations resembling the sequence of Sierra de los Organos instead of those typical of the sequences of Sierra del Rosario (Pszczółkowski *in*: Pszczółkowski & al. 1975, Pszczółkowski 1976b). The metamorphic rocks of southern Sierra de los Organos (*cf.* Millán 1972, Piotrowski 1976) are to be considered as equivalent to some formations known from the sequence of Sierra de los Organos, even although they contain some volcanic intercalations (cf. Piotrowski *in*: Pszczółkowski & *al.* 1975, Piotrowski 1976). A general description of the sequences of Sierra del Rosario is given by Pszczółkowski (1976a).

The stratigraphic sequences of Cordillera de Guaniguanico represent various parts of the geosynclinal basin of western Cuba. The Jurassic to Paleogene lithostratigraphic units involved in the sequences of Cordillera are described below. D. Danilewski, Dr. J. Grodzicki, Dr. G. Haczewski, Dr. R. Myczyński, and Dr. K. Piotrowska have also contributed to the lithostratigraphic pattern of Sierra de los Organos, as presented herein. Their data were used to the comment to the geological map (scale 1 : 250,000) of Pinar del Río Province (*cf.* Myczyński *in*: Pszczółkowski & *al.* 1975).

SAN CAYETANO FORMATION

HISTORY

The original name Cayetano Formation as given by De Golyer (1918) has been subsequently replaced by the name San Cayetano Formation (Dickerson & Butt 1935, Schuchert 1935, Imlay 1942, Bermúdez & Hoffstetter 1959, Bermúdez 1961). At the moment, the latter name is commonly used.

LITHOLOGY

Lithologically monotonous San Cayetano Formation is strongly dominated by shales, sandstones, and siltstones. However, there are also some intercalations of conglomerates and limestones (mostly at the top part). The deposits are dark-gray or black (*cf.* Khudoley & Meyerhoff 1971, Haczewski 1976). They are rhytmically bedded but their sedimentary structures vary among distinct tectonic units (Meyerhoff & Hatten 1974). With this respect, there is a sharp difference between the Sierra de los Organos and Sierra del Rosario (*cf.* Haczewski 1976).

Up to the recent investigations, this lithostratigraphic unit was but very poorly known in the Sierra del Rosario when compared to the Sierra de los Organos. In the former area, the deposits of San Cayetano Formation represent either sandy-silty-shaly sediments, or thick-bedded sandstones (Fig. 3). The sandy-silty-shaly sediments resemble the facies *G* and *H* of Haczewski (1976). At Cinco Pesos, such sediments are 70 meters thick and occur in the uppermost part of the formation. They have also been recorded in the tectonic units of La Paloma, Loma del Puerto, and El Mameyal (*cf.* Myczyński & Pszczółkowski 1976, Figs 1—3). Sometimes, these sediments contain also micritic limestones up to 15 meters thick. In La Baría section, the limestones occurring 30 meters below the top of the formation include a thin coquina with ammonites and singular ostreids.

The sandy-silty-shaly sediments include commonly (Dr. B. Łącka, *personal information*) clayey-ferruginous quartz sandstones with fairly abundant flaky minerals such as the muscovite, hydromuscovite, hydrobiotite, chlorites, and kaolinite. There are also a few fragments of siliceous and clayey rocks, and granitoid debris. The plagioclases are uncommon. However, some sandstone beds comprise abundant detrital grains of kaolinite that might have originated from a kaolinitisation of feldspars. Sporadically, alkalic lavas debris have been recorded. Polymictic sandstones occur commonly in La Paloma tectonic unit; besides quartz and clayey rocks debris, these sandstones contain also fragments of quartz-sericitic shales.

The thick-bedded, coarse-grained sandstones, sometimes with pebbles up to 5 cm. long, have been recorded in the environs of Cinco Pesos, in Los Palacios River valley (at a slope of Sumidero hill), south to San Cayetano, and in Los Palacios River bed (et El Aguacate, by Pinar fault). These sandstones resemble the facies *I* of Haczewski (1976). At Cinco Pesos, deposits belonging to this lithofacies occur in the upper part of the formation, some 70 meters below the contact with Francisco Formation; whereas in Sumidero section these deposits attain the very top of San Cayetano Formation (Fig. 3).

The petrography of thick-bedded sandstones has been studied in samples derived at Cinco Pesos and in Los Palacios River valley. These deposits represent clayey-ferruginous quartz sandstones, clayey polymictic sandstones, conglomeratic sandstones, and conglomerates (Dr. B. Łącka and Dr. B. Wierzchołowski, *personal informations*). The sandstones consist mostly of rounded quartz grains; among the minor components, there are fragments of clayey rocks, cherts, quartzites, sparitic carbonates (dolomites among others), cataclasites, granitoids, and volcanic rocks. The sandstones display usually graded bedding with coarser pebbles occurring at the base. The detrital material is dominated by quartz, often vein quartz; pebbles of quartz or quartzitic sandstones are also common. There are also some pebbles of cherts, quartzites, quartz-feldspar rocks, quartz-micaceous shales, and clayey rocks. Anhydrite was also recorded. The matrix is of quartz-silica-clayey character.

FAUNA AND AGE

Rare fossils reported from San Cayetano Formation include bivalves (Krömmelbein 1956, Torre 1960, Imlay *in:* Judoley & Furrazola-Bermúdez 1968) and flora remains (Vachrameev 1936). Recently, the following bivalves were found in the Sierra de los Organos: *Eocallista (Hemicorbula)* spp., *Vaugonia (V).* spp., *Gervillaria* sp., and *Neocrassina* spp. (Pugaczewska *in press*). The specimens assigned to the genera *Gervillaria* and *Neocrassina* were found by Dr. G. Haczewski in calcareous sandstones of the upper part of the Formation in Ancón tectonic unit (near the place called El Abra). The specimens of *Vaugonia* come from sandstones occurring a few tens to some hundred meters below the top of San Cayetano Formation; they are also reported from Ancón tectonic unit and the area of Alturas de Pizarras del Norte (*cf.* Fig. 1), south to Baja (Dr. K. Piotrowska, *personal information*).

In the Sierra del Rosario, some ammonites were found in Mogote Simón (Myczyński & Pszczółkowski 1976) and La Baría sections (Fig. 3). From La Baría, the following species are reported (Myczyński *in:* Pszczółkowski & *al.* 1975): *Perisphinctes spathi* Sánchez Roig, *Glochiceras* cf. *subclausum* (Oppel), and *Ochetoceras* sp. In Cinco Pesos section, a decapod crab fragment was found (identified by Docent A. Radwański). There are also abundant plant remains in San Cayetano Formation in the Sierra del Rosario (*cf.* Haczewski 1976). The ammonites collected in the uppermost part of San Cayetano Formation indicate the Oxfordian age, probably Middle Oxfordian (Myczyński & Pszczółkowski 1976). Geological age of the lower part of San Cayetano Formation remains unclear.

In the Sierra de los Organos, geological age of San Cayetano Formation has not insofar been directly determined. As judged from its relation to Zacarías Member of Jagua Formation (*cf.* Wierzbowski 1976), its uppermost part can be regarded as the Oxfordian deposits. The older deposits of San Cayetano Formation were assigned to the Lower to Middle Jurassic (Furrazola-Bermúdez & *al.* 1964, Judoley & Furrazola-Bermúdez 1968, Khudoley & Meyerhoff 1971). Some deposits were assigned to the Middle Jurassic on the basis of their bivalve fauna (Krömmelbein 1956, Torre 1960).

The terrigenous sedimentation of San Cayetano Formation persisted somewhat longer in some sections of the Sierra del Rosario than it did in the Sierra de los Organos. In fact, the San Cayetano Formation reaches locally limestones of Artemisa Formation thus, replacing the Francisco Formation (Table 1). Presumably, the time differences were but slight and hence, the position of the upper boundary of San Cayetano Formation appears approximately constant all over the Cordillera (*cf.* Kutek & *al.* 1976, Myczyński 1976, Wierzbowski 1976).

LATERAL EXTENT AND THICKNESS

In the Cordillera de Guaniguanico, the San Cayetano Formation spreads between the localities Guane and Soroa, that is over some 150 km. The maximum thickness of San Cayetano Formation has recently been estimated to approximate 3,000 m in the sequence of Sierra de los Organos (*cf.* Meyerhoff *in:* Khudoley & Meyerhoff 1971). However, other estimates are also cited (over 5,000 m — Khudoley *in:* Khudoley & Meyerhoff 1971; 1,500 m in the northern Sierra de los Organos — Haczewski 1976). The deposits attain 1,500 m in thickness in the mine of Matahambre (Khudoley & Meyerhoff 1971). In the southern sequence of Sierra del Rosario, the maximum thickness of San Cayetano Formation does not exceed 1,000 m. The thickness varies among tectonic units, mostly because of tectonic reasons. The San Cayetano Formation occurs in the northern sequence of Sierra del Rosario but exceptionally and under the form of sandy-shaly patches up to a dozen meters thick, at the base of Naranjo tectonic unit by Soroa.

JAGUA FORMATION

HISTORY

The Jagua Formation has been recognized by Palmer (1945) in the environs of Jagua Vieja, Sierra de los Organos. Its type section has not been described by Palmer (1945) or other students (Hatten 1967, Herrera 1961). During the recent geological mapping of Pinar del Rio Province, calcareous and clayey deposits occurring between the San Cayetano and Guasasa Formations were assigned to Jagua Formation (Pszczółkowski & *al.* 1975, Kutek & *al.* 1976, Myczyński 1976, Wierzbowski 1976). This definition of Jagua Formation is also accepted in this paper.

LITHOLOGY

The Jagua Formation consists of limestones and shales clustered into 4 distinct lithological sets that have been recognized for distinct members within the Formation. There are micritic, coquinitic, bioclastic, and marly limestones in Jagua Formation. In the middle part of the Formation, clayey and silty shales comprise fossiliferous calcareous concretions. More detailed lithological descriptions of the members are given below.

SUBDIVISION

The Jagua Formation has been subdivided into 4 members, *viz.* Pan de Azúcar, Zacarías, Jagua Vieja, and Pimienta Members (Herrera 1961, Kutek & *al.* 1976, Myczyński 1976, Wierzbowski 1976).

AGE

The Jagua Formation ranges in age since the Middle Oxfordian through the earliest Late Oxfordian (Myczyński 1976, Wierzbowski 1976).

LATERAL EXTENT AND THICKNESS

The Jagua Formation occurs exclusively in the sequence of Sierra de los Organos. It attains 160 m in thickness in places. In the southern Sierra de los Organos (Mestanza area), the deposits of Jagua Formation are metamorphosed and intercalated with thin tuffite layers and alkalic or neutral lavas (Piotrowski 1976).

PAN DE AZÚCAR MEMBER

NAME

The name has been derived from Mogote Pan de Azúcar, northwestern part of Pinar del Río Province. It was proposed by the Institute of Geology and Paleontology of the Cuban Academy of Sciences ("Criterios sobre las unidades litoestratigraficas", May 21, 1973) to replace the previously used names Azúcar Formation (Hatten 1957, 1967) and Pan Formation (Herrera 1961). This reduction of Azúcar Formation of Hatten (1957) to the rank of a member in the lower part of Jagua Formation has been substantiated by its limited lateral extent, difficulties in its field recognition, and its lithological variability. In fact, the limestones of Pan de Azúcar Member appear sometimes quite similar to other deposits of Jagua Formation, when studied in poorly exposed sections. The type section of Pan de Azúcar Member is at the southwestern slope of Mogote Pan de Azúcar. The same section represented stratotype of Azúzar Formation of Hatten (1957, 1967).

LITHOLOGY

The Pan de Azúcar Member consists of well-bedded (0.1 to 1.5 m), compact, shelly and bioclastic limestones. They are dark-gray to black; the weathered surfaces are light-gray. There are also sandy limestones, mostly in the lower part of the Member. The limestones include beds or lenses of silicified limestones up to 20 cm thick. The silicification was epigenetic. The silicified limestones contain usually a lot of bivalve shells (Fig. 4). Microfacies characteristics of the limestones of Pan de Azúcar Member are given later on.

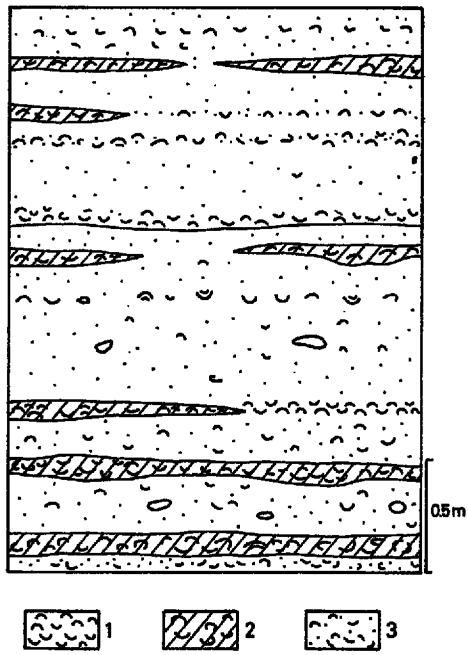


Fig. 4
Limestones of Pan de Azúcar Member of Jagua Formation at a locality between San Vicente and Valle del Ancón
1 bivalve coquinas, 2 silicified coquinas, 3 bioclastic deposits

FAUNA AND AGE

Some brachiopods (*cf.* Judoley & Furrazola-Bermúdez 1968, Khudoley & Meyerhoff 1971), ostreids (*cf.* Herrera 1961), and bivalves *?Posidonomya* sp. (*cf.* Khudoley & Meyerhoff 1971) were reported from the deposits of Pan de Azúcar Member. Specimens assigned by Pugaczewska (*in press*) to the bivalve genus *Gryphaea* Lamarck occur abundantly in Pan de Azúcar Member. The forams *Conicospirillina basiliensis* Mohler represent the only identified microorganisms (*cf.* Seiglie 1961) of this lithostratigraphic unit.

As judged from its relation to the adjacent units of Jagua Formation, the Pan de Azúcar Member is probably to be assigned to the Middle Oxfordian (*cf.* Kutek & *al.* 1976; Wierzbowski 1976).

LATERAL EXTENT AND THICKNESS

The Pan de Azúcar Member occurs but in the Sierra de los Organos, namely in the tectonic units of Ancón, Vinales, and some units of Alturas de Pizarras del Sur and del Norte. In its type section, the Member attains some 40 meters in thickness (36 m according to Hatten 1957); whereas in other places its thickness ranges from 0 to 78 meters (*cf.* Wierzbowski 1976).

CORRELATION

The Pan de Azúcar Member is time equivalent to the Zacarias Member of Jagua Formation (Wierzbowski 1976). Its stratigraphic correlation with the Oxfordian sediments of Sierra del Rosario appears less unequivocal. It may be equivalent to the lower part of Francisco Formation (Kutek & *al.* 1976, Myczyński 1976, Wierzbowski 1976) and sometimes also to the uppermost part of San Cayetano Formation, just as it is in the Sierra de los Organos (Fig. 3).

Outside the Sierra de los Organos, any time equivalents resembling the Pan de Azúcar Member in their lithology have not been recorded. In fact, the Smackover Formation, southeastern United States, was claimed by Meyerhoff & Hatten (1974) to be equivalent to Hatten's (1957) Azúcar Formation. However, its lithological and microfacies characteristics appear different from those typical of Pan de Azúcar Member (cf. Murray 1961, Bishop 1968, J. L. Wilson 1975).

ZACARÍAS MEMBER

HISTORY

This lithostratigraphic unit has been distinguished and formally described by Wierzbowski (1976) who recognized it for the lowest member of Jagua Formation.

LITHOLOGY AND THICKNESS

The Zacarías Member consists of clayey shales with thin intercalations of mudstones and bivalve coquinas. It attains 40 m in thickness. The Zacarías Member overlies directly the San Cayetano Formation and occurs in the Sierra de los Organos in the tectonic units of Ancón and Vinales.

FAUNA AND AGE

The deposits of Zacarías Member contain fairly abundant but poorly preserved ammonite imprints (Pszczółkowski 1971a; Nuez 1972, 1974; Wierzbowski 1976). Bivalves of the genera *Liostrea, Ostrea, Exogyra,* and *Plicatula* occur less commonly (Pugaczewska *in:* Wierzbowski 1976). There are also some imprints of small pieces of wood, and organic burrows (Wierzbowski 1976). The ammonites indicate that the Zacarías Member represents the Middle Oxfordian Wierzbowski 1976).

JAGUA VIEJA MEMBER

HISTORY

Jagua Vieja Member has been formally distinguished by Herrera (1961) who recognized it for the middle part of Jagua Formation.

LITHOLOGY

This member comprises shales and marly limestones. They are usually horizontally laminated. This lamination is especially apparent in some calcareous concretions typical of this lithostratigraphic unit. The deposits are black.

FAUNA AND AGE

The calcareous concretions contain well preserved ammonites, fish (*cf.* Gregory 1923), bivalves, and wood fragments. The list of ammonites is given by Wierzbowski (1976). The Jagua Vieja Member has recently been assigned to the Middle Oxfordian (Wierzbowski 1976).

LATERAL EXTENT, THICKNESS, AND CORRELATION

The Jagua Vieja Member occurs in the Sierra de los Organos. The maximum thickness is 60 m. In the sequences of Sierra del Rosario, its facies and partly time equivalent is the Francisco Formation (*cf.* Kutek & *al.* 1976, Myczyński 1976, Wierzbowski 1976).

PIMIENTA MEMBER

HISTORY

The Pimienta Member has been distinguished by Herrera (1961) in the upper part of Jagua Formation. The type section has been determined in the environs of Pimienta, southeastern margin of the Sierra de Cabezas (Herrera 1961; Fig. 2).

LITHOLOGY

According to Herrera (1961), the Pimienta Member consists of well-bedded, dark-gray to black limestones intercalated with calcareous siltstones. The present author has not recorded siltstones but thin shale intercalations do, indeed, occur in the lower part of the Member. The limestones are micritic, sometimes marly. Their layers range in thickness from a dozen to 60 cm. There are no calcareous concretions. At the top part, horizontally laminated limestones occur commonly; they are 2 to 4 meters thick.

FAUNA AND AGE

Myczyński (1976) described from Pimienta Member several ammonite species and assigned this unit to the upper part of Middle Oxfordian and perhaps also to the lower part of Upper Oxfordian. The limestones contain also poorly preserved planktic forams (Torre 1972—1975) and *Globochaete alpina* Lombard (*cf.* Seiglie 1961).

LATERAL EXTENT, THICKNESS, AND CORRELATION

The Pimienta Member occurs in all tectonically undisturbed sections of Jagua Formation in the Sierra de los Organos. It is absent from all the other stratigraphic sequences of Cordillera de Guaniguanico. The Member attains sometimes up to 60 m in thickness (*cf.* Myczyński 1976, Figs 5—7). In the Sierra del Rosario, the upper part of Francisco Formation and the lowermost part of Artemisa Formation are time equivalent to the Pimienta Member (Kutek & *al.* 1976, Myczyński 1976, Wierzbowski 1976).

FRANCISCO FORMATION

HISTORY

This formation has been distinguished in the Sierra del Rosario (Pszczółkowski *in:* Kutek & *al.* 1976). The deposits assigned to the Francisco Formation were described previously as transitional between San Cayetano and Artemisa Formations (Pszczółkowski 1971b). The type area has been determined by Cinco Pesos (Pszczółkowski *in:* Kutek & *al.* 1976, Figs. 2—3).

LITHOLOGY

The Francisco Formation consists of clayey and silty shales, micritic limestones, and thin sandstone intercalations. Sometimes there are calcareous concretions in the shales. In the type area, a volcanic rock half a meter thick occurs within laminated limestones intercalated with sandstones (Pszczółkowski *in* Kutek & *al.* 1976, Fig. 2, outcrop 3). According to Dr. A. Nowakowski (*personal information*), this rock is close to basalt with alkalic feldspars replaced by albite.

FAUNA AND AGE

The deposits contain some ammonites, rare bivalves, fish and plant remains. Sporadically, aptychi and spores *Globochaete* occur. The ammonites indicate the upper part of Middle Oxfordian (Kutek & Wierzbowski *in:* Kutek & *al.* 1976) and perhaps also the lower part of Upper Oxfordian (Myczyński 1976).

CORRELATION AND THICKNESS

The stratigraphic position of Francisco Formation appears close to that of most deposits of Jagua Formation (*cf.* Kutek & *al.* 1976, Myczyński 1976, Wierzbowski 1976). In places, the Francisco Formation is, however, replaced by sandstones and shales of the uppermost part of San Cayetano Formation. The Francisco Formation ranges from 0 to 25 m in thickness.

VIÑALES GROUP

HISTORY

The Vinales Group has been formally recognized by Herrera (1961) who has also distinguished two subunits, *viz.* Guasasa and La Mina Formations. However, De Golyer (1918) had previously proposed the name Vinales Limestone for the carbonate rocks building up mogotes of the central part of Sierra de los Organos and the Sierra del Rosario range. De Golyer (1918) did not determine any type section for this widely defined formation. Later on, Hatten (1957) restricted this definition to carbonate rocks occurring in the Sierra de los Organos, while Herrera (1961) raised Vinales Formation to the rank of a group.

In this study, the Vinales Group comprises also some formations occurring in the sequences of Sierra del Rosario. That is the definition appears close to the original definition given by De Golyer (1918), while it is wider than that given by Herrera (1961).

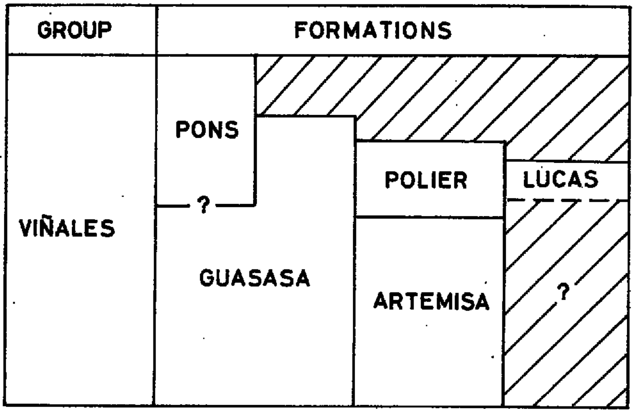
LITHOLOGY

The Vinales Group consists of limestones, sometimes with minor amounts of other sedimentary rocks (sandstones, cherts, and shales).

SUBDIVISION

The Vinales Group comprises 5 formations. These are: Guasasa, Artemisa, Polier, Lucas, and Pons Formations (Table 2).

Table 2
Viñales Group subdivision



AGE

In the Sierra de los Organos, the Vinales Group deposits range in age since the Oxfordian through Early Paleocene. In the sequences of Sierra del Rosario, the accumulation of Vinales Group deposits lasted since the Late Oxfordian through Albian time.

BOUNDARIES

The lower boundary of Vinales Group is equivalent to the lower boundary of Guasasa Formation in the Sierra de los Organos, and to the lower boundary of Artemisa Formation in the southern and northern sequences of Sierra del Rosario. The upper boundary of Vinales Group is marked by the upper boundaries of Guasasa and Pons Formations in the Sierra de los Organos, Artemisa Formation in the southern sequence of Sierra del Rosario, Polier Formation in the northern sequence of Sierra del Rosario, and Luças Formation in Quinones sequence of the Sierra del Rosario.

LATERAL EXTENT AND THICKNESS

The Vinales Group deposits are represented all over the Cordillera de Guaniguanico. They have insofar not been reported from any other facies-structural zone of Pinar del Río Province. The maximum (measured) thickness of Vinales Group attains 800 m.

CORRELATION

Limestones close to the deposits of Vinales Group have been recorded in Habana Province. These are the limestones of Martin Mesa Formation of Herrera (1961), containing the Tithonian to Lower Cretaceous microfauna (*cf.* Khudoley & Meyerhoff 1971).

The Tithonian to Lower Cretaceous limestones, mostly pelagic ones (Khudoley & Meyerhoff 1971), of Central Cuba were called Vinales Limestone or Aptychus beds (*cf.* Bermúdez & Hoffstetter 1959, Bermúdez 1961). These deposits can also be correlated with Vinales Group.

GUASASA FORMATION

HISTORY

Guasasa Formation has been distinguished by Herrera (1961) who determined its type area in the Sierra la Guasasa. Hatten (1957) had previously described and defined the Vinales Formation; however, the latter lithostratigraphic unit could not be accepted because of the existence of Vinales Group.

LITHOLOGY

The Guasasa Formation consists of massive and bedded limestones. The limestones are gray to black, sometimes dolomitized. There are chert nodules in the lower part, and chert intercalations at the top of the Formation. The limestones are variable in microfacies.

At the base of Guasasa Formation, a limestone breccia occurs in some sections. Some authors regarded it as tectonic in origin (Knipper & Puig-Rifa 1967). However, one may claim that it originated owing to purely sedimentary processes (*cf.* Hatten 1957, Meyerhoff *in:* Khudoley & Meyerhoff 1971). It often includes clasts of laminated limestones of Pimienta Member of Jagua Formation. In Mogote la Mina section (Fig. 5B), a similar breccia occurs within the uppermost layers of Jagua Formation.

SUBDIVISION

The Guasasa Formation has been subdivided into 5 members. These are: San Vicente, El Americano, Tumbadero, Tumbitas, and Infierno Members (Table 3). Transitional deposits occur sometimes between San Vicente and El Americano Members (Fig. 5A).

AGE

The Formation can be assigned to the Upper Oxfordian to Cenomanian or ?Turonian. The biostratigraphic reasons for such an age assignment are given below.

Table 3

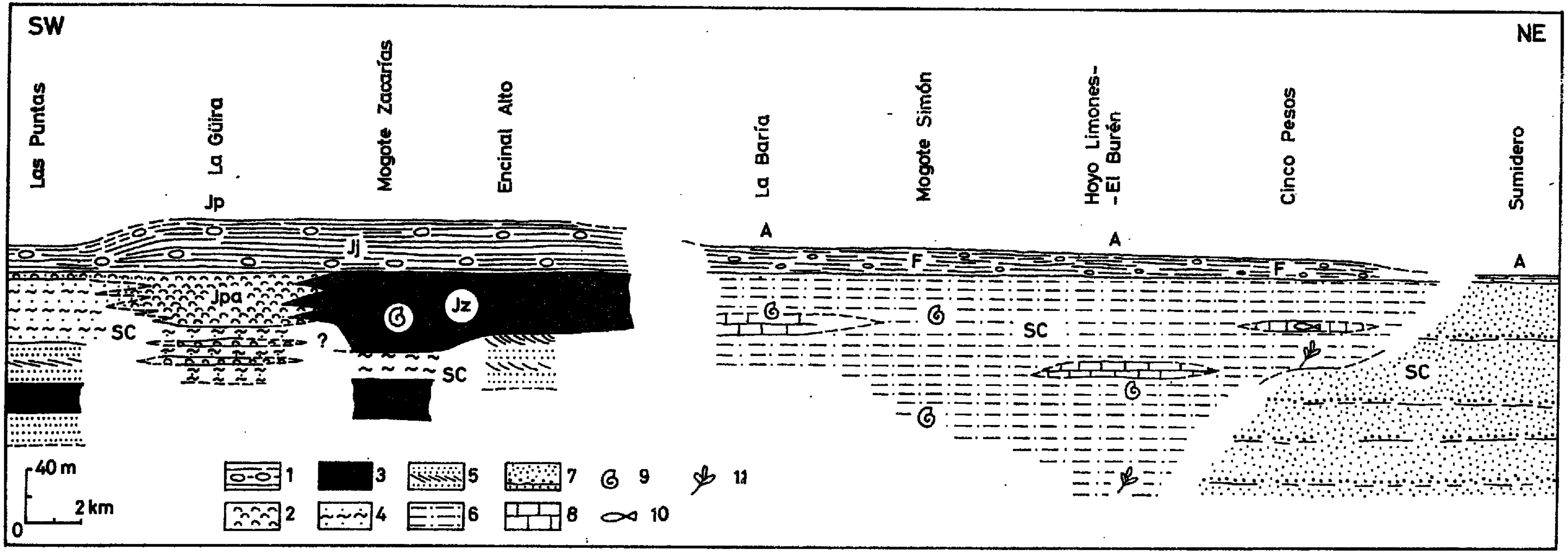
Guasasa and Artemisa Formations subdivision

GUASASA FORMATION	ARTEMISA FORMATION	
Infierno Member		
Tumbitas Member	Sumidero Member	
Tumbadero Member		
El Americano Member		La Zarza Member
San Vicente Member		

LATERAL EXTENT AND THICKNESS

The Guasasa Formation occurs exclusively in the Sierra de los Organos. It ranges in thickness from 300 meters in Ancón tectonic unit to 800 meters in Vinales unit.

Composite section through the Oxfordian deposits in various tectonic units of the Cordillera de Guaniguanico; Las Puntas to Encinal Alto (Sierra de los Organos), La Baría to Sumidero (Sierra del Rosario)



SC — San Cayetano Formation (the uppermost part), *Jpa* — Pan de Azúcar Member of Jagua Formation, *Jz* — Zacarías Member of Jagua Formation, *Jj* — Jagua Vieja Member of Jagua Formation, *F* — Francisco Formation, *A* — Artemisa Formation

1 shales with calcareous concretions and limestone intercalations, *2* coquinas and bioclastic limestones, *3* clayey shales, *4* sandstones and shales, *5* sandstones, siltstones, and shales, *6* thin-bedded sandstones, mudstones and shales, *7* thick-bedded sandstones, frequently graded, *8* limestone intercalations within San Cayetano Formation in the Sierra del Rosario, *9* ammonites, *10* fish remains, *11* plant remains

SYNONYMY AND CORRELATION

The Guasasa Formation is synonymous with Vinales Formation of Hatten (1957, not 1967). Its partial time equivalents are the Artemisa, Polier, Lucas, Buenavista, and Sierra Azúl Formations in the Sierra del Rosário, and Pons Formation in the sequence of Sierra de los Organos (Tables 1—2).

SAN VICENTE MEMBER

HISTORY

This lithostratigraphic unit has been recognized by Herfera (1961). It occurs in the lower part of Guasasa Formation. The type section is at the eastern margin of Sierra de Vinales.

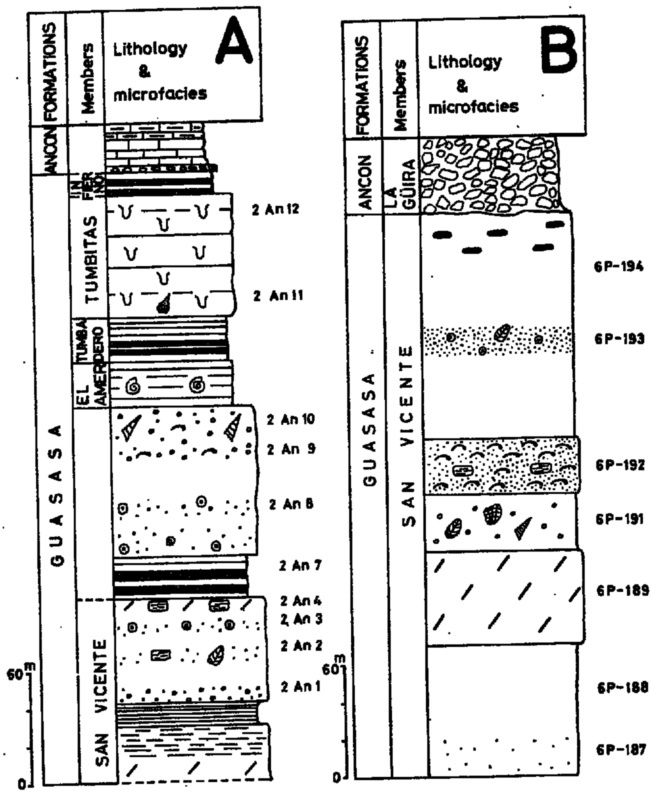


Fig. 6

A — Lithologic-microfacies section of Guasasa Formation at Valle del Ancón (*cf.* section *3* in Text-fig. 2*A*); *EL AMER.* — El Americano Member; for other explanations *see* Text-fig. 5*A*

B — Lithologic-microfacies section of Guasasa Formation in the western Sierra de la Güira (*cf.* section *8* in Text-fig. 2*A*); for explanation of the symbols *see* Text--fig. 5*A*

LITHOLOGY

San Vicente Member consists of massive or thick-bedded, light-gray to black limestones; they are usually strongly karstified. In places, the limestones are horizontally stratified. Often, they are partly or totally dolomitized. There are also gray to black chert nodules and lenses. Micritic limestones dominate in the lower part of the Member, while calcarenites occur usually in the top part (Figs 5—7). At the top, there are also well-bedded, horizontally laminated limestones up to a dozen meters thick (Figs 6A, 7A). The San Vicente Member includes a sedimentary limestone breccia separating the massive limestones of Guasasa Formation from Jagua Formation. The description of the basic microfacies types of San Vicente Member is given below.

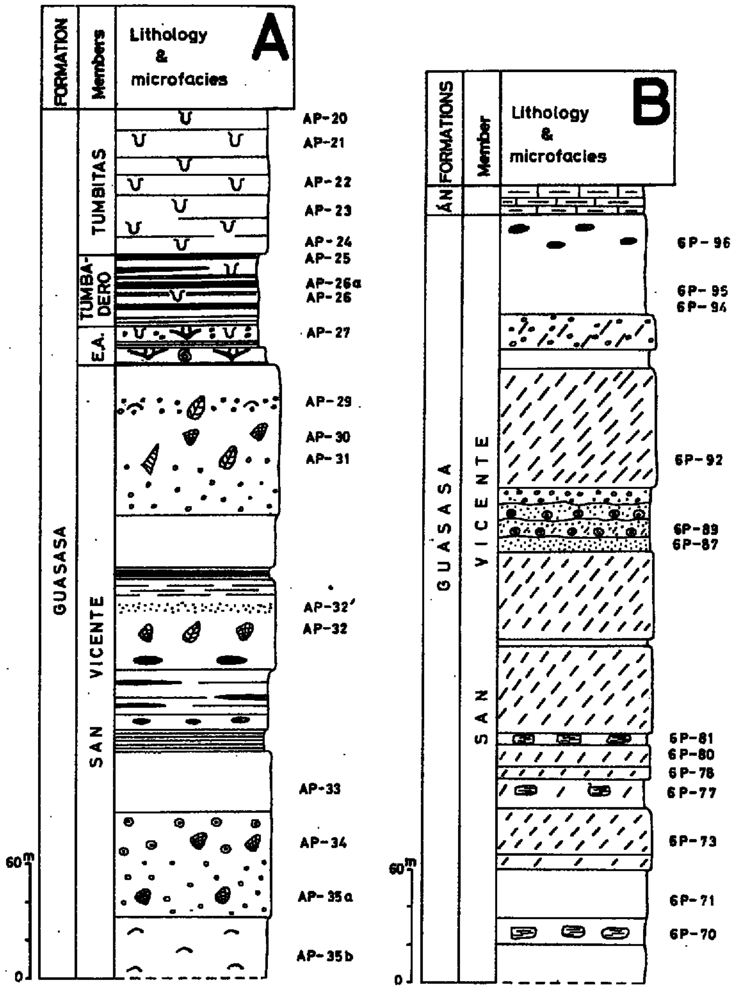


Fig. 7

A — Lithologic-microfacies section of Guasasa Formation at San Vicente (*cf.* section 5 in Text-fig. 2*A*); *E.A.* — El Americano Member; for other explanations *see* Text-fig. 5*A*

B — Lithologic-microfacies section of Guasasa Formation at El Abra (*cf.* section 2 in Text-fig. 2*A*); *AN* — Ancón Formation; for other explanations *see* Text-fig. 5*A*

FAUNA AND AGE

The limestones of San Vicente Member comprise bivalves, gastropods (*Nerinea* sp. among others — Myczyński *in:* Pszczółkowski & al. 1975), algal debris, echinoid spines, and benthic forams (Textulariidae among others). As judged from its relation to the adjacent lithostratigraphic units (Jagua Formation and El Americano Member), the San Vicente Member can be assigned to the Late Oxfordian to earliest Tithonian (*cf.* Table 4).

Table 4

Correlation of the Upper Jurassic to Lower Cretaceous lithostratigraphic units of the Cordillera de Guaniguanico

Tintinnid zonation follows Remané (1971), Allemann & al. (1971), and in part Kreisel & Furrazola-Bermúdez (1971); the recent studies on ammonite faunas (Houša 1974b, Myczyński in: Pszczółkowski & *al.* 1975, Kutek & Wierzbowski in: Kutek & *al.* 1976, Myczyński 1977) are taken into account; the Tithonian and Berriasian boundary is traced accordingly to the solution 2 of the Colloque sur la limite Jurassique-Crétacé (Lyon-Neuchâtel 1973)

SYSTEM	Stage	Tintinnid zones	Sequence of Sierra de los Organos: Formations	Members	Southern sequence of S. del Rosario: Formations	Members	Northern sequence of S. del Rosario: Formations	Members	Quiñones sequence: Formations	Members
CRETACEOUS	Albian-Aptian		PONS ?	Infierno	BUENAVISTA	Sabanilla	BUENA-VISTA	Sabanilla, Roble	SIERRA AZUL	
CRETACEOUS	Barremian-Hauterivian			Infierno	BUENAVISTA	Sabanilla	POLIER		LUCAS	
CRETACEOUS	Valanginian	Calpionellites darderi	GUASASA	Tumbitas	ARTEMISA	Sumidero	ARTEMISA	Sumidero		
CRETACEOUS	Berriasian Upper	Calpionellopsis	GUASASA	Tumbitas / Tumbadero	ARTEMISA	Sumidero	ARTEMISA	Sumidero		
CRETACEOUS	Berriasian Lower	Calpionella	GUASASA	Tumbadero	ARTEMISA	Sumidero / La Zarza	ARTEMISA	La Zarza		
JURASSIC	Tithonian Upper	Crassicollaria	GUASASA	El Americano	ARTEMISA	La Zarza	ARTEMISA	La Zarza		
JURASSIC	Tithonian Middle	Chitinoidella	GUASASA	El Americano	ARTEMISA	La Zarza	ARTEMISA	La Zarza		
JURASSIC	Tithonian Lower		GUASASA	El Americano	ARTEMISA	La Zarza	ARTEMISA	La Zarza		
JURASSIC	?—Kimmeridgian-Upper Oxfordian		GUASASA	San Vicente	ARTEMISA	La Zarza, San Vicente				

LATERAL EXTENT AND THICKNESS

The San Vicente Member has been recorded in all complete sections of the sequence of Sierra de los Organos, and in some sections of the southern sequence of Sierra del Rosario. The thickness ranges from 300 to 650 m in the Sierra de los Organos, the upper boundary of this lithostratigraphic unit being sometimes erosional (Figs 5*B*, 6*B*, 7*B*). In the Sierra del Rosario, the San Vicente Member does not exceed some tens meters in thickness.

SYNONYMY AND CORRELATION

San Vicente Member is synonymous with Vinales Formation as conceived by Judoley & Furrazola-Bermúdez (1965, 1968), Hatten (1967), Khudoley & Meyerhoff (1971) and Meyerhoff & Hatten (1974). Its time equivalent is the lower part of La Zarza Member of Artemisa Formation (Tables 3—4).

EL AMERICANO MEMBER

HISTORY

This lithostratigraphic unit has been recognized in the upper part of Guasasa Formation (Houša & Nuez 1972). The type section is at Hacienda El Americano, eastern margin of the Sierra de los Organos.

LITHOLOGY

The deposits comprise well-bedded, granular, dark-gray to black limestones, sometimes with intercalations of shaly marly limestones. Dolomitic limestones and dolomites do also occur. In the upper part of the Member, there are some sedimentary discontinuities indicating post-lithification erosion (Houša 1974b). Such discontinuities occur also at the top of San Vicente Member in the type section of El Americano Member.

FAUNA AND AGE

Numerous ammonites indicate the Tithonian age of El Americano Member (Houša & Nuez 1972, Houša 1974b). The co-occurrence of ammonite genera *Mazapilites* and *Pseudolissoceras* near the lower limit of the Member, and the lack of *Hybonoticeras* may indicate that the El Americano Member does not comprise the lowermost part of the Tithonian (Houša 1974b).

In the middle part of the Member, there are tintinnids *Chitinoidella* spp. (Fig. 8, sample *6P-128*; Table 5). These microfossils were previously recorded by Kreisel & Furrazola-Bermúdez (1971) in Valle del Ancón section in the Middle Tithonian deposits (? El Americano Member). In the upper part of the Member, there are tintinnids *Crassicollaria* and *Calpionella*, and abundant fragments of *Saccocoma* sp. The upper boundary of El Americano Member (Table 4) approximates the boundary between Crassicollaria and Calpionella Zones (*cf.* Remané 1971).

Apart from the ammonites, the deposits of El Americano Member contain also some brachiopods, gastropods, reptile bone fragments (Houša 1974b), bivalves *Buchia* sp. (Myczyński *in:* Pszczółkowski & *al.* 1975), fish teeth and vertebrae (Fig. 8).

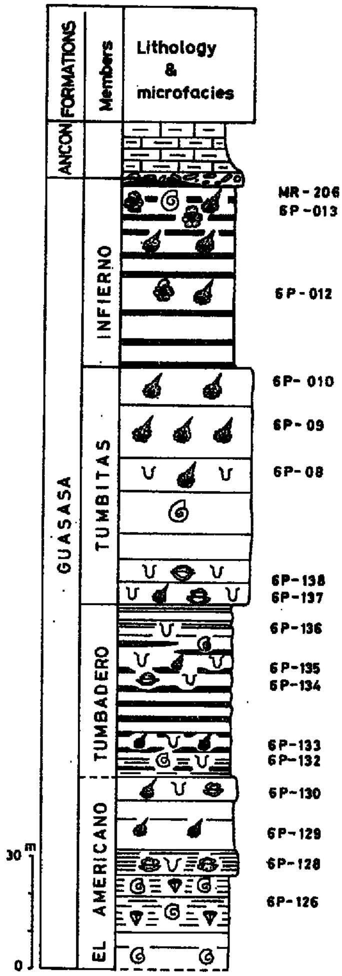


Fig. 8

Lithologic-microfacies section of the upper part of Guasasa Formation in the Sierra del Infierno (*cf.* section *1* in Text-fig. 2*A*); for explanation of the symbols *see* Text-fig. 5*A*

LATERAL EXTENT AND THICKNESS

The El Americano Member occurs but in the sequence of Sierra de los Organos, namely in Infierno and Viñales tectonic units. In the type section, the thickness is 40 m (Houša & Nuez 1972). The thickness ranges elsewhere from 20 to 45 m.

CORRELATION

In part, the El Americano Member is time equivalent to the upper part of La Zarza Member of Artemisa Formation of the southern and northern sequences of Sierra del Rosario (Table 3).

TUMBADERO MEMBER

HISTORY

This member has been distinguished for the first time by Herrera (1961) in the upper part of Guasasa Formation. It was thereafter redefined by Houša & Nuez (1972) to include the limestones between El Americano and Tumbitas Members.

Table 5

Tintinnids and forams of Guasasa Formation

The table contains data from thin sections studied by Torre (1972—1975): *6P-08, 6P-029, 6P-030, 6P-033, 6P-034, 6P-138, 6P-646, MR-205, MR-211*; and by the present author: *2A4, 2A5, AP-20—27, 2An11, 2An12, 6P-26, 6P-26a, 6P-27, 6P-128, 6P-130, 6P-132-135, 6P-137, MR-206* (*cf.* sample numbers in Text-figs 5—9)

Members	El Americano			Tumbadero			Tumbitas				Infierno		
Taxa / Sections	Sierra de Viñales	San Vicente	Sierra del Infierno	San Vicente	Hacienda El Americano	Sierra del Infierno	San Vicente	Valle del Ancón	Hacienda El Americano	Sierra del Infierno	Sierra del Infierno	Sierra de Pan de Azúcar	Laguna de Piedra
Tintinnids: sample symbols	2A	AP	6P	AP	6P	6P	AP	2An	6P	6P	MR	6P	MR
Chitinoidella bermudezi /Furrazola-Bermúdez/ Chitinoidella cf. cubensis /Furrazola-Bermúdez/ Chitinoidella cf. boneti Doben			128 128 128										
Calpionella alpina Lorenz Calpionella cf. alpina Lorenz		27	130	26	27	135 133 134		11		137			
Calpionella elliptica Cadisch Calpionella sp.	4 5	27			27	133 133	24 23			138			
Crassicollaria brevis Remané Crassicollaria sp.	4 5	27	130			132							
Tintinnopsella carpathica /Murg. & Filip./					27		24 22 20		26				
Tintinnopsella cf. carpathica /Murg. & Filip./							23	11 12		137			
Tintinnopsella longa /Colom/									26	137 138			
Tintinnopsella sp.				26a 26		134	23						
Remaniella cadischiana /Colom/ Remaniella dadayi Knauer				26		134	21	11	646				
Calpionellopsis simplex /Colom/ Calpionellopsis cf. simplex /Colom/						136	23	11	26	137 138			
Calpionellopsis oblonga /Cadisch/				26a			23 22			?138 08			
Calpionellopsis cf. oblonga /Cadisch/ Calpionellopsis sp.				25			24 20		646	137			
Calpionellites darderi /Colom/							22 20	12		08			
Calpionellites cf. darderi /Colom/							21						
Amphorellina sp.				26a				12					
Foraminifers:													
Hedbergella delrioensis /Carsey/ Hedbergella cf. delrioensis /Carsey/ Hedbergella sp.											206	?029 033 ?034 030	211
Planomalina buxtorfi /Gandolfi/											206		
Globigerinelloides sp.											206		
Ticinella sp.											?206		
Rotalipora sp.											?206 205		
Heterohelix cf. moremani /Cushman/ Heterohelix sp.													211 211

LITHOLOGY

This lithostratigraphic unit consists of well-bedded (0.1 to 0.3 m), often laminated, micritic limestones and calcilutites with black chert intercalations. Hence, Tumbadero Member appears quite distinct from the adjacent members.

FAUNA AND AGE

Ammonites occur but sporadically. In Sierra del Infierno section, they indicate the Tithonian age of the lowermost part of the Member (Myczyński *in:* Pszczółkowski & *al.* 1975). In the same section, tintinnid assemblage is dominated by *Calpionella* spp., while *Calpionellopsis* spp. appear but at the top (some 10 m below the upper boundary of the Member; *cf.* Fig. 8, Table 5). However, in San Vicente section, *Calpionellopsis* appears already in the middle of Tumbadero Member (Fig. 7A, Table 5). Thus, one may conclude that the Tumbadero Member represents Calpionella Zone and the lower part of Calpionellopsis Zone, that is the Berriasian.

LATERAL EXTENT AND THICKNESS

The Tumbadero Member occurs in all studied sections of Guasasa Formation, except of those reduced tectonically or eroded previously to the Late Paleocene time. In Hacienda El Americano section, the thickness of Tumbadero Member approximates 38 m (Houša & Nuez 1972), while it ranges elsewhere from 20 to 50 meters.

TUMBITAS MEMBER

HISTORY

This lithostratigraphic unit has been distinguished in the upper part of Guasasa Formation, above the deposits of Tumbadero Member (Houša & Nuez 1972). The type section is at Hacienda El Americano.

LITHOLOGY

The Tumbitas Member consists of thick-bedded, compact, light-grey micritic limestones, with some thin intercalations of darker limestones. The deposits are often mottled due to bioturbation.

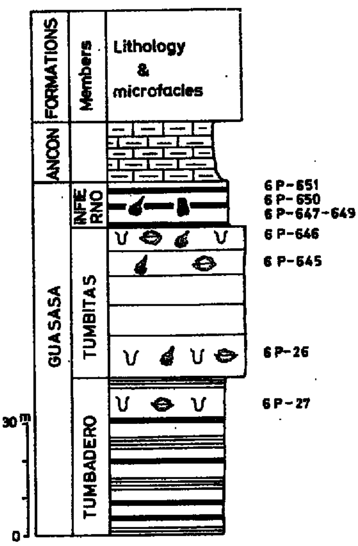


Fig. 9

Lithologic-microfacies section of the upper part of Guasasa Formation at Hacienda El Americano (*cf.* section 7 in Text-fig. 2*A*); for explanation of the symbols *see* Text-fig. 5*A*

FAUNA AND AGE

Ammonites are poorly preserved and occur very rarely. However, tintinnids are abundant (Figs 7*A*, 8, 9); they have been identified by Torre (1972—1975) and the present author. In the lower part of the Member, there are: *Calpionellopsis oblonga* (Cadisch), *C. simplex* (Colom), *Calpionella alpina* Lorenz, *Remaniella dadayi* Knauer, *Tintinnopsella carpathica* (Murgeanu & Filipescu), and *T. longa* (Colom). In the upper part, there are: *Calpionellites darderi* (Colom), *Calpionellopsis oblonga* (Cadisch), *Tintinnopsella carpathica* (Murgeanu & Filipescu), *Remaniella cadischiana* (Colom), and *Amphorellina* sp. Furthermore, *Stomiosphaera* cf. *moluccana* Wanner, *Globochaete alpina* Lombard, and *Nannoconus* sp. do also occur.

Calpionellites darderi (Colom) appears commonly in the middle of Tumbitas Member, and abundantly at the top part. *Calpionellites coronata* Trejo and *C. caravacaensis* Allemann (*cf.*

Allemann & Trejo 1975) may occur among the poorly preserved specimens but the investigated material could not be identified more precisely than as *Calpionellites* sp. As judged from the modern studies, *C. darderi* (Colom) appears in the Lower Valanginian well above the Berriasian boundary (Allemann *in:* Allemann & *al.* 1975) or even in the Middle Valanginian (Trejo 1975).

As judged form the tintinnid assemblage, the Tumbitas Member ranges in age since the latest Berriasian through Early Hauterivian (*cf.* Remané 1971). The boundary between Calpionellopsis and Calpionellites darderi Zones is situated in the middle part of Tumbitas Member (Table 4).

LATERAL EXTENT AND THICKNESS

The deposits of Tumbitas Member have been recorded in Vinales and Infierno tectonic units. In the type section, the thickness is 30 m (according to Houša & Nuez 1972) or 40 m (according to the present author), while it attains elsewhere up to 80 m.

INFIERNO MEMBER

NAME

This is a new member proposed by Myczyński & Pszczółkowski (*in:* Pszczółkowski & *al.* 1975) for limestones and cherts occurring at the top of Guasasa Formation; its description has insofar not been published. The name has been derived from the limestone range Sierra del Infierno, west to Vinales (Fig. 10*A*).

At Hacienda El Americano, Houša (1974b) distinguished above the deposits of Tumbitas Member his Mina Member, as equivalent to Herrera's (1961) Mina Formation of Pons area. However, the Mina Member cannot be accepted since the lithology differs from that ascribed to Mina Formation; moreover, the latter unit cannot be accepted either (*see* below).

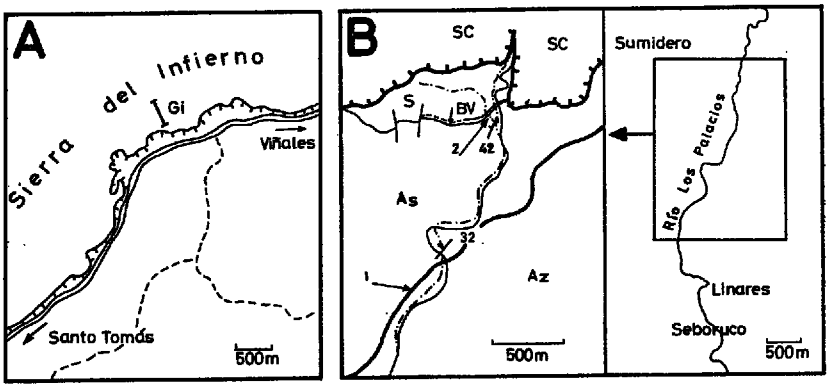


Fig. 10

A — Location map of the type section of Infierno Member of Guasasa Formation; *Gi* — type section of Infierno Member

B — Location map of the type section of Sumidero Member of Artemisa Formation; *1* type section beginning, *2* type section end

Az — La Zarza Member of Artemisa Formation, *As* — Sumidero Member of Artemisa Formation, *SC* — San Cayetano Formation, *BV* — Buenavista Formation, *S* — serpentinites

TYPE SECTION

The type section of Infierno Member is in the Sierra del Infierno, north to the road between Viñales and Pons localities (Fig. 10*A*). The section starts on some 50 meters above the valley bottom, and its topographic coordinates are 212.750 and 310.550.

Limestones of Tumbitas Member are overlaid by well-bedded (0.01 to 0.4 m) gray to dark-gray, micritic limestones intercalated with light-gray, micritic limestones and black cherts. These deposits represent the lower part of Infierno Member. In the lower part, the chert intercalations are not common. The upper part of the Member consists of dark-gray to black, micritic limestones with numerous intercalations of black cherts. The bed thickness remains approximately constant throughout the section. Some limestones appear mottled. The total thickness of Infierno Member is 50 m. Ammonites occur but sporadically and are poorly preserved. Thin sections contain mostly radiolarians (Pl. 5, Fig. 5) and a few forams.

In the type section, the deposits of Infierno Member are undisturbed tectonically.

BOUNDARIES

The type of the lower boundary of Infierno Member is at the base of the first chert intercalation in the type section, above the limestones of Tumbitas Member. The transition of the latter Member into the former one is gradational.

The type of the upper boundary of Infierno Member is in the same section, at the base of nearly black micritic limestones with numerous cherts often inflected and disturbed. These limestones pass laterally into a breccia typical of La Güira Member of Ancón Formation and hence, cannot be attributed to Infierno Member of Guasasa Formation. East and west to the type section, the deposits of Infierno Member are overlaid directly by limestone and chert-bearing breccias (La Güira Member of Ancón Formation; Paleocene). The upper boundary of Infierno Member is erosional.

AGE

The planktic forams recorded at the top of Infierno Member (Torre 1972—1975) indicate the Albian to ?Lower Turonian. The lower part of the Member contains a few forams (*Hedbergella* sp.) but any tintinnids have not been found therein. These biostratigraphic reasons and the lithostratigraphic position of Infierno Member permit to assign this unit to the Hauterivian to ?Lower Turonian (Myczyński & Pszczółkowski *in:* Pszczółkowski & *al.* 1975).

LATERAL EXTENT AND THICKNESS

The Infierno Member occurs in the Sierra de los Organos, namely in the sections of Viñales and Infierno tectonic units. In places, the upper part of the Member contains less cherts than it does in the type section. The bed thickness may vary. Some intercalations of shaly limestones may appear. The total thickness ranges from 0 to 50 m.

CORRELATION

The Infierno Member is facies equivalent and in part time equivalent to the Pons Formation of Valle de Pons tectonic unit, the sequence of Sierra de los Organos.

PONS FORMATION

HISTORY

The Pons Formation has been distinguished by Hatten (1957) in the Valle de Pons, Sierra de los Organos. The deposits attributed to this formation are overlaid by very similar limestones of Penas Formation of Hatten (1957). Later on, Herrera (1961) described from the same area his Mina Formation, with lithological characteristics very close to that of Hatten's (1957) Pons Formation.

Actually, the Pons (or Mina) Formation does not include significant amount of terrigenous rocks, as claimed by Hatten (1957) and Herrera (1961); moreover, Pons and Penas Formations differ in lithology but insignificantly (Piotrowska *in:* Pszczółkowski & *al.* 1975). Hence, Pons

Formation is here considered as including all the limestones with cherts, assigned by Hatten (1957) to either Pons, or Penas Formations. The name Mina Formation has not been accepted, as it was introduced later and the original description by Herrera (1961) appears insufficient.

TYPE SECTION

The type section of redefined Pons Formation is in the bed of Las Piedras River, 1.5 km south to Pons (Piotrowska *in*: Pszczółkowski & *al.* 1975). There are exposed in the river bed well-bedded, light-gray to nearly black, micritic limestones interbedded with cherts. In the lower part, thick-bedded, light-gray or mottled limestones are dominant; however, dark-gray varieties do also occur. In the upper part (= Peñas Formation of Hatten 1957), the beds are thinner and the rocks become dark-gray to black, although light-gray layers do also occur. Chert intercalations are more common in this part of the Formation. The total thickness of Pons Formation is 200 meters in its type section.

BOUNDARIES

There is no type of the lower boundary of Pons Formation, as its base is not exposed in the Valle de Pons. The type of the upper boundary is in the type section, at the top of the light-gray limestones with chert intercalations, bordering sharply upon marly limestones of Ancón Formation.

AGE

On the basis of planktic forams, Hatten (1957) assigned his Pons and Penas Formations to the Albian to Upper Campanian. New micropaleontological data (Torre 1972—1975) indicate that redefined Pons Formation ranges in age since the ?Hauterivian through Danian. The limestones and cherts of Pons Formation contain also abundant radiolarians; at the base, poorly preserved tintinnids and *Nannoconus* occur (Torre 1972—1975).

LATERAL EXTENT

The Pons Formation has been reported from the Pons and Pica Pica Valleys, central part of the Sierra de los Organos (Piotrowska *in:* Pszczółkowski & *al.* 1975). The limestones of Pons Formation may also occur in the environs of La Legua, southwestern part of the Sierra de los Organos. The uppermost part of the Formation resembles deposits of Infierno Member of Guasasa Formation.

CORRELATION

The lower part of Pons Formation is time equivalent to the Infierno Member of the Sierra de los Organos. In the southern and northern sequences of Sierra del Rosario, the Polier and Buenavista Formations are time equivalent to Pons Formation (*cf.* Table 1). In Quinones stratigraphic sequence, the Lucas, Sierra Azúl, and Guajaibón Formations represent jointly almost the same age interval as Pons Formation does.

ARTEMISA FORMATION

HISTORY

The Artemisa Formation (Lewis 1932) occurs in the lower part of Vinales Group (Table 2). Herrera (1961) raised this formation to a group rank and subdivided it into Aptychus and Yaya Formations. However, the name Aptychus Formation is not allowed by the rules of lithostratigraphic nomenclature (*cf.* Hedberg & *al.* 1972, Hedberg 1976); moreover, no type section has been designated by Herrera (1961) for this unit. The criteria proposed to recognize Aptychus and Yaya Formations in the field appear insufficient. Therefore, these units cannot be accepted.

The type area of Artemisa Formation has been designated by Cinco Pesos (Judoley & Furrazola-Bermúdez 1968). This was also the type section of Rosario Limestone of Hatten (1957)

LITHOLOGY

The Artemisa Formation comprises well-bedded micritic limestones, calcilutites, calcarenites, and some calcirudites. In places, there are also thin intercalations of radiolarian cherts, and clayey and marly shales. At the base of the Formation, siltstones and fine-grained sandstones occur sporadically.

SUBDIVISION

The Formation is partly subdivided into 3 members (Table 3), *viz.* San Vicente, La Zarza, and Sumidero Members (*cf.* Pszczółkowski 1978a). The San Vicente Member can be recognized but in a few sections (*e.g.* Fig. 11*A*). The La Zarza Member is also absent from some sections of Artemisa Formation.

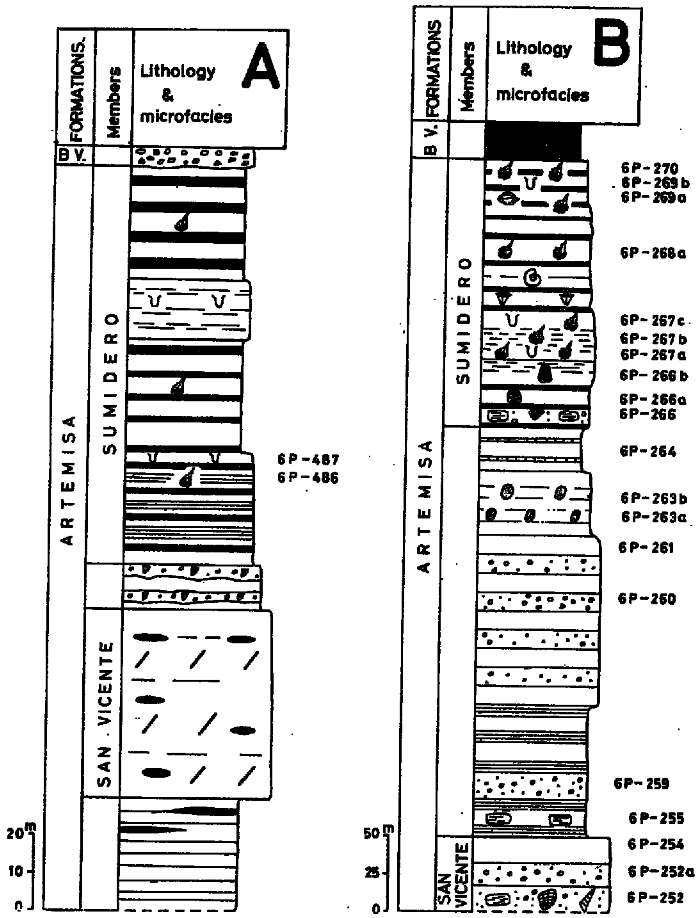


Fig. 11

A — Lithologic-microfacies section of Artemisa Formation at Los Bermejales (*cf.* section *9* in Text-fig. 2*A*); *BV* — Buenavista Formation; for other explanations *see* Text-fig. 5*A*

B — Lithologic-microfacies section of Artemisa Formation at Mil Cumbres de Catalina (*cf.* section *10* in Text-fig. 2*A*); *BV* — Buenavista Formation; for other explanations *see* Text-fig. 5*A*

AGE

As judged from ammonites (Imlay 1942, Judoley & Furrazola-Bermúdez 1968, Houša 1974b, Myczyński *in:* Pszczółkowski & *al.* 1975, Kutek & Wierzbowski *in:* Kutek & *al.* 1976) and tintinnids (Torre 1972—1975), the Artemisa Formation can be assigned to the upper Middle Oxfordian or lower Upper Oxfordian to Hauterivian.

BOUNDARIES

The lower boundary of the Formation is in the type section, at the base of thin-bedded micritic limestones overlying the shales of Francisco Formation. This boundary is exposed by the road towards Bahía Honda, at a slope of the Altos de San Francisco. Despite a slight décollement at the contact, it can be recognized for the type of the lower boundary of Artemisa Formation.

The upper boundary of the Formation is of tectonic nature in the type section. However, it can be observed at El Derrumbadero in La Zarza tectonic unit. The limestones with chert intercalations of Artemisa Formation border upon cherts of Buenavista Formation. In other sections of the southern sequence of Sierra del Rosario, the upper boundary of Artemisa Formation resembles that at El Derrumbadero, except of some places where the deposits of Artemisa Formation border upon red shales (Seboruco-Linares section) or a breccia of Buenavista Formation (*cf.* Pszczółkowski 1978a, Fig. 8). In the northern sequence of Sierra del Rosario, the upper boundary of Artemisa Formation is at the base of the first sandstone layer; the transition to deposits of Polier Formation is gradational therein.

LATERAL EXTENT AND THICKNESS

The Artemisa Formation occurs in the southern and northern sequences of Sierra del Rosario (Pszczółkowski 1978a). In the northern sequence, this formation occurs complete but in a few sections. The thickness is 300 m in the type section, while in some other sections it attains up to 700—800 m.

CORRELATION

The Artemisa Formation is time equivalent to the most of Guasasa Formation (except of Infierno Member; Tables 3—4) and the uppermost part of Jagua Formation (*cf.* Kutek & *al.* 1976, Table 1). The synonymy has been given by Judoley & Furrazola-Bermúdez (1968).

LA ZARZA MEMBER

NAME

The La Zarza Member has been distinguished in the lower part of Artemisa Formation (Pszczółkowski 1978a) but its formal description has insofar not been published. The name is after La Zarza hill, by the road between San Cristobal and Bahía Honda.

TYPE SECTION

The type section is by the road between La Zarza hill and Altos de San Francisco. Thus, it represents the lower part of Artemisa Formation type (*cf.* Judoley & Furrazola-Bermúdez 1968). At the base of the section, there are thin-bedded, gray, micritic limestones with some thin shale intercalations. The limestones contain rare aptychi. Higher in the section, there are similar rocks disturbed tectonically over a dozen or so meters. Then, the lower part of the Member is ended with somewhat thicker beds (0.2 to 0.5 m) of gray, micritic limestones. In the upper part of the Member, there are micritic limestones and calcilutites interbedded with dark-gray to black, bioclastic limestones and coquinas composed of ammonite shells and aptychi. At the top, the layers are up to 0.3 m thick. The total thickness of La Zarza Member attains 150 m in its type section.

BOUNDARIES

The type of the lower boundary of La Zarza Member is in the type section, at the base of micritic limestones overlying shales of Francisco Formation. The latter deposits comprise some thin intercalations of micritic limestones and hence, the transition appears gradational even despite a slight décollement.

The type of the upper boundary is at a turn of the road between La Zarza and Cinco Pesos (*cf.* Kutek & *al.* 1976, Fig. 2), at the top of the last layer of bioclastic limestones underlying light-gray, micritic limestones with chert intercalations assigned to the Sumidero Member. The transition is gradational.

FAUNA AND AGE

The oldest ammonite assemblage of La Zarza Member has been reported from Cinco Pesos section (Kutek & Wierzbowski *in:* Kutek & *al.* 1976). The ammonite genera *Cubasphinctes* and *Mirosphinctes* derived from the lowest layers of La Zarza Member, indicate the upper Middle Oxfordian or lower Upper Oxfordian. A younger ammonite asemblage occurs in the bioclastic and coquinitic limestones of the top part of the Member. It has been recorded in many sections in the southern part of Sierra del Rosario, between La Palma and Cayajabos localities. In some sections, the ammonite-bearing limestones up to 60 m thick occur some 50 to 70 m above the base of La Zarza Member (*e.g.* at Cinco Pesos); while in others, they occur some 70 to 120 m above the Member base. The most typical are the following ammonite genera: *Pseudolissoceras, Butticeras* (this genus has been erected for the Cuban species assigned previously to *Parodontoceras; cf.* Houša & Nuez 1973, 1975), *Corongoceras, Protancyloceras, Dickersonia,* and *Vinalesites.* This assemblage indicates the Tithonian (Imlay 1942, Judoley & Furrazola-Bermúdez 1968, Houša 1974b, Myczyński *in:* Pszczółkowski & *al.* 1975).

The tintinnids reported from the upper part of La Zarza Member at Cinco Pesos (Furrazola-Bermúdez 1965) are attributed to the genus *Chitinoidella,* typical of the Middle Tithonian (Kreisel & Furrazola-Bermúdez 1971). As indicated by the presence of *Calpionella* sp. (Fig. 12, sample *6P-385b;* Table 6), the top part of the Member represents the Upper Tithonian or Berriasian in some sections. The limestones of the upper part of La Zarza Member may contain radiolarians and *Saccocoma* fragments (Furrazola-Bermúdez 1965), some minute fragments of fish and a few reptile bones.

In Seboruco-Linares section (Fig. 13A), the limestones of the top part of La Zarza Member contain *Protancyloceras* and *Vinalesites,* and may already represent the Berriasian, as they are overlaid by Sumidero Member limestones with tintinnids (*cf.* Torre 1972—1975, sample *6P-289*) typical of the Upper Berriasian to Valanginian (Pszczółkowski *in:* Pszczółkowski & *al.* 1975, Myczyński 1977).

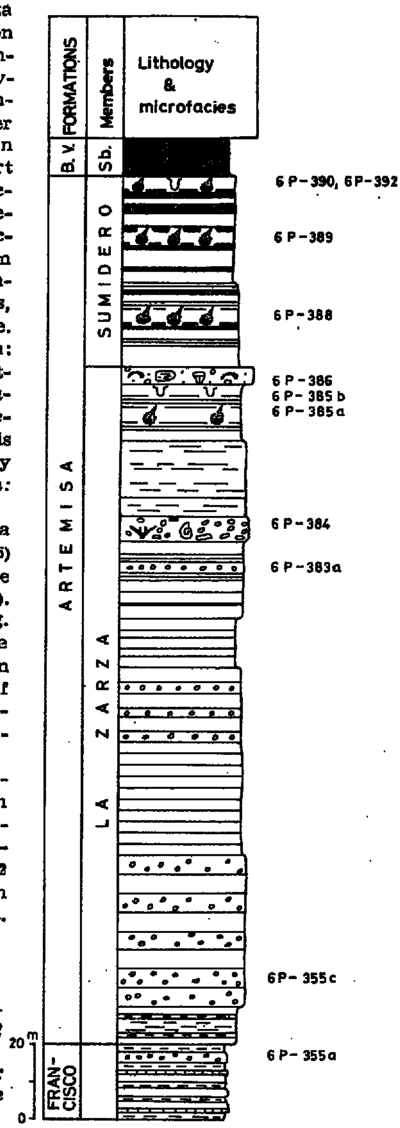


Fig. 12

Lithologic-microfacies section of Artemisa Formation by the locality Sabanilla (*cf.* section *12* in Text-fig. 2*A*)

BV — Buenavista Formation, *Sb.* — Sabanilla Member of Buenavista Formation; for other explanations see Text-fig. 5*A*

The limestones of the lower part of La Zarza Member have insofar not been biostratigraphically dated. They are 50 to 120 m thick and are situated between deposits containing the Oxfordian and Tithonian ammonites.

In general, the La Zarza Member can be attributed to the upper Middle Oxfordian or lower Upper Oxfordian to ?Lower Berriasian; however, the upper boundary is probably heterochronous.

LATERAL EXTENT AND THICKNESS

The La Zarza Member occurs commonly in the northern and southern sequences of Sierra del Rosario. Its best sections are in La Zarza and Cinco Pesos tectonic units (*cf.* Fig. 2A). The litology is variable; in some sections, there are calcarenites and limestone breccias (Fig. 12). The thickness ranges from 80 to 200 m. In the northern sequence, the sections are often incomplete because of tectonic reasons.

CORRELATION

The La Zarza Member is time equivalent to the upper part of Pimienta Member of Jagua Formation, and to the San Vicente, El Americano, and lower Tumbadero Members of Guasasa Formation (Table 4).

SUMIDERO MEMBER

NAME

The Sumidero Member has been distinguished (Pszczółkowski 1976a) in the upper part of Artemisa Formation (Table 3). Its formal description has insofar not been published. The name is after Sumidero hills, west to Los Palacios River, central part of the Sierra del Rosario (Fig. 10B).

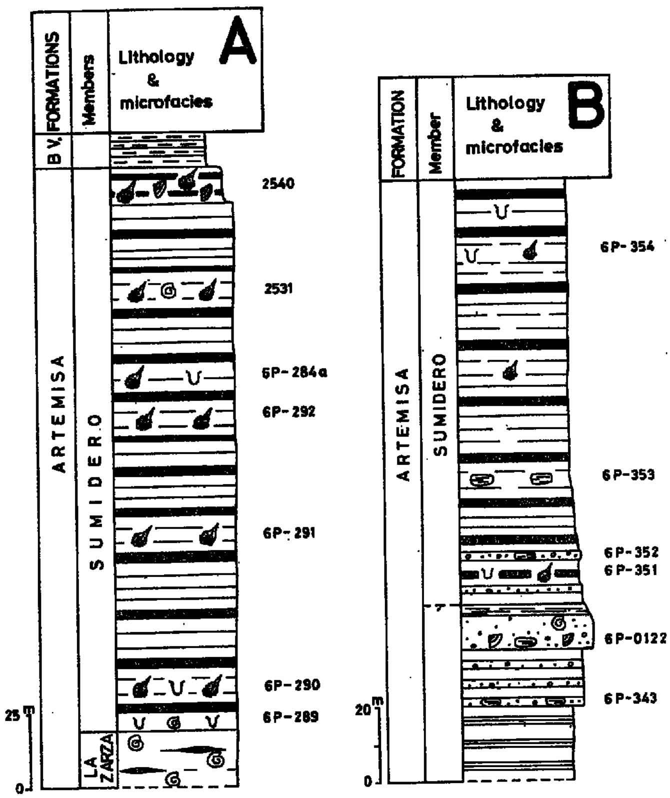


Fig. 13

A — Lithologic-microfacies section of Artemisa Formation at Seboruco-Linares (*cf.* section *11* in Text-fig. 2*A*); *BV* — Buenavista Formation; for other explanations *see* Text-fig. 5*A*

B — Lithologic-microfacies section of Artemisa Formation by the locality Niceto Perez (*cf.* section *13* in Text-fig. 2*A*); for explanation of the symbols *see* Text-fig. 5*A*

TYPE SECTION

The type section is in the valley of Los Palacios River, south to Sumidero hills, and north to the localities Seboruco and Linares (Fig. 10*B*). The section (Fig. 13*A*) starts on at the western slope of the valley, some meters above the terrace; its topographic coordinates are 260.100 and 323.640. The limestones of La Zarza Member are overlaid by light-brown and rose, micritic limestones with small unidentifiable ammonites. There are also radiolarians and tintinnids (Torre 1972—1975, samples *6P-289* and *6P-290*). Some meters above the base of these limestones, light-gray limestone intercalations and thin cherts appear. Higher in the section, there are gray to blue-gray, micritic limestones with thick intercalations of radiolarian cherts. The latter set is fairly thick and at the top, there are some layers of gray-violet, mottled limestones with a few poorly preserved ammonite imprints. There are also some horizontally laminated limestones in this part of the section. The type section of Sumidero Member ends with thin-bedded, dark-gray, micritic limestones intercalated with cherts. These limestones contain abundant aptychi and a few poorly preserved ammonite imprints. There are also abundant calcified radiolarians. The whole section attains some 190 m in thickness.

BOUNDARIES

The type of the lower boundary of Sumidero Member is in the type section, at the base of the light-brown and pink, micritic limestones. The boundary is sharp but without any suggestion of sedimentary discontinuity.

The type of the upper boundary is in the same section, at the top of the thin-bedded, micritic limestones with chert intercalations, bordering upon red shales of Buenavista Formation.

FAUNA AND AGE

The deposits of Sumidero Member contain some tintinnids (Torre 1972—1975) and a few ammonites (Myczyński 1977). At the base of the type section (Fig. 13*A*), there are tintinnids (samples *6P-289* and *6P-290*; Table 5) indicating the Upper Berriasian to Valanginian (*cf.* Remané 1971, Trejo 1975). In Mil Cumbres de la Catalina section (Fig. 11*B*), there are at the base tintinnids (sample *6P-267a, c*; Table 6) indicating Calpionellopsis Zone, that is the Upper Berriasian to Lower Valanginian. In Los Bermejales section (Fig. 11*A*, Table 6), there are no tintinnids at the base of the Member, except of *Calpionella alpina* Lombard which may indicate the Lower Berriasian. Tintinnids occurring at the top of Sumidero Member (Figs 11*B*, 12, 13 *B*; Table 6) indicate the Upper Berriasian to Lower Hauterivian.

Table 6
Tintinnids of Artemisa and Polier Formations

FORMATIONS	ARTEMISA								POLIER
Members	La Zarza	Sumidero							
Tintinnids / Sections	Sabanilla	Los Bermejales	Mil Cumbres de la Catalina	Seboruco-Linares	Nicato Perez	Sabanilla	Bano /Soroa/	Cayajabos	El Mangyal
Calpionella alpina Lorens		6P487						6P846	
Calpionella elliptica Cadisch							6P787		
Calpionella sp.	6P385b	6P487		?6P284a		?6P390			
Crassicollaria brevis Remane					56				
Tintinnopsella cf. carpathica /Murg. & Filip./				6P289					
Tintinnopsella longa /Colom/				6P289				?6P846	
Tintinnopsella sp.				?6P290					
Remaniella cadischiana /Colom/					57				
Calpionellopsis simplex /Colom/			?6P267a	?6P290					6P368a
Calpionellopsis cf. simplex /Colom/			6P267c						
Calpionellopsis oblonga /Cadisch/				6P289	?6P351	?6P388 6P392			6P368a
Calpionellopsis cf. oblonga /Cadisch/						6P390			
Calpionellopsis sp.			6P269a						
Calpionellites darderi /Colom/				?6P289		?6P390			
Calpionellites cf. darderi /Colom/					56				
Calpionellites cf. darderi /Colom/			6P269a						6P368a

Except of samples numbered *56* and *57* identified by Lupu (1973), all the taxa have been identified by Torre (1972—1975); *cf.* sample numbers in Text-figs 11—13

The present author has found in the environs of Soroa a few ammonites attributed to *Thurmanniceras* cf. *novihispanicus* (Imlay) suggesting the Valanginian age for the upper part of Sumidero Member (*cf.* Myczyński 1977). A single specimen assigned to ? *Karsteniceras* cf. *subtilis* (Uhlig) has been found near San Miguel; it might indicate the ? Lower Barremian (*cf.* Myczyński 1977) but so high the stratigraphic range of Sumidero Member has insofar not been confirmed in other sections.

In general, the Sumidero Member ranges in age since the Berriasian to Early Hauterivian.

LATERAL EXTENT AND THICKNESS

The deposits of Sumidero Member are present in all the sections of the upper part of Artemisa Formation in the northern and southern sequences of Sierra del Rosario. The lithology is relatively constant. However, the amount of cherts may vary, while at the base of the Member singular calcarenite intercalations may occur. The total thickness ranges from 50 to 200 m.

CORRELATION

Time equivalents of Sumidero Member are given in Table 3.

POLIER FORMATION

NAME

This is a new lithostratigraphic unit (*cf.* Pszczółkowski 1976a) recognized in the upper part of Viñales Group (Table 2). The name is after the hills Lomas de Polier, south to the locality Los Hoyos (Fig. 14A).

TYPE SECTION

The type section of the Formation is exposed in the western escarpment of the road between Soroa and San Diego de Nuñez; its topographic coordinates are 288.200 and 337.680. The section starts on at the tectonic contact with the Upper Cretaceous marly limestones. In the lower part of Polier Formation, there are thin-bedded, gray, micritic limestones intercalated with sandstones and shales. The sandstones are thin-bedded, gray-brown or gray, hard, fine-grained, and with calcareous cement. They display organic or inorganic casts on the soles (flute casts and groove casts among others). They may also display graded bedding and horizontal and cross lamination. Some sandstone layers are up to some tens centimeters thick. The sandstones consist mostly of sharp-edged quartz, with plagioclases and muscovite in minor amounts. The limestones of this part of Polier Formation contain some ammonite imprints.

The upper part of Polier Formation starts with light-gray, hard, micritic limestones interbedded with marly limestones and calcareous shales. The limestone set is up to 30 m thick. It is overlaid by flysch deposits 10 m thick, recognized for the Roble Member (see below). This lithostratigraphic unit consists of quartz sandstones with calcareous cement. The rocks display graded bedding; however, sometimes it may be rather indistinct. Flute casts (*cf.* Pl. 3, Fig. 2) occur commonly; there are both prod and groove casts. The sandstone layers are up to 0.3 m thick and separated by thin clayey shale intercalations. Higher in the section, these deposits border upon cherts of Buenavista Formation, the transition being gradational.

The total thickness of Polier Formation is almost 300 m in its type section.

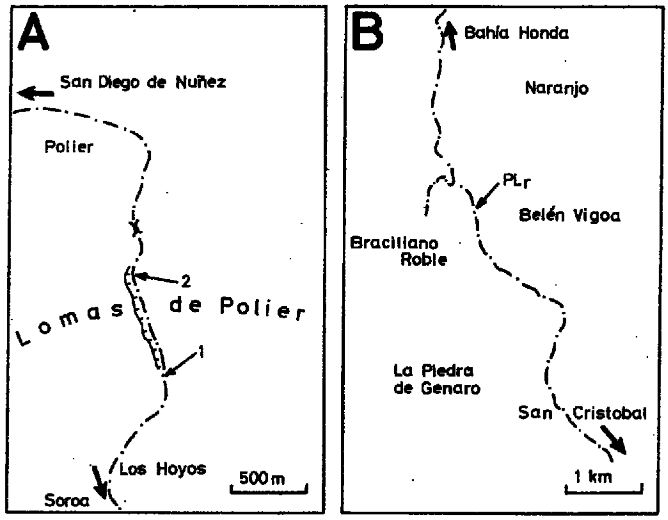


Fig. 14

A — Location map of the type section of Polier Formation; *1* type section beginning, *2* type section end

B — Location map of the type section of Roble Member of Polier Formation; *PLr* — the type section

BOUNDARIES

The type of the lower boundary of the formation has been designated in Naranjo tectonic unit (*cf.* Fig. 2*A*), between Rancho Manete and Naranjo elevations, by the road between San Cristobal and Bahía Honda. At that place, micritic limestones and cherts of Artemisa Formation pass into limestones with sandstone, shale, and sometimes chert intercalations of Polier Formation. The lower boundary of the latter formation is at the base of the first sandstone layer.

The type of the upper boundary of Polier Formation is in the type section, at the base of the first chert layer within the beds transitional between Polier and Buenavista Formations.

FAUNA AND AGE

There are tintinnids at the base of Polier Formation (*cf.* Table 6). Ammonite imprints are common in the middle part and rare at the top of the Formation. The list of ammonite species recorded in the Polier Formation is given by Myczyński (1977). These fossils indicate the Valanginian to ? Albian, although most deposits are attributed to the Hauterivian and Barremian (Myczyński 1977). The upper boundary of Polier Formation is probably heterochronous.

The fossil assemblage contains also radiolarians, while *Nannoconus* and planktic forams (*Hedbergella* sp.) occur less commonly. Aptychi are abundant all over the Formation.

SUBDIVISION

The Roble Member has been recognized by the present author (Pszczółkowski 1978a) for the uppermost part of Polier Formation.

LATERAL EXTENT AND THICKNESS

The Polier Formation occurs in the northern sequence of Sierra del Rosario (Fig. 2*B*; Table 1). The sandstones occur more abundantly in the tectonic units of Cangre and Sierra Chiquita.

The total thickness decreases southwards. It is about 300 m in Cangre unit (*e.g.* in Lomas de Polier section), while it attains but some tens meters in Belén Vigoa unit.

CORRELATION

In part, the Polier Formation is time equivalent to the Sabanilla Member of Buenavista Formation and to the top part of Sumidero Member of Artemisa Formation in the southern sequence of Sierra del Rosario. In the sequence of Quinones, its time equivalent is the Lucas Formation (Table 4). In part, the Polier Formation is also time equivalent to the Tumbitas and Infierno Members of Guasasa Formation and to the Pons Formation in the Sierra de los Organos.

ROBLE MEMBER

NAME

The Roble Member has been distinguished in the uppermost part of Polier Formation (Pszczółkowski 1976a). The name is after Braciliano Roble hill, central part of the Sierra del Rosario.

TYPE SECTION

The type section is designated in the western escarpment of the road between San Cristobal and Bahía Honda (Fig. 14B; Pl. 3, Fig. 1). Its topographic coordinates are 280.500 and 331.880. Micritic limestones with shale and sandstone intercalations of Polier Formation are overlaid by thick-bedded, medium-grained, quartz sandstones. Most sandstone layers display graded bedding and abundant casts on the soles. Thin intercalations of clayey shales occur between the sandstone layers. There are a few layers of micritic limestones in the middle of the Member while at the top, there is a single detritic limestone layer up to 1 m thick. The total thickness approximates 20 m. At the top, the deposits of this Member border upon cherts and shales of Sabanilla Member of Buenavista Formation (Pl. 3, Fig. 1).

BOUNDARIES

The type of the lower boundary of the Member is in the type section, at the base of the first layer of thick-bedded sandstone. The type of the upper boundary is in the same section, at the base of the first chert layer. Both boundaries are transitional, although the lower one is sharply marked.

FAUNA AND AGE

Apart from a few unidentifiable ammonite imprints, the sandstones of Roble Member contain but a few poorly preserved benthic and planktic forams. No microfauna has been reported from the shales. Then, the Roble Member can be but tentatively assigned to the Aptian to Albian on the basis of its lithostratigraphic position.

LATERAL EXTENT AND THICKNESS

The Roble Member occurs in the northern sequence of Sierra del Rosario; however, there are some sections of Polier Formation where this member cannot be recognized. The thickness of Roble Member deposits does never exceed 30 m but often it ranges from quite a few to 20 meters. Very similar Cretaceous sandstones occur also in the facies-structural zone of La Esperanza (Pszczółkowski & *al.* 1975).

LUCAS FORMATION

NAME

This is a new lithostratigraphic unit (Pszczółkowski 1976a) making part of Vinales Group (Table 2). The name is after Rancho Lucas, northern part of the Sierra del Rosario (Fig. 15).

TYPE SECTION

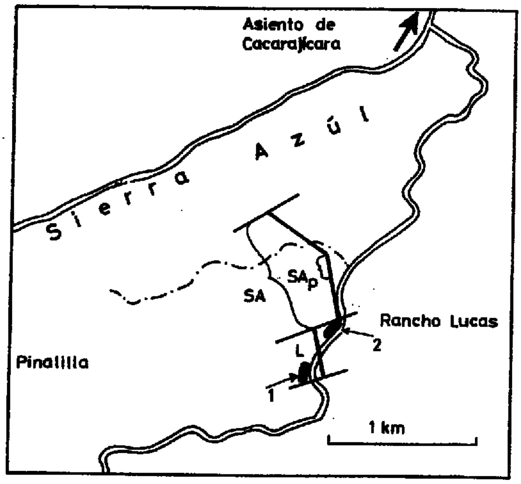


Fig. 15

Location map of the type sections of Lucas (*L*) and Sierra Azúl (*SA*) Formations, and Pinalilla Member (*SAp*) of Sierra Azúl Formation; *1*, *2* Lucas Formation exposures

The type section is in the environs of Rancho Lucas, south to the Sierra Azúl, by the road between Las Pozas and Pinalilla (Fig. 15). The best exposures are in the road escarpment. At the locality *1* (topographic coordinates 267.450 and 332.400), there are thin-bedded, gray to black, micritic limestones. They contain abundant aptychi and a few poorly preserved ammonite imprints. The rocks are strongly disturbed tectonically. The limestone thickness is 50 m at this locality.

The uppermost part of the Lucas Formation is exposed at the locality *2*, in the northern escarpment, 350 m NE to the former locality (Fig. 15). The deposits are represented by thin-bedded, gray, micritic limestones intercalated with hard, calcareous shales and marly shales (Pl. 2, Fig. 1). The limestones contain abundant aptychi and a few ammonite imprints. In thin sections, radiolarians occur commonly; they are usually calcified. The thickness of these deposits is 6 m. Higher in the section, the deposits of Lucas Formation pass gradationally into shales and cherts of Sierra Azúl Formation.

The type section of Lucas Formation starts on and ends just with the strata described from these localities. The total thickness of the Formation attains 200 m. Lithologically, the other deposits of the type section do not differ from those described above.

BOUNDARIES

The lower boundary of the Formation is unknown, as this unit borders upon other deposits along an overthrust surface. The type of the upper boundary of Lucas Formation is in the type section, at the top of the last limestone layer within the beds transitional to the Sierra Azúl Formation.

AGE

The ammonites made possible the assignment of Lucas Formation to the Upper Hauterivian to Lower Barremian (Myczyński 1977).

LATERAL EXTENT AND THICKNESS

The Lucas Formation occurs exclusively in Quiñones tectonic unit, northern part of the Sierra del Rosario (*cf*. Fig. 2A); it extends over some 20 km between the locality Pinalilla and Loma del Cable. There is a slight lithological variability in the amounts of thin chert or calcareous sandstone intercalations. However, these intercalations are never of any significance. The thickness of Lucas Formation attains its maximum (300 m) by Los Cayos.

CORRELATION

The Lucas Formation is time equivalent to the middle part of Polier Formation and to the Sabanilla Member of Buenavista Formation in the Sierra del Rosario; in the Sierra de los Organos, its time equivalents are the lower part of Pons Formation and the Infierno Member of Guasasa Formation (Table 4).

SIERRA AZÚL FORMATION

NAME

The Sierra Azúl Formation is a new lithostratigraphic unit (*cf.* Pszczółkowski 1978a). The name is after the limestone range Sierra Azúl, northern margin of the Sierra del Rosario.

TYPE SECTION

The type section of the Formation is at the southern slope of Sierra Azúl, between Pinalilla and Las Pozas (Fig. 15). The topographic coordinates are 267.675 and 332.800. The deposits of Lucas Formation are overlaid by shales and cherts therein (Fig. 16). At the base, there are hard, marly shales replaced by green cherts intercalated with thin layers of clayey shales, higher in the section. These cherts are overlaid by variegated clayey shales. At the top of these shales, there are green cherts with thin intercalations of fine--grained deposits including tuffite matter. The above described strata represent the lower part of Sierra Azúl Formation, and their thickness is 150 m.

The middle part of the Formation comprises massive or thick-bedded, gray-green, micritic limestone 170 m thick. This is the Pinalilla Member.

The deposits of the upper part of Sierra Azúl Formation vary in lithology (Fig. 16). They include marly limestones and shales (6 m), clayey shales with thin intercalations of siltstones and fine-grained sandstones (70 m), gray, marly limestones with shale intercalations (10 m), gray to brown, marly limestones (15 m), sandstones and shales (30 m), micritic limestones (4 m), interbedded calcareous shales, micritic limestones, and limestone breccias (approximately 100 m), and quartz sandstones (40 m). Thus, the upper part of the Formation attains 280 m in thickness. The total thickness of the Sierra Azúl Formation is 600 m in its type section.

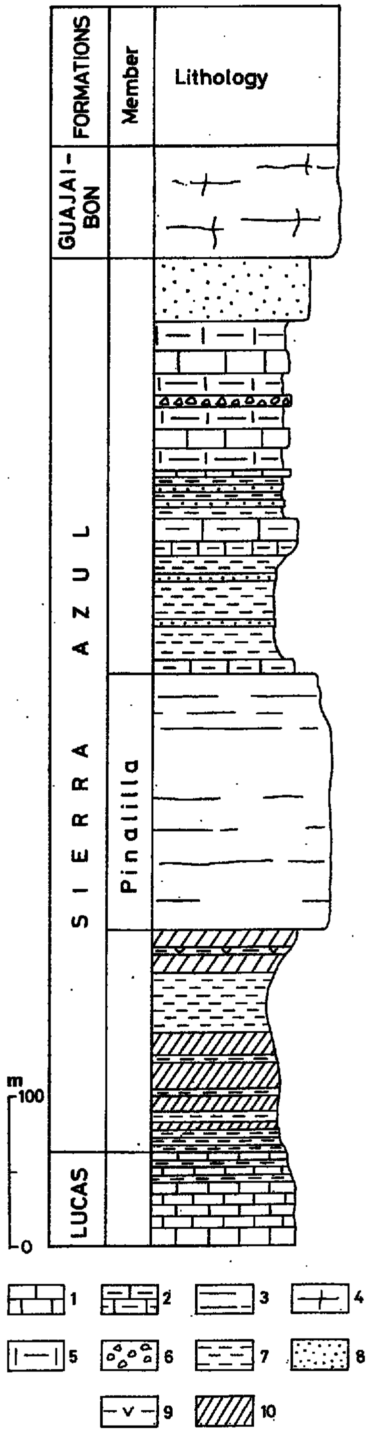


Fig. 16

Type section of Sierra Azúl Formation (for its geographic setting *see* Text-fig. 15)

1 well-bedded, micritic limestones, *2* marly limestones and marls, *3* massive and thick-bedded limestones of Pinalilla Member, *4* massive limestones of Guajaibón Formation, *5* hard, calcareous shales, *6* breccias, *7* clayey and marly shales, *8* sandstones, *9* fine-grained, tuffite-bearing rocks, *10* cherts

BOUNDARIES

The type of the lower boundary of the Formation is at the locality 2 (Fig. 15), at the top of the last limestone layer within the deposits transitional between Lucas and Sierra Azúl Formations.

The type of the upper boundary is in the type section, at the top of quartz sandstones underlying massive limestones of Guajaibón Formation. In other sections, the upper boundary of Sierra Azúl Formation is at the top of clayey shales or sandy-clayey deposits.

SUBDIVISION

The Pinalilla Member has been recognized for the middle part of the Formation (Fig. 16).

FAUNA AND AGE

The limestones of the middle part of Sierra Azúl Formation contain planktic forams indicating the Turonian age (Torre 1972—1975). At the top, the following foraminifer genera have been recorded: *Vaughanina*, *Orbitoides*, *Pseudorbitoides*, *Sulcorbitoides*, and *Sulcoperculina*, indicating the Campanian to Maastrichtian age (Torre 1972—1975). There are no forams typical of the Upper Maastrichtian. Thus, the Sierra Azúl Formation can be assigned to the Barremian to ? Lower Maastrichtian. This inference is also confirmed by the lithostratigraphic position of the formation.

LATERAL EXTENT AND THICKNESS

The Sierra Azúl Formation is present in the stratigraphic sequence of Quinones, northern part of the Sierra del Rosario. The terrigenous deposits in the upper part of the Formation may be thicker in some sections than they are in the type area. In the environs of El Pan de Guajaibón, there is a volcanic rock up to a dozen meters thick at the top of the Formation. The total thickness of Sierra Azúl Formation does not exceed 600 m.

CORRELATION

In part, the Sierra Azúl Formation is time equivalent to the upper part of Polier Formation and to the Buenavista Formation in the Sierra del Rosario; and to the Pons Formation (except of its lowermost and uppermost parts) and the Infierno Member of Guasasa Formation in the Sierra de los Organos. In the facies-structural zone of Bahía Honda, the Sierra Azúl Formation is partly time equivalent to the Felicidades Formation (Pszczółkowski & *al.* 1975).

PINALILLA MEMBER

NAME

The Pinalilla Member (Pszczółkowski 1976a) has been distinguished in the middle part of Sierra Azúl Formation. The name is after the locality called Pinalilla, north to the road towards Asiento de Cacarajícara (Fig. 15).

TYPE SECTION

The type section of the Member is in the neighborhood of Rancho Lucas (Fig. 15). It makes part of the type section of Sierra Azúl Formation. The Member consists of thick-bedded, sometimes massive, gray-green, micritic limestones without any intercalations.

BOUNDARIES

The type of the lower boundary of Pinalilla Member is in the type section, at the base of thick-bedded limestones overlying cherts of the lower part of Sierra Azúl Formation.

The type of the upper boundary is in the same section, at the top of a limestone layer underlying the first marly shale intercalation.

FAUNA AND AGE

Some samples derived from the limestones of Pinalilla Member contain planktic forams (Torre 1972—1975, Pszczółkowski *in:* Pszczółkowski & *al.* 1975). Moreover, there are also radiolarians, *Stomiosphaera*, and *Pitonella*. The microorganisms indicate the Turonian age for the Member (*cf.* Pszczółkowski 1978a).

LATERAL EXTENT AND THICKNESS

The Pinalilla Member occurs exclusively in the sequence of Quinones. The total thickness is 170 m in the type section, while it ranges from 100 to 140 m in other sections.

BUENAVISTA FORMATION

NAME

Buenavista Formation has been recognized in the southern and northern sequences of Sierra del Rosario (Pszczółkowski 1976a). Its detailed description (Pszczółkowski *in:* Pszczółkowski & *al.* 1975) has insofar not been published. The name is after a locality by the road between El Cuzco and Valdés, northeastern part of the Sierra del Rosario. The deposits assigned at present to the Buenavista Formation had previously not been regarded as a formal lithostratigraphic unit.

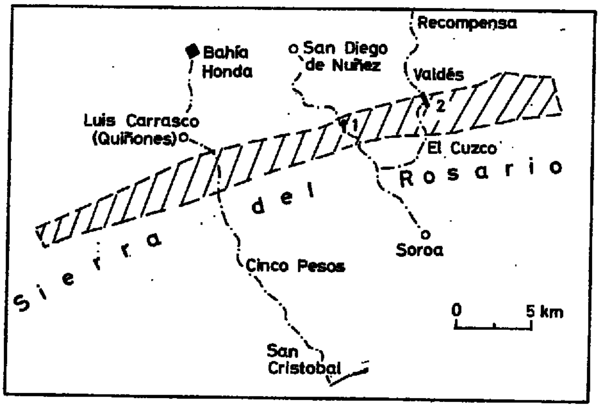


Fig. 17. Location map of the type area of Buenavista Formation
1 type section of the lower and middle parts of the Formation, *2* type section of the upper part of the Formation; *hachured* type area

TYPE AREA

Poor exposures, high lithological variability, and tectonic complexity do not allow for the moment to designate a single type section for the whole Buenavista Formation. The type area is in the northeastern part of Sierra del Rosario (Fig 17). The hypostratotype for the southern sequence of Sierra del Rosario is at the locality called Buena Vista, east to La Palma (Fig. 23).

The type section of the lower and middle parts of Buenavista Formation are by the road between Soroa and San Diego de Nuñez, crossing the hills Lomas de Polier (Fig. 17). Deposits of Roble Member of Polier Formation are overlaid by cherts and green shales, with some layers of silicified turbidite sandstones at the base (Fig. 18*A*). At the top of this chert set, there are some intercalations of red and green, micritic limestones; they are usually silicified. These deposits are 38 m thick and have been assigned to the Sabanilla Member. They are overlaid by a set of detritic and micritic limestones and cherts. The detritic limestones include calcirudites (limestone breccias), calcarenites, and calcilutites. They often display graded bedding and are sharply separated from the micritic limestones and cherts at the base. These deposits include also some clayey shales, sometimes silicified; however, the latter rocks occur but in minor amounts although they become more common in the uppermost part. These limestone- and chert-bearing deposits are

23 m thick. Higher in the section, they pass gradationally into sandstones and shales intercalated with limestones. The latter deposits are but poorly exposed in Lomas de Polier section.

The type section of the upper part of the Formation is by the road from Recompensa and Valdés towards El Cuzco (Fig. 17). The topographic coordinates are 294.000 and 339.650. Detritic and micritic limestones and cherts are overlaid by a coarse breccia 10 m thick (Fig. 18*B*), consisting of limestone and chert debris up to 1 m in size. This breccia underlies green, banded cherts. Higher in the section, there is another breccia 8 m thick consisting of limestone and chert debris up to 0.8 m in size. It is overlaid by cherts with rare intercalations of clayey-tuffite deposits. At the top, these cherts border upon detritic limestones of Cascarajícara Formation. The total thickness of the upper part of Buenavista Formation is about 80 m.

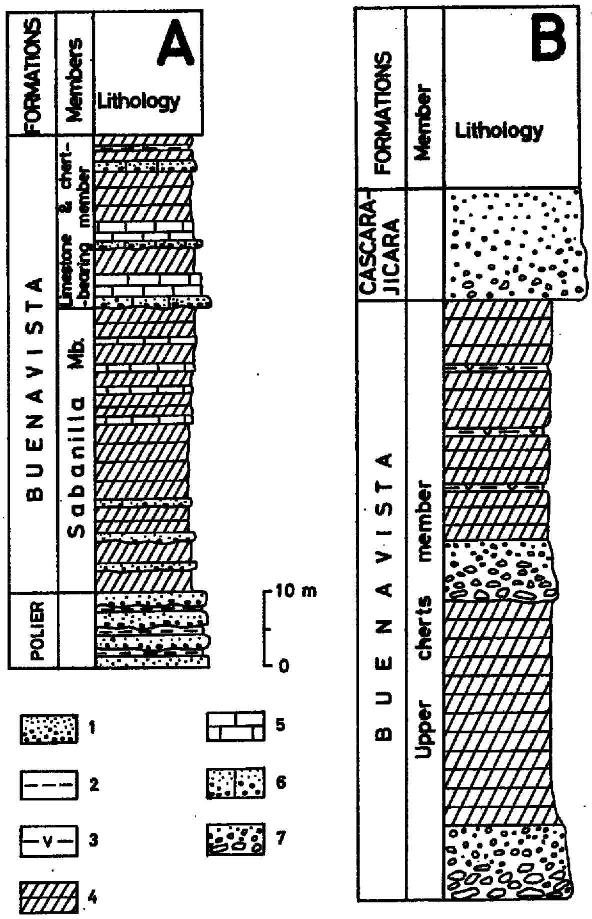


Fig. 18

Type sections of Buenavista Formation (for their geographic setting *see* Text-fig. 17)
A — type section of the lower and middle parts of the Formation at Lomas de Polier (simplified lithology)
B — type section of the upper part of the Formation south to Valdés (simplified lithology)
1 turbidite sandstones, *2* shales, *3* clayey-tuffite intercalations, *4* cherts, *5* micritic, sometimes silicified limestones, *6* graded detritic limestones, *7* breccias

HYPOSTRATOTYPE

The hypostratotype of the Formation (Fig. 19) is at the locality called Buena Vista, 4 km east to La Palma (*cf.* Fig. 23). It is exposed along a dirt road going northwards. Limestones of Artemisa Formation are overlaid by cherts exposed in some small outcrops; they represent the Sabanilla Member. Higher in the section, there is a breccia 15 m thick consisting of limestone and chert debris. It underlies micritic and marly limestones (10 m) passing at the top into a set of marly and clayey shales and wacky sandstones intercalated with thin layers of marly limestones (totally, 70 m). Higher in the section, there are thick-bedded, wacky sandstones including large amounts of volcanic debris (25 m). They are overlaid by volcanic rocks (100 m) of basalt or andesite composition (Skupiński *in:* Pszczółkowski & *al.* 1975), including volcanic breccias and agglomerates. These volcanics underlie micritic limestones (4 m) containing some microorganisms (Torre 1972—1975). Higher in the section, there are cherts (40 m), breccias (6 m) and at the top, yellow shales and cherts (20 m). The upper boundary of Buenavista Formation is of tectonic nature in this section.

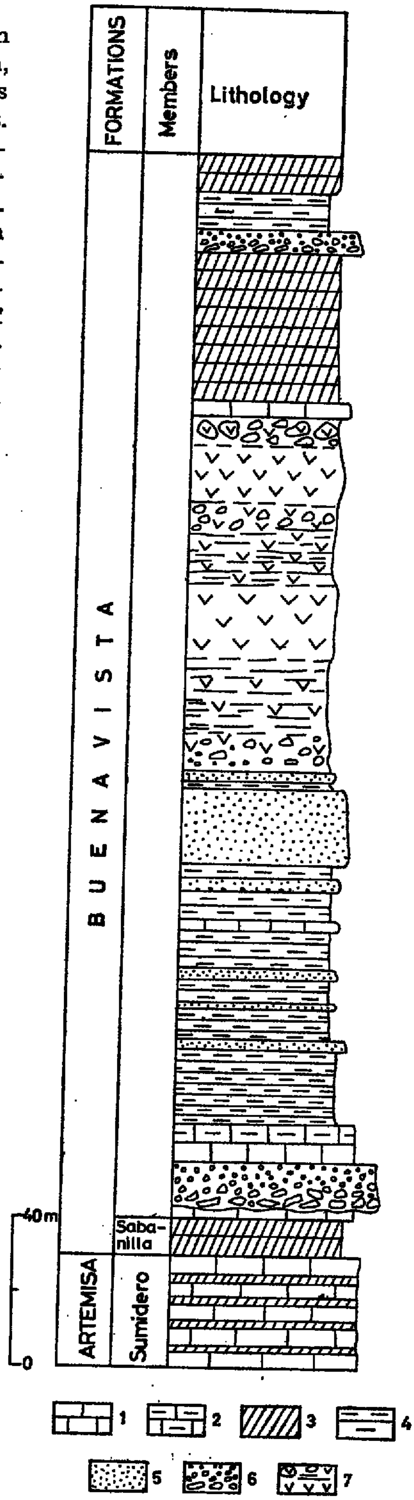


Fig. 19

Hypostratotype of Buenavista Formation at Buena Vista, east to La Palma (for its geographic setting *see* Text-fig. 23)

1 micritic limestones, 2 marly limestones, 3 cherts, 4 shales, 5 wacky sandstones, 6 breccias, 7 volcanic rocks

BOUNDARIES

The type of the lower boundary of the Formation is in the type section, at the base of the first chert layer within the beds transitional between Polier and Buenavista Formations.

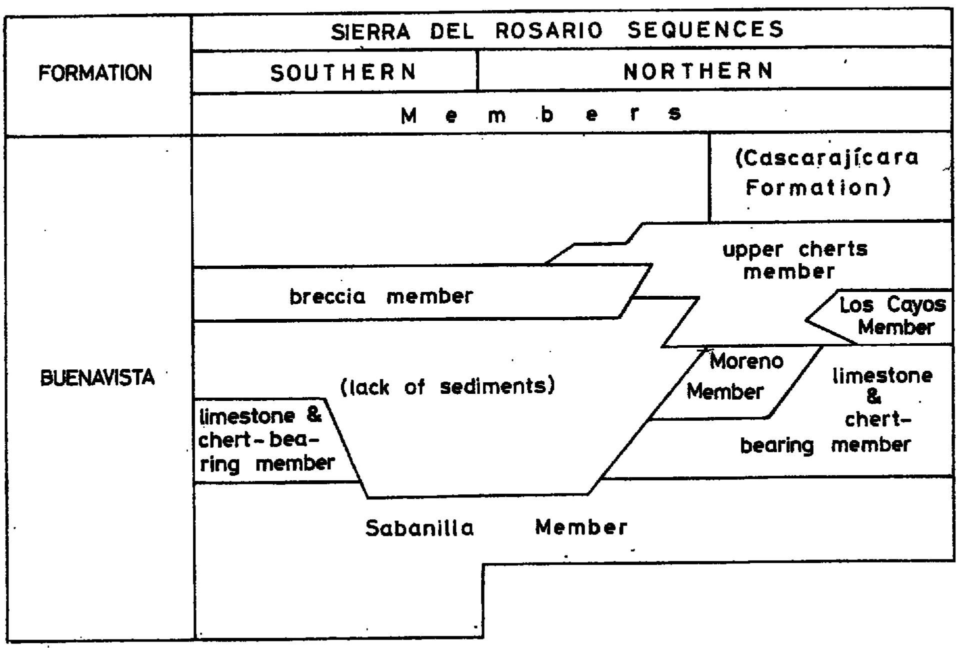
The type of the upper boundary is in the type section of the upper part of Buenavista Formation, at the top of the chert set and at the base of detritic limestones of Cascarajícara Formation. The upper boundary of Buenavista Formation is poorly exposed; nevertheless, one may claim that it is of erosional nature (submarine erosion).

SUBDIVISION

Three members have been distinguished within Buenavista Formation, *viz.* Sabanilla, Moreno, and Los Cayos Members. Furthermore, three informal lithostratigraphic units have also been recognized, *viz.* limestone- and chert-bearing, limestone breccia, and upper chert members (Table 7).

Table 7

Buenavista Formation subdivision in the southern and northern sequences of Sierra del Rosario



FAUNA AND AGE

Chronostratigraphic attribution of the deposits of Buenavista Formation is based on some forams identified in thin sections, mostly by Torre (1972—1975) and in a few cases by the present author (Pszczółkowski *in:* Pszczółkowski & al. 1975).

The base deposits of Buenavista Formation overlie the Valanginian to Lower Hauterivian limestones of Artemisa Formation in the southern sequence of Sierra del Rosario (*cf.* Table 4). However, in the northern sequence of Sierra del Rosario, the cherts of Sabanilla Member overlie the Barremian to Albian deposits of Polier Formation. Thus, the lower boundary of Buenavista Formation appears heterochronous.

In the northern sequence of Sierra del Rosario, the strata of lower and middle parts of Buenavista Formation are assigned to the ? Albian to Campanian, with the Cenomanian, Turonian, and Campanian documented micropaleontologically. There are no Santonian fossils in those deposits, while forams that may indicate the Coniacian have been recorded but in a single sample (Torre 1972—1975). The upper part of Buenavista Formation represents the Campanian to Maastrichtian in Sierra Chiquita and Cangre tectonic units; whereas in some southern sections of the northern sequence of Sierra del Rosario, the Upper Paleocene deposits have also been recorded (Torre 1972—1975). Moreover, Lower Eocene strata may also be present in the sections (Pszczółkowski *in:* Pszczółkowski & *al.* 1975).

There are usually no Turonian to Campanian deposits in the southern sequence of Sierra del Rosario, due to a pre-Maastrichtian erosion (*cf.* Pszczółkowski 1971b, 1976a). The Maastrichtian to Paleogene deposits of the upper part of Buenavista Formation are but in a few sections underlaid by the Turonian to ? Coniacian rocks. Some Late Paleocene micro-organisms have been reported from this stratigraphic sequence (Torre 1972—1975; for sample localization *see* Pszczółkowski *in:* Pszczółkowski & *al.* 1975). Lower Eocene deposits may also be present in this sequence.

In general, the Buenavista Formation ranges in age since the Hauterivian through ? Early Eocene. The upper boundary of the Formation is heterochronous.

LATERAL EXTENT, LITHOLOGIC VARIABILITY, AND THICKNESS

The Buenavista Formation occurs in the southern and northern sequences of Sierra del Rosario (Table 1). The lithology varies in the amount of terrigenous and volcanic deposits, and in the proportions of detritic deposits, cherts, and pelagic limestones. The high facies variability and the occurrence of stratigraphic hiatus in some sections may substantiate an eventual advancement of Buenavista Formation to the rank of a group; such an advancement requires, however, more detailed stratigraphic and mapping research.

The maximum thickness is almost 400 m. Nevertheless, it ranges usually from 80 to 300 m, mostly because of tectonic reasons.

CORRELATION

The Buenavista Formation is time equivalent to the uppermost part of Guasasa Formation and most of Pons Formation, and in part to the Ancón and Pica Pica Formations in the sequence of Sierra de los Organos (Table 1). In the Sierra del Rosario, time equivalents of diverse parts of Buenavista Formation are: Lucas, Sierra Azúl, and Guajaibón Formations in the sequence of Quiñones, and most of Polier Formation and the Cascarajícara Formation in the northern sequence. Lithologically, the Buenavista Formation resembles most closely some strata of Sierra Azúl and Pica Pica Formations.

SABANILLA MEMBER

NAME

This is a new lithostratigraphic unit recognized in the lower part of Buenavista Formation (Pszczółkowski 1976a). The name is after the locality Sabanilla, central part of the Sierra del Rosario.

TYPE SECTION

The type section of Sabanilla Member is by the road between Niceto Perez and Mil Cumbres, 2 km away from Sabanilla (Fig. 20*A*). The topographic coordinates are 265.175 and 326.440. Micritic limestones intercalated with cherts of Artemisa Formation are overlaid by cherts with thin intercalations of green, silicified, clayey shales at the base. In the lower part of the section, these cherts are green, sometimes horizontally laminated. They represent radiolarian cherts and radiolarites. At the top of the Member, the cherts become red or red-brown (in weathering zone), and shale intercalations disappear. The deposits of this member are 20 m thick in this section.

BOUNDARIES

The type of the lower boundary of Sabanilla Member is in the type section, at the base of a chert layer overlying the last limestone layer within the beds transitional between Artemisa and Buenavista Formations.

The type of the upper boundary is in the same section, at the top of the last chert layer underlying a breccia composed of limestone and chert debris. In the type section this boundary is of erosional nature. However, in some other sections the cherts pass gradationally into limestones intercalated with cherts (limestone- and chert-bearing member).

FAUNA AND AGE

The Albian to Cenomanian planktic forams have been recorded in some exposures of the southern sequence of Sierra del Rosario (Torre 1972—1975, Pszczółkowski *in:* Pszczółkowski & *al.* 1975). A sample derived at Los Tumbos, NE to Cinco Pesos, contains abundant specimens of the foraminifer genera *Rotalipora, Praeglobotruncana, Clavihedbergella, Schackoina,* and *Hedbergella,* the whole assemblage indicating the Cenomanian to lowermost ? Turonian. The forams come from the top part of Sabanilla Member; on the other hand, this member overlies the Artemisa Formation. Hence, one may conclude that the Sabanilla Member ranges in age since the Hauterivian through earliest ? Turonian in the southern sequence of Sierra del Rosario.

In the northern sequence of Sierra del Rosario, the only insofar recorded foraminifer genus is *Ticinella* indicating the Albian to Lower Cenomanian (*cf.* Bandy 1967). As judged from its position relative to Polier Formation, the Sabanilla Member is of younger than the Barremian age in this sequence. In some sections, it overlies even deposits of the Aptian and ?Early Albian age. On the other hand, the Sabanilla Member is overlaid by the Cenomanian to Turonian limestone- and chert-bearing member (*see* below). Hence, one may conclude that the Sabanilla Member ranges in age since the Aptian through Cenomanian in the northern sequence of Sierra del Rosario, although the stratigraphic range is usually shorter in each particular section.

Both the boundaries of the Member are heterochronous.

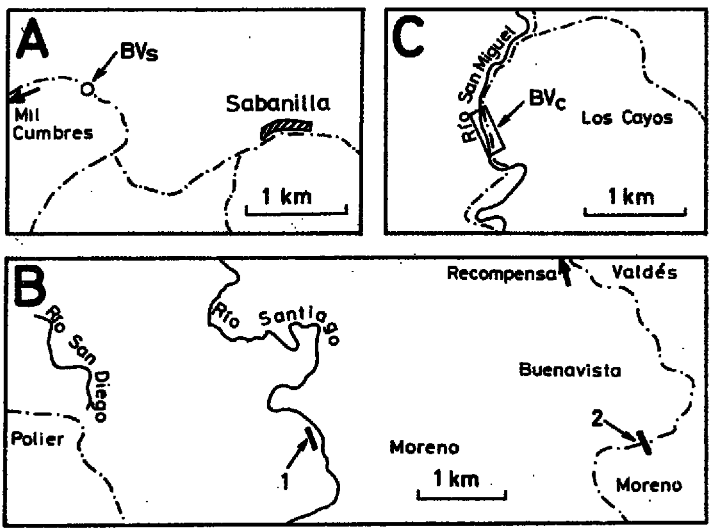


Fig. 20

A — Location map of the type section of Sabanilla Member of Buenavista Formation; *BVs* — type section

B — Location map of the type section of Moreno Member of Buenavista Formation; *1* type section of the lower part of the Member, *2* type section of the upper part of the Member

C — Location map of the type section of Los Cayos Member of Buenavista Formation; *BVc* — type section

LATERAL EXTENT AND THICKNESS

The Sabanilla Member occurs in most sections of Buenavista Formation in the southern and northern sequences of Sierra del Rosario (Tables 4, 7; *cf.* also Pl. 2, Fig. 3); however, it may be absent from some sections due to a pre-Maastrichtian erosion (*cf.* Pszczółkowski 1978a, Fig. 3). The lithological variability is but slight and expressed in the amounts of singular intercalations of micritic limestones (southern sequence) or silicified sandstones (northern sequence). The total thickness ranges from 0 to 40 m.

CORRELATION

The Sabanilla Member is time equivalent to shales and cherts of the lower part of Sierra Azúl Formation in the sequence of Quinones. The base part of the Member in the southern sequence is time equivalent to the Polier Formation in the northern sequence. In the sequence of Sierra de los Organos, time equivalents of Sabanilla Member are: the lower part of Pons Formation and most of Infierno Member of Guasasa Formation.

In the facies-structural zone of La Esperanza, cherts and shales of Panchita Formation (Danilewski *in:* Pszczółkowski & *al.* 1975) are equivalent in facies and partly in age to the Sabanilla Member.

LIMESTONE- AND CHERT-BEARING MEMBER

This informal lithostratigraphic unit recognized in the middle part of Buenavista Formation comprises micritic limestones, calcarenites, calcirudites (limestone breccias), cherts, and silicified shales overlying deposits of Sabanilla Member (Table 7). These deposits have already been described when discussing the type section of the lower and middle parts of the Formation (Fig. 18*A*). The pelagic components of this member (micritic limestones and cherts) are interbedded with detritic limestones of turbidite origin. The latter layers are a few centimeters up to 20 m thick. Some calcarenites and calcirudites comprise also significant amounts of terrigenous material, sharp-edged quartz, wacky sandstone debris, and plagioclases including. There are also fairly common bioclasts, *viz.* echinoderm and thick-shelled bivalve debris (Pl. 4, Fig. 6), algal detritus, and a few benthic and planktic forams. In thin sections, one may also see some ooids, shallow-water limestone, and pelagic, radiolarian-bearing limestone debris. The thickness of this member ranges from a few to 70 meters.

The limestone- and chert-bearing member appears typical of the northern sequence of Sierra del Rosario. It is especially well-developed in the tectonic units of Sierra Chiquita, Cangre, and La Serafina. In the southern sequence of Sierra del Rosario, it occurs but in a few sections and lacks usually both detritic limestones and shales.

The planktic forams (Torre 1972—1975) have allowed to assign the limestone- and chert-bearing member to the Cenomanian to Turonian but the member may range up into the Campanian at least in a single section (Pszczółkowski *in:* Pszczółkowski & *al.* 1975).

MORENO MEMBER

NAME

This is a new lithostratigraphic unit (*cf.* Pszczółkowski 1978a) recognized for the middle part of Buenavista Formation. The name is after Moreno locality, south to Valdés (Fig. 20*B*).

TYPE AREA

The type area of Moreno Member is in the northeastern part of Sierra del Rosario (Fig. 20*B*).

The type section of the lower part of the Member is by the road near Santiago River, in the environs of another Moreno locality (Fig. 20*B*); the topographic coordinates are 190.400 and 339.650. Deposits of Moreno Member overlie the limestone- and chert-bearing member. They comprise shales, polymictic sandstones, marly and detritic limestones. At the base, the section is dominated by clayey shales intercalated with limestones, while the limestones become lacking at the top. The detritic limestones may display graded bedding. In this section, the Moreno Member attains 30 m in thickness. The upper boundary of this part of the Member is of tectonic nature.

The type section of the upper part of Moreno Member is at Moreno, south to Valdés (Fig. 20*B*); the topographic coordinates are 293.700 and 339.250. The exposure is in a road escarpment. There are sandstones and conglomerates bearing a volcanic material, intercalated with clayey shales. There are also some intercalations of green, dacitic tuffites (Skupiński *in*: Pszczółkowski & *al.* 1975). These deposits attain 15 m in thickness. At the top, they border upon cherts of the upper part of Buenavista Formation.

BOUNDARIES

The type of the lower boundary of Moreno Member is in the type section of the lower part of the Member. It is designated at the base of the first layer of clayey shales, within the beds transitional between the limestone- and chert-bearing member and Moreno Member.

The type of the upper boundary of Moreno Member is in the type section of the upper part of the Member. It appears sharp and occurs at the base of the first chert layer overlying the terrigenous deposits of Moreno Member.

AGE

The marly limestones contain some planktic forams indicating the Cenomanian to Turonian (Torre 1972—1975, Pszczółkowski *in:* Pszczółkowski & *al.* 1975).

LATERAL EXTENT, LITHOLOGICAL VARIABILITY, AND THICKNESS

The Moreno Member occurs in the northern sequence of Sierra del Rosario (Table 7), *viz.* in the tectonic units of Sierra Chiquita, Cangre, and La Serafina. The deposits are lithologically variable. In some sections, detritic limestones are lacking, while volcanic deposits (mostly tuffites) up to 15 m thick appear. The polymictic and wacky sandstones may be rare or even absent from some sections. The total thickness of Moreno Member ranges from 0 to 60 m.

CORRELATION

The Moreno Member is time equivalent to the upper part of limestone- and chert-bearing member (Table 7). In fact, these units may intergrade laterally.

LIMESTONE BRECCIA MEMBER

A breccia composed of limestone and subordinately chert fragments is here regarded as an informal lithostratigraphic unit called limestone breccia member. It occurs commonly in sections of the middle part of Buenavista Formation, overlying deposits of the limestone- and chert-bearing member or Sabanilla Member. It has been recorded in the southern sequence of Sierra del Rosario and in some southern tectonic units of the northern sequence of Sierra del Rosario. It was previously described under the name of Limestone Breccia from the environs of Cinco Pesos and Soroa (Pszczółkowski 1971b), and under the name of limestone-chert breccia from the southern sequence of Sierra del Rosario (Pszczółkowski 1976a).

The Upper Campanian to Maastrichtian forams have been reported from clasts making part of the breccia (Torre 1972—1975, Pszczółkowski *in:* Pszczółkowski & *al.* 1975). One may claim that the breccia is of Maastrichtian age (Pszczółkowski *in:* Pszczółkowski & *al.* 1975), however it may be younger locally. The thickness ranges from 2 to 30 m.

LOS CAYOS MEMBER

NAME

This is a new lithostratigraphic unit recognized in the upper part of Buenavista Formation (*cf.* Pszczółkowski 1976a). The name is after Los Cayos locality, northern part of the Sierra del Rosario (Fig. 20*C*).

TYPE SECTION

The type section of Los Cayos Member is in the bed of San Miguel River and at the base of a slope of the adjacent hill (Fig. 20*C*); the topographic coordinates are 273.700 and 334.000. The Member consists exclusively of a coarse breccia (Pl. 2, Fig. 2) composed of chert and limestone debris. The largest blocks are up to 1.3 m in size. The debris is usually densely packed and the matrix is usually invisible. Among the components of the breccia, there are detritic, micritic, and marly limestones among others. The clasts of biodetritic limestones contain rudist debris and benthic forams. The thickness of Los Cayos Member is 180 m in the type section.

BOUNDARIES

Types of the boundaries of Los Cayos Member cannot be designated, as they are poorly exposed. In the type section, the breccia of Los Cayos Member overlies gray, marly limestones. The contact appears to be erosional. The upper boundary is inaccessible but there are cherts interbedded with thin layers of clayey shales, higher in the section. The latter deposits make part of the upper chert member of the Buenavista Formation.

AGE

No organic remains have been found in the breccia to allow a precise stratigraphic assignment. However, in a section east to the type area, detritic limestones underlying Los Cayos Member contain the Campanian to Lower Maastrichtian microfauna (Torre 1972—1975). On the other hand, Los Cayos Member is overlaid by the Campanian to Maastrichtian cherts and shales of the upper chert member. Hence, one may assign the deposits of Los Cayos Member to the Campanian to Maastrichtian.

LATERAL EXTENT AND THICKNESS

The Member occurs but is some sections of the northern sequence of Sierra del Rosario (Table 7), *viz.* in the tectonic units of Sierra Chiquita and Cangre (?). In Sierra Chiquita unit, the breccia of Los Cayos Member occurs under the form of three elongate lenses (Fig. 21) but these lenses may join themselves somewhere, in a section inaccessible for the moment. The total thickness of the Member ranges from 0 to 180 m. The largest clasts have been found in the bed of Santiago River (radiolarian chert blocks up to 5 m in length).

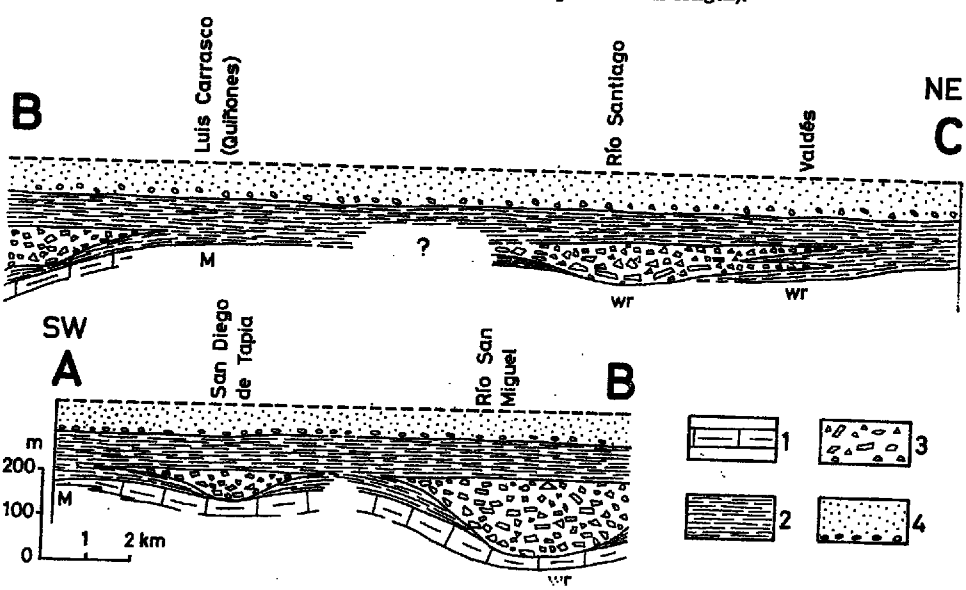


Fig. 21. Section of Buenavista Formation in Sierra Chiquita tectonic unit
1 marly limestones, *2* upper chert member, *3* breccia of Los Cayos Member of Buenavista Formation, *4* clastic limestones of Cascarajícara Formation, *wr* limestone- and chert-bearing member of Buenavista Formation, *M* Moreno Member of Buenavista Formation

UPPER CHERT MEMBER

In the northern sequence of Sierra del Rosario, the upper part of Buenavista Formation is dominated by radiolarian cherts. These deposits are here regarded as an informal lithostratigraphic unit called upper chert member (Table 7). Apart from the cherts, these strata comprise also breccias, clayey shales, fine-grained deposits with tuffogenic matter, and calcarenites. The breccias resemble closely in composition that one typical of Los Cayos Member. In some sections (*e.g.* Loma del Mulo), there are at least 10 breccia beds, each attaining quite a few meters in thickness, within the upper chert member. Some breccias and calcarenites display graded bedding. The total thickness of the upper chert member ranges from 70 to 100 m.

The upper chert member is overlaid by deposits of Cascarajícara Formation. As judged from some planktic forams recorded in thin sections of the cherts (*Globotruncanella* sp., ?*Globotruncana* sp., *Hedbergella* sp.), these strata are to be assigned to the Campanian to Maastrichtian. The breccia clasts contain some benthic forams attributed to the following genera: *Orbitoides*, *Sulcoperculina*, and *Lepidorbitoides* (accordingly to Van Gorsel (1972, 1975), the American representatives of the genus *Lepidorbitoides* should rather be assigned to *Orbitocyclina* Vaughan). These genera have insofar been reported only from the Campanian and Maastrichtian strata (*cf.* Brönnimann & Rigassi 1963, Seiglie & Ayala-Castanares 1963).

GUAJAIBÓN FORMATION

HISTORY

The name of the Formation is after the highest mountain of Pinar del Río Province, namely El Pan de Guajaibón, northern part of Sierra del Rosario. At present, the limestones of Guajaibón Formation are regarded as a formal lithostratigraphic unit (*cf.* Pszczółkowski 1976a). They were previously recorded by Herrera (1961) who mentioned „Guajaibón platform limestones" with abundant miliolid fauna but did not describe the strata or designate the type section. Pardo (1975) noticed „massive, shallow-water, Aptian to Coniacian limestones" of Cacarajícara belt.

TYPE SECTION

The type section of the Formation starts on by a road along the western margin of El Pan de Guajaibón (Fig. 22). The topographic coordinates are 254.500 and 330.200. Deposits of Sierra Azúl Formation are overlaid by light-gray, massive limestones comprising calcarenites and biomicrites, sometimes horizontally laminated. There are also some calcirudites and micritic limestones. In places, the limestones are dolomitized. Higher in the section, the gray, massive limestones may contain fragmented bivalve shells, algae, and corals.

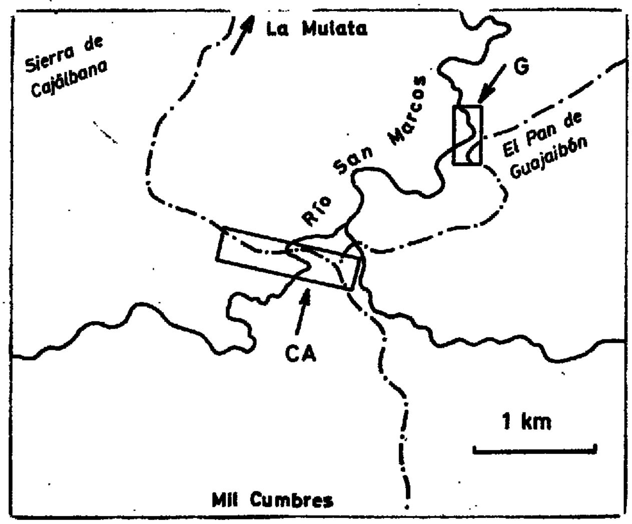


Fig. 22. Location map of the type sections of Guajaibón (G) and Cascarajícara (CA) Formations

BOUNDARIES

The lower boundary of the Formation has not been designated, as it is most probably of tectonic nature. The Guajaibón Formation borders upon the Sierra Azúl Formation along a sharp contact.

The upper boundary of the Guajaibón Formation is unknown, as this formation borders northward upon rocks of the facies-structural zone of Bahía Honda, the contact with which is tectonic (Pszczółkowski 1976a, b). In the type section there is a carbonate conglomerate at the top of the Formation.

HYPOSTRATOTYPE

The hypostratotype of Guajaibón Formation is exposed along the road between Cacarajícara and El Pan de Guajaibón, eastern margin of the Sierra Azúl. Clayey shales of Sierra Azúl Formation are overlaid by massive, micritic, partly dolomitized limestones. Higher in the section, there are calcarenites and biomicrites with algal debris among other components. The upper part of the Formation is represented in isolated "mogotes", NE to the range Sierra Azúl; it is dominated by calcarenites (Pl. 6, Fig. 1) and calcirudites consisting of intraclasts, ooids, onkoids, coral and algal bioclasts. Miliolid-bearing biomicrites are less common.

AGE

At the base of the Formation, there are but benthic forams (Miliolidae and Textulariidae) and fragmented algae, bivalves and gastropods. Some tens meters above the base, several planktic forams have been recorded (Pszczółkowski & *al.* 1975), and these indicate the Maastrichtian. The shallow-water limestones of the lower and the top parts of the Guajaibón Formation have not been dated micropaleontologically. Herrera (1961) suggested Aptian to Maastrichtian age of the „Guajaibón limestone", whereas Pardo (1975) noticed „Aptian to Coniacian limestones" in his Cacarajícara belt.

LATERAL EXTENT AND THICKNESS

The Guajaibón Formation occurs exclusively in the northernmost part of the Sierra del Rosario, east to the Sierra de Cajálbana. It is placed in the Quinones sequence (*cf.* Pszczółkowski 1976a), but most probably this formation is separated by the tectonic contact from the Sierra Azúl Formation, and it may be considered as a distinct tectonic unit. The thickness of the Guajaibón Formation is 380 m in the type section, and 350 m in the hypostratotype. The maximum thickness (500 m) is recognizeable in the central part of El Pan de Guajaibón.

CORRELATION

The Guajaibón Formation is probably time equivalent to some strata of the upper part of Buenavista Formation in the Sierra del Rosario, and Pons Formation in the Sierra de los Organos.

CASCARAJÍCARA FORMATION

NAME

This formation has been distinguished by Hatten (1957) who designated the type section by the road between La Mulata and Mil Cumbres (Fig. 22), western margin of the Sierra Chiquita (called previously Sierra de Cacarajícara).

LITHOLOGY

The Formation has been described in details by the present author (Pszczółkowski *in:* Pszczółkowski & *al.* 1975). It starts usually with a breccia up to 10 m thick, composed of chert and limestone clasts. In places, this breccia is very thin or lacks at all. Higher in the section, the breccia passes into coarse-grained calcarenites. They consist mostly of shallow-water limestone fragments (miliolid-bearing biomicrites, ooid-bearing calcarenites, and biocalcirudites with large forams, among others) and deep-water biomicrites with *Globotruncana* spp., *Rotalipora* spp., and Calcisphaerulidae. There are also clasts of radiolarian cherts, dolomites, shales, quartzitic sandstones and quartzites; volcanic rock fragments (Pl. 6, Fig. 2) assigned to trachites (Skupiński *in:* Pszczółkowski & *al.* 1975) do also occur. Among the minor components, there are fragments of alkalic rocks, tuffites, metamorphic rocks, and plagioclases. Detritic quartz appears fairly abundant. Bioclasts comprise forams, rudist (Pl. 6, Figs 2—3), algal, and echinoderm fragments. Poorly developed matrix consists of microsparite and fine detritic material.

The upper part of Cascarajícara Formation comprises fine-grained calcarenites and calcilutites including abundant planktic forams. At the top, the deposits may be horizontally bedded. The limestones of Cascarajícara Formation appear to be well indurated. The change in average grain size from the base to top of the Formation, appears usually gradational. However, there are also some exceptions (*e.g.* Sierra Chiquita section).

FAUNA AND AGE

The list of forams recorded by Torre (1972—1975) and the present author in the deposits of Cascarajícara Formation, is given by Pszczółkowski (*in:* Pszczółkowski & *al.* 1975). The species *Omphalocyclus* cf. *macroporus* (Lamarck) has been recorded at both the base and top of the Formation. This species is generally regarded as typical of the Upper Maastrichtian (Brönnimann & Rigassi 1963, Seiglie & Ayala-Castanares 1963). At the top of the Formation, there are *Globotruncana stuarti* (Lapparent), *Rugoglobigerina scotti* (Brönnimann), and *Globotruncanella havanensis* (Voorwijk) among others. *R. scotti* (Brönnimann) occurs in the Upper Maastrichtian (Bandy 1967, Postuma 1971). Most of the foraminifer fauna are redeposited (*cf.* Hatten 1957, Khudoley & Meyerhoff 1971).

No Paleogene microfauna has been recorded recently in the deposits of Cascarajícara Formation. Hatten (1957) mentioned, however, *Asterocyclina* sp. and some spinose globigerinids. On this basis, Bryant & *al.* (1969) and Khudoley & Meyerhoff (1971) assigned the Cascarajícara Formation to the Middle or Upper Eocene. Nevertheless, *Asterocyclina* has also been reported from the Lower Eocene strata of Capdevila Formation, Habana Province (Brönnimann & Rigassi 1963). The available data may indicate the Late Maastrichtian age for Cascarajícara Formation but the reported presence of Paleogene microfauna (Hatten 1957) may prompt to attribute this formation tentatively to the Upper Maastrichtian to ? Lower Eocene.

The Cascarajícara Formation resembles deposits of Penalver Formation, Habana Province, described by Brönnimann & Rigassi (1963). The latter Formation has been attributed to the Upper Maastrichtian (Brönnimann & Rigassi 1963) although some other authors (Seiglie & Ayala-Castanares 1963) assigned it to the Paleocene.

BOUNDARIES

There is no good exposure of the lower boundary of the Formation. Nevertheless, it seems obvious that the Cascarajícara Formation overlies the upper chert member of Buenavista Formation (*cf.* Fig. 21). There is probably no disconformity between these lithostratigraphic units although the contact may be of erosional nature. The upper boundary of Cascarajícara Formation is tectonic.

LATERAL EXTENT AND THICKNESS

The Cascarajícara Formation occurs exclusively in the northern sequence of Sierra del Rosario, in the tectonic units of Sierra Chiquita and Cangre. In the former unit, this formation forms a continuous range some 50 km long.

The Formation attains its maximum thickness (450 m) in the type section and in the northeastern part of Cangre unit. In Sierra Chiquita unit, the thickness decreases northeastwards down to 100 m.

CORRELATION

The most close facies equivalent to the Cascarajícara Formation is Penalver Formation, Habana Province (*cf.* Pszczółkowski 1978a).

ANCÓN FORMATION

HISTORY

This formation has been recognized for the first time by Hatten (1957) and described formally by Herrera (1961). The type section is in the Valle del Ancón.

LITHOLOGY

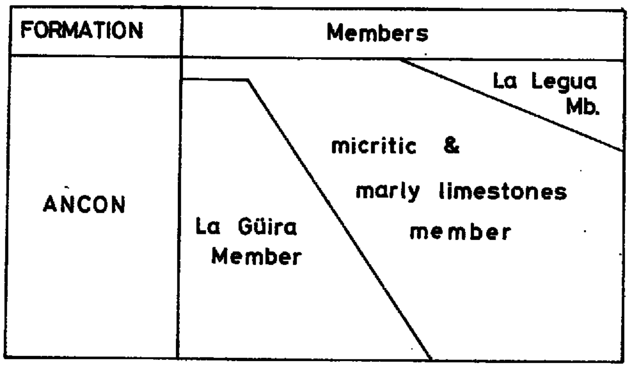
This formation includes well-bedded, often horizontally laminated, red, pink, green, and brown-yellow, marly and micritic limestones. The limestones may be intercalated with breccias. The breccias occurring at both the base and top of the marly and micritic limestones (Pszczółkowski & *al.* 1975) have also been attributed to Ancón Formation.

SUBDIVISION

Two members have been recognized within Ancón Formation, *viz.* La Güira and La Legua Members (Table 8). Moreover, one may also distinguish an informal lithostratigraphic unit called marly and micritic limestone member; the latter unit was previously recognized under another name (Myczyński *in:* Pszczółkowski & *al.* 1975).

Table 8

Ancón Formation subdivision



AGE

Planktic forams occur abundantly in the marly and micritic limestone member (Pl. 6, Fig. 4). On this basis, these strata were attributed to either the Upper Cretaceous to Lower Eocene (Herrera 1961), Upper Paleocene to Lower Eocene (Khudoley & Meyerhoff 1971), or Lower Eocene (Hatten 1957, Judoley & Furrazola-Bermúdez 1968). The present author is of the opinion that the new micropaleontological data (Torre 1972—1975) indicate the Late Paleocene age for the marly and micritic limestone member. In fact, the most abundant are *Globorotalia velascoensis* (Cushman), *Globorotalia* ex gr. *aequa* Cushman & Renz, and *Globigerina* cf. *pseudobulloides* (Plummer); less commonly occur *Globorotalia wilcoxensis* Cushman & Ponton, *G. brodermanni* Cushman & Bermúdez, *G.* cf. *acuta* Toulmin, *G. elongata* Glaessner, *G. occlusa* Loeblich & Tappan, *Planorotalia pseudomenardii* (Bolli), *Acarinina* cf. *soldadoensis* (Brönnimann), and *Globigerina velascoensis* (Cushman). The abundance of *Globorotalia velascoensis* (Cushman) indicates the Upper Paleocene Globorotalia velascoensis Zone (Bolli 1957, Caro & *al.* 1975, Salaj & *al.* 1976). The presence of *Planorotalia pseudomenardii* (Bolli) may indicate that the lower part of this member can also be assigned to Planorotalia pseudomenardii Zone (Postuma 1971, Caro & *al.* 1975).

The La Güira and La Legua Members are probably to be also attributed to the Upper Paleocene.

BOUNDARIES

There is a disconformity between the deposits of Ancón Formation and the underlying strata of Guasasa and Pons Formations (*see* below). The lower boundary of Ancón Formation is at the base of the marly and micritic limestone member or the breccia of La Güira Member.

The upper boundary of the Formation is at the top of the marly and micritic limestone member or the breccia of La Legua Member. The contact may be sharply marked (Pl. 1, Fig. 3) or it may occur within the beds transitional to Pica Pica Formation.

LATERAL EXTENT AND THICKNESS

The Ancón Formation occurs in all the tectonic units of the sequence of Sierra de los Organos. The total thickness ranges from 0 to 50 m.

CORRELATION

In part, the Ancón Formation is time equivalent to the strata of Buenavista Formation above the limestone breccia member (mostly in the southern sequence of Sierra del Rosario). Outside the Cordillera de Guaniguanico, the Ancón Formation may be time (but not facies) equivalent to the lowermost part of Capdevila Formation (*cf.* Pszczółkowski & *al.* 1975).

LA GÜIRA MEMBER

HISTORY

The La Güira Member has been recognized by Pszczółkowski (*in:* Pszczółkowski & *al.* 1975) in the lower part of Ancón Formation but its description has insofar not been published. The Member consists of a breccia related closely to the other deposits of Ancón Formation.

TYPE SECTION

The type section of the Member is by a road towards Caiguanabo, 550 m north to the road between La Palma and Entronque de Herradura; the topographic coordinates are 242.500 and 318.450. In this section, massive limestones of San Vicente Member of Guasasa Formation are overlaid by a breccia. The breccia consists mostly of limestone clasts (Pl. 1, Fig. 1) derived from diverse members of Guasasa Formation. Limestones of San Vicente Member and ammonite- and tintinnid-bearing limestones occur commonly among the clasts. There are also some chert fragments. The debris is poorly sorted. The clasts do not exceed 0.4 m in size. The matrix is poorly developed or invisible at all. The thickness of the breccia is 50 m.

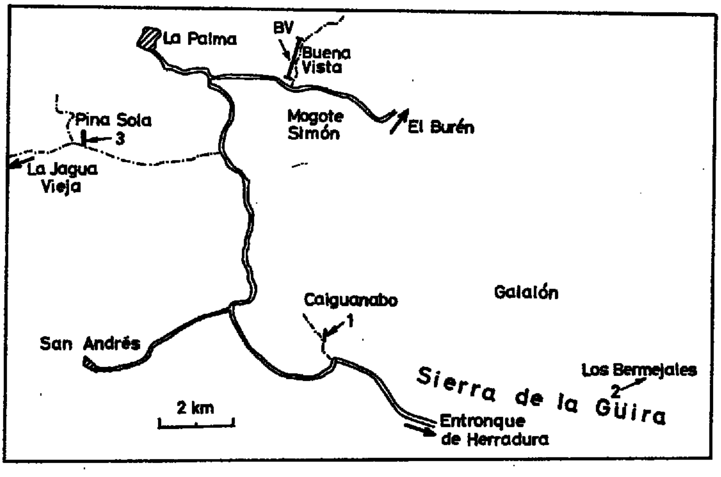


Fig. 23. Location map of the type section and hypostratotypes of La Güira Member of Ancón Formation, and the hypostratotype of Buenavista Formation
1 type section of La Güira Member, *2*, *3* hypostratotypes of La Güira Member, BV hypostratotype of Buenavista Formation

HYPOSTRATOTYPE I

This section is by the locality Los Bermejales, at the northern slope of the Sierra de la Güira (Fig. 23). There is exposed the upper part of La Güira Member represented by a limestone breccia, the clasts of which do not exceed a few centimeters in size. The clasts decrease in size upward and the breccia passes ultimately into micritic limestones of Ancón Formation (the marly and micritic limestone member). The thickness of the accessible section of the breccia is 15 m.

HYPOSTRATOTYPE II

This section is at Pina Sola, north to the road towards Jagua Vieja (Fig. 23). The breccia of La Güira Member is up to 16 m thick in this section and overlies directly massive limestones of San Vicente Member of Guasasa Formation, filling up fissures a dozen centimeters long. Lithologically, the breccia resembles closely that one exposed in the type section (*cf*. Pl. 1, Fig. 2); however in the upper part, a red micritic matrix appears, identical to red limestones of the marly and micritic limestone member of Ancón Formation.

BOUNDARIES

The type of the lower boundary of the Member is in the type section. An irregular surface of massive limestones of Guasasa Formation is overlaid by a coarse breccia of La Güira Member filling up erosional depressions. This boundary displays also the same characteristics in the hypostratotype II.

The type of the upper boundary of the Member has been designated in Pina Sola section (hypostratotype II), where La Güira Member is overlaid by red, micritic limestones of Ancón Formation. The contact is sharp.

AGE

Clasts of the breccia of La Güira Member contain the Tithonian, Berriasian to Valanginian, and Albian to Turonian microorganisms (Torre 1972—1975). There are also some Tithonian ammonites. As judged from the lithostratigraphic position, the breccia cannot be older than of the Maastrichtian to earliest Paleocene age, nor younger than of the Late Paleocene age. The occurrence of red matrix in Pina Sola section may indicate the Late Paleocene age of the clastic deposits.

The La Güira Member replaces laterally the marly and micritic limestone member of Ancón Formation (Table 8); nevertheless, it may locally be somewhat older than the latter deposits are.

LATERAL EXTENT AND THICKNESS

La Güira Member occurs in the sequence of Sierra de los Organos in the following tectonic units: Sierra de la Güira, Ancón, Vinales, Infierno, Valle de Pons, and La Legua (= ? Valle de Pons). Its thickness ranges from 0 to 50 m.

LA LEGUA MEMBER

This member has been distinguished in the upper part of Ancón Formation (Myczyński *in:* Pszczółkowski & *al.* 1975) after the data of Haczewski (1973). The following description of La Legua Member is but a brief account of the basic features.

The type section is by the locality La Legua, west to the Sierra de Mesa. There is a coarse breccia consisting mostly of limestone blocks up to 5 m in size. Chert clasts are less abundant. The matrix is usually invisible. The breccia attains 25 m in thickness.

The La Legua Member overlies the marly and micritic limestone member of Ancón Formation. In its turn, it underlies either marly limestones 0.5 m thick of Ancón Formation or yellow shales of Pica Pica Formation. The Member is of the Late Paleocene age. In other sections, the La Legua Member is less thick or lacking at all (*cf*. Table 8).

PICA PICA FORMATION

HISTORY

This lithostratigraphic unit has been proposed recently (Piotrowska, Pszczółkowski & Myczyński *in:* Pszczółkowski & *al.* 1975) to replace the Manacas and Canalete Cherts Formations of Hatten (1957) and the Pinos Formation of Herrera (1961). The name is after Valle de Pica Pica, southwestern part of the Sierra de los Organos.

The Manacas Formation of Hatten (1957) represents but a part of a variable lithologic succession comprising not only terrigenous deposits but also cherts and limestones. Moreover, the position of so-called „Vieja Wildflysch" appears in many sections quite different from that assumed by Hatten (1957). In fact, these deposits do not separate strata attributed by this author to the Manacas and Canalete Cherts Formations. The latter Formation cannot be accepted, as cherts occur commonly within diverse deposits in variable stratigraphic positions above the Ancón Formation. The Pinos Formation of Herrera (1961) is represented in the type section by chaotic rocks called previously „Vieja Wildflysch" (Hatten 1957) and described subsequently as a mélange (*cf.* Piotrowska 1976a, b). Thus, this formation does not include the strata assigned to the Manacas and Canalete Cherts Formations. Therefore, the Pica Pica Formation has been erected to replace the lithostratigraphic units created by Hatten (1957) and Herrera (1961). However, Manacas unit can be regarded as a member of Pica Pica Formation.

TYPE SECTION

The type section of Pica Pica Formation has been designated in Valle de Pica Pica, by the road between Sumidero and Gramales. It extends northwards to the place with topographic coordinates 294.300 and 197.700. The type section has been described in detail by Piotrowska (*in:* Pszczółkowski & al. 1975). Limestones of Ancón Formation are overlaid by sandstones and calcareous shales (10 m), polymictic sandstones intercalated with gray, micritic limestones (2.5 m), light-gray and red, micritic limestones (2.5 m), calcarenites with a tuffite material (3.0 m), yellow, tuffite shales (8.0 m), polymictic sandstones intercalated with shales and cherts (15.0 m), red, thin-bedded cherts (5.0 m), and at the top, polymictic sandstones with shale and breccia intercalations (20.0 m). Higher in the section, there are chaotic rocks. The total thickness of Pica Pica Formation is 85 m in its type section.

HYPOSTRATOTYPE

The hypostratotype of the Formation is by La Legua, west to the Sierra de Mesa. The topographic coordinates are 189.600 and 290.800. The La Legua Member of Ancón Formation is overlaid by yellow shales (5 m), gray, marly limestones intercalated with breccias (6 m), yellow shales and wacky sandstones (6 m), a breccia with limestone and chert clasts (2.5 m), shales and wacky sandstones (up to 30 m) and at the top, volcanics (diabases and andesites, according to Haczewski 1974) and tuffogenic rocks (30 m). The total thickness of Pica Pica Formation is almost 80 m in this section. These deposits are overlaid by chaotic rocks.

BOUNDARIES

The type of the lower boundary of the Formation, is in the type section, at the base of the first layer of polymictic sandstone overlying limestones of Ancón Formation. The transition is gradational.

The upper boundary of Pica Pica Formation is at the top of the last undisturbed layer, below chaotic rocks. In the hypostratotype, this boundary is at the top of volcanic rocks.

AGE

Deposits of Pica Pica Formation contain a lot of redeposited Upper Cretaceous forams, while the Paleogene microfauna is rather scarce. Hatten (1957) assigned his Manacas Formation to the Lower Eocene. Some planktic forams, *Globorotalia* ex gr. *velascoensis* (Cushman) among others, have been found in micritic limestones of the lower part of Pica Pica Formation (Torre 1972—1975, Pszczółkowski & *al.* 1975). As judged from these fossils, the lower part of the Formation can be assigned to the uppermost Paleocene and ? Lower Eocene. The overlying strata of Pica Pica Formation may represent the Lower and, in part, Middle Eocene.

SUBDIVISION

The deposits described by Hatten (1957) as Manacas Formation can be recognized for a member (Manacas Member) in the lower part of Pica Pica Formation. In the upper part of the Formation, two lithologic units occur most commonly, *viz.* the chert and sedimentary-volcanic ones (Myczyński *in:* Pszczółkowski & *al.* 1975); however, they are not treated as formal lithostratigraphic units.

LATERAL EXTENT AND THICKNESS

The Pica Pica Formation occurs in most tectonic units of the sequence of Sierra de los Organos. The lithology is highly variable among the units which makes the correlation of diverse stratigraphic sections very difficult. The maximum thickness may approximate 200 m but it is commonly tectonically reduced or repeated.

CORRELATION

The Pica Pica Formation may be time equivalent to the upper part of Buenavista Formation in the Sierra del Rosario. Outside the Cordillera de Guaniguanico, it is equivalent to most of Capdevila Formation, Pinar del Río Province.

REMARKS ON CHAOTIC ROCKS

Chaotic rocks occur in both the Sierra de los Organos ("Vieja Wildflysch" of Hatten 1957) and Sierra del Rosario ("Big Boulder Bed" of Pszczółkowski 1971b). Their nature has insofar not been unequivocally explained and hence, these rocks should for the moment not be regarded as a lithostratigraphic unit.

The considered rocks comprise considerable amounts of exotic material (*cf.* Hatten 1957, 1967; Pszczółkowski 1971b). They are usually tectonically disturbed which makes the interpretation difficult but may suggest that tectonic processes significantly contributed to their genesis or ultimate formation. A detailed description of the chaotic rocks of Sierra del Rosario is given by Pszczółkowski (*in:* Pszczółkowski & *al.* 1975). Some interpretive remarks are given further on, in this paper.

FACIES DEVELOPMENT AND PALEOGEOGRAPHY OF THE JURASSIC

SAN CAYETANO FORMATION

The sedimentary conditions of San Cayetano Formation have been recently recognized by Haczewski (1976) who distinguished 9 facies interpreted within a framework of deltaic and flysch sedimentation model. Some sections in both the Sierra de los Organos and Sierra del Rosario may provide further evidence of the sedimentary environment and facies variability of the upper part of San Cayetano Formation.

In the Sierra de los Organos, the San Cayetano Formation of Ancón tectonic unit (Fig. 2*A*) comprises but sporadically thick-bedded sandstones; while shales and siltstones occur commonly, intercalated with sandstones, often cross-bedded (Fig. 24). In both the cross-bedded sandstones and shales and siltstones, there are also some limestone intercalations ranging in thickness from a few centimeters up to 2 meters. These limestone beds contain a marine fauna, mostly trigoniid bivalves; the bivalves occur also in the sandstones. Such limestone intercalations appear already some 400 m below the top of San Cayetano Formation in Ancón tectonic unit

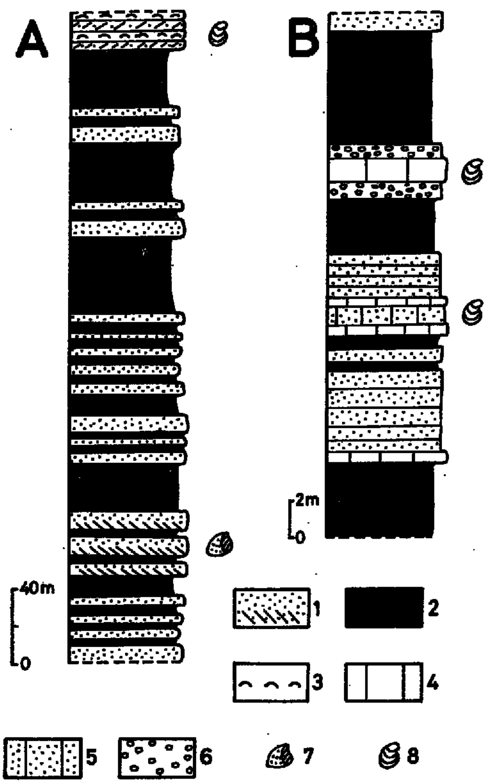


Fig. 24

The upper part of San Cayetano Formation in Ancón tectonic unit (Sierra de los Organos)

A — Composite section through the strata exposed south to El Abra

B — Section through the strata exposed south to the Sierra Ancón, some 400 m below the top of the Formation

1 sandstones, cross-bedded in places, *2* shales and siltstones, *3* bivalve coquina intercalations, *4* micritic and bioclastic limestones, *5* sandy limestones, *6* onkolitic, silicified limestones, *7* trigoniids, *8* ostreids

(Fig. 24*B*). There are some distinct microfacies among these limestones (throughout this paper, the terminology of Folk (1959) is generally used when describing carbonate microfacies).

Bivalve coquinas. Bivalve coquinas and bioclastic limestones composed mostly of fragmented bivalve shells appear as the most common microfacies. Sometimes, the shells are autochthonous; however, biocalcirudites and biocalcarenites occur more commonly. Ostreids may dominate among the fauna. There is usually sparitic to sandy-sparitic matrix. These bivalve shell accumulations might have originated owing to short sedimentary episodes, *e.g.* during storm periods. This interpretation may be especially plausible in the case of coquinas intercalating the cross-bedded sandstones (Fig. 25).

Biomicrites. These limestones represent usually thin intercalations. Apart from a micritic, somewhat marly matrix, they consist also of ostreid and other bivalve, gastropod, and sporadically echinoderm debris.

Oobiosparites and onkolitic limestones. Limestones with a considerable proportion of ooids are rather uncommon in San Cayetano Formation. They comprise fairly large amounts of bioclastic material. In Ancón

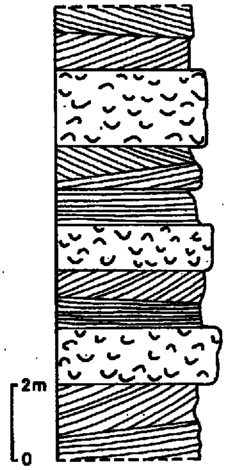


Fig. 25

Ostreid coquinas within cross-bedded sandstones of the uppermost part of San Cayetano Formation in Ancón tectonic unit, south to El Abra

tectonic unit, there are 2 layers of a silicified limestone composed mostly of onkoids (Pl. 4, Fig. 1; Fig. 24*B*) and ooids, some 400 m below the top of the Formation. The matrix was originally sparitic.

Sandy limestones. Detrital quartz may also significantly contribute to the limestones of San Cayetano Formation. It appears especially abundant in some micritic limestones.

The characteristics of terrigenous deposits intercalated with limestones, recorded in the upper part of San Cayetano Formation, Ancón tectonic unit, indicates a shallow-marine depositional environment. In fact, this area could be inundated by a shallow sea somewhat earlier than the rest of Sierra de los Organos delta. The investigated sediments were deposited in a shallow shelf environment influenced probably by a deltaic sedimentation. In the sequence of Sierra de los Organos, carbonate sedimentation started while the terrigenous deposits were still accumulating. Its extent could be dependent more upon terrigenous influx than upon other factors in the western Cuban sea at the beginning of the Late Jurassic.

SEDIMENTARY CONDITIONS OF THE OXFORDIAN CALCAREOUS AND ARGILLACEOUS DEPOSITS

The Oxfordian could not be delimited precisely in the Cordillera de Guaniguanico. Nevertheless, it seems reasonable to trace its lower boundary within the deposits of the upper part of San Cayetano Formation (Myczyński & Pszczółkowski 1976), and the upper boundary within the lower part of Artemisa Formation in the Sierra del Rosario and in the lowermost part of Guasasa Formation in the Sierra de los Organos (Kutek & *al.* 1976).

The depositional environment of Jagua Formation has been generally described by Hatten (1957) and Wierzbowski (1976), mostly after the fauna.

PAN DE AZÚCAR MEMBER

The Pan de Azúcar Member of Jagua Formation consists mostly of bioclastic limestones (Pl. 4, Fig. 3) and bivalve coquinas. They are usually dark-gray to black. The coquinas form layers 0.2 to 1.5 m thick and may occur in beds up to some meters thick; often, they represent but thin intercalations within the bioclastic deposits. The bivalve shells appear usually parallel to stratification. Cross-bedding occurs but occasionally and exclusively in sandy limestones. The mode of shell fragmentation and poorly developed stratification may suggest a significant bioturbation at least in some limestone layers. Coquina bases may be sharp and irregular thus, indicating the erosional nature (Fig. 4). The preservation state of bivalves is variable; disarticulated and somewhat rounded valves occur most commonly but some coquinas consist mostly of articulated shells. In some sections, a sandy calcirudite up to some tens centimeters thick occurs at the contact with San Cayetano Formation. This deposit comprises bioclastic, oosparitic, and pelsparitic limestone intraclasts and algal mat fragments (Pl. 4, Fig. 2). Thus, the onset of Pan de Azúcar Member accumulation was related to a temporary increase in hydrodynamic energy in the shallow-water zone of Sierra de los Organos basin.

Among the investigated deposits, some distinct microfacies can be recognized.

Bivalve-echinoderm microfacies. Apart from the bivalve coquinas, this is the most common deposit. Echinoderm debris are usually rather fine-grained. They are represented mostly by echinoid fragments, while crinoids occur but sporadically. Echinoderm debris may be more abundant than bivalve fragments.

Bivalve-gastropod microfacies. It is represented by some biocalcirudites comprising also ooids, aggregated lumps, peloids, intraclasts, detrital quartz, and some echinoderm debris.

Bivalve-algal microfacies. It occurs very rarely. The bioclastic matter is dominated by fragmented bivalve shells and algae (Dasycladaceae). Echinoderm debris occur in minor amounts.

Bivalve-pellet microfacies. It occurs but exceptionally, *e.g.* by the locality Mantua. The biopelmicrosparitic deposit contains abundant bivalve prodissoconchs and pellets less than 0.1 mm in size.

Bioclastic-oolitic microfacies. It appears fairly common in the deposits of Pan de Azúcar Member. Ooids occur usually in subordinate amounts. They are poorly developed and fairly small (less than 0.3 mm). There are also some lumps composed of agglutinated ooids. Onkoids may also occur, forming locally onkolitic or bioclastic-onkolitic limestones. Organic remains consist mostly of bivalve fragments.

The dark color of the limestones of Pan de Azúcar Member has resulted from the abundance of organic matter. Nevertheless, it is clearly indicated by the facies characteristics of the deposits that the basin was not stagnant. Anaerobic conditions could occur locally beneath the sediment-water interface but they did not influence the epifauna represented mostly by the bivalves Gryphaeinae. The latter animals lived on a soft bottom stabilized locally by shell accumulations. The bivalve shells are commonly encrusted by agglutinated forams resembling the genera *Tolypammina* or *Glomospira*; such forams occur usually in environments of slow sedimentation and a terrigenous influx (Dr. J. Kaźmierczak, *personal information*). Organic remains may also be encrusted with blue-green algae although such coatings occur less commonly.

The bioclastic sediments were probably deposited owing to short sedimentary episodes following destruction of autochthonous bivalve accumulations. The commonness of ooids indicates that there were also some areas with an oolitic sedimentation, nearby. In fact, ooids are commonly transported out to adjacent depositional environments in areas of present-day carbonate sedimentation (*cf.* Purdy 1963, Loreau & Purser 1973).

One may conclude that the bioclastic limestones and coquinas of Pan de Azúcar Member were deposited at a depth ranging from a few up to some tens meters, among some areas of sandy and clayey-silty sedimentation (*cf.* Fig. 3), in proximity of some oolitic shoals.

ZACARÍAS AND JAGUA VIEJA MEMBERS

The lithological characteristics of clayey-silty deposits of Zacarías Member of Jagua Formation indicates a relatively low-energy, offshore sedimentary environment, outside the area of a coarse terrigenous sedimentation. These deposits appear to have been but slightly bioturbated (*cf.* Wierzbowski 1976), which may suggest a fairly high sedimentation rate. The occurrence of wood imprints indicates that the environment was influenced by a deltaic sedimentation. Then, one may conclude that the deposits of Zacarías Member were probably accumulated in an outer shelf environment adjacent to a delta (*cf.* Walker 1971).

Deposits of Jagua Vieja Member have been recently studied by Wierzbowski (1976) who claimed that they had originated under reducing conditions and in proximity of shoreline. As judged from a change in the ammonite assemblage, the depositional environment was somewhat deeper than that of Zacarías Member sediments, although the water depth did not exceed 100 m.

One can see a horizontal lamination in limestones and shales of Jagua Vieja Member, especially in large calcareous concretions. The limestone set comprises micrites and biomicrites. The calcareous concretions often include ammonite accumulations (another macrofauna and wood fragments occur less commonly), the shells being undamaged and unabraded. The concretions are of early diagenetic origin (Wierzbowski 1976). They do not display any borings. There are also no indications of the concretions being excavated by hydrodynamic agents. Then, one may claim that the concretions originated within the deposit and their genesis was related to an organic matter decomposition (*cf.* Weeks 1953, Berner 1968, Kaźmierczak 1974). The calcium carbonate migrated toward places with high contents of organic matter (ammonite accumulations, fish, large wood fragments, bones). The fish remains are usually flattened which shows that concretions originated under a deposit cover; however, this burden was insufficient to damage the ammonite shells. In fact, the shells are filled with either a homogenous, dark calcareous sediment, or a coarse sparitic calcite with minor amounts of bituminous matter. Thus, some shells were void during the cementation, while others comprised still soft tissues (*cf.* Zangerl & *al.* 1969).

The sedimentation rate of Jagua Vieja Member could not be very high, as indicated by considerable accumulations of ammonite shells in the deposits. Moreover, the occurrence of wood appears as the only reflection of a relation to deltaic sedimentation; one should, however, keep in mind that wood pieces may float pretty far away. In summary, the deposits of Jagua Vieja Member were probably accumulated in a deep shelf environment. This facies occurs commonly in the Sierra de los Organos, while its equivalents occur also in some tectonic units of the Sierra del Rosario (Francisco Formation).

PIMIENTA MEMBER AND THE LOWERMOST ARTEMISA FORMATION

Terrigenous matter is almost lacking in limestones of Pimienta Member of Jagua Formation. There are a few ammonites (Myczyński 1976) and planktic microorganisms (forams and *Globochaete*) in these strata. Thus, one may claim that these deposits were accumulated in a deep shelf environment. This conclusion is also supported by the lack of perisphinctid ammonites and the preponderance of aspidoceratids (*cf.* Ziegler 1967, 1971).

The upper part of Pimienta Member is time equivalent to limestones of the lowermost Artemisa Formation in the Sierra del Rosario (Kutek & *al.* 1976). The latter deposits may be intercalated with thin shale layers and sporadically with sandstones and breccias comprising shallow-water limestone debris, quartz, and quartz sandstone pebbles. These intercalations indicate that the depositional environment was influenced (even although slightly) by a terrigenous influx. At the base of the Formation, there are some possible polychaete borings and accumulations of phosphate coprolites (Pszczółkowski *in*: Kutek & *al.* 1976).

OUTLINE OF THE OXFORDIAN PALEOGEOGRAPHY

There are both terrigenous and calcareous deposits in the Jurassic of the Cordillera de Guaniguanico. Terrigenous deposits occur mostly in San Cayetano Formation in the Sierra de los Organos; they originated under the conditions of a deltaic sedimentation that persisted over a fairly long span of time (Khudoley *in*: Khudoley & Meyerhoff 1971, Pszczółkowski 1971b, Haczewski 1976). The lower part of the Formation comprises fluvial deposits (*cf.* Haczewski 1976). In the upper part, terrigenous-calcareous sediments occur commonly with an abundant marine fauna. Thus, there is a transgressive facies succession in the sequence of Sierra de los Organos.

The petrography of clastic deposits of the upper part of San Cayetano Formation in the Sierra del Rosario may indicate that the terrigenous material came from a land built up by terrigenous, calcareous, metamorphic, and igneous rocks. The debris of vitreous volcanics and alkalic lavas might be derived from the Jurassic volcanic rocks reported from the southern part of Sierra de los Organos (Piotrowski 1976). There are no indications of terrigenous influx from any source areas occurring just within the depositional basin of San Cayetano Formation, as it was suggested by some authors (Meyerhoff *in*: Khudoley & Meyerhoff 1971, Meyerhoff & Hatten 1974).

In general, the present author approves the paleogeography as proposed by Haczewski (1976, Fig. 14) for San Cayetano Formation. Haczewski (1976) based his conclusions on the assumption of terrigenous influx mostly from the south or SSW, whereas all paleocurrent observations indicating a transport from the northeast have been regarded as insignificant. Post-sedimentary, tectonic bias in the present facies distribution has not been considered by that author. Moreover, some deposits of the southern sequence of the Sierra del Rosario resembling the facies G of Haczewski (1976) do not represent flysch. Proximal turbidites occur in some tectonic units in the Sierra del Rosario, up to 70 km one place away from another.

The uppermost San Cayetano Formation in the Sierra del Rosario comprises a scarce ammonite fauna attributed to the Perisphinctidae and Oppeliidae. There are also some decapods, ostreids, and fish remains. This fauna may indicate that the water depth did not exceed 300 m in the sedimentary environment of these deposits (*cf.* Ziegler 1967, 1971). This was also the case for the lowermost part of Francisco Formation. At the Oxfordian time, the basin of Sierra del Rosario did not receive any considerable amounts of bioclastic material. Coquinitic and bioclastic deposits of the sequence of Sierra de los Organos were not affected by those processes transporting deltaic and shallow-water sediments towards deeper parts of the basin, that is toward the sequence of Sierra del Rosario (*cf.* Haczewski 1976). As judged from the present-day, post-tectonic facies distribution, the paleoslope declined towards the north or NE. Flysch and flysch-like deposits of the sequence of Sierra del Rosario are bordering along a tectonic contact upon shallow-water sediments of the sequence of Sierra de los Organos. Transitional deposits may be partly covered by the overthrust units of the Sierra del Rosario (Pszczółkowski 1976b).

During the later Middle Oxfordian, a calcareous and argillaceous sedimentation spread over the whole area of Sierra de los Organos (Fig. 26). At that time, sediments of Jagua Vieja and lowermost Pimienta Members were deposited in the Sierra de los Organos. The same sedimentation type prevailed also in the southern sequence of Sierra del Rosario (Francisco Formation) although psammitic influx continued in some places. Here and there in the Sierra del Rosario, the terrigenous deposition of San Cayetano Formation still persisted (Fig. 26).

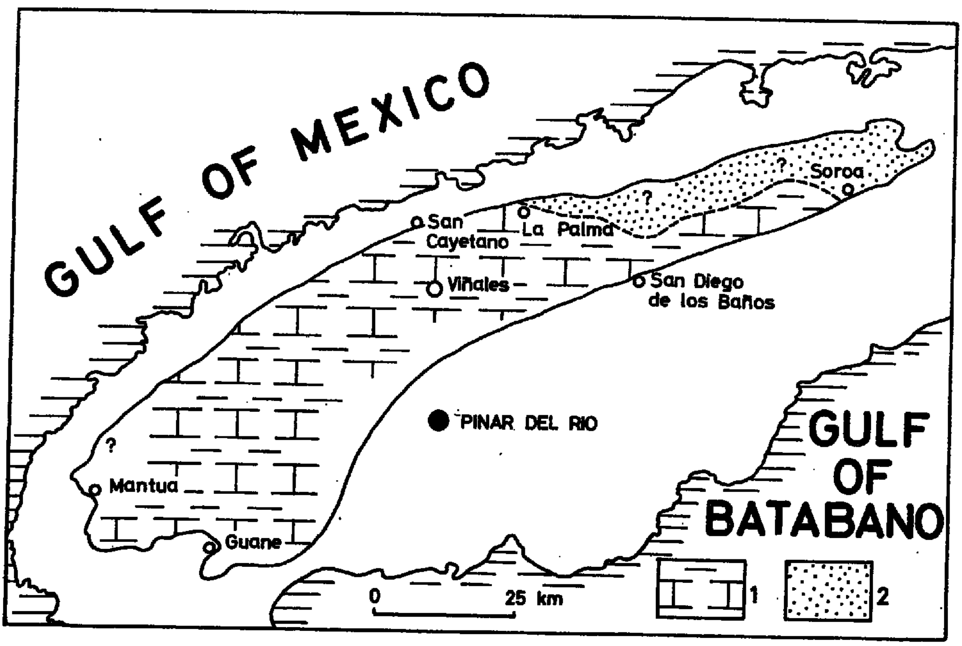


Fig. 26. Lithofacies pattern of the upper Middle Oxfordian of the Cordillera de Guaniguanico

1 calcareous and clayey deposits, *2* terrigenous, mostly sandy deposits

At the Middle and Late Oxfordian boundary, limestones of Pimienta Member accumulated in the sequence of Sierra de los Organos, while the deposition of Artemisa Formation started in the southern sequence of Sierra del Rosario (*cf.* Kutek & *al.* 1976, Myczyński 1976). Terrigenous intercalations in Artemisa Formation limestones originated owing tc either a redeposition of older deltaic and/or shallow-water sediments, or an influx from an active delta outside the area covered by the sequences of Sierra de los Organos and Sierra del Rosario.

MICROFACIES OF SAN VICENTE MEMBER LIMESTONES

Limestones of San Vicente Member of Guasasa Formation are described after a thin-section analysis. The samples have been derived from 6 lithostratigraphic sections (Figs 5—7) in the tectonic units of Viñales, Sierra de la Güira, and Ancón, eastern part of the Sierra de los Organos (*cf.* Fig. 2*A*).

COPROLITE-BEARING MICRITES

Micrites and pelmicrites of San Vicente Member comprise commonly coprolites of decapod crustacean origin (Fig. 5*A*, samples *1V5* and *1V9*; Fig. 5*B*, sample *6P-118*; Fig. 6*A*, sample *2An4*; Fig. 7*B*, sample *6P-70*, *6P-77* and *6P-81*). These coprolites are elongate (up to 1.5 mm) and their cross-sections are ovate. Seiglie (1961) attributed them to *Favreina*. However, at least some of these coprolites may be assigned to *Parafavreina* Brönnimann, Caron & Zaninetti, as the channels are usually obscured by a recrystallization of their calcium carbonate fill (*cf.* Brönnimann 1972). Some specimens resemble also those described by Brönnimann (1955) as *Favreina joukovskyi* Brönnimann, and attributed later on (Brönnimann 1976) to *F. salevensis* (Paréjas). Throughout this paper, all these crustacean coprolites (*cf.* Paréjas 1948, Brönnimann & Norton 1960) will be termed favreines or *Favreina*-form coprolites, because of their similarity in both the form and occurrence. Apart from the favreines, the micrites comprise also sparitic or microsparitic structures regular in form and up to 1.5 mm in size. These are either some other coprolites or entirely recrystallized favreines, the former possibility being more plausible.

The coprolite-bearing micrites occur mostly at the base of the Member, sometimes in limestone units up to some tens meters thick.

PELSPARITES

These deposits consist of peloids cemented by a sparite or microsparite. Most peloids appear to be fecal pellets. Apart from the peloids, this microfacies comprises also some *Favreina*-form coprolites, and ooids or onkoids. Some of these pelsparites were originally pelmicrites. The pelsparitic deposits have been recorded in the sections of Sierra de Viñales (Fig. 5*A*, sample *2V1*) and El Abra (Fig. 7*B*, sample *6P-87*), among others.

OOMICRITES AND OOSPARITES

Sediments dominated by ooids (Pl. 5, Fig. 2) are rather uncommon in San Vicente Member. In Sierra de la Güira section, there are oomicrites at the base of the Member (Fig. 6*B*, sample *6P-187*), with small ooids of a radial structure, dispersed in a micritic matrix. The ooid nuclei are also micritic. This sediment accumulated probably in less agitated environment. In the Sierra de Viñales section there are oosparites at the top of the Member, consisting of ooids 0.5 to 1.0 mm in diameter and concentric in structure. Actually, some of these ooids may be small onkoids ("pisonkolites" *sensu* Kutek & Radwański 1965). There are also some aggregated lumps (Pl. 5, Figs 3—4) and bioclasts in these sediments. The bioclasts are represented mostly by gastropod and echinoderm fragments and small benthic forams (Textulariidae among others). The matrix is coarse sparitic.

Some oolitic deposits represent actually oobiosparites or favreine-bearing oosparites (Fig. 6*A*, sample *2An2*).

BIOSPARITES

Deposits consisting exclusively or mostly of organic remains are rather uncommon in San Vicente Member. This microfacies (Fig. 5*B*, samples *6P-109* and *6P-110*; Fig. 7*A*, samples *AP-30* and *AP-32*) is dominated by algal biosparites consisting mostly of fragmented blue-green and codiacean algae (up to 60%) with minor amounts of coralline, solenoporacean and dasycladacean debris. There are also some echinoderm fragments and benthic forams. Sporadically, bivalve coquinitic limestones do also occur (Fig. 6*B*, middle part of the section and sample *6P-192*).

BIOMICRITES

Some micritic limestones contain fragmented bivalve shells (Fig. 7*A*, sample *AP-35b*). There are also some darker, micritic grains of unclear origin. These deposits represent low-energy sedimentary environment. They could be considerably reworked by burrowing organisms.

ONKOLITIC LIMESTONES

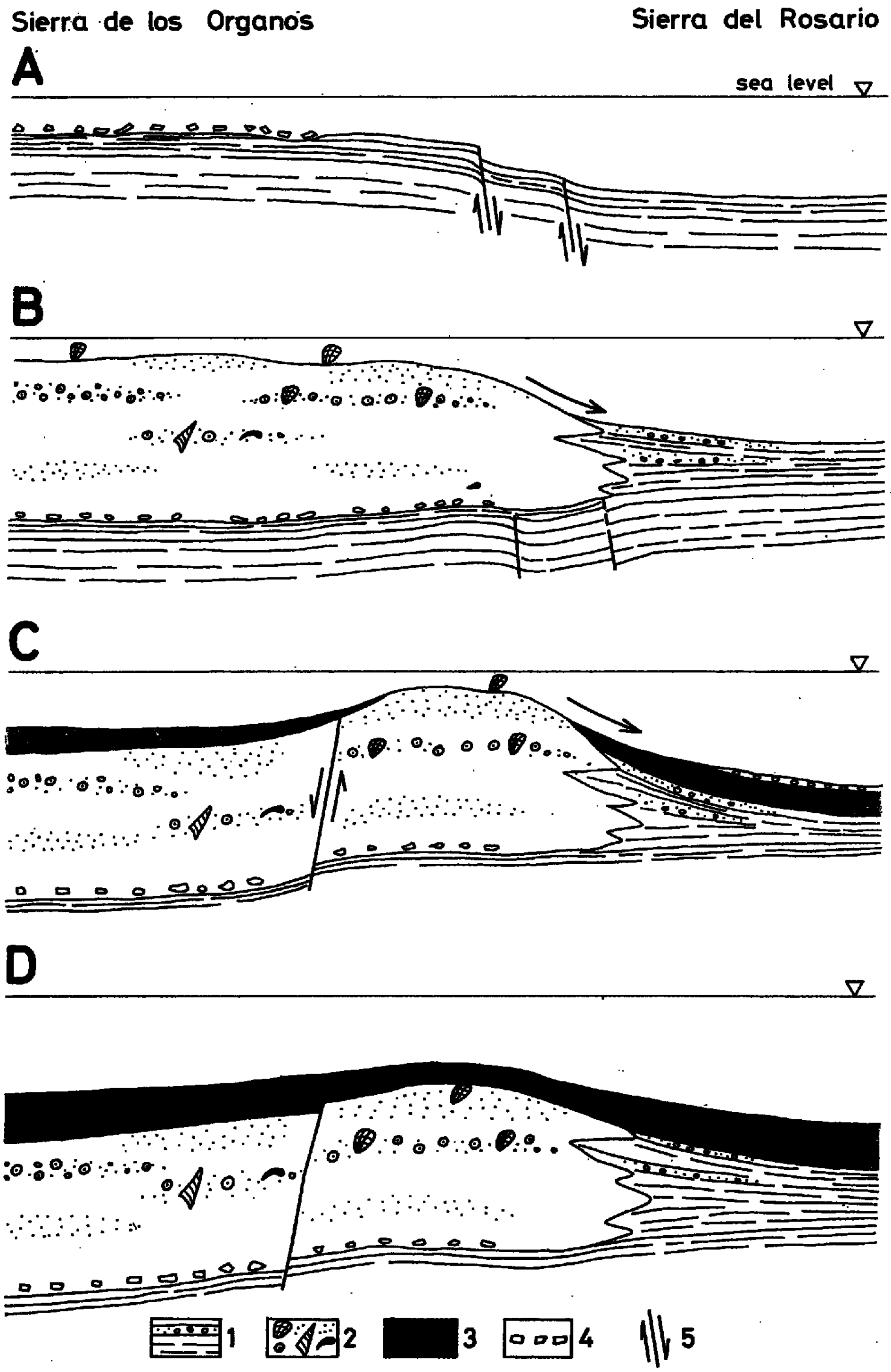
In the upper part of San Vicente Member, there are calcarenites and calcirudites. They include also onkolitic limestones. The onkoids attain some 4 mm in size and their concentric structures are usually poorly developed. In some cases, the onkolitic envelops are developed around organic remains. Limestones with larger onkoids are quite easily distinguishable in the field. They occur in beds up to some tens centimeters thick or in limestone units up to a dozen meters thick (Fig. 7*A*).

INTRAOOSPARITES AND INTRABIOSPARITES

Calcarenites composed of intraclasts, ooids, and bioclasts occur fairly commonly at the top of San Vicente Member (Fig. 5*A*, samples *2V2* and *1V14*; Fig. 6*A*, sample *2An1*; Fig. 7*A*, sample *AP-35a*). The ooids represent but a subordinate constituent of the deposit. The intraclasts are usually micritic, often rounded. There are some aggregated lumps composed of intraclasts and/or ooids. The intrabiosparites contain fragmented echinoderms and bivalves, some benthic forams, and algal debris (*Clypeina* sp. among others). The matrix is sparitic or microsparitic. The deposits are poorly sorted. In places (*e.g.* at the top of the Member in San Vicente section), these intrabiosparites and intraoosparites indicate high-energy depositional environment.

SEDIMENTARY CONDITIONS AND PALEOGEOGRAPHY OF THE KIMMERIDGIAN AND LOWERMOST TITHONIAN

At the Oxfordian decline, a shallow-water carbonate sedimentation started in the Sierra de los Organos. In some sections, these deposits attain 500 to 650 m in thickness. Their sedimentary structures and


Sierra de los Organos
Sierra del Rosario
A
sea level
B
C
D
1
2
3
4
5

microfacies characteristics indicate moderately agitated depositional environment. This is evidenced by the occurrence of crustacean-coprolite bearing micrites, pelmicrites, and calcarenites with abundant lumps and onkoids, and on the other side by the scarcity of pure oolitic limestones and the lack of large-scale cross bedding. The present-day micritic sediments with abundant fecal pellets are usually formed in fairly deep and/or sheltered areas (Purdy 1963).

The lower part of San Vicente Member is often dominated by these low--turbulence deposits, whereas the calcarenites and calcirudites prevail at the top. Both these sediment types may occur in the middle of the Member but that part of the section is often obscured by a considerable dolomitization. Such a sedimentary sequence reflects a change in hydrodynamic regime during the deposition of San Vicente Member. This change could have resulted from some changes in bottom configuration rather than from a gradual decrease in water depth. At the top of San Vicente Member and within the strata transitional to El Americano Member, there are some units of thin-bedded micrites and calcilutites (Figs 6*A* and 7*A*), which may indicate a lateral interfingering of the shallow-water and basin facies.

The San Vicente Member does not comprise any biolithites (*sensu* Folk 1959). There are also no detritic deposits comparable to the coralgal facies of Great Bahama Bank (Purdy 1963) or the coralgal sands of Persian Gulf (Wagner & Todt 1973). In fact, there are some algal biosparites in San Vicente Member which might be regarded as facies equivalent to these modern coralgal sediments; however, they occur rather uncommonly. Then, one may conclude that the shallow-water Upper Jurassic carbonates of the Sierra de los Organos were not related to any active coral or coralgal reefs.

There are horizons with abundant organic burrows in some sections of San Vicente Member (Fig. 5B). They separate the micritic and pelmicritic deposits from the overlying calcarenites and calcirudites. The contacts appear to be erosional and the detritic sediments display sometimes graded bedding. One may claim that these sediments resulted from singular sedimentary events related to sudden resuspension and subsequent deposition of the sediments in adjacent portions of the basin. Such effects may be caused by storms, cyclones, or tsunami (Kelling & Mullin 1975).

Rocks assigned to San Vicente Member do also occur in some sections of the southern sequence of Sierra del Rosario. However, these strata do not exceed some tens meters in thickness which contrasts to their facies equivalents in the Sierra de los Organos. The lower part of Artemisa Formation may comprise some calcarenite and calcirudite intercalations (Figs 11*B* and 12) with echinoderm, bivalve and algal bioclasts, and ooids, onkoids, pellets, and lithoclasts typical of shallow-water

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Fig. 27. Development of the carbonate bank of Sierra de los Organos

A — Late Oxfordian: uplift (or tilt) at the end of the deposition of Jagua Formation (*cf.* Hatten 1957)

B — Oxfordian decline to earliest Tithonian: shallow-water carbonate sedimentation; the arrow indicates transport direction of the shallow-water deposits to the Sierra del Rosario basin (Artemisa Formation)

C — Tithonian: submergence of most of the carbonate bank; the shallow-water sediments were still redeposited in the Sierra del Rosario basin

D — Berriasian: end of the shallow-water sedimentation all over the Sierra de los Organos

1 well-bedded, micritic limestones intercalated with calcarenites and calcirudites comprising shallow-water material, *2* shallow-water carbonate deposits comprising algal detritus, gastropods, bivalves, ooids, and onkoids, *3* pelagic limestones, *4* breccias at the base of shallow--water limestones, *5* hypothetic synsedimentary faults carbonate deposits. Some of these layers display graded bedding and border sharply upon the underlying micrites or shales. The latter deposits contain a few poorly preserved planktic microorganisms (*Globochaete alpina* Lombard, radiolarians, and ?stomiosphaerids). Thus, one may conclude that the detiritic limestones are allochthonous, their clasts being derived mostly from some shallow-water areas; while most micritic limestones and shales are autochthonous. The latter deposits prevail by far in the lower portions of La Zarza Member, in Cinco Pesos and La Zarza tectonic units. Their small thickness relative to their time equivalents in the sequence of Sierra de los Organos (San Vicente Member) indicates that the sedimentation rate was lower in the southern sequence of Sierra del Rosario than in the Sierra de los Organos.

The pre-Tithonian deposits of Artemisa Formation in the Sierra del Rosario do not exhibit any features typical of true pelagic sediments (*cf.* Garrison & Fischer 1969). They could be accumulated in a zone transitional to deeper basin or within a partly isolated environment. In fact, this is suggested by the lack of autochthonous benthic or nektic fauna, except of two base beds of the Formation.

All the preceding paragraphs show that a shallow-water carbonate sedimentation developed since the Late Oxfordian to Early Tithonian time all over the Sierra de los Organos (Fig. 27), under the conditions of a subsidence allowing the accumulation of deposits 500 to 650 meters thick. As judged from the present-day extension of these deposits in the Sierra de los Organos (with tectonic bias taken into account), this sedimentation type spread over an area of 125 km in length and 70 to 120 km in width. Furthermore, the boundaries of the region of Sierra de los Organos are of tectonic nature and hence, one may claim that actually the carbonate bank was much larger; in fact, facies and time(?) equivalents of considered sediments may also be present at Isla de Pinos (*cf.* Somin & Millán 1972).

This shallow-water carbonate bank of the Sierra de los Organos was somewhat different from the Mediterranean Mesozoic carbonate "platforms" (*cf.* Aubouin 1965, Jenkyns 1970, Catalano & al. 1974) and the modern Bahama Banks (*cf.* Purdy 1963, Hine & Neumann 1977). The lack of terrigenous material in the considered deposits of the Sierra de los Organos shows that the carbonate bank was separated from eroded land masses (*cf.* Meyerhoff *in*: Khudoley & Meyerhoff 1971).

The calcareous and argillaceous deposits of Sierra del Rosario were accumulated in somewhat deeper and less agitated environment at the margins of Sierra de los Organos carbonate bank (Fig. 27). This marginal zone was at least 30 to 50 km wide.

SEDIMENTARY CONDITIONS OF THE TITHONIAN

At the Early Tithonian time, the shallow-water carbonate sedimentation stopped in the Sierra de los Organos. The calcarenites are commonly overlaid by deposits of El Americano Member comprising the Tithonian ammonites and microorganisms, and a scarce benthic fauna.

Shallow-water limestones of Guasasa Formation are usually overlaid by *Saccocoma*- and/or tintinnid-bearing biomicrites and biocalcilutites (Figs 5*A*, 7*A*, 8). The latter sediments may be replaced laterally with intrasparites and intramicrosparites with tintinnids, planktic crinoids, and cephalopod and fish remains. Sedimentary discontinuities, traces of submarine erosion, and considerable accumulations of ammonites and other organic remains appear typical of slow sedimentation conditions. The uppermost Tithonian strata are dominated by tintinnid- and radiolarian-bearing biomicrites and biocalcilutites.

The above-described Tithonian deposits are equivalent to micrites intercalated with shales and calcarenites, and ammonite-bearing bioclastic limestones of the upper part of La Zarza Member in the Sierra del Rosario. The micrites contain poorly preserved planktic microorganisms such as *Cadosina* sp., radiolarians, stomiosphaerids, and *Saccocoma* sp. The ammonite-bearing limestones are biomicrites and biosparites often literally filled up with ammonite shells and aptychi. Some Tithonian biosparites of the Sierra del Rosario are composed mostly of fragmented fish remains and/or bivalve shells. In some sections of the southern sequence of Sierra del Rosario, micritic and pelagic deposits comprise intercalations of polymictic calcarenites (Fig. 13B) with ooids, *Favreina*-form coprolites (usually under the form of bioclasts), and other shallow-water constituents. These calcarenites attain up to 10 m in thickness and are distinctly allochthonous. They could originate due to a redeposition of shallow-water sediments to a deeper basin; turbidite currents contributed probably to this process. In fact, this inference is also confirmed by the occurrence of submarine-slump breccias (Fig. 28) in the Tithonian strata (*cf.* also Fig. 12). The breccias are composed of shallow-water matter intermixed with limestone clasts containing ammonites and aptychi. These redeposited shallow-water sediments occur most commonly in Taco Taco tectonic unit (Fig. 13B), while they are less common or lacking at all in other units. This

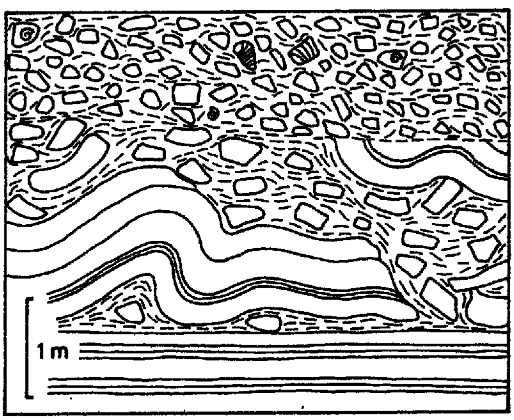

Fig. 28

Slump breccia in the Tithonian limestones of the southern sequence of Sierra del Rosario by the locality Sabanilla (*cf.* Text-fig. 12); at the top, there are both ammonite-bearing and shallow-water fauna-bearing clasts

may be the cause of a distinct variability in thickness of the Tithonian deposits in the Sierra del Rosario. In most sections, these deposits range in thickness from 40 to 60 m, whereas they attain 100 to 120 m in Taco Taco tectonic unit. The Tithonian sediments do usually not exceed 60 m in thickness in the Sierra de los Organos; moreover, they are quite similar to those exposed in the Sierra del Rosario. Hence, one may claim that the carbonate bank of Sierra de los Organos had drowned at the Middle Tithonian time which resulted in a considerable uniformity of facies all over the Cordillera de Guaniguanico (Fig. 27). The end of the shallow-water sedimentation was not followed by an accumulation of typical condensed deposits reported commonly from the Mediterranean (*cf.* Jenkyns & Torrens 1971). At the Tithonian time, the sedimentation rate of the pelagic and related deposits (8 to 10 m/my) was some ten times lower than that of the older shallow-water carbonates (80 to 100 m/my).

The shallow-water sedimentation ended at differential moments all over the carbonate bank, as evidenced by the occurrence of allochthonous shallow-water deposits throughout the whole Tithonian in the southern sequence of Sierra del Rosario. The paleoslope remained probably declined toward the Sierra del Rosario at the Tithonian time, just as it had been previously.

FACIES DEVELOPMENT AND PALEOGEOGRAPHY OF THE LOWER CRETACEOUS

PELAGIC SEDIMENTS OF THE BERRIASIAN AND VALANGINIAN

At the Jurassic and Cretaceous boundary, a pelagic sedimentation spread all over the Cordillera de Guaniguanico, except of the sequence of Quiñones. At the Berriasian time, tintinnid- and radiolarian-bearing limestones were deposited in the Sierra de los Organos (Tumbadero and Tumbitas Members of Guasasa Formation); similar sediments were also accumulated in the Sierra del Rosario (Sumidero Member). In the sequence of Sierra de los Organos, the Berriasian to lowermost Valanginian deposits are dominated by tintinnid and tintinnid-globochaete microfacies; while other microfacies are less common (*e.g.* the radiolarian-tintinnid one; Pl. 4, Fig. 4). The Valanginian limestones are dominated by tintinnid and tintinnid-nannoconus microfacies. At the same time, radiolarian and radiolarian-tintinnid microfacies have been prevalent in the Sierra del Rosario. The radiolarians are usually calcified in the Berriasian to Valanginian limestones and hence, their structures appear usually obscure. The radiolarian biocalcilutites may be horizontally laminated. The pelagic deposits of the Berriasian to Valanginian age resemble in microfacies the Late Jurassic to Early Cretaceous limestones of the Austrian Alps (*cf.* Garrison 1967).

In general, these pelagic limestones do not contain any terrigenous material. However, in some sections of the southern sequence of Sierra del Rosario (Fig. 13*B*), there are in the Berriasian some singular beds of allochthonous, ooid-bearing calcarenites. Aptychi may occur abundantly in some strata. The maximum extension of calcareous pelagic sedimentation was but at the Berriasian and earliest Valanginian time. At the Late Valanginian time, a terrigenous influx (turbidite sandstones) did already occur in the northern sequence of Sierra del Rosario.

Such pelagic facies have usually been interpreted as deep-water sediments, deposited approximately at the CCD (*cf.* Colom 1967, Garrison 1967); however, according to some authors, this interpretation might be wrong (Hallam 1971, Kuhry & *al.* 1976). In the case of the pelagic sediments of Sierra de los Organos, the Upper Jurassic to Lower Cretaceous facies succession and a decrease in sedimentation rate (5 to 7 m/my) relative to the Tithonian, may suggest a gradual sinking of the basin during the Berriasian and Valanginian (Fig. 27).

FACIES PATTERN AND PALEOGEOGRAPHY OF THE HAUTERIVIAN TO BARREMIAN

At the Valanginian and Hauterivian boundary, calcareous and siliceous pelagic deposits of the southern sequence of Sierra del Rosario were successively passing into radiolarian cherts and shales of Sabanilla Member of Buenavista Formation. At that time, the pelagic limestone sedimentation persisted still in the sequence of Sierra de los Organos. Thin intercalations of turbidite sandstones appear within pelagic limestones of the northern sequence of Sierra del Rosario; they are uncommon at the beginning but become more numerous during the Hauterivian. Radiolarian or radiolarian-spicule microfacies are typical of the Hauterivian to

Barremian pelagic limestones. There are a few planktic forams in these microfacies in Polier Formation. In the latter deposits, ammonites are usually preserved but under the form of imprints, which indicates that the shells have been dissolved. These pelagic sediments contain also some rhyncholites and abundant aptychi (Houša 1969, 1974a). All these characteristics may suggest that one deals here with deep-water sediments deposited above the CCD.

Radiolarian cherts, radiolarites, and silicified shales of the southern sequence of Sierra del Rosario could be deposited in the deepest part of the Early Cretaceous basin of the Cordillera de Guaniguanico. Their origin does not appear to have been directly related to a submarine volcanism, as there are no volcanic rocks in Sabanilla Member itself or its time equivalents. These siliceous-clayey deposits are rather thin due to a very low sedimentation rate. They could be accumulated near the CCD that ranged at the Cretaceous time probably from 3500 to 4500 m in the Atlantic and Pacific Oceans (*cf.* Hesse & Butt 1976).

The Hauterivian to Barremian terrigenous sediments of the northern sequence of Sierra del Rosario display the features typical of a distal flysch (*cf.* Walker 1967). However, in some sections, they may rather resemble a normal flysch. The transport directions are very difficult to determine because of intense tectonic deformations and poor exposures. They have been reliably measured but in a dozen or so exposures, mostly after flute casts and cross-lamination; *viz.* at Lomas de Polier (120°—180°), El Aguacatillo (100°—130°), Loma Caldoso (135°—180°), Loma del Mulo (135°), south to Mango Bonito (120°), by the locality Plan Rosario (along the N—S axis), at Moreno by Santiago River (along the WNW—ESE axis), and at Braciliano Roble (WNW—ESE). These data suggest that the terrigenous influx was mostly from NW and partly from the north. However, this inference cannot be regarded as conclusive, since it is based on but a few measurements.

The flysch sandstones occur most commonly in the tectonic units of Sierra Chiquita and Cangre (*cf.* Fig. 2*A*), while their amounts decrease markedly southwards. There are no sandstones at all in the Hauterivian to Barremian deposits of the southern sequence of Sierra del Rosario. In the sequence of Quiñones, turbidite sandstones occur but sporadically. Then, the terrigenous influx could not be from the areas of Sierra de los Organos, Quiñones sequence, or southern Sierra del Rosario. The terrigenous material might be transported along the NE—SW (*cf.* Pszczółkowski 1971b) or NW—SE axes. As the matter of fact, the determined transport directions may rather indicate that the influx was from the northwest (*cf* Fig. 29*A*). However, a subsequent tectonic rotation might also occur in the Sierra del Rosario, changing the original transport directions of clastic material.

Thus, one may conclude that there was a distinct variability in sedimentation type among the stratigraphic sequences of Cordillera de Guaniguanico at the Hauterivian to Barremian time. The Early Cretaceous basin was split into some facies zones (Fig. 29*A*). The sequences of Sierra de los Organos and Quiñones represented an area of calcareous pelagic sedimentation; a subordinate terrigenous influx did also occur in the latter sequence. This area was somewhat shallower than the zone of siliceous-clayey sediments of the southern sequence of Sierra del Rosario. The latter zone must have been separated at that time from a turbidite basin of the northern sequence of Sierra del Rosario. Such a barrier might be the narrow zone of calcareous pelagic deposition.

PALEOGEOGRAPHY AT THE EARLY CRETACEOUS DECLINE

The Aptian to Albian deposits occur in the sequence of Sierra de los Organos (Hatten 1957, 1967; Khudoley & Meyerhoff 1971; Torre 1972—

1975) within a continuous succession of pelagic limestones of Pons and Guasasa Formations. In the southern sequence of Sierra del Rosario, the contemporaneous strata include cherts and shales of Sabanilla Member; whereas, both flysch and pelagic (mostly siliceous) sediments of that age occur in the northern sequence of Sierra del Rosario. In the sequence of Quiñones, the Late Barremian to Albian strata are represented by shales and radiolarian cherts of the lowermost part of Sierra Azúl Formation.

Flysch deposits of the northern sequence of Sierra del Rosario resemble a normal flysch and in places a proximal one. There are tool casts indicating the currents along the WNW—ESE axis. The Early Cretaceous basin of the Sierra del Rosario received terrigenous clastic material from a land emerged over at least 25 my, and built up by terrigenous sedimentary rocks and/or metamorphic and acid igneous rocks. A few measurements of the transport directions suggest that the land occurred northwest to the depositional basin of Polier Formation.

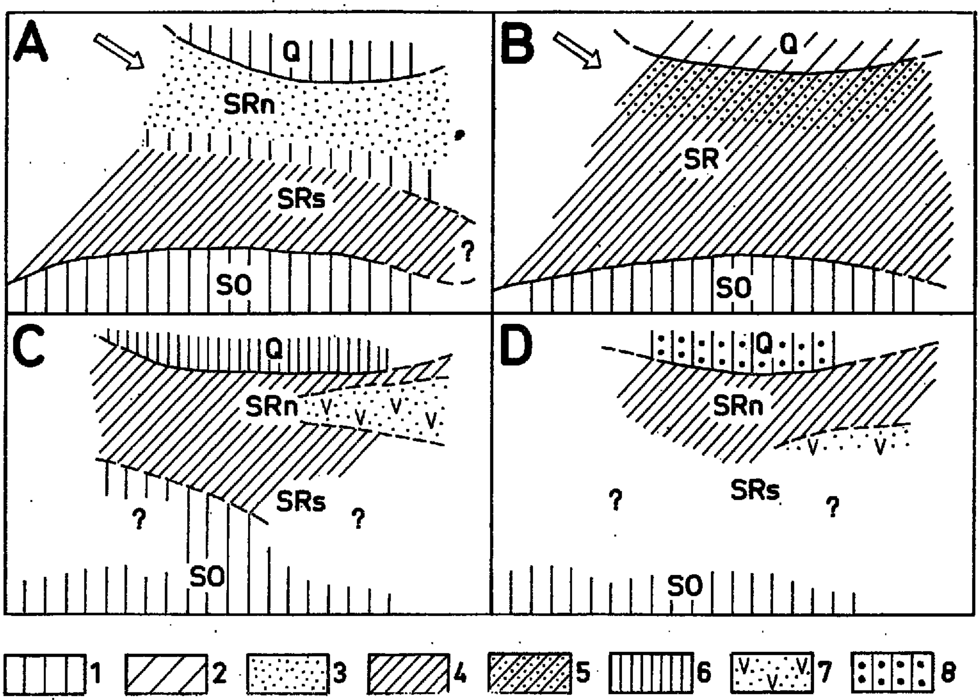

Fig. 29. Facies zones of the Cordillera de Guaniguanico at the Cretaceous time
A — Hauterivian to Barremian
B — Early Cretaceous decline
C — Turonian
D — Campanian to Maastrichtian

SO — sequence of the Sierra de los Organos, *SRs* — southern sequence of the Sierra del Rosario, *SRn* — northern sequence of the Sierra del Rosario, *SR* — both the sequences of Sierra del Rosario, *Q* — Quinones sequence of the Sierra del Rosario

1 pelagic calcareous facies, *2* clayey-siliceous facies, *3* pelagic-flysch facies (limestones, sandstones, and shales), *4* pelagic, mostly siliceous facies (radiolarian cherts), *5* pelagic siliceous and/or flysch facies, *6* pelagic calcareous facies of Quinones sequence, *7* mixed calcareous-terrigenous-volcanic facies, *8* carbonate shallow-water facies of ?Quinones sequence

The arrows indicate the inferred direction of terrigenous influx

In general, the Hauterivian to Barremian paleogeographic pattern had persisted up to the end of the Early Cretaceous when the facies became uniform all over the sequences of Sierra del Rosario (Fig. 29*B*). There were two distinct areas of pelagic sedimentation at the Early Cretaceous decline; *viz.* calcareous pelagic zone in the Sierra de los Organos, and siliceous-clayey one in the Sierra del Rosario. At that time, the sedimentary basin of Sierra del Rosario could reach its deepest point (Pszczółkowski 1971b).

There is no unequivocal evidence for the bathymetry of Mesozoic radiolarian muds and related deposits (*cf.* Garrison 1974). The Lower Cretaceous siliceous-clayey sediments of the Sierra del Rosario were deposited in a basin widening successively since the Hauterivian through Albian time (*cf.* Fig. 29). At the Early Cretaceous decline, this area was at least 50 km wide (when measured perpendicularly to the tectonic structures) but it was probably still wider, as some tectonic units are covered by other units. The area of slow radiolarian-mud sedimentation was a bottom depression delimited by the calcareous pelagic zone of Sierra de los Organos, and on the other side by the clayey-siliceous basin of Quiñones sequence. The radiolarian deposits of the northern and southern sequences of Sierra del Rosario may have originated in proximity of (or temporarily below?) the CCD, in relation, at least in part, to calcium carbonate dissolution (*cf.* Garrison & Fischer 1969). In fact, this interpretation may be confirmed by the symmetrical occurrence of calcareous-chert deposits with respect to the radiolarian sediments in some sections of the southern sequences of Sierra del Rosario (Pszczółkowski 1976a); there are also such cases although under a less typical form in the northern and Quiñones sequences.

SEDIMENTATION RATE OF THE LOWER CRETACEOUS DEPOSITS

The sedimentation rate of radiolarian cherts and shales of the southern sequence of Sierra del Rosario can be estimated to approximate 0.8 to 1.0 m/my, when their subsequent compaction (*cf* Garrison 1967) and deposition span (the Hauterivian or Barremian to Albian, *i.e.*, 25 to 19 my; *cf.* Hinte 1976) are taken into account. This rate is almost identical to that determined for the Jurassic radiolarites of Eastern Alps (Garrison & Fischer 1969); while it is considerably lower than that determined for the Jurassic radiolarites of Polish Tatra Mountains (Lefeld 1974).

The sedimentation rate of the Aptian to Cenomanian clayey-siliceous deposits of Quiñones sequence appears to have been 7 to 10 time higher than that determined for radiolarian cherts and shales of the Sierra del Rosario. This results probably from the significant amounts of clayey matter.

The Lower Cretaceous pelagic and flysch sediments of the northern sequence of Sierra del Rosario were deposited at an average rate of 8 to 12 m/my.

In its turn, the sedimentation rate of calcareous pelagic deposits of the Sierra de los Organos was approximately twice as high as that of radiolarian cherts and shales of the Sierra del Rosario.

FACIES DEVELOPMENT AND PALEOGEOGRAPHY OF THE UPPER CRETACEOUS AND LOWER PALEOGENE

CENOMANIAN TO TURONIAN

FACIES DIVERSITY

The Cenomanian strata of the Cordillera de Guaniguanico are but indistinctly separated from the Lower Cretaceous ones, as the facies are identical or closely related one to the other. At the Cenomanian to Turonian time, a pelagic, mostly calcareous sedimentation persisted in the sequence of Sierra de los Organos (Pons Formation and Infierno Member of Guasasa Formation). This area remained also isolated from both land masses and active volcanic belts. Pelagic sediments were deposited all over the Sierra de los Organos, as indicated by the occurrence of their clasts within the Paleocene breccias present in sections where the Cenomanian to Turonian strata are lacking.

There is a considerable variability in the Cenomanian to Turonian facies among the sequences of Sierra del Rosario. Thus, deep-water, clayey-siliceous sediments were still deposited in the southern sequence at the Early Cenomanian time; later on (sometimes but at the Turonian time), these sediments became replaced with limestones intercalated with radiolarian cherts. These limestones attributed to the limestone- and chert-bearing member of Buenavista Formation (*cf.* Table 7) are preserved but in a few sections (*cf.* Pszczółkowski 1976a). In Quiñones sequence, the Cenomanian has not been evidenced paleontologically; nevertheless, such an age attribution is highly probable in the case of the upper part of shales and cherts of Sierra Azúl Formation. Presumably, a significant change in facies occurred in this sequence at the Cenomanian and Turonian boundary. The Turonian is represented by micritic limestones with planktic forams, assigned to Pinalilla Member (Fig. 29C). The Cenomanian to Turonian facies variation appears to have been the highest in the northern sequence of Sierra del Rosario. In fact, there are micritic and detritic limestones intercalated with radiolarian cherts. Turbidite currents derived the detritic material from some shallow-water areas and deposited it in a pelagic basin. Some radiolarian-bearing chert clasts do also occur in the detritic limestones due to an erosion of bottom highs by the currents. These calcareous-siliceous deposits appear to be best developed in the tectonic unit of Sierra Chiquita, while their amounts in sections of Buenavista Formation decrease southwards. This indicates that the clastic influx was neither from the southern sequence of Sierra del Rosario, nor from the Sierra de los Organos. The Cenomanian to Turonian deposits of the northern sequence of Sierra del Rosario include also shales, marly limestones, sandstones, and detritic limestones of Moreno Member, intercalated occasionally with rocks of volcanic origin (mostly tuffites). Turbidite currents contributed also to the accumulation of the clastic sediments of this facies. This facies is best developed in the eastern part of Sierra del Rosario (Fig. 29C).

PALEOGEOGRAPHIC CONNECTIONS

At the ? Late Cenomanian and Turonian time, the northern sequence of Sierra del Rosario was in part influenced by a volcanic activity. This is the case of a narrow area comprising Cangre tectonic unit and some related slices with dacitic and ryodacitic tuffites. Volcanogenic clastics make also part of sandstones and conglomerates of Moreno Member.

A volcanic activity may also have influenced a part of the southern sequence of Sierra del Rosario, as indicated by a trachyandesite exposed in an inlier west to Soroa (Pszczółkowski 1976a).

An intense volcanic activity took place at the Cenomanian to Turonian time mostly in the southern Cuba; in fact, there are andesites, dacites, and ryolites within sedimentary rocks of that area (Khudoley & Meyerhoff 1971). In Pinar del Río Province, volcanic rocks of andesite composition occur in the Bahía Honda zone (Skupiński *in:* Pszczółkowski & *al.* 1975); these rocks were previously termed porphyrites (Furrazola-Bermúdez 1969). These rocks occur within the Felicidades Formation (Pszczółkowski & Skupiński *in:* Pszczółkowski & *al.* 1975) of the Cenomanian to ?Campanian age.

The Bahía Honda zone makes part of the eugeosynclinal Zaza facies--structural zone (Furrazola-Bermúdez & *al.* 1964). As judged from the volcanic contribution to the Cenomanian to Turonian strata of Zaza zone and the northern sequence of Sierra del Rosario, some paleogeographic connections between the two areas appear quite possible at that time.

Clastic material present in the Cenomanian to Turonian deposits of the northern sequence of Sierra del Rosario could be derived from two source areas. One area occurred in a zone of intense volcanic activity, the other was related to a shallow-water carbonate sedimentation. Carbonate material of sandy calcarenites of Moreno Member came partly from the same source areas. In fact, some rudist banks occurred locally in Zaza zone at the Turonian time (Khudoley & Meyerhoff 1971). They could encircle little volcanic islands; metamorphic rocks could also contribute to these islands. Then, the Cenomanian to Turonian paleogeography might resemble the present-day geography of Greater Antilles, north to Hispaniola. Today, mixed calcareous-terrigenous deposits are being transported off Hispaniola (and probably Cuba, too) to the deep-water basin of Hispaniola-Caicos (Bennetts & Pilkey 1976). However, these Late Cretaceous islands could rather be compatible in size with some of the Lesser Antilles displaying volcanic activity.

There are, however, no indications in the Sierra del Rosario (and so more in the Sierra de los Organos) of a large emerged area south to Pinar del Río Province at the Cenomanian to Turonian time (*cf.* Khudoley *in*: Khudoley & Meyerhoff 1971; J. L. Wilson 1975, Fig. XI-5).

At the Turonian time, the area of pelagic limestone deposition of Quiñones sequence (Pinalilla Member) was outside the deep-water basin of Sierra del Rosario. As the matter of fact, this may reflect a trend to successively rise the bottom in deep-water basin margins.

CONIACIAN TO SANTONIAN

In the Sierra de los Organos, the Coniacian to Santonian deposits occur in the uppermost part of Pons Formation (=Peñas Formation of Hatten 1957). Pelagic limestones of this formation contain the Santonian micro-organisms among others (*cf.* Khudoley & Meyerhoff 1971). Coniacian to Santonian strata have insofar not been documented within Guasasa Formation but one may claim that the pelagic facies had been spread all over the Sierra de los Organos prior to the erosion that took place at the Paleocene time.

The Coniacian to Santonian deposits are poorly evidenced in the Sierra del Rosario. The possible Coniacian marly limestones with some planktic forams occur in Buenavista Formation in the northern sequence of Sierra del Rosario, *viz.* in Cangre tectonic unit. Similar limestones and in a similar lithostratigraphic position have also been recorded in some other sections of Buenavista Formation. The Coniacian to Santonian strata occur surely in some sections of the uppermost part of Moreno Member and in the upper part of the limestone- and chert-bearing member of Buenavista Formation. In Quiñones sequence, the Coniacian to Santonian is probably represented by marly deposits of the upper part of Sierra Azúl Formation.

The Coniacian to Santonian paleogeography remains unclear because of the poor biostratigraphic data. One may but claim that in the deepest part of the Late Cretaceous basin (northern sequence of Sierra del Rosario) the sedimentary conditions did generally not change. In the southern sequence, the Coniacian to Santonian deposits have been eroded previously to the Maastrichtian to Paleocene time. Nevertheless, they may be in places preserved at the base of the limestone breccia member of Buenavista Formation. In fact, Judoley & *al.* (1963) noticed the Aptian to Santonian deposits by the locality Soroa (*cf.* also Khudoley & Meyerhoff 1971); these strata may, however, be pre-Coniacian.

CAMPANIAN TO MAASTRICHTIAN

FACIES ZONES

At the Campanian to Maastrichtian time, the area of pelagic calcareous sedimentation persisted in the Sierra de los Organos (Fig. 29*D*), separated from terrigenous, volcanic, and shallow-water carbonate material sources. At present, deposits of this age occur exclusively in the environs of Pons (Pons Formation) but one may claim that they were spread all over the investigated area previously to the Paleocene time.

In the northern sequence of Sierra del Rosario, the Campanian to Maastrichtian sediments are best developed in the tectonic units of Sierra Chiquita, Cangre, and La Serafina (*cf.* Fig. 2*A*). They include radiolarian cherts intercalated with breccias, with marly limestones present in a few sections. Then, one may conclude that a deep-water sedimentation area existed at that time in the northern sequence of Sierra del Rosario (Fig. 29*D*), receiving temporarily carbonate clastic material.

Pelagic and clastic deposits of the upper part of Buenavista Formation are 200 to 300 m thick and hence, they could not fill up the deep-water basin of the northern sequence of Sierra del Rosario. Then, this basin existed also during the deposition of detritic limestones of Cascarajícara Formation.

In Dolores tectonic unit, the Maastrichtian is represented by shales, marls, and sandstones intercalated with tuffites. In the tectonic units of Naranjo and Belén Vigoa and in the southern Sierra del Rosario, the Maastrichtian is usually represented by a breccia ranging in thickness from 2 to 30 meters (the limestone breccia member of Buenavista Formation).

In Quiñones sequence, the Campanian to Maastrichtian strata include the upper part of Sierra Azúl Formation (200 m in thickness). Terrigenous deposits of the lowermost part of this set of strata resemble the flysch, while clayey shales and sandstones higher in the section lack any flysch analogies.

CALCAREOUS CLASTIC DEPOSITS OF THE SIERRA DEL ROSARIO

The Campanian to Maastrichtian strata commonly comprise clastic deposits, mostly breccias, of the northern sequence of Sierra del Rosario. Their lithology, thickness, and lithostratigraphic position are variable but, nevertheless, there are also some common characteristics. The thickest are the clastic limestones of Cascarajícara Formation (up to 450 m), not older than of Late Maastrichtian age.

In the northern sequence of Sierra del Rosario, breccias occur within the upper chert member of Buenavista Formation. They include commonly clasts and blocks of shallow-water limestones with rudists, corals, and benthic forams, and on the other hand some radiolarian chert and pelagic limestone fragments. There are also some clasts of limestones bearing both planktic and benthic microfauna. Apart from the breccias, there are also some calcarenite intercalations displaying graded bedding and sharp lower boundaries. The shallow-water limestone clasts are obviously allochthonous. The fragments of pelagic limestones and cherts can be regarded as almost indigenous.

Los Cayos Member breccia (Pl. 2, Fig. 2) appears unusual in its large thickness and very coarse detritic material. It occurs in Sierra Chiquita tectonic unit (*cf.* Fig. 21), overlying gray, marly limestones, some blocks of which occur at the base of the breccia. The mode of occurrence of this particular breccia and related but less thick deposits may indicate that the coarse clastic material was received only by some parts of the deep-water basin of Sierra del Rosario. The fragments of coral-foraminifer and rudist limestones have been derived from some Upper Cretaceous reef (?) areas. When transported to the basin of Sierra del Rosario, the shallow-water carbonate debris intermixed with eroded deposits of various sedimentary zones, deep-water ones including. The breccias originated due to debris flow-like processes (*cf.* Moran 1976) of a considerable eroding ability. The coarse clastic material was deposited probably at the base of a paleoslope (carbonate shelf ?), maybe at the mouth of some submarine canyons (*cf.* Schlager & Schlager 1973, Carter & Lindqvist 1975, Walker 1975). Further away from the paleoslope, the clastic material was transported by turbidite currents (breccias and calcarenites displaying graded bedding).

The considered paleoslope must have occurred outside the area of both the northern and southern sequences of Sierra del Rosario. When considering the present-day structural pattern, the amounts of carbonate clastic material significantly increase northwards in the Campanian to Maastrichtian deposits, and the maximum has been attained in some sections of Sierra Chiquita tectonic unit. Therefore, one may conclude that the source area occurred probably north to this tectonic unit.

The origin of detritic deposits of Cascarajícara Formation was considered by some authors (Hatten 1957, Bryant & *al.* 1969, Khudoley & Meyerhoff 1971). Hatten (1957) regarded these detritic limestones as a talus at the base of an active carbonate-bank edge, maybe a fault scarp. According to him, the clasts indicate that the Cretaceous shallow-water carbonates of the Central Cuba extended also over the area of today Pinar del Río Province, and provided the investigated detritic material. Khudoley & Meyerhoff (1971) claim that the shallow-water zone of Remedios recalled by Hatten (1957) occurs today north to Pinar del Rio Province. These conclusions were partly based on the assumed lack of terrigenous and volcanic material in Cascarajícara Formation (*cf.* also Bryant & *al.* 1969). However, thin sections derived from diverse parts of the Formation indicate clearly (Pszczółkowski *in*: Pszczółkowski & *al.* 1975) that apart from the most abundant limestone clasts and bioclasts, a terrigenous material and volcanic debris do also occur rather commonly (Pl. 6, Fig. 2). Then, one may conclude that a shallow-water sedimentation area north or northeast to the Sierra del Rosario could not be the only source area for the clastic material. There are no Cretaceous pelagic, terrigenous, or volcanic rocks in the zone of Remedios (*cf.* Khudoley & Meyerhoff 1971). In fact, Knipper & Cabrera (1974) reported the Cenomanian to Campanian pelagic sediments from that area, but their notion of Remedios zone appears wider than that accepted in other publications. The investigated clastic material includes terrigenous, volcanic (neutral and alkalic), and metamorphic rocks (quartzites and quartz slates) and pelagic and shallow-water carbonates. There is no facies-structural zone that might have provided such a variable detritic material. It seems quite possible that all these diverse constituents of the detritic limestones of Cascarajícara Formation have somewhere become mixed previously to their deposition in the deep-water basin of Sierra del Rosario; subsequently, a single or some turbidite currents transported them to the latter area. A gradual decrease in grain size from the base to the top of the Formation may indicate that the deposits originated owing to a single sedimentary event; however, in such a case a turbidite current should have been loaded with some 40 to 70 km^3 of the clastic material.

The Cascarajícara Formation occurs but in two tectonic units which suggests that its depositional area represented a rather narrow, elongate depression. The presence of both shallow-water carbonate debris and volcanic and pelagic limestone fragments indicates a mutual proximity of some volcanic-sedimentary sequences and shallow-water carbonate zone, the latter represented probably by the marginal part of the Florida-Bahama (or Yucatan?) platform.

LATE CRETACEOUS TECTONIC MOVEMENTS

The lack of Campanian and pre-Campanian Cretaceous deposits in the Sierra del Rosario has resulted from some pre-Maastrichtian tectonic movements (Pszczółkowski 1971b, 1976a). In the southern sequence of Sierra del Rosario and in the southernmost tectonic units of the northern sequence, the Late Cretaceous tectonic movements took place during the Turonian through Maastrichtian time. Sometimes, one may be unable to make a clear distinction between the effects of the "Subhercynnian" and "Early Laramide" tectonic movements in Cuba. The stratigraphic hiatus including the Turonian through Campanian has insofar been related to the Subhercynnian events (Furrazola-Bermúdez & *al.* 1964, Khudoley 1967); or the Turonian to Santonian Subhercynnian movements have been distinguished from the Campanian to Maastrichtian Laramide ones (Hatten 1967, Khudoley & Meyerhoff 1971). The present author prefers to treat these movements in more general time terms (*i.e.*, the Late Cretaceous) or to determine precisely, if possible, their time range.

In some sections of the Sierra del Rosario, one may be unable to determine precisely the time range of the limestone breccia member of Buenavista Formation and therefore, a possibility of a local overlap of the effects of the Late Cretaceous and Paleocene movements cannot be rejected. Neither the Late Cretaceous tectonic movements in the Sierra del Rosario, nor the Paleocene ones in the Sierra de los Organos (Pszczółkowski & *al.* 1975), have resulted in any overthrusts or folds. Furthermore, the Late Cretaceous erosion did not penetrate deeply into the Jurassic strata of Sierra del Rosario, if reached them at all. It took off deposits 30 to (?) 120 m thick, depending on particular section in the southern sequence (Pszczółkowski 1976a, Fig. 3). The maximum value could even be less than 120 m, be a part of the lacking strata deposited under the conditions of very low sedimentation rate.

In some tectonic units of the northern sequence of Sierra del Rosario, the Late Cretaceous tectonic movements did not disturb the deep-water sedimentation. Nevertheless, they changed significantly the geosynclinal basin of the Sierra del Rosario as a whole. Vast area of the Cenomanian to Turonian deep-water deposition was reduced at the Maastrichtian time to the northern sequence, mostly to the tectonic units of Sierra Chiquita and Cangre. The basin of Quiñones sequence was taken to a shallow-water sedimentation area at the Maastrichtian time.

Approximately the same intensity was also attained by the Campanian and Maastrichtian tectonic movemens all over Pinar del Río Province (Pszczółkowski *in*: Pszczółkowski & *al.* 1975). As judged from the facies development and disconformity characteristics, there were no intense orogenic movements in La Esperanza, San Diego de los Baños, and Bahía Honda zones; consequently, no overthrusts or considerable folds were formed at the Late Cretaceous time. The most remarkable disconformities occur at the base of Capdevila Formation which represents the Upper Paleocene to Lower Eocene (Myczyński & Piotrowski *in*: Pszczółkowski & *al.* 1975), and at the base of Loma Candela Formation (Middle Eocene).

PALEOCENE OF THE SIERRA DE LOS ORGANOS

At the Early Paleocene time, pelagic limestones and cherts were still accumulating in the Sierra de los Organos. These deposits are preserved today but in a single tectonic unit, namely Valle de Pons unit. There is a hiatus in the sequence of Sierra de los Organos (Hatten 1967) comprising diverse Upper Jurassic and Cretaceous stages and most of the Paleocene (*cf.* Table 1). The Paleocene erosion was rather limited in the base tectonic units (Valle de Pons and Infierno units) and hence, deposits of Ancón Formation overlie the Upper Cretaceous or Lower Paleocene limestones. In the top tectonic units (Viñales, Ancón, Sierra de la Güira), the Upper Paleocene deposits overlie massive limestones of Guasasa Formation (Kimmeridgian to lowermost Tithonian) or the Tithonian to Valanginian strata; this indicates that carbonate sediments 80 to 450 m thick were eroded prior to the Late Paleocene time. There has insofar not been recorded any section with the Paleocene erosion that had reached the lowermost part of Guasasa Formation or Jagua Formation. The La Güira

Member attains its maximum thickness in sections displaying a considerable Paleocene erosion. Be the Cretaceous or Lower Paleocene limestones overlaid directly by micritic and marly limestones of Ancón Formation, the latter deposits comprise fairly abundant chert clasts. The disconformity recognized between the Guasasa and Pons Formations and the Ancón Formation has not been observed in the field. In fact, it is less than 2° and it can be demonstrated only by means of a comparison among the sections of successive tectonic units (Pszczółkowski *in*: Pszczółkowski & *al.* 1975). Thus, one may conclude that the top surface of the pre-Late Paleocene strata is of erosional-tectonic nature in the Sierra de los Organos.

Presumably, the Paleocene erosion of the Jurassic to Cretaceous limestones took place mostly under the submarine conditions. In fact, the uppermost Cretaceous to Lower Paleocene deposits of Pons Formation represent pelagic facies just as micritic and marly limestones of Ancón Formation do. There is no evidence of any change in facies towards more shallow-water sedimentary conditions. The La Güira Member breccia does not comprise any shallow-water limestone clasts or bored debris. Any Paleocene karst structures have not been recorded at the top of

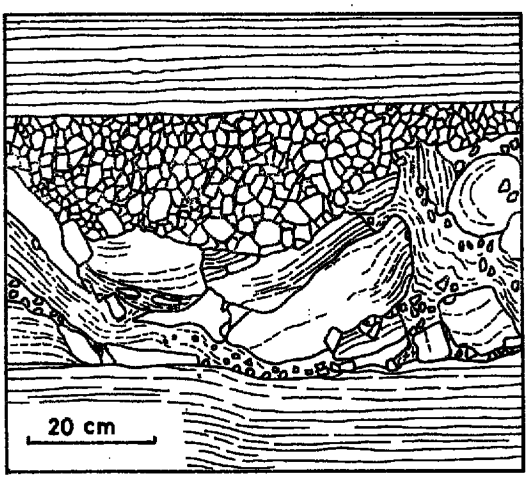

Fig. 30
Breccia within pelagic limestones of Ancón Formation, western margin of the Valle del Ancón; the limestone clasts have been derived mostly from the older formations of the sequence of Sierra de los Organos

Pons and Guasasa Formations, even although the top surface may be somewhat irregular. A calcareous matrix occurs in both the uppermost part of La Güira Member breccia in Pina Sola section and breccia intercalations within pelagic limestones of Ancón Formation (Fig. 30), which indicates that the detritic deposits were commonly (if not exclusively) accumulated in a pelagic sedimentation area.

The disconformity at the base of the Upper Paleocene deposits has resulted from tectonic movements that took place after the Maastrichtian to Early Paleocene but prior to the Late Paleocene time. These movements represented probably a differential uplift of some blocks built up by the Jurassic to Cretaceous deposits (Fig. 31). There is, however, no evidence of any folds or overthrusts of the pre-Late Paleocene age. In fact, such deformations should have exposed some strata older than Guasasa Formation and caused their clasts to occur in the breccia of La Güira Member of Ancón Formation.

Along with uplifted blocks, some submarine depressions resulted also from the Paleocene tectonic movements. These depressions were partly filled by the sediments of Ancón Formation. This is indicated by the facies and thickness variability exhibited by Ancón Formation. Limestone and chert debris were deposited in proximity of hypothetic faults attaining some tens to hundreds meters in displacement. Some especially uplifted areas supplied a carbonate debris throughout the deposition period of Ancón Formation. Olistostrome-like breccia intercalations were deposited in bottom depressions due to submarine slump processes. La Legua Member breccia appears unusual because of its large thickness (up to 25 m) and block size (up to 5 m). It originated probably in relation to an intense uplift in the Late Paleocene of an adjacent block built up by the Jurassic to Cretaceous limestones.

Some red, marly limestones occur in the uppermost part of Ancón Formation in Viñales and Infierno tectonic units. Their silica content approximates 27%, that is it exceeds considerably that typical of the limestones of the lower part of Ancón Formation (6 to 7.5%). This difference resulted from an increased clayey influx by the end of the sedimentation period of the considered deposits. At that moment, the area represented today by Viñales and Infierno tectonic units (the latter one but in part) was probably localized in a deeper zone of the basin. This was also the case at the beginning of the Late Paleocene sedimentation (cf. Fig. 31).

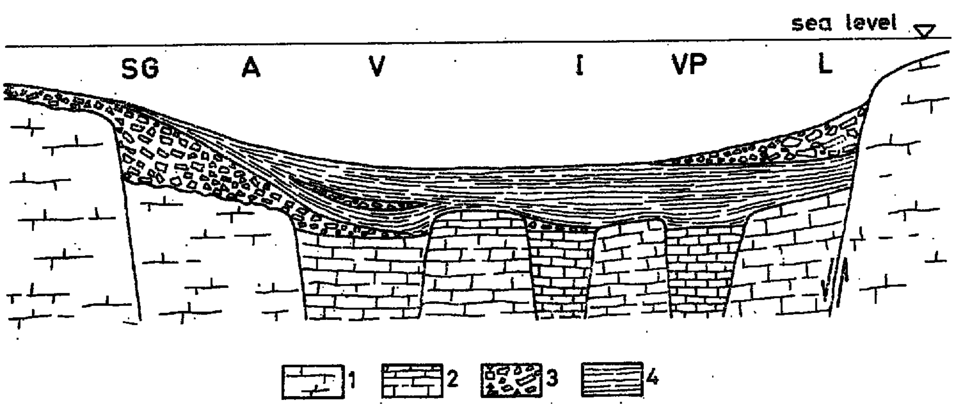

Fig. 31. Sedimentation pattern of the Upper Paleocene Ancón Formation in the Sierra de los Organos (not to vertical scale)

Tectonic units: *SG* — Sierra de la Güira, *A* — Ancón, *V* — Viñales, *I* — Infierno, *VP* — Valle de Pons, *L* — La Legua

1 the Upper Jurassic massive limestones (Guasasa Formation), *2* the Tithonian to Lower Paleocene pelagic limestones (Guasasa and Pons Formation), *3* breccia of Ancón Formation, *4* pelagic, micritic and marly limestones of Ancón Formation

The sedimentation rate of the Upper Paleocene micritic and marly limestones was much higher than that of the Cretaceous pelagic deposits in the Sierra de los Organos.

LOWER EOCENE OF THE SIERRA DE LOS ORGANOS

At the Late Paleocene decline, a general change in facies occurred in the Sierra de los Organos. Pelagic and clastic deposits of Ancón Formation were replaced successively by terrigenous sediments of Pica Pica Formation. Lithological variability in the latter formation resulted from the clastic influx coming from some distinct source areas. Most clastic material is of volcanic origin. The volcanic debris could be derived in part from reworked diabases and andesites to dacites present within the Paleogene strata of the Sierra de los Organos (*cf.* Grodzicki 1972, Haczewski 1974). However, most volcanic clasts are allochthonous. This is, indeed, indicated by the associated bioclasts including abundant fragments of the Campanian to Maastrichtian rudists, algae and large forams, and the lithoclasts including biomicrites with planktic microorganisms (Calcisphaerulidae among others). Such a detritic material might be derived from a sedimentary-volcanic sequence resembling the zone of Bahía Honda. Carbonate fragments were also supplied by the Guasasa and Pons Formations. In fact, limestone debris of these two formations occur commonly in breccias, and less abundantly in wacky sandstones and calcarenites of Pica Pica Formation.

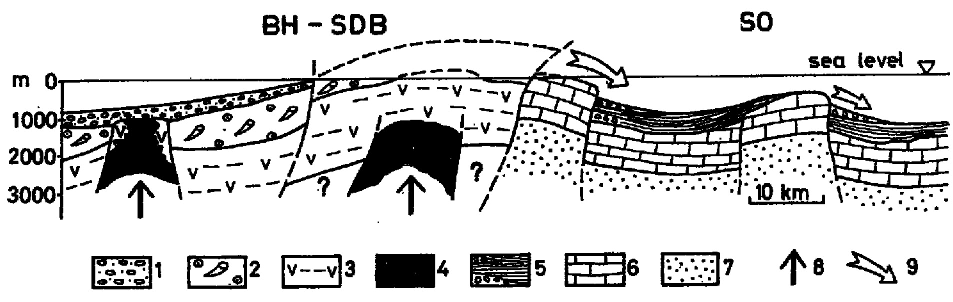

Fig. 32. Hypothetic section showing the Lower Eocene paleogeography of the western Pinar del Río Province

BH-SDB — facies-structural zones of Bahía Honda and San Diego de los Banos, *SO* — sequence of the Sierra de los Organos

1 Capdevila Formation, *2* the Campanian to Maastrichtian limestones with rudists and benthic forams, and clastic deposits (San Juan Formation), *3* the Cenomanian to Campanian sedimentary and volcanic rocks, *4* serpentinites, gabbroids, and ultrabasic rocks, *5* clastic deposits of Pica Pica Formation and the underlying deposits of Ancón Formation, *6* the Upper Jurassic to Lower Paleocene limestones, *7* San Cayetano Formation, *8* uplift of some areas in Bahía Honda and San Diego de los Banos zones, *9* clastic influx to the Sierra de los Organos basin

Turbidite currents transported the terrigenous clastics to the basin of Sierra de los Organos (Hatten 1957); whereas the detritic carbonate material was transported mostly by submarine slumping off some local elevations built up mainly by Guasasa Formation limestones (Fig. 32). The lavas resulted from a volcanic activity related mostly to the upper part of Pica Pica Formation (*cf.* Myczyński *in*: Pszczółkowski & *al.* 1975). By the locality La Constancia, an andesitic agglomerate lava (Malinovsky & Carassou 1974) occurs within the chaotic rocks; however, one may suppose that it was originally related to the uppermost Pica Pica Formation.

The Lower Eocene thickness and facies variability in the sequence of Sierra de los Organos resulted from a tectonic instability at the deposition time (Hatten 1957). The flysch sediments range in thickness from merely 30 m to more than 100 m. Despite the tectonic deformations, one may claim that this variation was in part determined by the basin topography. The reduced flysch sections occur commonly over those Jurassic limestones that had already been uplifted at the Paleocene time. In places, the flysch deposits overlie directly such Jurassic rocks and then, comprise abundant pebbles and angular clasts taken off the substrate.

The Lower Eocene flysch and pelagic deposits of the Sierra de los Organos can be regarded as synorogenic sediments. However, the very meaning of such an attribution is quite different from that accepted previously by Hatten (1967). It is proposed here that most clastic material came from some degraded sedimentary-volcanic sequences rather than from the sequence of Sierra de los Organos itself. During the Early Eocene flysch deposition, sediments of Jagua and San Cayetano Formations remained covered by the overlying strata, which indicates that the tectonic units of the Sierra de los Organos could not be at that time overthrust to any significant extent. Presumably, the Lower Eocene rocks were deposited in front of a rock sequence being uplifted or even overthrust; the latter sequence can be claimed to have resembled the rocks of Bahía Honda facies-structural zone. In Pinar del Río Province, the uplift and/or overthrusting of that hypothetic rock sequence had started at the Paleocene time, and increased in intensity at the Early Eocene time.

LOWER PALEOGENE OF THE SIERRA DEL ROSARIO

No deposits younger than of the Late Paleocene age have insofar been documented in the Sierra del Rosario. Nevertheless, the Lower Eocene sediments may also occur in some sections of Buenavista Formation. The Paleogene of Sierra del Rosario comprises clayey and tuffite shales, wacky sandstones, polymictic calcarenites, cherts, micritic and marly limestones, marls, and breccias. In La Paloma tectonic unit (east to La Palma), limestones, shales, and sandstones are separated from radiolarian cherts and breccias by a unit of volcanic rocks some 100 m thick (Fig. 20).

In the southern sequence of Sierra del Rosario, rather thin pelagic limestones were deposited along with some shales and sandstones at the Paleocene time. Thereafter, the general facies development became somewhat similar to that typical of the sequence of Sierra de los Organos. A clastic sedimentation prevailed at first; sandstones and calcarenites composed mostly of bioclasts and volcanic debris were deposited at that time. A more intense volcanic activity disturbed or interrupted the clastic sedimentation, supplying some basalt- or andesite-type rocks. The volcanic activity persisted also at the deposition time of radiolarian cherts and tuffite shales (? Early Eocene time) although its intensity decreased. The latter sediments are also intercalated with some olistostrome-like, coarse-clastic rocks (up to 15 m thick) and occasional olistolites.

In some tectonic units, the Paleogene deposits of Buenavista Formation are overlaid by chaotic rocks. There are blocks of diverse sedimentary rocks (*cf.* Pl. 1, Fig. 4) within the chaotic mass, those unknown in the sequence of Sierra del Rosario including. This indicates that sedimentary processes contributed to the formation of chaotic rocks; the rocks underwent also considerable tectonic deformations (Pszczółkowski 1971b, *in*: Pszczółkowski & *al.* 1975). The autochthonous constituents represent usually but a minor component of the chaotic rocks of the Sierra del Rosario. Furthermore, there are no blocks derived from Artemisa and San Cayetano Formations. Then, one may conclude that the chaotic rocks were formed prior to the oldest formations of the Sierra de Rosario being widely exposed. In fact, blocks and pebbles derived from those formations appear in large amounts but in the upper Middle Eocene, *viz.* in Loma Candela Formation (Pszczółkowski *in*: Pszczółkowski & *al.* 1975, Pszczółkowski 1976b) south to Pinar fault, that is outside the Cordillera de Guaniguanico.

The chaotic rocks of Sierra del Rosario comprise serpentinites, diabases, and diverse metamorphic rocks (*cf.* Pszczółkowski 1971b). There are also blocks of shallow-water sedimentary rocks up to some hundred meters in size (*e.g.*, by the locality Soroa). When assuming that Bahía Honda facies-structural zone has been overthrust from the south over the tectonic units of the Sierra del Rosario (*cf.* Meyerhoff *in*: Khudoley & Meyerhoff 1971, Pardo 1975, Pszczółkowski 1976b), the serpentinites and their associates in chaotic rocks can be treated as precursory olistostromes of Elter & Trevisan (1973). However, the occurrence of some shallow-water limestone and dolomite blocks may require another explanation. Some other hypotheses have, indeed, been proposed to explain the facies-structural relations in Pinar del Río Province (Piotrowska & Pszczółkowski *in*: Pszczółkowski & *al.* 1975, Piotrowska 1976a). The very nature of the Sierra del Rosario chaotic rocks is complex which is here meant that their origin resulted from both sedimentary and tectonic processes acting penecontemporaneously (*cf.* Hsü 1973) or at differential time.

GUANIGUANICO SEQUENCES COMPARED TO OTHER FACIES-STRUCTURAL ZONES OF CUBA

Any unmetamorphosed equivalents of the sequence of Sierra de los Organos have not been reported from outside the Cordillera de Guaniguanico. Metamorphosed terrigenous-carbonate rocks of the Sierra del Escambray and Isla de Pinos may be equivalent, at least in part, to the Jurassic strata of the Sierra de los Organos (Brown & O'Connell 1922; Lewis 1932; Khudoley 1967, *in*: Khudoley & Meyerhoff 1971; Somin & Millán 1972). However, there is no unequivocal interpretation of Isla de Pinos section (*cf.* Millán 1974). Metamorphosed deposits of San Cayetano

Formation (or Arroyo Cangre Formation ?) may also occur in a submarine ridge southwest to Guanahacabibes Peninsula (*cf.* Pyle & *al.* 1973).

LA ESPERANZA AND BAHIA HONDA ZONES

There are some deposits in La Esperanza zone close in their facies to the Polier and Buenavista Formations of the northern sequence of Sierra del Rosario. They represent probably the Lower and Upper Cretaceous (Pszczółkowski & *al.* 1975). They comprise larger amounts of turbidite sandstones than many sections of the Sierra del Rosario do, which may support the inference about the terrigenous influx from the northwest at the Early Cretaceous time.

No formation typical of the sequences of Sierra del Rosario has been recorded in Bahia Honda zone. Nevertheless, there are some Upper Cretaceous deposits comparable in both the areas. One may recall the occurrence of volcanic rocks in Buenavista and Sierra Azúl Formation (Sierra del Rosario) and Felicidades Formation (Bahía Honda); or cherts and breccias in Buenavista and Felicidades Formations. On the other hand, there are also some important differences, especially in the Maastrichtian to Lower Eocene facies. The Upper Paleocene to Lower Eocene sediments (Bratu 1973) of Capdevila Formation have been deposited in places directly onto gabbroid rocks (Pszczółkowski *in:* Pszczółkowski & *al.* 1975). Then, the gabbroids (probably along with serpentinites) appeared in Bahía Honda zone prior to the latest Paleocene or earliest Eocene time, but following the deposition of the Upper Maastrichtian strata. There are no transitional facies from the Cordillera de Guaniguanico to Bahía Honda zone in the northeastern part of Pinar del Río Province, which makes the tectonic nature of the contact (Pszczółkowski 1976b) even more pronounced.

PRE-MIDDLE EOCENE SEQUENCE OF LA HABANA SURROUNDINGS

The Cretaceous to Paleogene deposits of La Habana surroundings[a] have been described in detail by Brönnimann & Rigassi (1963). The Cenomanian to Turonian sediments ("Pre-Vía Blanca beds") of this sequence appear compatible with some of their time equivalents in the northern sequence of Sierra del Rosario. However, the Campanian to Lower Maastrichtian deposits are distinctly different. La Habana sequence comprises terrigenous flysch deposits intercalated with conglomerates, calcarenites, tuffites, and andesites, attaining up to 500 m in thickness; the conglomerates may be dominated by igneous rock debris, volcanic material including (Brönnimann & Rigassi 1963). In its turn, the time equivalent part of the northern sequence of Sierra del Rosario comprises radiolarian cherts intercalated with breccias that lack usually any material of igneous or volcanic origin. The Maastrichtian deposits of Dolores tectonic unit resemble, however, the facies of Vía Blanca Formation of La Habana sequence. The Cascarajícara and Peñalver Formations have already been compared in this paper. Their clastic material could be derived from the same source areas; both these formations were deposited under similar paleogeographic conditions. The Upper Paleocene and Lower Eocene appear to be incompatible in both the sequences.

In summary, the northern sequence of Sierra del Rosario may resemble the Cretaceous strata of La Habana surroundings rather than the Bahía Honda zone.

CAMAJUANI AND PLACETAS ZONES

There are several facies-structural zones parallel to the northern coast of the island in Central Cuba, 300 km east to Pinar del Río Province. There are the Jurassic, Cretaceous, and Paleogene strata south to platform zones of Cayo Coco and

Remedios (Furrazola-Bermúdez & *al.* 1964). They have been attributed to either Camajuaní and Placetas zones (Ducloz & Vuagnat 1962, Knipper & Cabrera 1974), Las Villas, Placetas, and Cifuentes zones (Pardo 1975), or Las Villas zone itself (Khudoley 1967, *in:* Khudoley & Meyerhoff 1971); this is the miogeosyncline and median welt of A. A. Meyerhoff. From the south, these strata border upon serpentinites, gabbroids, and volcanic and sedimentary rocks of Santa Clara zone (Ducloz & Vuagnat 1962) called also Zaza zone (Furrazola-Bermúdez & *al.* 1964). The latter zone has commonly been conceived as eugeosynclinal part of the Greater Antilles orthogeosyncline in Cuba (Hatten 1967, Khudoley & Meyerhoff 1971, Knipper & Cabrera 1974, Pardo 1975).

The equivalents of the Upper Jurassic to Cretaceous deposits of the southern sequence of Sierra del Rosario occur in Camajuaní zone (Pszczółkowski 1976a). Moreover, the Upper Jurassic limestones and dolomites of Camajuaní zone comprising a shallow-water fauna and ooids may be facies (and time ?) equivalent to the San Vicente Member of Guasasa Formation; nevertheless, it would probably be unreasonable to claim that any direct paleogeographic connections did exist. The Tithonian to Berriasian ammonite-bearing limestones of Camajuaní zone may also be roughly correlated with their time equivalent deposits of Artemisa Formation. Shallow-water conditions were replaced by a pelagic sedimentation at approximately the same time in Camajuaní zone and the Sierra de los Organos. The Lower Cretaceous limestones of the southern sequence of Sierra del Rosario resemble very closely those of Camajuaní zone. The Lower Cretaceous to Cenomanian cherts and radiolarian shales of the southern sequence of Sierra del Rosario resemble the Albian to (?) Turonian cherts, radiolarites, and limestones of Placetas zone (*cf.* Ducloz & Vuagnat 1962, Pardo 1975). In Camajuaní zone, the Maastrichtian is represented by cherts intercalated with detritic limestones. Somewhat similar deposits occur in the Campanian to Maastrichtian in the northern sequence of Sierra del Rosario. The Paleogene is quite differential in both the considered areas. There is a stratigraphic hiatus between the Cenomanian to Turonian and Maastrichtian, typical of both Camajuaní zone (Knipper & Cabrera 1974) and the southern sequence of Sierra del Rosario. In general, the northern sequence of Sierra del Rosario resembles Placetas zone; whereas the sequence of Quiñones appears to have no counterpart in Central Cuba.

As demonstrated by the above paragraphs, there are some vague analogies between the sequences of Cordillera de Guaniguanico (except of Quiñones sequence) and some other facies-structural zones of Central Cuba. Nevertheless, the latter zones cannot be extended automatically over the area of Pinar del Río Province, as there are also some important differences. One may recall here the abundance of terrigenous Jurassic deposits in the Cordillera de Guaniguanico, carbonate pelagic Cretaceous sediments in the Sierra de los Organos, and volcanogenic rocks in the Sierra del Rosario.

PRETECTOGENIC PATTERN OF GUANIGUANICO SEQUENCES

The Middle Eocene orogenesis changed significantly the original pattern of the western Cuban geosynclinal zones. The boundaries of Guaniguanico facies-structural zone are tectonic in nature (Piotrowska 1976a, Pszczółkowski 1976b), which makes difficult to unequivocally reconstruct the original paleogeographic position of this zone. The sequence of Sierra de los Organos does also border upon the sequences of Sierra del Rosario along a tectonic contact; nevertheless, there are some transitional links making possible to find out the original facies and paleogeographic relationship between both the major parts of the Cordillera de Guaniguanico. Moreover, there are also some sections transitional between the southern and northern sequences of Sierra del Rosario. The Hauterivian to Cenomanian deposits of Quiñones sequence resemble their time equivalent strata of the other sequences of Sierra del Rosario; whereas the younger deposits are distinctly different.

The stratigraphic sequences of the Cordillera de Guaniguanico have been partitioned tectonically and overthrust; nevertheless, the original facies sequence has not been changed to any considerable extent. This holds true mostly for the most important facies, present in some tectonic units at least. The present-day distribution of these facies in the nappe--slice units of the Cordillera reflects generally their pretectogenic pattern. This correlation is quite apparent in the case of the Jurassic facies and to a less degree the Lower Cretaceous ones.

The tectonic units of the Cordillera de Guaniguanico are almost parallel to the main Mesozoic facies zones. Then, one may claim that the tectonic transport was more or less perpendicular to the strike of these facies zones. A comparison made between the lithologies of metamorphic rocks of the Sierra de los Organos and Isla de Pinos (Millán 1972) may suggest that previously to the Middle Eocene time, the sequence of Sierra de los Organos was the southernmost one in the Cordillera de Gauniguanico (*cf.* Piotrowska 1976a). This inference is also supported by the paleogeographic pattern proposed by Haczewski (1976) for San Cayetano Formation. Then, one may claim that the sequences of Sierra del Rosario occurred originally to the north or northeast of the sequence of Sierra de los Organos. In fact, Rigassi-Studer (1963) assessed already that the sedimentation area of the Jurassic to Cretaceous strata of the Sierra del Rosario occurred to the north or northwest of the Sierra de los Organos.

It seems improbable that Guaniguanico sequences were originally more or less perpendicular to the W—E axis. One might argue that the tectonic units of the Sierra del Rosario occur mostly in the eastern part of the Cordillera de Guaniguanico, but the counterargument is that nobody knows how far to the east are there the tectonic units of the Sierra de los Organos under the cover of Sierra del Rosario units. The problem is also obscured by Pinar fault cutting obliquely the tectonic units between San Diego de los Baños and Soroa. The original pattern of Guaniguanico stratigraphic sequences as suggested by Rigassi-Studer (1963) and confirmed by the present author, implies also some constraints on possible interpretations of the overthrust direction of the units of Sierra del Rosario (*cf.* Pszczółkowski 1976b).

GUANIGUANICO SEQUENCES AND THEIR PALEOTECTONICS

The Cordillera de Guaniguanico was regarded as an intrageanticline making part of an eugeosyncline, mostly because the Cretaceous to Paleogene deposits were claimed to contribute but insignificantly to Guaniguanico sections (Furrazola-Bermúdez & *al.* 1964, Khudoley 1967). However, somewhat more detailed investigations have demonstrated that the Cretaceous to Paleogene strata are widespread in both the Sierra de los Organos (Hatten 1957, 1967; Pszczółkowski & *al.* 1975) and Sierra del Rosario (Pszczółkowski 1976a, Myczyński 1977). In the northern units of the Sierra del Rosario, many sections comprise exclusively deposits of that age (Pszczółkowski 1976a, b).

The substrate of San Cayetano Formation is unknown in Pinar del Río Province. The source area for the deposits of San Cayetano Formation must have comprised sedimentary, metamorphic, and igneous rocks. It could be related to the substrate of the Jurassic delta of the Sierra de los Organos. One may claim that it was of continental crust type (*cf.* MacGillavry 1970). The substrate of the sequences of Sierra del Rosario represented either a continental crust, or a modified oceanic crust (*sensu* Iturralde-Vinent 1975). Any ultimate solution remains impossible for the moment. At the Eocene time, Guaniguanico sequences have become separated from their substrate by a décollement (Piotrowska 1975, 1976b) or shears (Rigassi-Studer 1963, Pszczółkowski 1976b).

Miogeocline and exogeosyncline stages (*cf.* Dickinson 1972) can be recognized within the sequence of Sierra de los Organos. The miogeocline stage comprises the Jurassic to Cretaceous strata, while the exogeosyncline stage comprises the Upper Paleocene to Lower Eocene deposits. Three main sedimentation phases can be distinguished within the miogeocline stage, *viz.* deltaic, shallow-water, and pelagic phases. The first two phases are usually conceived as the most typical of these sequence because of large thickness of their deposits. However, the pelagic sedimentation was the most persistent one (over 80 my) and hence, appears as the most important one. The exogeosyncline stage of the sequence of Sierra de los Organos was significantly influenced by synsedimentary tectonics and volcanism; the latter under the form of both effusions and a considerable supply of volcanic debris.

There are no equivalents of the sequence of Sierra de los Organos in either the northwestern Caribbean, or Central America. Some facies analogies can be claimed but in the case of the Jurassic terrigenous deposits of the Sierra de los Organos and Honduras or Guatemala; according to Meyerhoff (*in:* Khudoley & Meyerhoff 1971), sediments of El Plan Formation of Honduras resemble those of San Cayetano Formation. However, the El Plan Formation is much thinner (300 to 500 m) than the San Cayetano Formation; moreover, the former has been tentatively attributed to the Upper Triassic to Lower Jurassic (Mills & Hugh 1974). A large part of San Cayetano Formation may be time equivalent to the Todos Santos Formation, but the facies characteristics appears to be quite different (*cf.* Mills & *al.* 1967, Mills & Hugh 1974). Nevertheless, one may claim that previously to the Oxfordian time, the deposition area of the sequence of Sierra de los Organos was close to the area of present-day Honduras and maybe Guatemala (*cf.* Iturralde-Vinent 1975, Haczewski 1976). The Central America made probably part of a large continent extending to the south or southeast. The Lower Jurassic continental deposits have, indeed, been reported from the northwestern margin of Guiana shield (Geyer 1973). According to some paleogeographic reconstructions, the northwestern part of South America could be linked with the area of Honduras and Nicaragua at the Early to Middle Jurassic time (Freeland & Dietz 1971, Fig. 5). The differential Upper Jurassic facies indicate that the sequence of Sierra de los Organos became separated at that time. The pre-Oxfordian Jurassic volcanism recorded in the Sierra de los Organos may reflect a rift formation south to the exposures of San Cayetano Formation deposits and their metamorphosed equivalents (Piotrowski 1976). At the Oxfordian time, this volcanic activity decreased, as demonstrated by thin volcanic intercalations in Jagua Formation (*cf.* Piotrowski 1976); this inference is also supported by the scarcity of volcanic material in Francisco Formation, Sierra del Rosario. The Jurassic volcanism related to disjunctive tension tectonics has also been recorded in deposits of Todos Santos Formation (Mills & Hugh 1974). Ladd (1976) claimed that since the Early Jurassic through Valanginian time, the South America had removed far away to the southeast from the North America. This relative movement of South America could have resulted at the Jurassic time in the tension conditions reflected all over the area of present-day Central America and northwestern Caribbean (*cf.* Ladd & *al.* 1973).

The psammitic influx to the sequence of Sierra de los Organos was inhibited at the Middle Oxfordian time (*cf.* Wierzbowski 1976) because of either an inundation of the source area (Haczewski 1976), or a separation of the basin from eroded areas. In Central America, a terrigenous deposition persisted in some places up to the Late Jurrasic and Early Cretaceous time (Mills & *al.* 1967, Mills & Hugh 1974, H. H. Wilson 1974). The Yucatán Peninsula was probably emerged at the Late Jurassic time (Lopez Ramos 1975). Then, one has to accept that the sequence of Sierra de los Organos and other sequences of the Cordillera de Guaniguanico have been separated from the Central America at the Oxfordian decline, that is earlier than it was proposed by some authors (*cf.* Freeland & Dietz 1971, Iturralde-Vinent 1975).

An intense shallow-water sedimentation developed in the sequence of Sierra de los Organos at the Kimmeridgian and earliest Tithonian time. The lack of terrigenous material may indicate that the carbonate bank of Sierra de los Organos was quite efficiently separated from large land masses. The shallow-water sedimentation was counterbalanced by a subsidence. The homogeneity in the Tithonian to Berriasian facies all over the Cordillera de Guaniguanico suggests that disjunctive tectonics contributed to the submergence of the carbonate bank (Fig. 27). In Camajuaní zone, Central Cuba, a shallow-water carbonate sedimentation did also disappear at the Tithonian time, which suggests that tectonic processes affected somewhat larger area than the sequence of Sierra de los Organos itself. A possible hypothesis of widespread Tithonian transgression is less probable because of the occurrence of the Tithonian to Lower Cretaceous shallow-water deposits on Bahamas platform (*cf.* Khudoley & Meyerhoff 1971).

The Tithonian to Cretaceous pelagic, mostly carbonate deposits of the sequence of Sierra de los Organos are rather thin. In this respect, the Sierra de los Organos may be related to the Central Cuban leptogeosyncline, as conceived by Knipper & Cabrera (1974) and Iturralde-Vinent (1975). Nevertheless, recognition of the Sierra de los Organos area for an oceanic trough (*cf.* Iturralde-Vinent 1975) appears disputable, especially in the case of the Jurassic to Cretaceous strata. In fact, oceanic troughs display thick turbidite deposits underlaid by rocks typical of the oceanic crust (Mitchell & Reading 1971, Scholl & Marlow 1974, Prince & Kulm 1975, Moore & Karing 1976). The pelagic sediments of the Sierra de los Organos were probably deposited on a submarine plateau bordering upon some deeper basins. The siliceous and clayey-siliceous deposits of the sequences of Sierra del Rosario were accumulated in these basins. These basins restricted the influx of terrigenous and shallow-water material; moreover, they separated the pelagic deposition area of the sequence of Sierra de los Organos from volcanic regions.

The Upper Cretaceous breccias and Cascarajícara Formation can also be treated in terms of the Late Cretaceous to Paleogene tectonics (*cf.* Pszczółkowski 1976a). At least some of the carbonate clasts making part of the considered deposits were derived from some shallow-water areas free of volcanic activity. At the Late Cretaceous time, there was such an area at the southern margin of Bahamas platform and the platform itself (Khudoley & Meyerhoff 1971, Pardo 1975). Then, the composition of the rudaceous deposits indicates that at that time the northern sequence of Sierra del Rosario was by the southern margin of Florida-Bahamas or ? Yucatán platforms. The Lower Cretaceous sediments of the Sierra del Rosario do not comprise any significant amount of redeposited shallow--water carbonate material. When interpreted within the mobilistic framework (*cf.* Freeland & Dietz 1971, Malfait & Dinkelman 1972, Nagy 1972), these data may indicate that the Caribbean (or Caribbean-East Pacific) plate came close to the Florida-Bahamas (and Yucatán?) platform at the Late Cretaceous time. It was previously assessed that this could have happened but at the Paleocene time (Malfait & Dinkelman 1972), with the Eocene collision (Iturralde-Vinent 1975). An immobilistic hypothesis would explain the composition of the rudaceous deposits of the Sierra del Rosario by an uplift and consequent erosion of the edge of Florida-Bahamas platform.

Presumably, the Paleocene disjunctive tectonics in the Sierra de los Organos was related to some major tectonic processes. According to Malfait & Dinkelman (1972), the Cuban segment of Caribbean plate came close to the Bahamas platform at the Paleocene time; furthermore, it changed its movement direction from the northeast to ENE. This reorientation could be related to either the Early Tertiary formation (Perfit 1977) of the rift of Cayman trough (Holcombe & *al.* 1973), or rejuvenation of the Late Cretaceous fault zones of Cayman trough and Cauto structure (*cf.* Khudoley & Meyerhoff 1971). The structure and development of Yucatán Basin are rather poorly known but one may claim that the faults delimiting this abyssal plain (Fahlquist & Daives 1971, Uchupi 1973) were also active at the beginning of the Tertiary. The activity of the major fault zones of the northwestern Caribbean resulted in partitioning the Jurassic to Cretaceous strata of the Sierra de los Organos into several blocks. The relatively sunken blocks received more intense terrigenous influx at the beginning of the Eocene.

The Paleogene basaltic-andesitic-dacitic volcanism of the Cordillera de Guaniguanico might be related to an active subduction zone along the western Cuban part of the Greater Antilles geosyncline. Such a volcanic activity is usually related to Benioff zones (*cf.* Dickinson & Hatherton 1967) and oceanic crust subduction (Ringwood 1974). According to Malfait & Dinkelman (1972), the volcanic activity decreased succesively and the Benioff zone disappeared since the latest Cretaceous to Miocene time, following cutting off successive segments of the northwestern part of Caribbean plate. However, such an interpretation cannot be approved because of tectonic reasons (Meyerhoff & Meyerhoff 1973) as well as the volcanism development in western Cuba.

This problem has to be analysed once more, with the recent geologic and petrologic data on the whole Cuba taken into account. Then, one will be allowed to determine the position of the subduction zone relative to the major Cuban facies-structural zones at the Late Cretaceous to Early Paleogene time. In fact, this crucial point was subjected to diversified interpretations (Malfait & Dinkelman 1972; Mattson 1973, 1974; Iturralde-Vinent 1975), most of which have subsequently been criticized (Meyerhoff & Meyerhoff 1972, 1974; Khudoley & Meyerhoff 1974; Iturralde-Vinent 1974).

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A. PSZCZÓŁKOWSKI

SUKCESJE GEOSYNKLINALNE KORDYLIERY GUANIGUANICO NA KUBIE: ICH LITOSTRATYGRAFIA, ROZWÓJ FACJALNY I PALEOGEOGRAFIA

(Streszczenie)

Przedmiotem pracy są zagadnienia litostratygrafii, rozwoju facjalnego i paleogeografii jurajsko-paleogeńskich sukcesji Kordyliery Guaniguanico w zachodniej części Kuby (fig. 1 i 2). Schemat litostratygraficzny tej kordyliery obejmuje grupę Viñales z 5 formacjami (tabele 1 i 2) oraz 9 innych formacji. W pracy opisane są nowe formacje (Polier, Buenavista, Lucas, Sierra Azúl i Guajaibón — *por.* fig. 14—17 oraz 19—23) sygnalizowane we wcześniejszej publikacji (Pszczółkowski 1976a). Ponadto opisane są: nowa formacja Pica Pica wyróżniona (*in*: Pszczółkowski & *al.* 1975) w miejsce dotychczasowej formacji Pinos proponowanej przez Herrerę (1961), oraz nowe ogniwa wydzielone w obrębie niektórych formacji (*por.* tabele 3—8, fig. 10, 18 oraz pl. 1—3). Dokonana została korelacja poszczególnych jednostek litostratygraficznych Kordyliery Guaniguanico, a jednostki te zostały omówione na tle 4 sukcesji stratygraficzno-facjalnych (jednej w Sierra de los Organos oraz 3 w Sierra del Rosario: południowej, północnej i Quiñones). Wymienione sukcesje kordyliery odpowiadają różnym strefom facjalnym zachodniej części geosynkliny Wielkich Antyli.

W sukcesji Sierra de los Organos po utworzeniu osadów deltowych (*por.* Haczewski 1976), rozpoczęła się w oksfordzie sedymentacja osadów wapiennych i ilastych (*por.* fig. 3—4 oraz 24—26) facjalnie różnorodnych (pl. 4). Pod koniec oksfordu w sukcesji Sierra de los Organos utworzyła się płycizna, na której składane były płytkowodne osady węglanowe (fig. 5—9 oraz pl. 5). Równowiekowe wapienie górnojurajskie sukcesji południowej (fig. 11—13) i północnej Sierra del Rosario zostały osadzone na ogół w spokojniejszych i zapewne głębszych partiach zbiornika. Płycizna węglanowa została pogrążona w dolnym tytonie (*por.* fig. 27), a następnie we wszystkich sukcesjach kordyliery rozwinęła się wapienna facja pelagiczna, niekiedy z wkładkami redeponowanych osadów płytkowodnych (fig. 28). W kredzie dolnej w Sierra del Rosario (fig. 29) pojawiła się głębokowodna facja krzemionkowo-ilasta (sukcesja południowa), podczas gdy w sukcesji północnej składane były osady fliszowe i pelagiczne. W tej ostatniej sukcesji sedymentacja głębokowodna trwała do końca kredy. Brekcje i kalkarenity kampanu-mastrychtu zawierają także materiał płytkowodny, co wskazuje na sąsiedztwo płycizn węglanowych platformy florydzko-bahamskiej lub jukatańskiej. Osadzenie biomikrytów otwornicowych (fig. 30 oraz pl. 6) i brekcji z materiałem lokalnym formacji Ancón, w górnym paleocenie sukcesji Sierra de los Organos, poprzedzone było w wielu miejscach silną erozją utworów jurajsko-kredowych (*por.* fig. 31). W dolnym eocenie do obszaru depozycji tej sukcesji dostarczany był materiał obcy (*por.* fig. 32). W starszym paleogenie w niektórych sukcesjach kordyliery zaznaczył się synsedymentacyjny wulkanizm.

W świetle dokonanych porównań sukcesji Kordyliery Guaniguanico z niektórymi strefami tektoniczno-facjalnymi środkowej części Kuby zasadnym jest pogląd, że te ostatnie nie powinny być mechanicznie przedłużane na obszar prowincji Pinar del Río. Następstwo paleogeograficzne sukcesji Kordyliery Guaniguanico przed tektogenezą eoceńską było zbliżone do aktualnie obserwowanego w zespole jednostek płaszczowinowo-łuskowych omawianej strefy. Rozwój facjalny sukcesji Sierra de los Organos w jurze górnej był częściowo uwarunkowany przerwaniem połączeń z obszarami lądowymi już w oksfordzie, co można wiązać z powstawaniem hipotetycznego ryftu na południe od Kuby. Przejawy wulkanizmu zarejestrowane w osadach starszego paleogenu Kordyliery Guaniguanico mogą stanowić podstawę do reinterpretacji niektórych koncepcji geodynamicznych w odniesieniu do północno-zachodniej części regionu karaibskiego.

A. PSZCZÓŁKOWSKI

LAS SECUENCIAS GEOSINCLINALES DE LA CORDILLERA DE GUANIGUANICO EN CUBA: LITOESTRATIGRAFIA, DESARROLLO DE FACIES Y PALEOGEOGRAFIA

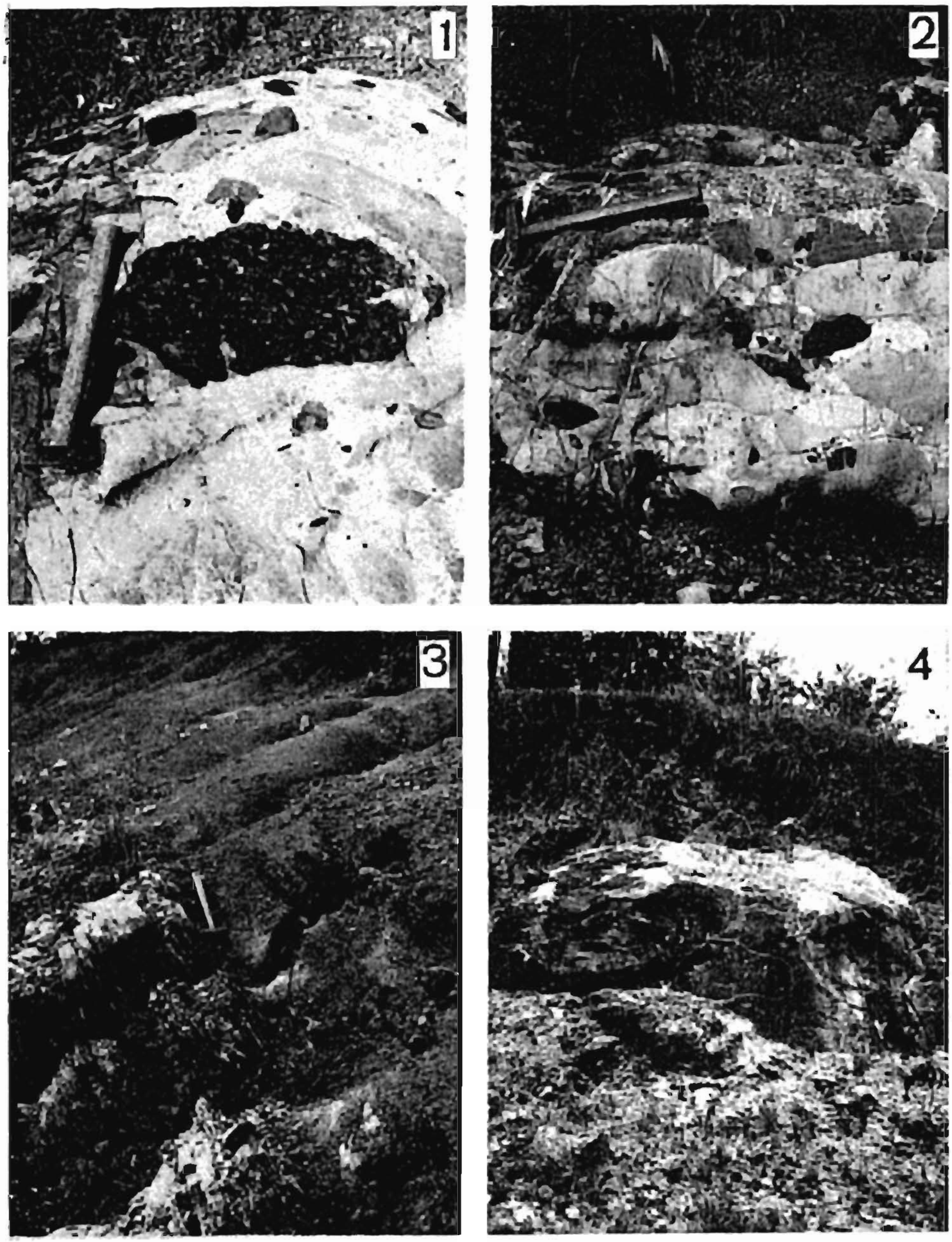
(Resumén)

El trabajo trata los problemas de la litoestratigrafía, desarrollo de facies y paleogeografía de las secuencias del Jurásico-Paleógeno de la Cordillera de Guaniguanico, en la parte occidental de Cuba (fig. 1 y 2). El esquema litoestratigráfico de la Cordillera de Guaniguanico contiene al Grupo Viñales con 5 formaciones (tab. 1 y 2), así como también otras 9 formaciones. Se describen nuevas formaciones (Polier, Buenavista, Lucas, Sierra Azúl y Guajaibón — *ver.* fig. 14—17 y 19—23), señaladas en la publicación anterior (Pszczółkowski 1976a). Además, estan descritos: Formación Pica Pica (*ver* Pszczółkowski & al. 1975), establecida en lugar de la Formación Pinos propuesta anteriormente por Herrera (1961), y nuevos miembros distinguibles dentro de algunas formaciones (*ver.* tab. 3—8, fig. 10, 18 y lám. 1—3). Las unidades litoestratigráficas de la Cordillera están correlacionadas mútuamente. Las unidades litoestratigráficas están presentadas en los marcos de 4 secuencias estratigráficas. En la Sierra de los Organos hay una sola secuencia estratigráfica, mientras que en la Sierra del Rosario existen 3 secuencias: meridional, septentrional y Quiñones. Dichas secuencias de la Cordillera de Guaniguanico corresponden a las variadas zonas faciales de la parte occidental del geosinclinal de las Antillas Mayores.

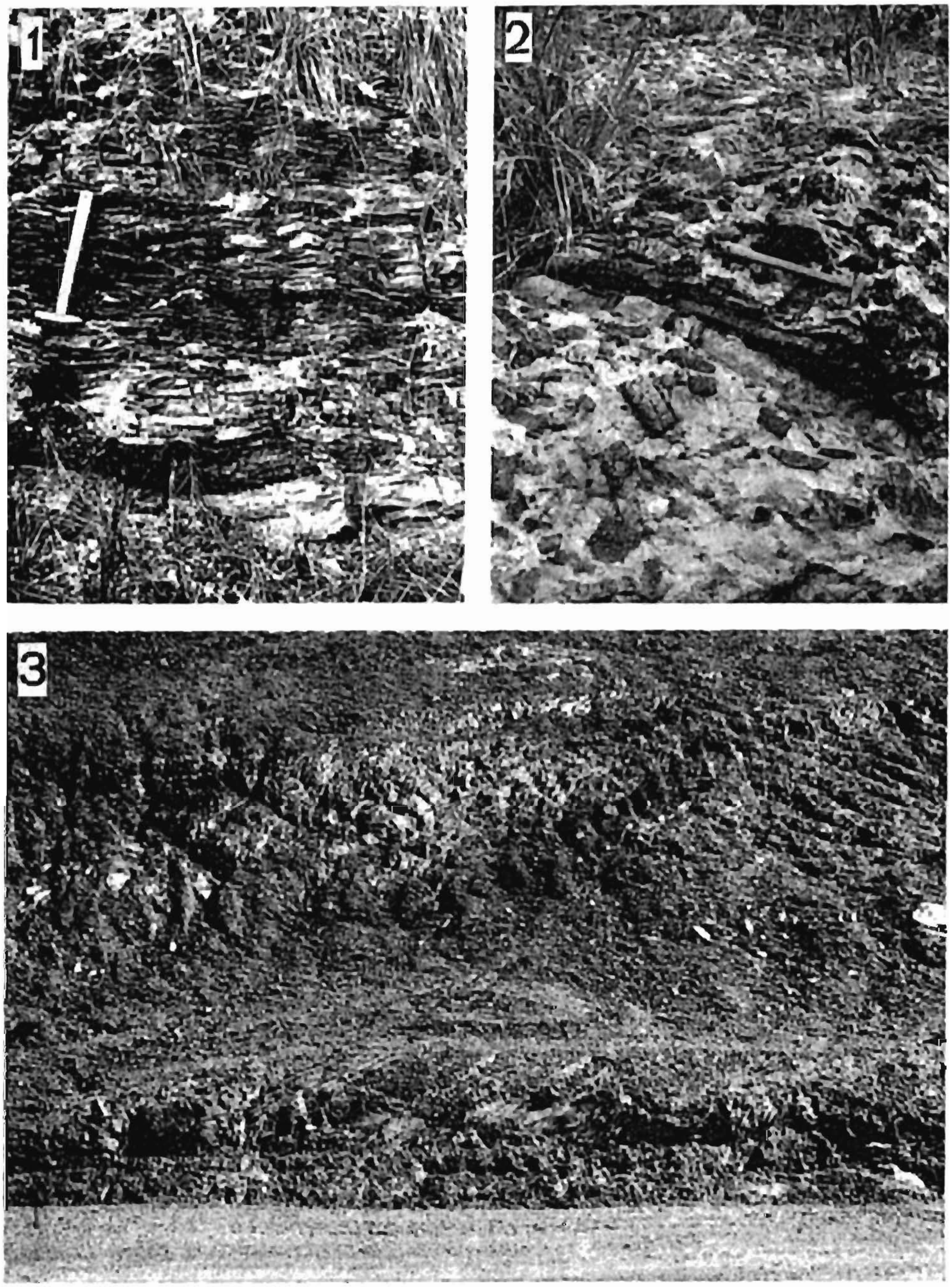
En la secuencia de la Sierra de los Organos, en el Oxfordiano, después de la acumulación de los depósitos de delta (*ver* Haczewski 1976) se inició la sedimentación de los depósitos calcáreos y arcillosos (*ver* fig. 3—4 y 24—26) facialmente diversos (lám. 4). Al final del Oxfordiano en la misma secuencia se formó un banco, donde se acumulaban los depósitos carbonatados de mar poco profundo (fig. 5—9 y lám. 5). Las calizas de edad equivalente (Oxfordiano Tardío-Titoniano Temprano) en las secuencias meridional (fig. 11—13) y septentrional de la Sierra del Rosario han sido depositadas por lo general en la parte más tranquila y, al parecer, más profunda de la cuenca marítima que cubría el área estudiada. El banco carbonatodo se hundió en el Titoniano Inferior (*ver* fig. 27) y después en casi todas las secuencias de la Cordillera se extendió la facies de calizas pelágicas, a veces con las intercalaciones de sedimentos redepositados de ambiente poco profundo (fig. 28). Durante el Cretácico Inferior en la Sierra del Rosario (fig. 29) se desarrolló la facies silícico-arcillosa (en la secuencia meridional), mientras que en la secuencia septentrional se acumulaban los depósitos pelágicos carbonatados y de flysch. En la última secuencia mencionada la sedimentación de aguas profundas duraba hasta el final del Cretácico. Las brechas y calcarenitas del Campaniano-Maestrichtiano contienen también el material clástico proveniente de las zonas de la sedimentación poco profunda, lo que señala la proximidad de los bancos carbonatados de la plataforma de Florida--Bahamas o de Yucatán. Las biomicritas con foraminíferos planctónicos (lám. 6) y brechas, con material local, de la Formación Ancón (fig. 30), en la secuencia de la Sierra de los Organos se acumularón en el Paleoceno Superior, después de la etapa de erosión considerable de los depósitos del Jurásico y Cretácico (*ver* fig. 31). Durante el Eoceno Inferior la cuenca de sedimentación de esta secuencia obtenía ante todo un material clástico ajeno (*ver* fig. 32). En el Paleógeno, en algunas secuencias de la Cordillera de Guaniguanico se manifestó la actividad volcánica (rocas efusivas de composición diabasica y de andesito-dacita).

La comparación de las secuencias estratigráficas de la Cordillera de Guaniguanico con algunas zonas estructuro-faciales de la parte central de Cuba revela la existencia de los rasgos generales semejantes de ambas regiones; no obstante — según la

opinión del autor — no se debe prolongar automáticamente las zonas mencionadas de Cuba Central hasta la Cordillera de Guaniguanico. La sucesión paleogeográfica de las secuencias estratigráficas de esta Cordillera antes de la orogénesis eocénica fue, por lo general, aproximadamente igual al cuadro actual de dichas secuencias en las unidades tectónicas de la zona estudiada. El desarrollo de las facies en la secuencia de la Sierra de los Organos durante el Jurásico Superior fue la consecuencia (en parte por lo menos) de la separación del área de sedimentación desde la tierra firme, ya en el Oxfordiano, que fue el resultádo de la formación del rift hipotético al sur de Cuba. La presencia de las manifestaciones de la actividad volcánica en el Paleógeno de la Cordillera de Guaniguanico necesita una nueva interpretación de algunos esquemas geodinámicos referidos a la parte noroeste de la región del Caribe.



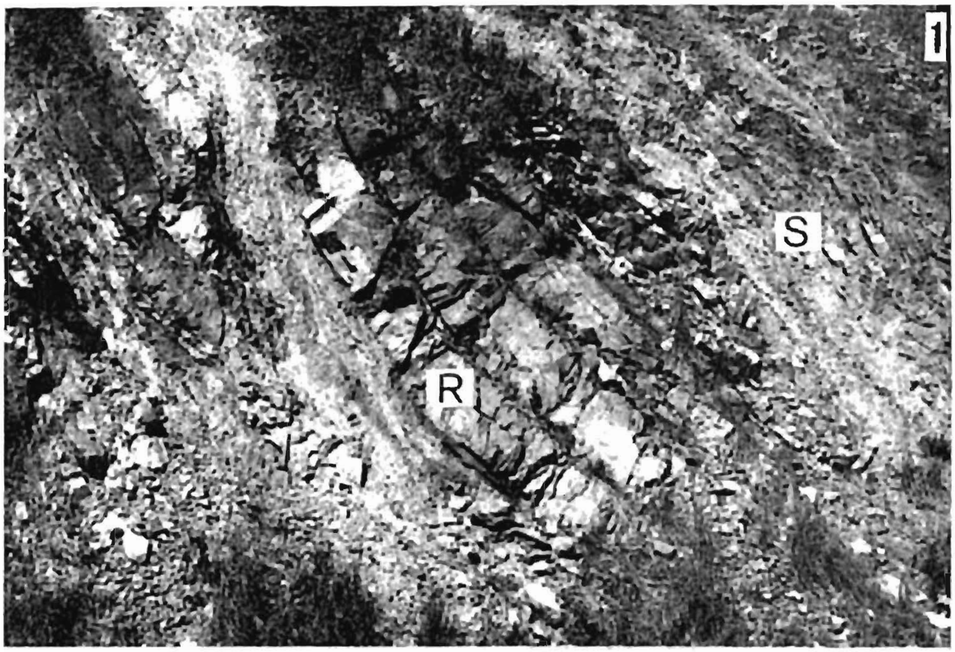
1 — Breccia of La Güira Member (Ancón Formation) in the type section at Caiguanabo (*cf.* Text-fig. 23)
2 — Breccia of La Güira Member (Ancón Formation) in Pina Sola section, east to Jagua Vieja (*cf.* Text-fig. 23)
3 — Tectonically undisturbed contact of limestones of Ancón Formation (at left) and shales of Pica Pica Formation at Jagua Vieja
4 — Breccia block (3×3.5×5 m in size) within the chaotic rocks at the western slope of Mango Bonito, Sierra del Rosario



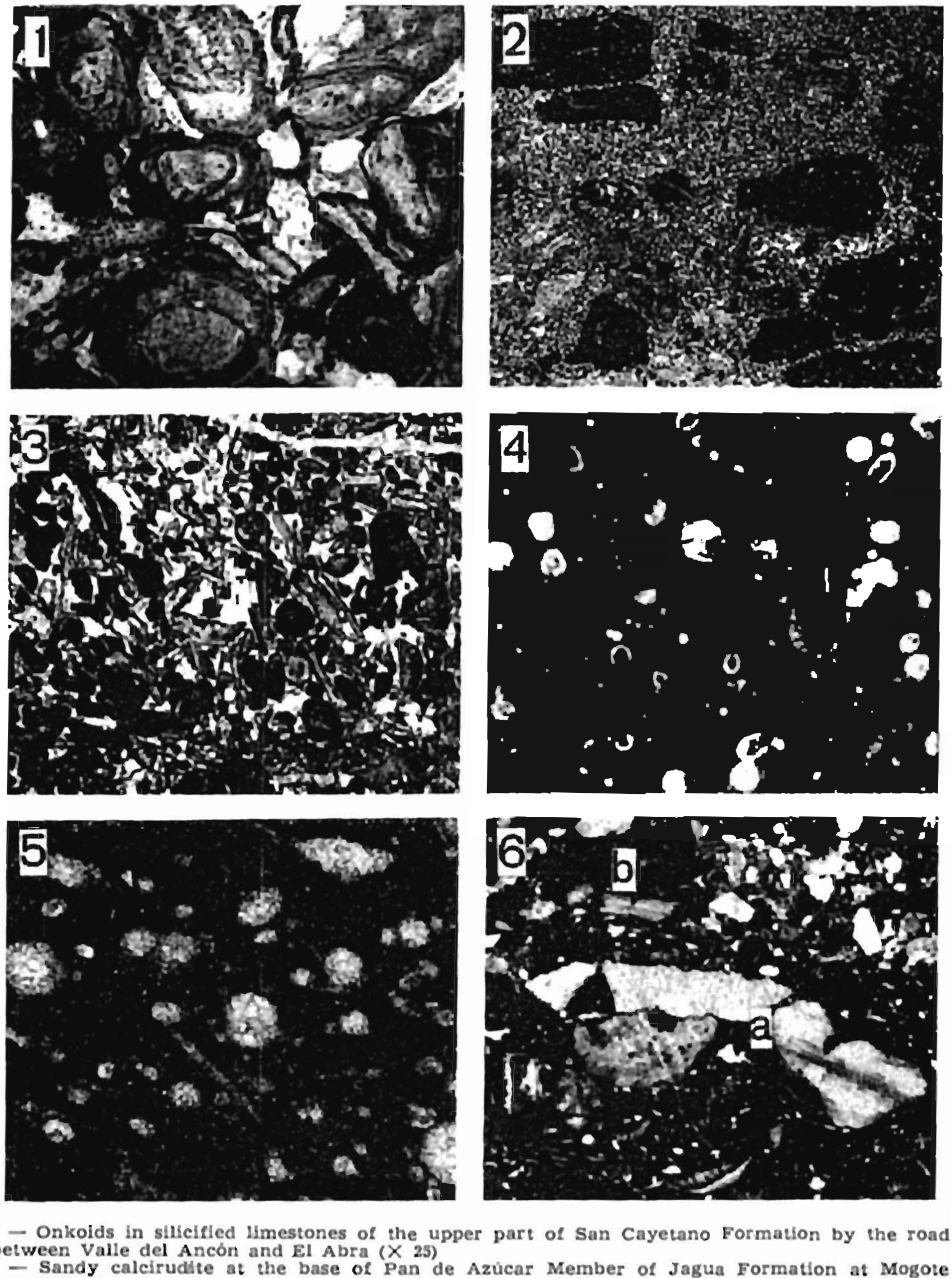
1 — Micritic limestones intercalated with shales; the uppermost part of Lucas Formation in the type section at Rancho Lucas (*cf.* Text-fig. 15)

2 — Breccia of Los Cayos Member (Buenavista Formation) in the type section at Río San Miguel (*cf.* Text-fig. 20C)

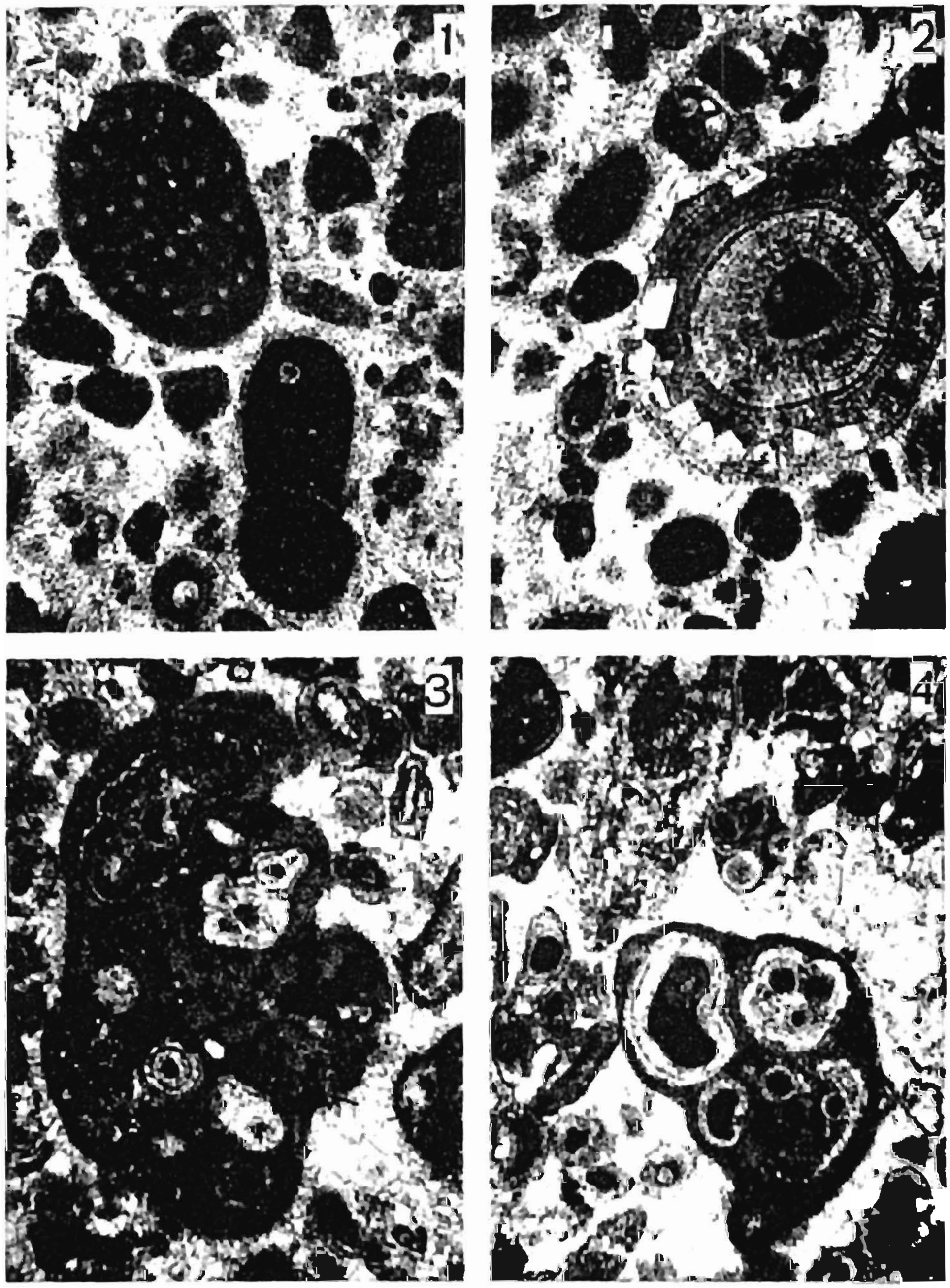
3 — Cherts and shales of Sabanilla Member (Buenavista Formation) at a slope of Belén Vigoa hill, by the road between San Cristobal and Bahía Honda



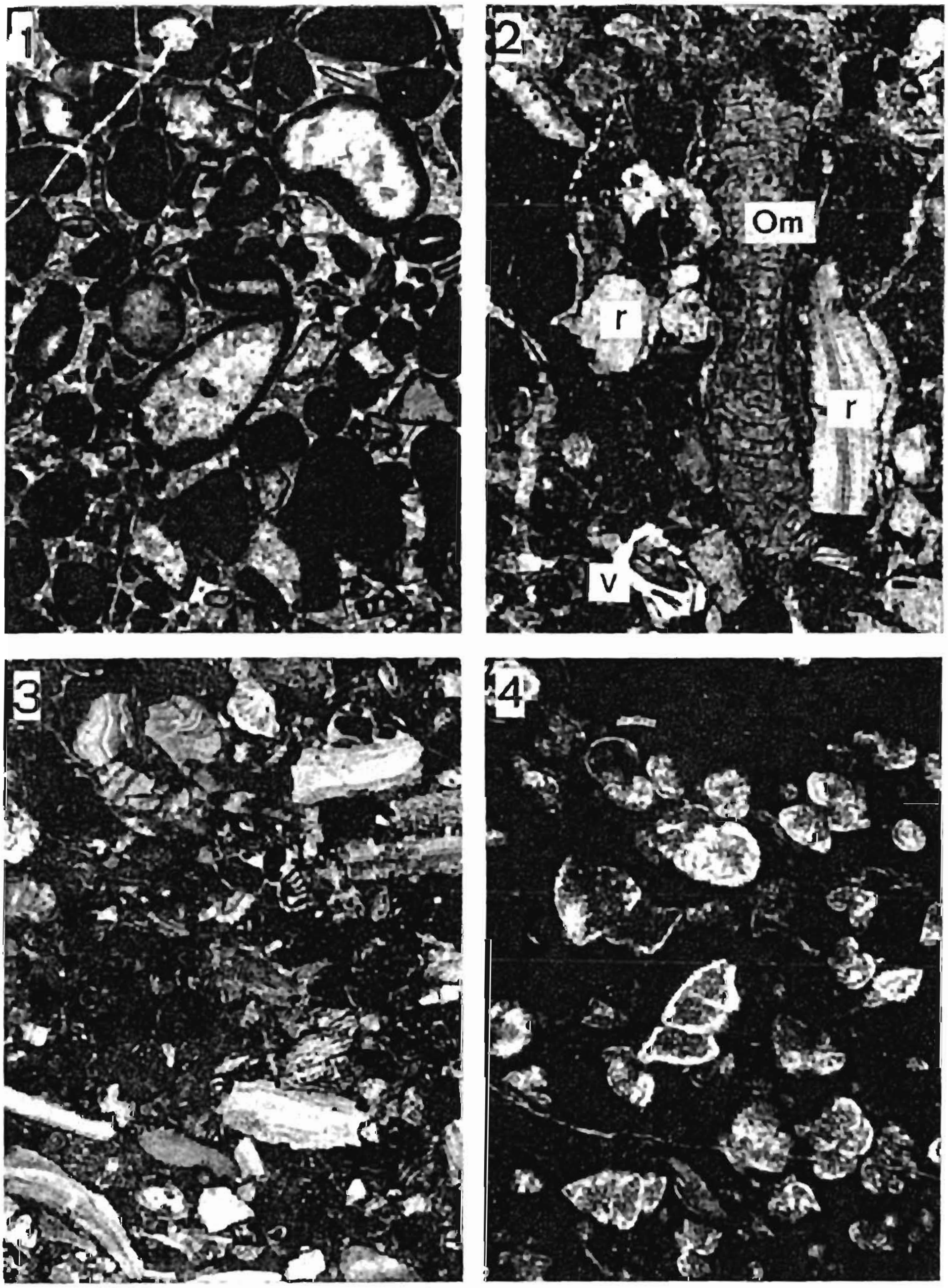
1 — Deposits of Roble Member (Polier Formation) in the type section (*cf.* Text--fig. 14*B*); *R* — sandstones and shales of Roble Member, *S* — cherts and shales of Sabanilla Member (Buenavista Formation)
2 — Flute casts on the sole of a sandstone in Roble Member of Polier Formation



1 — Onkoids in silicified limestones of the upper part of San Cayetano Formation by the road between Valle del Ancón and El Abra (X 25)
2 — Sandy calcirudite at the base of Pan de Azúcar Member of Jagua Formation at Mogote Pan de Azúcar (X 10)
3 — Biocalcarenite; upper part of Pan de Azúcar Member of Jagua Formation in the Sierra Ancón (X 3)
4 — Radiolarian-tintinnid microfacies of Tumbitas Member of Guasasa Formation; visible are *Tintinnopsella carpathica* (Murgeanu & Filipescu) and *Calpionella* sp. (X 40)
5 — Radiolarian biomicrite of Infierno Member of Guasasa Formation in the Sierra del Infierno; the radiolarians are entirely calcified (X 50)
6 — Sandy calcarenite of turbidite origin of the limestone- and chert-bearing member of Buenavista Formation (X 15); *a* echinoderm fragments, *b* thick-shelled bivalve fragments



1 — *Favreina*-form coprolites in an oosparite of the upper part of San Vicente Member of Guasasa Formation; Valle del Ancón section (×60)
2 — Ooid with a peloid as its nucleus (dolomite rhombs are also visible); oosparite of the upper part of San Vicente Member of Guasasa Formation; Valle del Ancón section (×60)
3—4 — Lumps in calcarenites of the uppermost part of San Vicente Member of Guasasa Formation; the ooids and onkoids are cemented by microsparitic calcium carbonate, probably of bluegreen-algal origin (×25)



1 — Ooid-bearing intrabiosparite of the uppermost part of Guajaíbón Formation; the Sierra Azúl (×10)
2 — Calcirudite of the lower part of Cascarajícara Formation in Lomas de Polier section (×15); *Cm* forams *Omphalocyclus* sp., *r* rudist debris, *v* volcanic debris
3 — Coarse calcarenite of the lower part of Cascarajícara Formation in Lomas de Polier section; the rock is composed of rudist debris, benthic forams, and shallow-water limestone clasts (×10)
4 — Foraminifer-bearing biomicrite (with *Globorotalia* spp.) of the lower part of Ancón Formation; Hacienda El Americano section (×60)