ELSEVIER

Sedimentary Geology 135 (2000) 295–320

Sedimentary Geology

www.elsevier.nl/locate/sedgeo

Origin of the Peñalver Formation in northwestern Cuba and its relation to K/T boundary impact event

H. Takayama[a,*], R. Tada[a], T. Matsui[b], M.A. Iturralde-Vinent[c], T. Oji[a], E. Tajika[a], S. Kiyokawa[d], D. Garcia[e], H. Okada[f], T. Hasegawa[g], K. Toyoda[h]

[a]*Geological Institute, University of Tokyo, Tokyo, 113-0033, Japan*
[b]*Department of Earth and Planetary Physics, University of Tokyo, Tokyo, 113-0033, Japan*
[c]*Museo Nacional de Historica Natural, Obispo no. 64, Plaza de Armas, La Habana, 10100, Cuba*
[d]*National Science Museum, Department of Geology, Tokyo, 169-0073, Japan*
[e]*Instituto de Geología y Paleontología, San Miguel del Pardon, La Habana, Cuba*
[f]*Department of Earth and Planetary Sciences, Hokkaido University, Sapporo, 060-0810, Japan*
[g]*Department of Earth Sciences, Faculty of Science, Kanazawa University, Kanazawa, 920-1192, Japan*
[h]*Graduate School of Environmental Earth Science, Hokkaido University, Sapporo, 060-0810, Japan*

Received 3 March 1999; accepted 28 July 1999

Abstract

The uppermost Maastrichtian Peñalver Formation and its equivalents in northwestern Cuba are characterized by thick, normal-graded calcarenite with distinct basal conglomerate, and have been suspected as deposits related to the Cretaceous–Tertiary boundary (K/T) impact. However, its depositional mechanism is not well understood and clear evidence for its association with the impact has never been presented. In this study, detailed field survey and petrographic observations were carried out on the Peñalver Formation in order to clarify its sedimentary processes and to test its relation to the K/T impact. The Peñalver Formation at the type locality near Havana is approximately 180 m thick and is subdivided into the Basal, Lower, Middle, Upper and Uppermost Members based on its lithology. The Basal Member consists of massive, poorly sorted, calcirudite with grain-supported fabric, which contains abundant macrofossils of shallow-marine origin and occasional large intraclasts of the underlying strata, suggesting that it was formed by a grain flow from a carbonate platform on the Cretaceous Cuban Arc. The Lower to Uppermost Members consist of massive calcarenite and calcilutite that show upward fining. Composition of the calcarenite is distinctly different from that of the Basal Member. The homogeneous appearance, coarse-tail normal grading, abundant water escape structures, and abundant reworked fossils in these members are consistent with those of the Mediterranean "homogenite", a deep sea tsunami-induced deposit that was formed by settling from a high density suspension. Repetition of thin conglomerate beds in the Lower Member that contain well-sorted, well-rounded mud clasts and shallow marine fossils is considered to reflect intermittent lateral flow possibly induced by a series of tsunami waves during an early stage of settling of grains from a high density suspension. Altered vesicular glass of probable impact origin and shocked quartz are discovered in the Basal and the Lower to Upper Members, respectively. Together with the biostratigraphically constrained age of the Peñalver Formation, this evidence suggests that the Peñalver Formation has a genetic relation to the K/T impact. Distribution of altered vesicular glass and shocked quartz grains can explain the sequence of the initial grain flow and the following tsunami waves. © 2000 Published by Elsevier Science B.V.

Keywords: Peñalver Formation; Cretaceous–Tertiary (K/T) boundary layer; lithology

* Corresponding author. Japan Broadcasting Corporation, Nagoya Station, Hagashi-ku, 1-13-3 Higashikura, Nagoya, 461-8725, Japan. Fax: +81-52-952-7022.

E-mail address: hdtakayama@aol.com (H. Takayama).

0037-0738/00/$ - see front matter © 2000 Published by Elsevier Science B.V.
PII: S0037-0738(00)00079-8

1. Introduction

Based on abnormally high Ir concentration in the Cretaceous–Tertiary (K/T) boundary layer in the deep-sea limestone of Italy, Alvarez et al. (1980) proposed that the cause of Ir anomaly and associated mass extinction was an impact of an asteroid up to 10 km in diameter on the earth. Eleven years later, Hildebrand et al. (1991) identified a circular subsurface structure about 180 km diameter in the northwestern part of the Yucatan Peninsula, Mexico as the crater formed by the K/T boundary impact. Sandstone units probably formed by the impact-related event also have been reported in the Gulf of Mexico and Caribbean regions since late 1980s (e.g. Bourgeois et al., 1988; Smit et al., 1992). Smit et al. (1996) called these layers a K/T sandstone complex and pointed out that most of these layers are characterized by: (1) a fining-upward sequence several meters in thickness; (2) a poorly sorted and coarse-grained basal part with altered glass spherules; (3) a coarse- to fine-grained middle part with shocked quartz grains (Leroux et al., 1995); and (4) a fine-sand to silt-grained upper part with an Ir and Ni-rich spinel concentrated layer. Several studies proposed that large tsunami waves generated by the K/T impact was the cause of these sandstone layers (e.g. Bourgeois et al., 1988; Albertão and Martins, 1996). Especially, Smit et al. (1996) showed repetition of cross laminated beds with alternate paleocurrent directions in the middle part of the K/T sandstone complex at La Lajilla, Mexico and insisted that they were generated by a series of large tsunami waves caused by the impact. On the other hand, Bohor (1996) argued that these siliciclastic deposits were of sediment gravity flow origin triggered by the K/T impact rather than tsunami origin since lithology of a tsunami-generated deep-water deposit should be single graded and monotonous such as a “homogenite” (e.g. Kastens and Cita, 1981). Recently, a distinctive mixture of reworked fossils, impact-induced materials and lithic fragments, termed the K/T boundary “cocktail”, has been identified at several deep sea K/T boundary sites in the basinal Gulf of Mexico and Caribbean areas (Bralower et al., 1998). Although the origin of the K/T boundary “cocktail” was interpreted to be a result of giant gravity flows triggered by the impact (Bralower et al., 1998), enough evidences has not yet been presented to support this hypothesis.

The uppermost Cretaceous marine strata in the Havana area were originally described by Brönnimann and Rigassi (1963) as the Peñalver Formation. They suggested that it was a single event deposit up to 180 m thick with a basal erosional contact and monotonous upward fining. Pszczolkowski (1986) correlated the Cacarajicara Formation in western Cuba and the Amaro Formation in central Cuba with the Peñalver Formation, and proposed that these formations were originated by a tsunami wave triggered by either a large earthquake or a K/T impact. Later, Bohor and Seitz (1990) suggested the possibility of an impact ejecta blanket origin of these deposits. However, detailed investigation of their origin and relationship with the K/T impact has not yet been done. Iturralde-Vinent (1992) proposed that these formations represent a mega-turbidite related to tsunami, and if they were related to impact the impact may have taken place south of the Cuban Maastrictian carbonate platform. He also proposed that the age of the Peñalver Formation may be older than the K/T boundary. However, an alternative interpretation is that the fossils are of reworked origin and give apparent ages older than the true age of deposition. Consequently, further investigations on the sedimentary process, depositional age, and relation with the K/T impact are necessary.

The purpose of this study is (1) to investigate whether the uppermost Cretaceous strata in Cuba have a genetic relation with the K/T boundary impact and (2) to clarify the sedimentary processes of these formations, based on the study of the Peñalver Formation and its equivalents. The K/T boundary layers previously reported from the Gulf of Mexico and Caribbean region are, at most, several meters in thickness and those with thickness exceeding 100 m have never previously been reported. If the 180 m thick Peñalver Formation has a genetic relation with the K/T boundary impact, understanding its sedimentary processes is essential to understanding the consequences of the impact.

2. Geological setting

In Cuba, the latest Eocene to Recent autochthonous

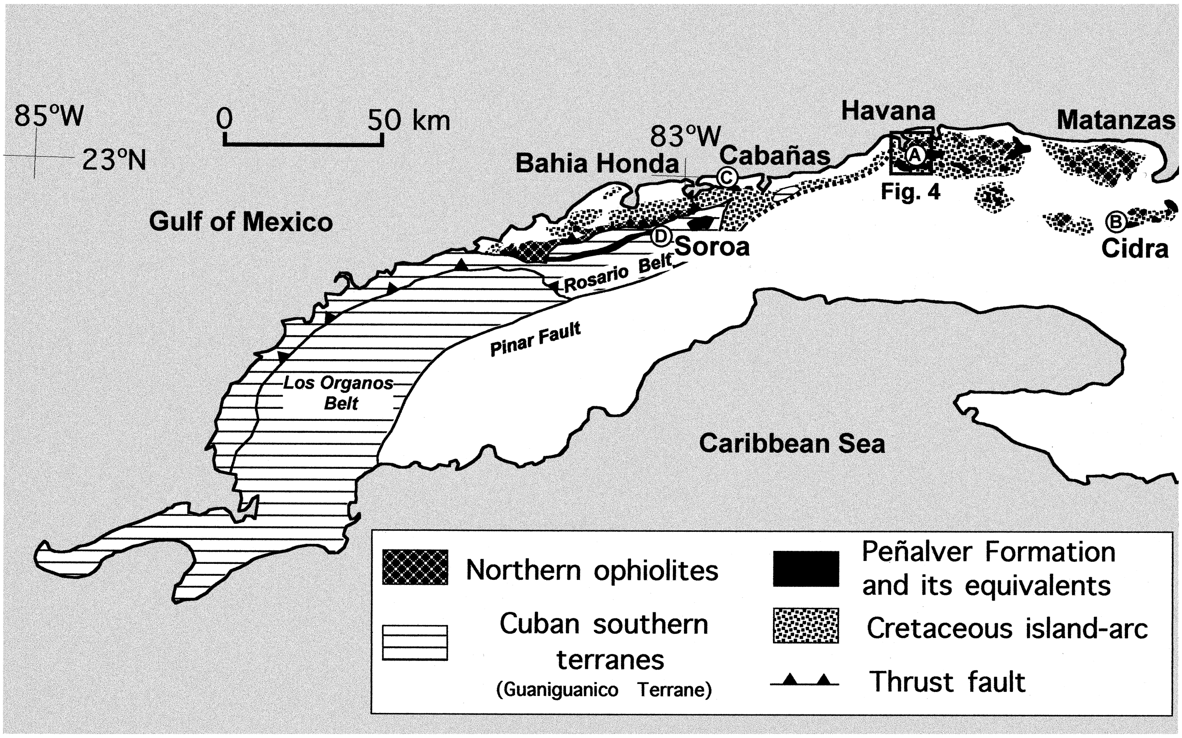

Fig. 1. Location map for the studied sites: (A) Havana; (B) Matanzas; (C) Bahia Honda; and (D) Soroa in northwestern Cuba. The area of Fig. 4 is indicated as a square. Distribution of strata is based on Gutiérrez et al. (1989) and division of the tectonic units is based on Iturralde-Vinent (1994).

strata cover several allochthonous tectonic units which were assembled by a collision between the North American passive margin and the Cretaceous Cuban volcanic arc, a presently extinct arc located between North and South America during Paleocene to latest Eocene (Bralower and Iturralde-Vinent, 1997). The tectonic units include: (1) the Bahamas Platform, (2) the Cuban Southwestern terranes, (3) the Northern ophiolitic melange, (4) the Cretaceous island-arc, and (5) the Paleogene volcanic arc of southeastern Cuba (Iturralde-Vinent, 1994). Among the above units, the Bahamas Platform and the Cuban Southwestern terranes are regarded as continental elements of the North American plate whereas the other three are regarded as oceanic elements related to the Greater Antilles Arc (Bralower and Iturralde-Vinent, 1997).

The Peñalver Formation is considered to have been deposited in a late Campanian to Eocene oceanic basin on the northwestern slope of the Cretaceous Cuban Arc. In the studied area (Fig. 1), the Cretaceous island-arc is composed of extrusive and volcaniclastic rocks of Albian through Campanian Chirino, Encrucijada, La Trampa and Orozco Formations (Albear et al., 1985), which are overlain by late Campanian to Maastrichtian clastic and carbonate rocks (Iturralde-Vinent, 1994). Studied sites, Matanzas, Havana and Bahia Honda, extend in the east–west direction over a distance of approximately 140 km (Fig. 1) and this area corresponds to the northwestern part of the allochthonous Cretaceous island-arc unit. According to Brönnimann and Rigassi (1963), the late Campanian to Eocene basinal deposits of this area are defined as the Via Blanca, Peñalver, the Apolo, Alkazar, Capdevila and Universidad Formations in ascending order (Fig. 2). These formations generally represent hemipelagic facies with intercalations of turbidite.

According to Brönnimann and Rigassi (1963), the Via Blanca Formation is approximately 500 m thick and Campanian to early Maastrichtian in age. It is

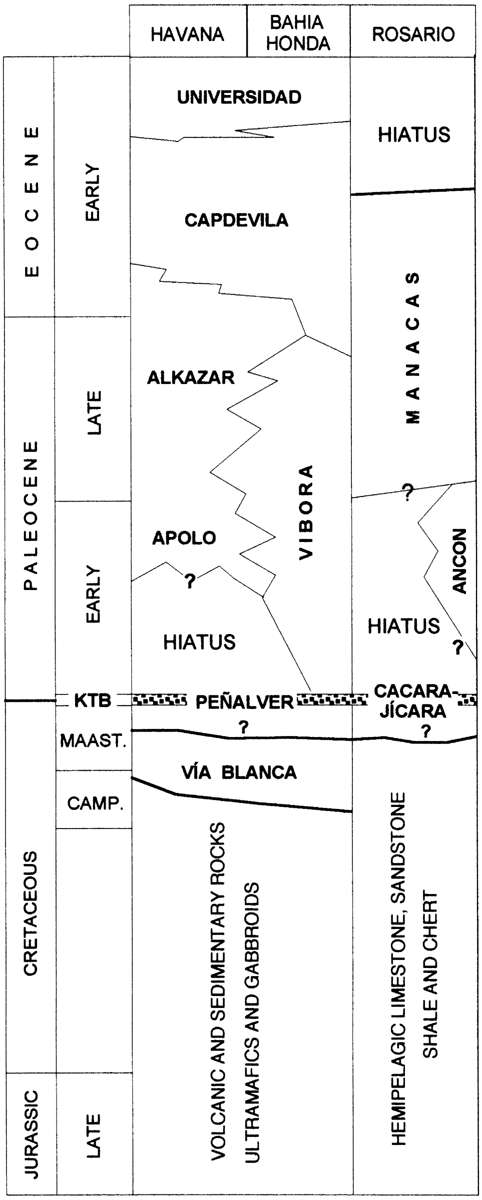

Fig. 2. Stratigraphic division of upper Jurassic to lower Eocene sequence in northwestern Cuba discussed in the text. Correlation with related formations is based on Albear et al. (1985) and Bralower and Iturralde-Vinent (1997).

mainly composed of hemipelagic sediments occasionally interrupted by turbidites, which contain fragments derived from sedimentary reef and igneous lithic fragments derived from the Cretaceous Cuban Arc. The depositional depth of the Via Blanca Formation is estimated as 600 m to 2000 + m based on the ratio of planktonic to benthic foraminifera in pelitic layers (Brönnimann and Rigassi, 1963). The Via Blanca Formation overlies with a major angular unconformity the deformed Cretaceous volcanic arc sequences and the Jurassic to lower Cretaceous ultramafics of the Northern ophiolitic melange unit which was probably derived from the northern oceanic crust (Albear et al., 1985). The Apolo Formation, which overlies the Peñalver Formation with a disconformity, is approximately 100 m thick and the age was estimated as early Eocene based on its planktonic foraminiferal assemblage (Brönnimann and Rigassi, 1963). Lithology of this formation resembles the Via Blanca Formation and grades upward into chalky limestone of the Lower Eocene Alkazar Formation (Brönnimann and Rigassi, 1963).

The Cacarajicara Formation is exposed within the thrust sheet of the Rosario Belt which belongs to one of the Cuban Southwestern terranes and it is thicker in the northern part of the belt (Iturralde-Vinent, 1994). Distribution of the Cacarajicara Formation is controlled by on E–W trending major thrust system and extend in the east–west direction over a distance of up to 60 km. The outcrop near Soroa is located in the middle of its distribution in the Rosario Belt (Fig. 1).

The Greater Antilles Arc is considered to have been located about 1000 km south of the present location during Campanian (e.g. Ross and Scotese, 1988; Pindel and Barrett, 1990), which continuously moved northeastward during the late Cretaceous and early Tertiary. At the end of Cretaceous, the Cretaceous Cuban Arc which is considered to have been a part of the Greater Antilles Arc, was probably located 400–500 km to the south of the present Cuban island according to Rosencrantz (1990) (Fig. 3). The Rosario Belt, on the other hand, is considered to have been originally located along the southeastern margin of the Yucatan Peninsula before Tertiary time (Iturralde-Vinent, 1994; Hutson et al., 1998). The Cretaceous formations in the northern part of the Rosario Belt, including the Cacarajicara Formation, are considered to have been deposited on the eastward-dipping slope in the distal part of the Yucatan block to the west of the depositional site of the Peñalver Formation (Iturralde-Vinent, 1994; Hutson et al., 1998). Northeastward movement of the Cretaceous Cuban Arc

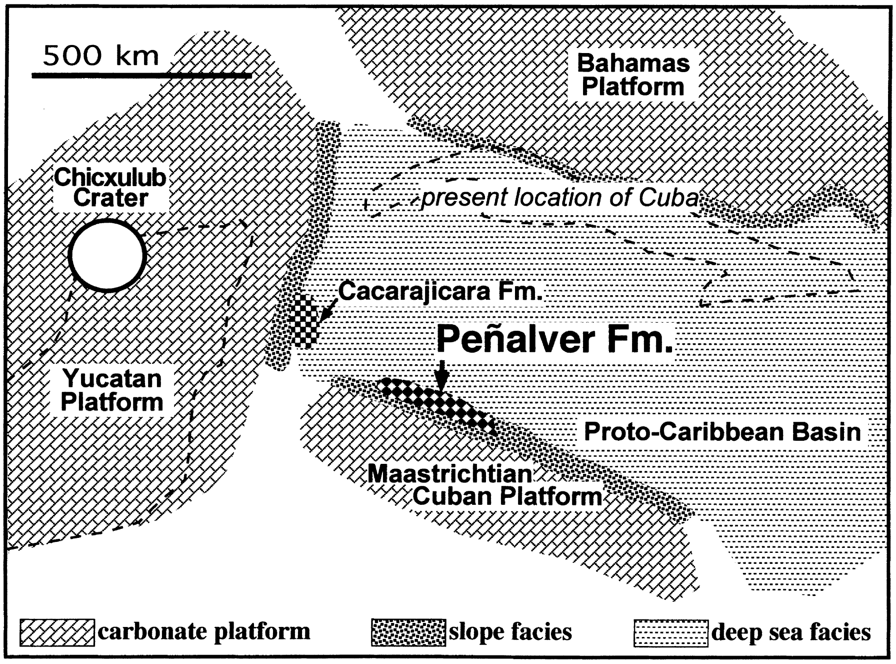

Fig. 3. Paleogeographic reconstruction of the Cretaceous Cuban Arc the during latest Maastrichtian time with the probable distribution of the Peñalver and Cacarajicara Formations adapted from Iturralde-Vinent (1992) and Rosencrantz (1994).

during late Cretaceous to early Tertiary is considered to have caused the detachment of the Guaniguanico terrane, which includes the Rosario Belt, from the southeastern margin of Yucatan block and transported it northward to its present position (Hutson et al., 1998). By late Paleocene to early Eocene, the terrane is considered to have been imbricated into a thrust-bounded belt and the Cretaceous Cuban Arc and Northern ophiolitic melange units obducted on the top of thrust-faulted continental margin sedimentary belts during late Paleocene to early Eocene (Bralower and Iturralde-Vinent, 1997).

3. Studied localities

Geological fieldwork was mainly conducted at Havana, Matanzas, Bahia Honda, and Soroa. A detailed investigation was made around the type locality of the Peñalver Formation in Havana and a preliminary survey was carried out in the other three sites.

Havana area is located in the western end of the east–west trending Habana-Matanzas anticlinorium, and the Peñalver Formation is exposed repeatedly by folds with wave lengths of several kilometers (Fig. 4). The Peñalver Formation is well exposed in an abandoned quarry about 100 m to the east of the type locality (Brönnimann and Rigassi, 1963) in a southeastern suburb of La Habana (Fig. 5). The quarry is situated on the northern flank of the east–west trending syncline, where the upper part of the Via Blanca Formation and the major part of the Peñalver Formation are continuously exposed. However, the uppermost part of the Peñalver Formation is cut by a east–west trending south-dipping normal fault along the syncline axis. The uppermost part of the Peñalver Formation is partly exposed on the southern flank of the syncline. The uppermost part on the southern flank is in fault contact with the major part of the formation on the northern flank. The stratigraphic interval in the upper part of the Peñalver Formation, not exposed around the type locality due to the fault, is exposed at Reparto San Pedro, approximately 6 km to the southwest of the type

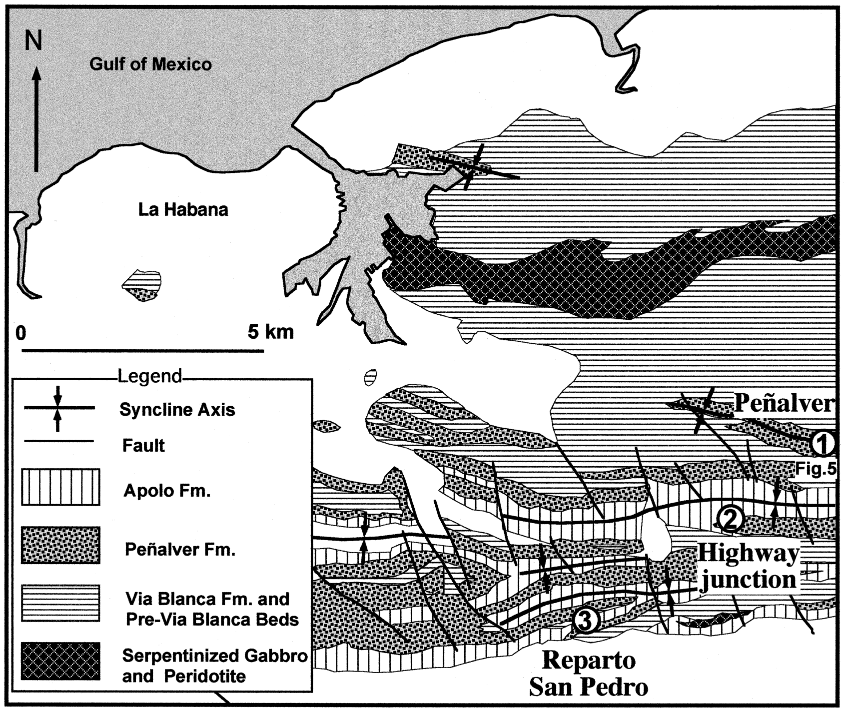

Fig. 4. Distribution of the Peñalver Formation and related formations in Havana area: (1) type locality of the Peñalver Formation; (2) highway junction; and (3) Reparto San Pedro.

locality (Fig. 4). The overlying Apolo Formation is not distributed around the type locality of the Peñalver Formation, but a part of it is exposed at a highway junction about 3 km to the southwest of the type locality of the Peñalver Formation (Fig. 4). The contact with the Peñalver Formation is not exposed there although it used to be exposed before construction of the highway.

The Peñalver Formation in Matanzas province was investigated at Jesus María in an abandoned quarry 5 km to the west of Cidra, Matanzas. The sequence from the upper part of the Via Blanca Formation to the middle part of the Peñalver Formation is continuously exposed at the quarry located within the east–west trending Habana-Matanzas anticlinorium. The upper part of the Peñalver Formation and the Apolo Formation are not exposed at this quarry. The exposed thickness of the Peñalver Formation there is approximately 70 m.

Several outcrops were examined in Bahia Honda, but exposures of the Peñalver Formation are limited. In the eastern part of Cabanas, 60 km to the west of Havana, the over 15 m thick lower part of the Peñalver Formation is exposed.

The Cacarajicara Formation is exposed in the road cut between Soroa and San Diego de Nunez, 80 km to the west of Havana. The basal contact with the underlying units and the upper contact with the overlying Ancon Formation are exposed, but some of the upper parts of the Cacarajicara Formation are not exposed.

4. Lithostratigraphy

4.1. The Peñalver Formation

The Peñalver Formation consists of calcareous clastic rocks which show upward fining from calcirudite to calcilutite (Fig. 6). Thickness of the formation reaches more than 180 m at the type locality. One of the conspicuous characteristics of the Peñalver Formation is lack of any evidence of bioturbation. The Peñalver Formation overlies the Via Blanca Formation with an erosional surface. At the type

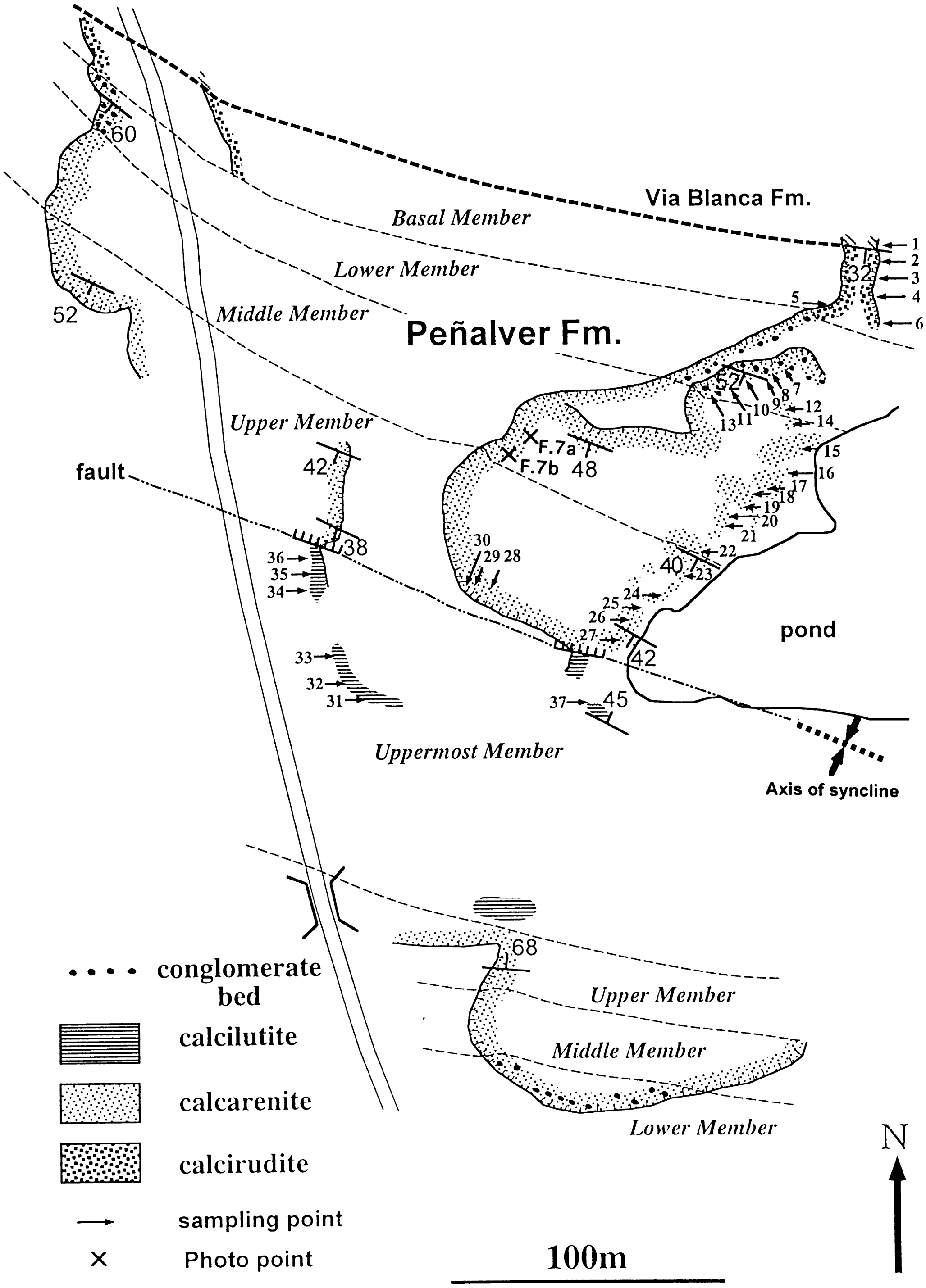

Fig. 5. Route map with sampling localities of the Peñalver Formation around the type locality.

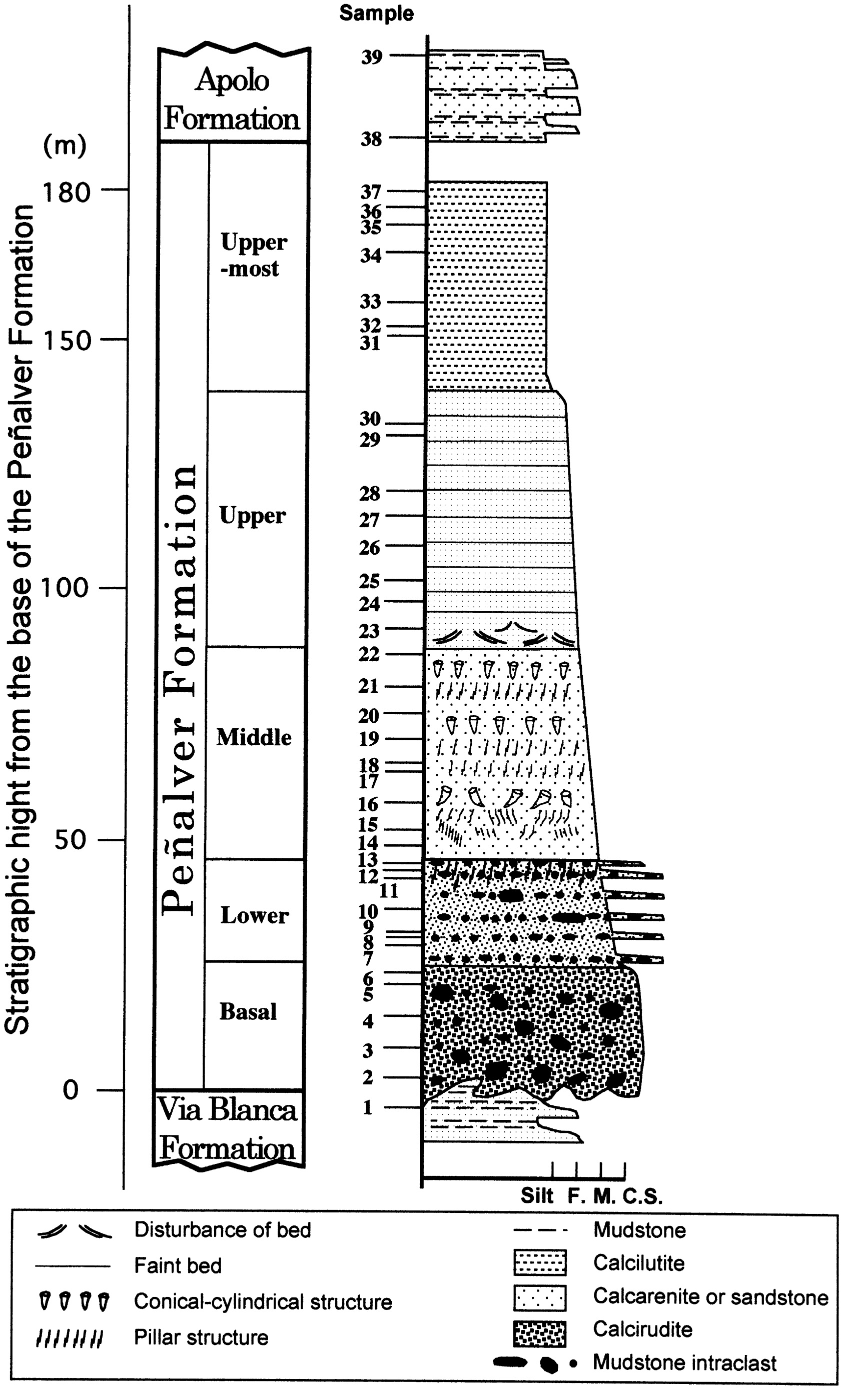

Fig. 6. A lithologic columnar section of the Peñalver Formation and its subdivision into members.

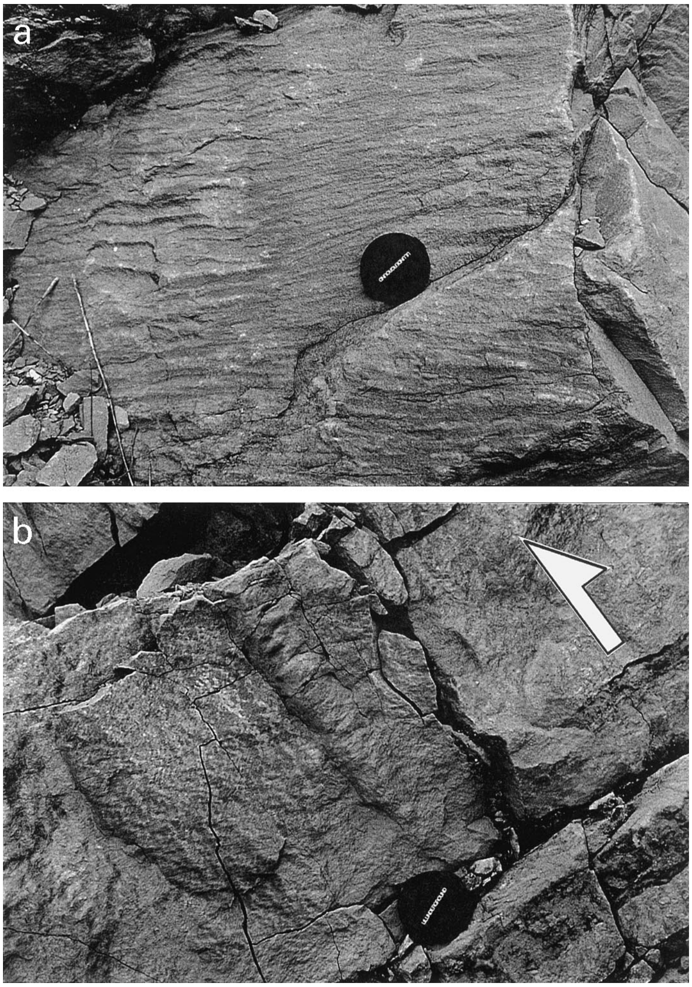

Fig. 7. (a) Pillar structures on the bedding surface in the Middle Member of the Peñalver Formation at the type locality (Photo point: F. 7a in Fig. 5). (b) A conical–cylindrical structure (enclosed by a white line) on the surface vertical to the bedding in the Middle Member of the Peñalver Formation at the type locality (Photo point: F. 7b in Fig. 5). The arrow indicates the upward direction of stratigraphy. Diameter of the lens cap is 5 cm.

locality, only the uppermost part of the Via Blanca Formation is exposed with a thickness of about 1 m that consists of decimeter scale alternations of gray to reddish brown fine-grained turbidite sandstone and hemipelagic mudstone. At Matanzas, the uppermost part of the Via Blanca Formation is exposed for about 3 m in thickness and consists of decimeter-scale alternations of greenish shale and white marl. According to Brönnimann and Rigassi (1963), the Peñalver Formation is overlain by the Apolo Formation with a disconformity. However, the contact between the two formations is not observed at the studied sites. At the highway junction in Havana (Fig. 4), the Apolo Formation is exposed over a stratigraphic interval about 15 m thick, and consists of decimeter-scale alternations of gray graded calcareous turbidites, greenish marlstones, and whitish to greenish calcareous shales. Abundant trace fossils are observed in marlstone and calcareous shale.

The Peñalver Formation is subdivided into five members (Fig. 6). The Basal Member consists of light gray to gray, poorly sorted, massive calcirudite that shows grain-supported fabric. Granule to pebble size grains are mainly composed of subangular to angular fragments of gray to whitish limestone, megafossil fragments such as rudists and large calcareous benthonic foraminifera, and brown to green mudstone fragments. In the basal part of this member, the underlying Via Blanca Formation is entrained as rip-up intraclasts up to 1 m in diameter. Small pieces of mud clasts are disseminated around larger clasts and part of the Via Blanca mud is completely disaggregated and mixed into the matrix of the calcirudite, suggesting that erosion occurred by intense flow when the underlying Via Blanca Formation was not completely lithified. Rip-up clasts of the Via Blanca Formation over 1 m in diameter occur throughout this member and no size grading is recognized. The thickness of this member is 25 m at the type locality, 15 m at Matanzas, and more than 10 m at Bahia Honda.

The Lower Member consists mainly of gray coarse- to medium-grained calcarenite with frequent intercalations of thin conglomerate beds. The thickness of this member is 20 m around the type locality and 13 m at Matanzas. The calcarenite in the lowest part of this member gradually changes from the underlying Basal Member and shows gradual upward fining with the sorting also becoming better upward. Fragments of carbonate bioclasts and greenish gray mudstone are occasionally contained in the lower part of the calcarenite but their amount decreases upward. The several centimeters thick conglomerate beds repeat 14 times with intervals of 25 cm to 4 m. The conglomerate is mainly composed of rounded spherical to ellipsoidal pebbles of greenish gray mudstone accompanied with a small amount of rudist fragments. The pebbles are 1–2 cm in diameter, and generally well-sorted except for a few mudstone intraclasts of over several tens centimeters in diameter. The ellipsoidal pebbles tend to show a preferred orientation with their major axes in an east–west direction parallel to the bedding. At another quarry 2 km to the northwest of the type locality, large intraclasts about 3 m long occur in the upper part of this member. These intraclasts consist of alternations of greenish gray very fine-grained sandstone and blackish gray mudstone which is similar to that of the underlying Via Blanca Formation. A pillar structure occurs in calcarenite of the upper part of this member (Fig. 7a). The pillar structure is composed of white streaks approximately 1 cm wide and about 1 m long that are arranged parallel to each other at intervals of several centimeters in the cross section parallel to the bedding. In a vertical view, the height of pillars ranges from 5 to 10 cm and runs vertical to the bedding. Thin-section observation of the pillar structure does not reveal any obvious difference in composition, grain size, and grain orientation between inside and outside of the structure except for color of the matrix which is lighter within the structure. This structure is similar to stress pillars that are classified by Lowe (1975) as one type of fluidization channels and he consider they are formed during settling of a dense sediment cloud.

The Middle Member consists of gray, massive, well-sorted, medium- to fine-grained calcarenite that shows upward fining. The thickness of this member is 40 m around the type locality. The calcarenite of this member is harder than that of the Lower Member and grain composition does not change significantly throughout the member. The base of this member is defined as the last occurrence of the thin conglomerate bed of the Lower Member. Water escape structures, such as pillar structure and conical–cylindrical structure, previously reported as a cone-shaped concretion by Palmer (1945), are abundant throughout this member. This is a funnel-shape structure of 5–10 cm across and 20–50 cm long, and generally developed perpendicular to the bedding (Fig. 7b). In cross section parallel to the bedding, it is circular. Their distribution is random, separated by a distance of 5–20 cm, although they are locally arranged on line in the lower part of the Middle Member. Like the pillar structure, no obvious difference in composition and grain size is observed except that the color of matrix inside this structure is lighter than that outside. At the type locality, the conical–cylindrical structure is concentrated in three horizons, approximately 11, 23 and 38 m above the base of the this member. On the other hand, pillar structure tends to concentrate several meters below these horizons. As Brönnimann and Rigassi (1963) already mentioned, the conical–cylindrical structure is considered to be a water escape structure, similar in shape and occurrence to a structure formed by spouting, a kind of aggregative fluidization, described by Allen (1982). Aggregative fluidization is considered to occur under conditions of high solids density, low fluid viscosity and large sediment diameter in the absence of intergranular cohesive forces (Allen, 1982). In the lower part of this member, the conical–cylindrical structures and pillar structures are deformed plastically suggesting soft-sediment deformation.

The Upper Member consists of gray to light gray fine-grained calcarenite that shows several tens of centimeters to several meters thick faint bedding. At the type locality, this member is in fault contact with the overlying Uppermost Member and the exposed thickness is up to 40 m. Water escape structures are not recognized except for small scale disturbance of faint beds in the basal part of this member. The calcarenite of this member is similar in composition and fabric to that of the Middle Member. The only major difference is the grain size which is smaller and gradually decreases upward. The base of this member is defined as the horizon where faint bedding becomes recognizable at the outcrop. Within each bed, changes in color, grain size and composition are not obvious, and faint lamination occasionally recognized within the beds does not accompany obvious changes in grain size and fabric in thin-sections. Therefore, bedding and lamination in this member are reflecting, at most, only subtle changes in grain composition or fabric that are possibly enhanced by diagenesis and weathering.

The Uppermost Member consists of massive, gray calcilutite. The estimated thickness of this member is at least 40 m around the type locality. The calcilutite of this member rarely contains angular clasts of black shale of less than 3 cm in diameter and greenish gray rounded mudstone clasts of less than 1 cm in diameter. Although this member is in fault contact with underlying member at the type locality, the contact with the underlying Upper Member is observed at the Reparto San Pedro where massive, gray calcilutite of this member overlies the fine-grained calcarenite of the Upper Member with sharp contact. On the other hand, the contact with the overlying Apolo Formation is not exposed at the studied sites. Evidence of bioturbation is absent throughout this member.

4.2. The Cacarajicara Formation

For comparison with the Peñalver Formation, lithology of the Cacarajicara Formation is briefly described here. The Cacarajicara Formation also consists of clastic rocks that show upward fining from boulder-size breccia to calcilutite. It is up to 400 m thick in Soroa. Although exposure of the upper part is not good, this formation is temporarily subdivided into three members.

The Lower Member consists of massive, boulder- to pebble-size breccia that is mainly composed of fragments of black chert, bioclastic micritic limestone, sparry limestone and greenish altered volcano-clastic rocks. The breccia is clast-supported with its matrix filled with dark gray micritic calcite grains. Fragments of rudists are rarely recognized. This member unconformably overlies several older units from the Campanian up to the Tithonian. The basal part of this member is chaotic and mainly composed of cobble to boulder-size fragments of black chert, micritic and sparry limestone and greenish altered volcanoclastic rocks with their maximum diameter exceeding 2 m. The size of rubble decreases upward and grades into the coarse-grained calcarenite of the Middle Member. Thickness of the Lower Member is about 150 m. The Lower Member of the Cacarajicara Formation is characterized by large fragments derived from deep sea facies such as chert and limestone, and is distinctly different from that of the Basal Member of the Peñalver Formation.

The Middle Member mainly consists of dark gray, massive, well-sorted, coarse- to fine-grained, calcarenite that shows upward fining. In the upper part of calcarenite of this member, several meters thick faint bedding is recognized. This member is tentatively correlated with the calcarenite of the Lower to Upper Members of the Peñalver Formation, although intercalations of thin conglomerate beds and water escape structures are not recognized. The thickness of this member is estimated as about 150 m near Soroa.

The Upper Member is not exposed well. But several meters thick gray massive calcilutite that is conformably overlain by calcareous shale of the Paleocene Ancon Formation is exposed. The Upper Member is estimated as over 50 m in thickness and is correlated with the Uppermost Member of the Peñalver Formation based on its lithology.

5. Biostratigraphy

The depositional age of the Via Blanca Formation was previously considered as Campanian to early Maastrichtian based on planktonic foraminifera, such as *Globotruncana linneiana* and *Rugotruncana gansseri* (Brönnimann and Rigassi, 1963). Rudists of probable reworked origin are found from turbiditic conglomerates and identified as *Barrettia monilifera* fauna of Campanian age and *Titanosarcolites giganteus* fauna of latest Campanian to Maastrichtian age (Rojas et al., 1995). In this study, calcareous nannofossils, *Micula murus*, *Nannoconus* spp., and *Thoracosphaera* spp. are found in a turbiditic siltstone approximately 10 cm beneath the Peñalver Formation at the type locality. *Micula murus* is useful marker of the Late Maastrichtian and its first occurrence is estimated as 66.2 Ma (Bralower et al., 1995). On the other hand, *Nannoconus* was extinct long before the Maastrichtian (Perch-Nielsen, 1985) and its presence is considered as due to reworking. Tertiary species are not found in this formation. Thus, the upper age limit of the Via Blanca Formation is estimated as after 66.2 Ma and before the first appearance of Tertiary species at 65.0 Ma.

Rudist fragments commonly occur in the Basal and Lower Members of the Peñalver Formation. The rudists include *Barrettia monilifera* and *Titanosarcolites giganteus* faunas that are found in the Maastrichtian carbonate platform deposits of Central Cuba (Rojas et al., 1995). In this study, a calcareous nannofossil, *Micula prinsii*, is found in a large mudstone intraclast in the Lower Member of the Peñalver Formation. The first occurrence of the *Micula prinsii* is estimated as 65.4 Ma in the latest Maastrichtian (Bralower et al., 1995). Tertiary species are not found in the Peñalver Formation. Consequently, the Peñalver Formation is inferred to be younger than 65.4 Ma and older than 65.0 Ma, the first occurrence of Tertiary species. The fact that *Micula murus* is found throughout the Peñalver Formation is consistent with its constrained age. However, *Nannoconus* sp. is found from the Uppermost Member, and *Reinhardtities levis* whose last occurrence is estimated as 69.2 Ma (Erba et al., 1996), *Tranolithus orinonatus* whose last occurrence is estimated as 69.6 Ma (Bralower et al., 1995), *Eiffelithus eximius* whose last occurrence is estimated 75.0 Ma (Erba et al., 1996) are found in the Upper and Uppermost Member. Their presence is inconsistent with the constrained age of the Peñalver Formation and these species are considered as having been reworked, as is the case for the Via Blanca Formation.

Planktonic foraminiferas, *Globotruncana rosetta*, *G. linneiana*, *G. arca*, *G. stuartiformis*, *G. bulloides*, *Rugoglobogerina* sp., *Globotruncanita stuarti*, *Contusotruncana plicata* are also found in the Lower to Upper Member. *G. linneiana*, *G. Ventricosa*, *G. bulloides*, and *Heterohelix* sp. are found in the Uppermost Member. Tertiary species are not found. Among these foraminiferas, *G. linneiana*, *G. Ventricosa*, and *G. bulloides* are considered to have been extinct within the *Gansserina gansseri* zone (Sliter, 1989), and the assemblage of foraminifera found in the Lower to the Uppermost Member of the Peñalver Formation in this study is the same as that of the *Gansserina gansseri* zone (72.8–68.25 Ma; Shipboard Scientific Party, 1997). On the other hand, Iturralde-Vinent (1992) reported planktonic foraminifera *Abathomphalus mayaroensis* (68.25–65.0 Ma; Shipboard Scientific Party, 1997) from the uppermost part of the Peñalver Formation. A life span of this species is consistent with the estimated age of the Peñalver Formation based on calcareous nannofossils. Consequently, the species that were extinct before first appearance of *A. mayaroensis* are considered to be of reworked origin and the assemblage of planktonic foraminifera found throughout the Peñalver Formation in this study suggests reworking from the underlying Via Blanca Formation.

Planktonic foraminifera, *Morozovella aequa* (56.5–53.6 Ma; Berggren et al., 1995) is found from the lowest part of the Apolo Formation. Calcareous nannofossils from a sample in the upper part of the Apolo Formation characterized by an assemblage typical for Late Paleocene (CP8, 56.2–55.0 Ma; Shipboard Scientific Party, 1997). Therefore, the deposition of the Apolo Formation is at least earlier than the last occurrence of *M. aequa* (53.6 Ma, Berggren et al., 1995). Recently, an earliest Paleocene (NP1) nannofossil assemblage with a substantial contribution of late Cretaceous assemblage was reported from the lower part of the Vibora Group near Bahia Honda, which also overlies the Peñalver Formation (Bralower and Iturralde-Vinent, 1997). This is consistent with our inference that the upper limit age of the Peñalver Formation is before the age of the first occurrence of a Danian species (65.0 Ma).

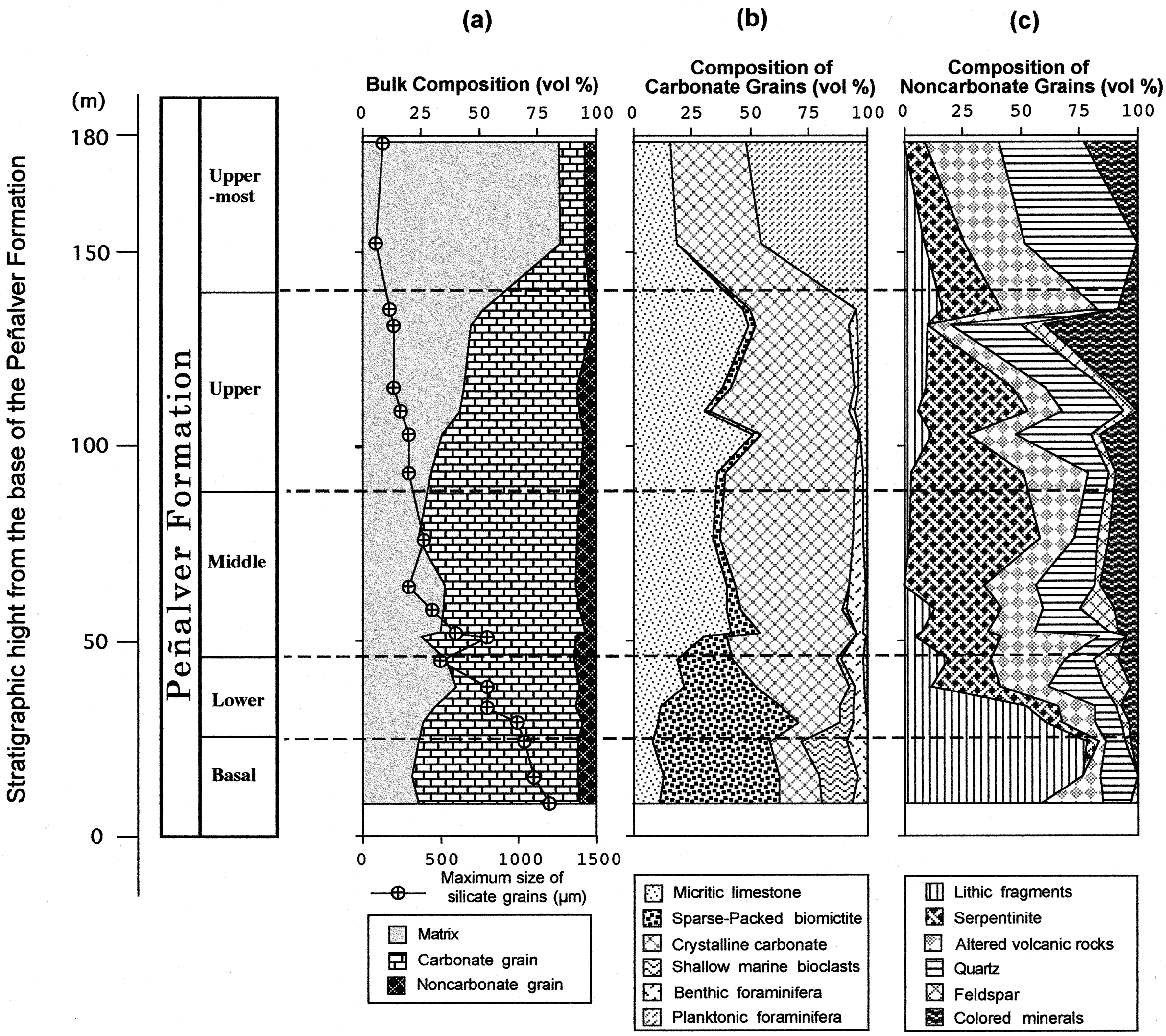

Fig. 8. Vertical variation in grain composition of the Peñalver Formation based on point counting: (a) composition of bulk samples; (b) composition of carbonate grains; and (c) composition of noncarbonate grains.

6. Composition of the Peñalver Formation

6.1. Analytical methods

Petrographical, mineralogical, chemical and grain-size analyses of the Peñalver Formation were conducted in order to clarify its sedimentary process, source, and relation with the K/T impact event. Seventy-five thin-sections representing every 2 m stratigraphic interval were observed under the petrographic microscope to examine grain composition, texture and sizes. Composition of grains larger than 32 μm in diameter for 20 selected samples were determined by a point counting method. Five hundred points were counted on each thin section at 0.5 mm point spacing along transverse lines of 0.6 mm apart. Chemical composition of major component grains was determined by electron probe microanalyzer JEOL JCMA-733MK II at Geological Institute, University of Tokyo. Analytical procedures are similar to those given by Nakamura and Kushiro (1970) with correlation procedure of Bence and Albee

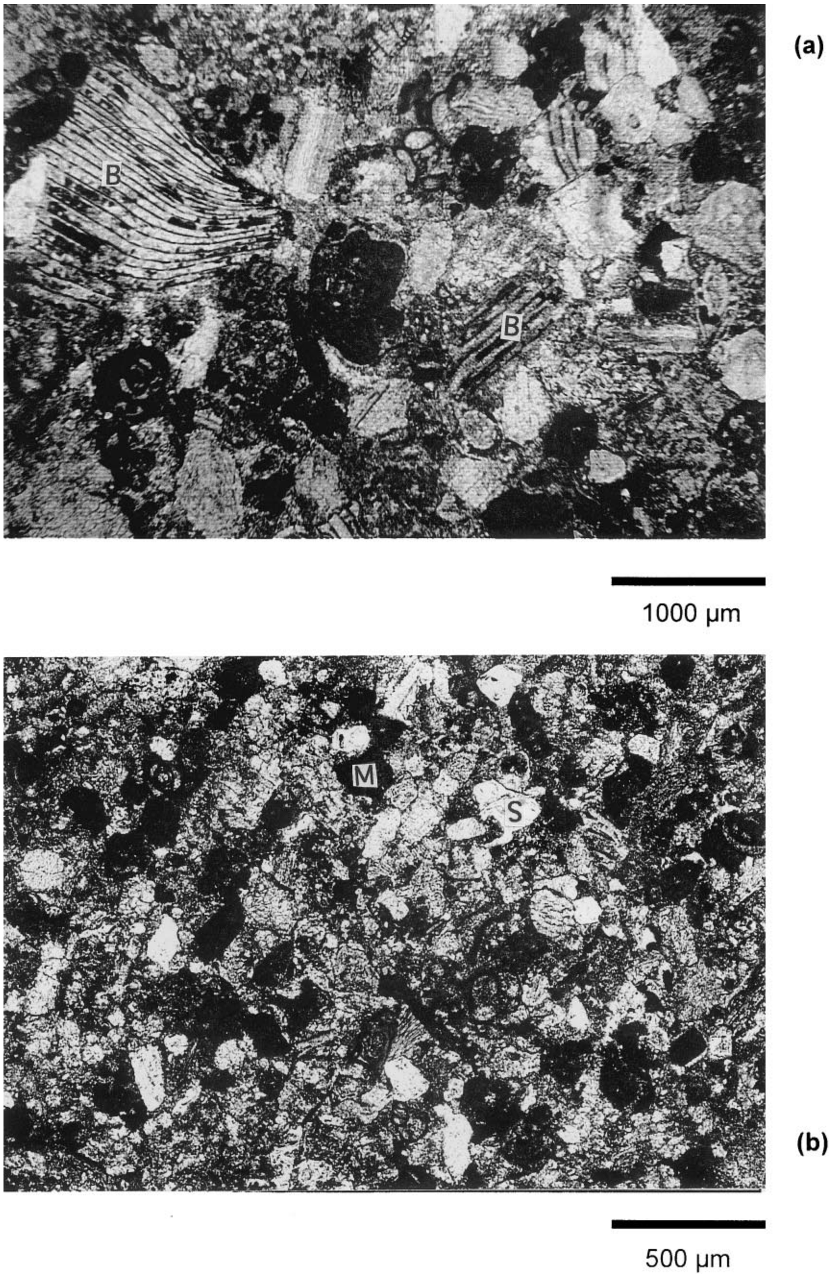

Fig. 9. (a) A thin section photomicrograph of calcirudite of the Basal Member (sample no. 3). Abundant fragments of bioclasts (B) are observed and fabric is grain-supported. Thin section photomicrographs of (b) calcarenite of the Middle Member (sample no. 16) and (c) calcarenite of the Upper Member (sample no. 26). Micritic limestone (M) and serpentine grains (S) are abundant. In the Upper Member, average diameter of grains becomes smaller and sorting becomes better than those of the Middle Member. (d) A thin section photomicrograph of calcilutite of the Uppermost Member (sample no. 31). Grains are mainly composed of planktonic foraminifera (P) and fabric is matrix-supported.

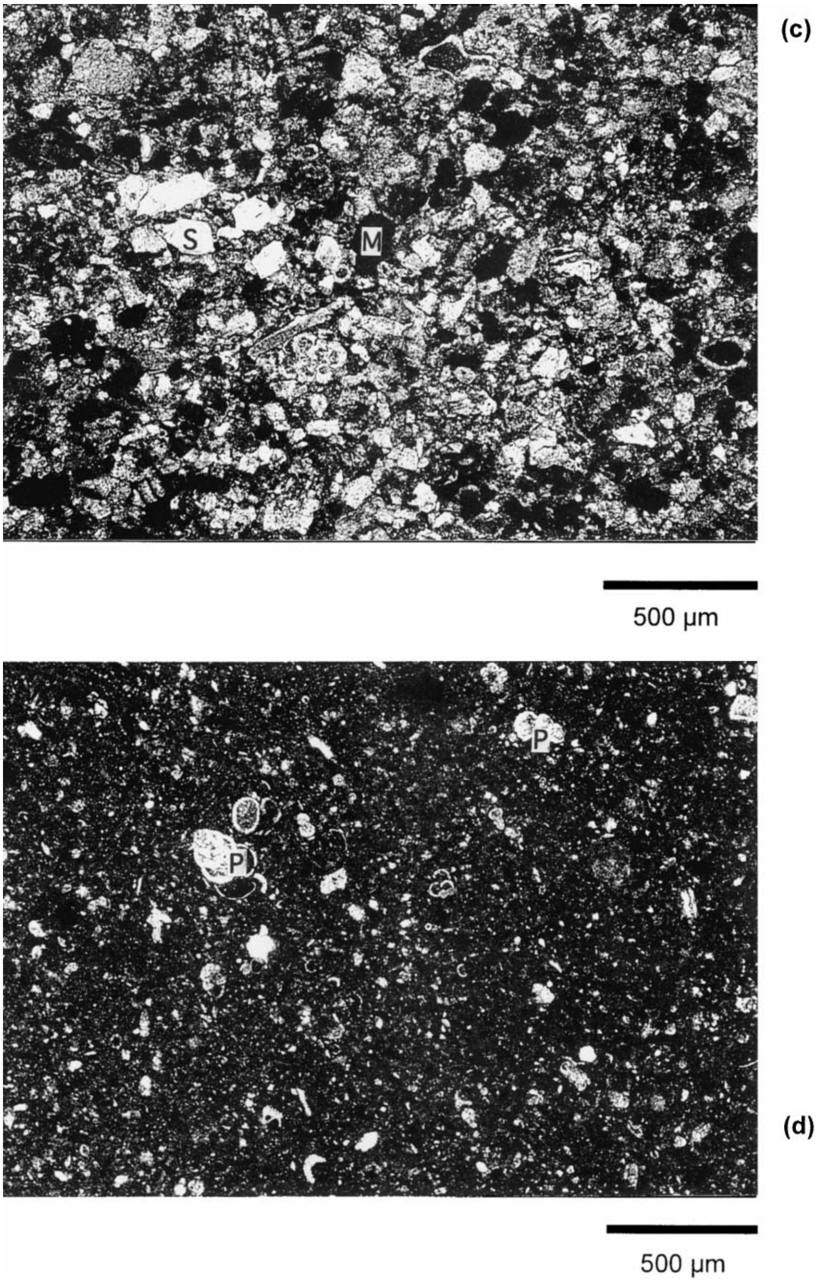

Fig. 9. (*continued*)

(1968). Operating condition was 15 kV accelerating voltage and 12 nA beam current and 5 s counting time with 1–20 μm broad beam.

We also examined shocked quartz from 12 selected samples. Concentration of quartz was attempted before examination. As a first step, about 10 g samples were weighted and treated with 6 N HCl for 12 h, then the HCl insoluble residue was decanted to isolate the lager than 32 μm fraction based on the procedure of Leroux et al. (1995). The fraction larger than 32 μm of residue was then treated with hydrofluorosilicic acid (H_2SiF_6) to dissolve silicate minerals other than quartz (Sridhar et al., 1975). The final residue, mostly composed of quartz, was dried, weighed and then examined by microscopic observation.

Semiquantitative bulk mineral composition analysis of 21 selected samples was conducted by a MAC Science MXP3 X-ray powder diffractometer at Geological Institute, University of Tokyo. Measurements were conducted at 40 kV and 20 mA with variable slit system of 25 mm beam width, scanning speed of $4°2\theta$/min and data sampling step of $0.02°2\theta$. Since bulk samples are highly calcareous, samples were treated with acetic acid for 12 h to remove calcite. This treatment does not dissolve dolomite. The insoluble residue was subject to mineral composition analysis under the same analytical conditions except for scanning speed of $2°2\theta$/min.

The contents of Cr and Ni of 16 selected samples were determined on pressed-powder bulk samples using a Philips PW-1480 X-ray fluorescence spectrometer at Geological Institute, University of Tokyo. Analytical procedures and error ranges are similar to those by Yoshida and Takahashi (1997). Iridium was determined for 12 selected bulk samples by radiochemical neutron activation analysis (RNAA) using a high-resolution Ge detector at Central Institute of Isotope Science, Hokkaido University, after samples in quartz tubes were irradiated in the JRR-3M reactor of the Japan Atomic Energy Research Institute (JAERI) at a neutron flux of 1.2×10^{14} cm^{-2} s^{-1} for 6 h. RNAA procedures and analytical precision are essentially the same as those of Kyte et al. (1992).

The grain-size distribution of the nine insoluble residues, which were treated with 10% HCl and then with 10% H_2O_2, was analyzed by a Horiba LA-920 laser diffraction grain-size analyzer at Geological Institute, University of Tokyo.

6.2. Results

6.2.1. Petrography

Calcirudite of the Basal Member mainly consists of poorly sorted angular to subangular granules 1–2 mm in diameter. Over 90% of total grains are calcareous including fragments of biomictite, micritic limestone, rudists and other mollusks, large calcareous benthic foraminifera, echinoderm, and other bioclasts of probable shallow-marine origin (Fig. 8). Noncarbonate grains are composed of lithic fragments of mudstone and altered andesitic volcanic rocks, monocrystalline and polycrystalline quartz, feldspar, and biotite. Altered vesicular glass shards are rarely observed as will be described later. Mudstone fragments are as much as 2 cm in diameter and probably derived from the underlying Via Blanca Formation as rip-up clasts. The matrix is less than 30% in volume and composed of micrite and small bioclastic fragments. Fabric is completely grain-supported (Fig. 9a).

Calcarenite of the Lower Member mainly consists of subangular calcareous grains of micritic limestone and biomictite fragments, crystalline calcite or dolomite, and bioclasts such as benthic foraminifera and rudist, which comprise 82–88% of total grains. Noncarbonate grains are composed of lithic fragments of mudstone, shale, serpentine, and altered andesitic volcanic rocks, and quartz, feldspar and colored minerals such as amphibole and spinel. Sorting improves upward from poor to moderate and average grain size decreases upward from 1 mm to 500 μm. Matrix is up to 40% in volume and composed of micritic calcite particles. The fabric is grain-supported. Fragments of large bioclasts and biomictite of probable shallow marine origin decrease whereas the content of agglutinated benthic foraminifera increases upward through this member. Serpentine grains increase upward from 10 to about 30% of the noncarbonate grains whereas mudstone fragments decrease upward.

Calcarenite of the Middle to Upper Members mainly consists of moderate- to well-sorted, subangular to subround grains 400–100 μm in diameter that gradually decreases upward. Sorting also becomes better upward. Approximately 90% of grains are calcareous composed of micritic limestone fragments, crystalline calcite or dolomite and bioclasts. Large macrofossil fragments of shallow-water origin are nearly absent. Agglutinated benthic foraminifera become common (up to 90% of bioclasts) at the base of Middle Member but decrease upward in the Upper Member. Conversely, planktonic foraminifera contribute less than 10% of bioclasts at the base of the Middle Member but increase upward to 70% in the upper part of the Upper Member. As for non-carbonate grains, fragments of serpentine reach 50% of

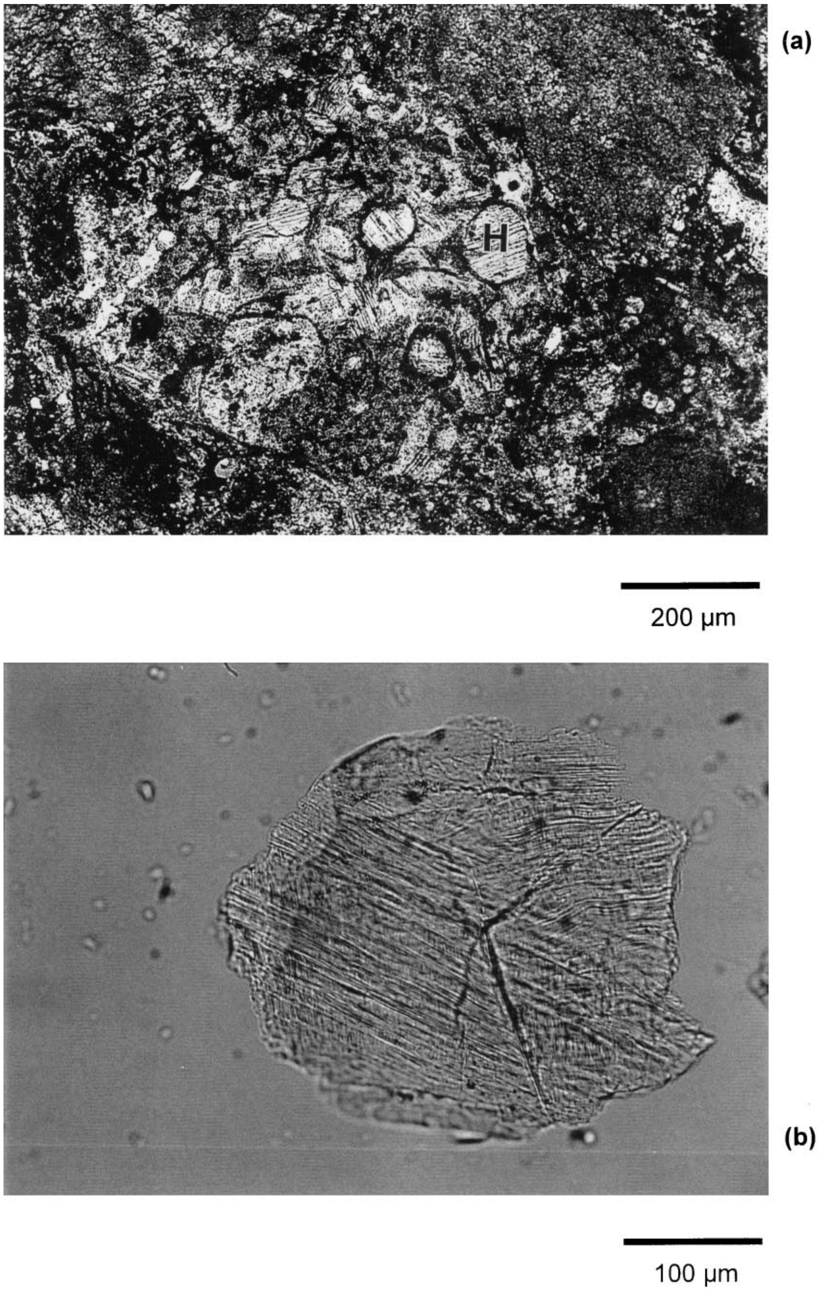

Fig. 10. (a) A thin-section photomicrograph of altered vesicular glass in the Lower Member of the Peñalver Formation (sample no. 8). Heulandite group zeolite (H) fills cavities. Thin section with open nicol. (b) A photomicrograph of a quartz grain with planar deformation feature in the Lower Member of the Peñalver Formation (sample no. 13). Open nicol.

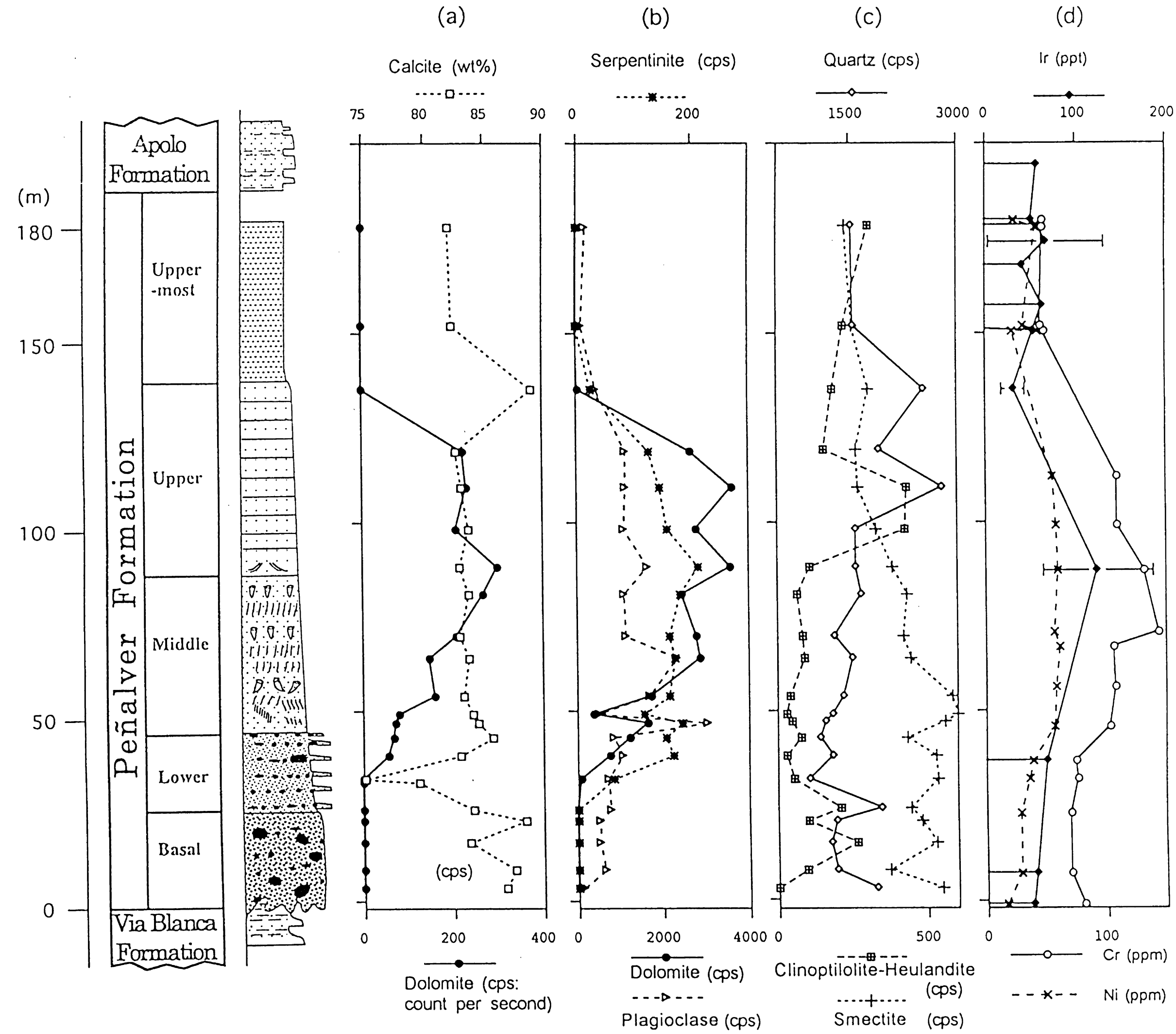

Fig. 11. (a) Vertical variations in weight % of calcite and X-ray diffraction peak heights of dolomite in bulk samples, (b) vertical variations in X-ray diffraction peak heights of dolomite, serpentine and plagioclase in the residue samples, (c) quartz, heulandite group zeolite and smectite in the residue samples, and (d) concentration of Ir, Cr, and Ni in bulk samples of the Peñalver Formation.

non-carbonate grains in the Middle Member and lower part of the Upper Member. In the upper part of the Upper Member, fragments of serpentine slightly decrease and quartz slightly increases upward. Other lithic fragments are small in amount (<10%). Altered volcanic lithic fragments and colored minerals such as spinel and amphibole are present in small amounts. Matrix is usually micritic although sparry calcite fill veins are occasionally found. In the Middle Member, calcarenite is grain-supported with the matrix comprising 30–40% (Fig. 9b) that gradually increases upward. In the Upper Member, the matrix comprises up to 50% and the fabric becomes matrix-supported (Fig. 9c).

Serpentine grains in the Peñalver Formation usually show mesh texture that suggests alteration from olivine, whereas some of them have bastite texture which suggests alteration from orthopyroxene. Reddish brown detrital spinel is also present in small amounts within the Lower to Uppermost

Member. The existence of the serpentine and the chemical composition of spinel suggest that they were derived from upper mantle peridotite of island arc or fore-arc setting (Arai, 1990), most likely from the ultramafic body of the Northern Ophiolites.

Calcilutite of the Uppermost Member is composed of more than 80% matrix and less than 20% grains of up to 50 μm in diameter, showing a matrix-supported fabric (Fig. 9d). The grains consists of planktonic foraminifera (30–40%), limestone fragments (~30%), and quartz (10–15%). A small amount of serpentine, altered volcanic fragments and chromian spinels are found. In addition, mudstone fragments over 100 μm in diameter are occasionally found. The matrix consists dominantly of calcareous nannofossils.

6.2.2. Identification of ejecta

Possibility of the presence of impact ejecta such as altered tektite and shocked quartz was also examined during the course of petrographic observation. A small number of altered vesicular glass fragment (83 grains out of 84 cm^2 of examined area) were found in the Basal Member and the lower part of the Lower Member (Fig. 10a). Their concentration is the highest in the basal part of the Basal Member and decreases upward. They are not present above the middle part of the Lower Member. The maximum size of these grains reaches 2 mm. The vesicular glass grains show bubbly texture with bubble diameter of several tens μm. The bubbles generally have smectite rim-cement and are cemented either with sparry calcite, heulandite or smectite. Locally heulandite replaces part of the glass. Most of the vesicular glass grains are angular. The vesicular glass grains are similar in shape and texture to bubbly spherules of probable ejecta origin described from the lower part of the K/T boundary layers in the Gulf Coast (Smit et al., 1996), or vesicular glass shards described in probable ejecta blanket deposits in Belize (Ocampo et al., 1996). Apart from vesicular glass grains, one brownish altered glass spherule of up to 280 μm in diameter, which is considered to be an altered microtektite, is found from the middle part of the Basal Member.

In addition, quartz grains with visible planar structure are found from the Lower, Middle and Upper Members (Fig. 10b). This planar structure is composed of lamellae that are less than 1 μm in width and run parallel to each other with a spacing of 5 μm. One of the grains shows at least four sets of planar lamella with different orientations. These features agree with planar deformation features (PDFs) of shocked quartz produced under pressure above 12 GPa that is only possible by impact (Koeberl and Anderson, 1996). Thus, these quartz grains are identified as shocked quartz formed by asteroid impact. The maximum size of shocked quartz grains reaches 380 μm in the Lower Member and their sizes decrease upward in the similar way as other grains. The existence of these materials strongly suggests that the Peñalver Formation was formed in association with the K/T boundary impact.

6.2.3. Mineral composition

The Peñalver Formation is dominantly composed of calcite of which content is of 75–87% and slightly decreases upward (Fig. 11a). Small amounts of dolomite, quartz and plagioclase are also detected. Dolomite is present through the Lower to Upper Member and reaches a maximum (about 8%) in the upper part of the Middle Member. Dolomite is not detected in the Basal and Uppermost Members. Quartz and plagioclase contents are less than 1% throughout the formation. In order to examine mineral composition of non-calcite fraction, acetic acid insoluble residue samples were examined. The residue is mainly composed of dolomite, quartz, plagioclase, smectite, clinoptilolite-heulandite, and serpentine. Illite, amphibole and microcline are minor components and contained only in some samples.

The major component minerals are divided into three groups based on their interrelationships and similarities in the pattern of vertical variation. First group is serpentine, plagioclase and dolomite (Fig. 11b). In the Basal Member, dolomite and serpentine are not present and plagioclase is present only in small amount. Their contents increase upward in the Lower Member, stay high in the Middle Member, decrease upward in the upper Member, and are nearly absent in the Uppermost Member. Second group is clinoptilolite-heulandite and quartz (Fig. 11c). Their contents are moderate in the Basal Member, low in the Lower Member, slightly increase upward in the upper part of the Lower Member to the Middle Member, high in the lower part of the Upper Member, and moderate in the Uppermost Member. Third group is smectite that is

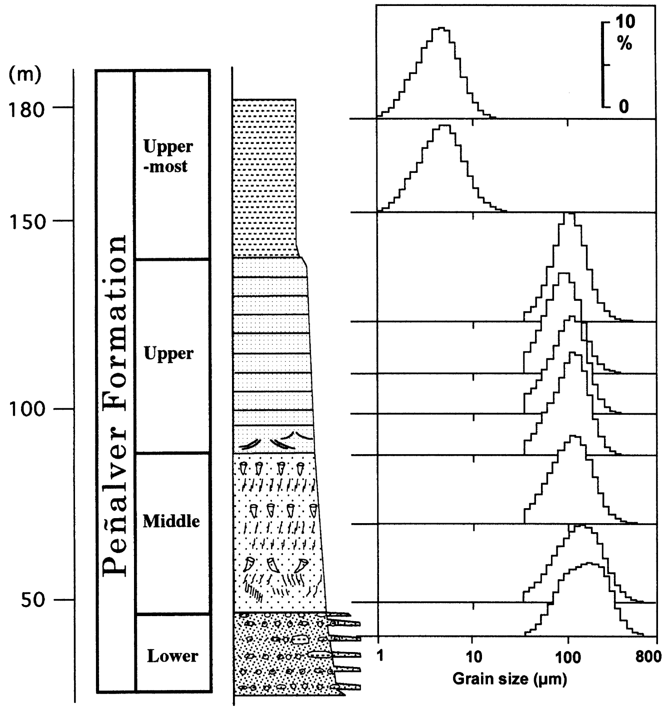

Fig. 12. Vertical variation of grain size distributions of >32 μm HCl insoluble residues of the calcarenite from the Lower to the Upper Member and HCl insoluble residues in the calcilutite of the Uppermost Member of the Peñalver Formation. The upward fining for calcarenite with increasing kurtosis shows the coarser grains decrease rapidly whereas modal position dose not change significantly.

high in the Basal and Lower Members, decreases upward in the Middle to the lower part of the Upper Member and is low in the upper part of the Upper Member and the Uppermost Member (Fig. 11c).

Illite is present in small to negligible amounts throughout the formation, whereas amphibole is present in several samples from the Basal, Lower, and Upper Members but not present in the Middle and Uppermost Members. Microcline is present throughout the Basal Member except in its lowest part, and is also present in small amounts at several horizons throughout the formation. Serpentine and dolomite are also detected in the calcarenite of the Cacarajicara Formation of which the composition is similar to that of the Lower to the Upper Member of the Peñalver Formation. On the other hand, the residue of the sand layers of the Via Blanca Formation and the Apolo Formation are mainly composed of smectite, clinoptilolite-heulandite, quartz, and plagioclase but do not contain serpentine and dolomite. The mineralogical composition of the mudstone pebbles of the conglomerate bed of the Lower Member resembles that of mudstone in the Via Blanca Formation.

Mineral compositions of the Peñalver Formation suggest that contribution of serpentine is significant within the interval from the base of the Lower Member to the top of the Upper Member. Whereas, its contribution is absent in the Basal Member. Consequently, the composition changes distinctly between the Basal and the Lower Members.

6.2.4. Minor elements

Many previous studies (e.g. Doehne and Margolis, 1990) showed that platinum group elements such as Ir, together with Cr and Ni are concentrated in the K/T boundary layer all over the world. Especially, high concentration of Ir is considered as one of the best evidences to support the contribution of extraterrestrial material. In the Peñalver Formation, the contents of Cr and Ni varies between 46.5 and 144 ppm and 24.1 and 61.6 ppm, respectively (Fig. 11d). The contents of Cr and Ni in the siltstone of the underlying

Via Blanca Formation are 81.0 and 16.3 ppm. In the Peñalver Formation, their contents show strong positive correlation with MgO and also serpentine content.

Ir content in the Peñalver Formation is compared with that in the underlying Via Blanca Formation and the overlying Apolo Formation. Ir content of the Peñalver Formation is less than 68 ppt except one sample from the Middle Member which is 124 ± 61 ppt (Fig. 11d). The Ir content of less than 68 ppt in the Peñalver Formation is not significantly different from those of the Via Blanca and the Apolo Formations (<57 ppt). A high Ir sample also shows high concentration of Cr and Ni. Ir (ppb)/Cr (ppm) ratio of the sample is $0.94 \pm 0.46 \times 10^{-3}$ that is similar to other samples from the Peñalver Formation (0.79×10^{-3} to 1.22×10^{-3}). These values are similar to the average value of upper mantle (0.94×10^{-3}; Wänke et al., 1984), or crust (0.54×10^{-3}; Taylor and McLennan, 1985), and much smaller than that of C1 chondrites (0.18; Anders and Grevesse, 1985). Thus, slightly high contents of Ir, Cr and Ni in the Middle and Upper Members do not necessarily indicate contribution from extraterrestrial material but can be explained by larger contribution of upper mantle materials such as peridotite.

6.2.5. Grain size distribution

HCl and H_2O_2 treated residue of calcarenite from the Lower to Upper Member shows bi-modal grain size distributions with finer and coarser modal positions at about 10 μm and about 100 μm, respectively. Grain size distribution of residue from calcilutite in the Uppermost Member shows a uni-modal distribution with its modal position at about 6 μm. Finer mode of the calcarenite residue is interpreted to be contributed from fine fraction contained in limestone fragments, which was the major component of the calcarenite. Consequently, we examined grain-size distribution of coarser mode using the larger than 32 μm fraction, which was extracted from the residue by decantation (Fig. 12). The median diameter of these fractions decreases from 139 μm in the Lower Member to 83 μm in the Upper Member, showing upward-fining except one sample from the Upper Member. The maximum diameter decreases from 600 to 340 μm that results in a leptokurtic distribution upward. This upward fining with increasing kurtosis is called coarse-tail grading that shows a decrease in the amount of the coarsest few percent grains whereas modal position does not change significantly (Allen, 1982). The coarse-tail grading is considered to be caused by settling from high concentration suspensions with solid fraction of greater than 30% by volume (Middleton, 1966).

7. Discussion

7.1. Depositional process

The Basal Member of the Peñalver Formation is composed of poorly sorted massive calcirudite with grain supported fabric, which overlies the Via Blanca Formation with an irregular erosional surface and contains large intraclasts throughout the member. These features are consistent with those of grain flow deposits clarified by many sedimentologists (e.g. Stauffer, 1967). The Basal Member contains a large amount of bioclasts such as rudists as well as bioclastic limestone fragments that were derived from shallow-marine carbonate platform. The taxa of rudists in the Basal Member suggests a carbonate platform on the Cretaceous Cuban Arc to the south of the depositional site as a source area (Iturralde-Vinent, 1992). The Basal Member is distributed throughout the northwestern basin of the Cretaceous Cuban Arc terrane over a distance of 150 km direction from Matanzas area to Bahia Honda, and the grain flow was considered to have occurred simultaneously over a wide area along the northwest slope of the Cretaceous Cuban Arc. On the other hand, the basal breccia of the Cacarajicara Formation in the Cuban Southwestern terranes has a distinctly different grain composition from that of the Basal Member of the Peñalver Formation. This is the same for the basal part of the Amaro Formation in the Bahamas Platform, which contains the rudist taxa different from that of the Basal Member of the Peñalver Formation (Rojas et al., 1995). Therefore, distribution of the grain flow deposit of the Peñalver Formation is restricted to the basin on the north side of the Cretaceous Cuban Arc.

Excluding thin conglomerate beds, calcarenite of the Lower to Upper Member of the Peñalver Formation shows a monotonous upward fining. The calcarenite also shows upward improvement of sorting, coarse-tail grading, and abundant water escape structures suggestive of rapid sedimentation in its lower part. No trace fossils are found throughout these members. All these features of the calcarenite suggest it was formed by rapid settling of grains from a high density suspension. Calcilutite of the Uppermost Member can be regarded as representing continuous upward fining from the calcarenite of the Upper Member. Most of planktonic foraminifera and calcareous nannofossils in the Middle to Uppermost Member of the Peñalver Formation suggest ages ranging from Campanian to late Maastrichtian that are older than the stratigraphically constrained age of the Peñalver Formation. On the other hand, these ages are in good agreement with the depositional age of the underlying Via Blanca Formation. Thus, *Nannoconus* sp. and other pre K/T fossils in the Middle to Uppermost Member of the Peñalver Formation can be either reworked from the underlying Via Blanca Formation (which also contains reworked fossils from older strata) or from both the Via Blanca Formation and older units. The dominance of agglutinated benthic foraminifera in the Lower Member, and its upward decrease as well as upward increase of planktonic foraminifera in the Upper and Uppermost Members are considered to be caused by size-sorting effect. The textural and compositional analyses of serpentine grains in the Lower to Upper Member suggest the presence of ultramafic rocks of probable Northern Ophiolites origin in the source area. These materials are not found in the Basal Member and the underlying Via Blanca Formation. On the other hand, bioclast from a shallow-marine platform which are abundant in the Basal Member decrease upward in the Lower Member and are nearly absent above the Middle Member. These facts suggest difference in source area between the Basal and the Lower to Upper Members.

Concentration of the high density suspension is estimated as approximately 330–100 g/l if we assume that the sediments were completely dispersed throughout the water column 600–2000 m thick based on the estimated water depth of the underlying Via Blanca Formation (Brönnimann and Rigassi, 1963). Settling time can be estimated by using settling velocities calculated by Impact and Stokes laws (e.g. Gibbs et al., 1971). Assuming that the maximum grain size at the top of the Lower Member indicates the size of grains completely settled out from the dispersed suspension at the time of deposition of that horizon, the time required for deposition of the Lower Member is estimated as approximately 2–8 h after generation of the dispersed suspension. In similar manner, deposition of the Middle, Upper, and Uppermost Members requires approximately 4–13 h, 3–12 days, and at least 11–38 months, respectively, after generation of the suspension.

Then, what is a cause of the high density suspension? One possibility is a turbidity current followed by the initial grain flow. Grain flows are usually accompanied by high density turbidity currents (e.g. Lowe, 1982). However, it is difficult to explain the massive calcarenite and calcilutite of the Lower to Uppermost Member as a turbidite, because: (1) grain composition of calcarenite of the Lower to Upper Member is different from that of the Basal Member; (2) structures such as current ripple and parallel lamination characteristic of turbidites are not observed; and (3) calcarenite of the Middle Member of the Cacarajicara Formation, which belongs to a different terrane (Yucatan block), is also similar in lithology and composition suggesting that the suspension extends to the base of the Yucatan margin, north of the depositional site of the Peñalver Formation.

Cita et al. (1996) described a structureless and monotonic upward fining sedimentary layer named “homogenite” from 2000 m and more deep sea floor of the eastern Mediterranean, which they believe to be a convincing deep-sea tsunami deposit. The homogenite in abyssal basins is formed by gravitational settling from a single suspended sediment cloud that was reworked from basin walls. The cause of suspension is considered to be a large tsunami wave on the basis of: (1) coincidence in timing with the collapse of the caldera of the Volcano Santorini at 3500 yr B.P.; (2) lack of sedimentary structure; (3) simultaneous deposition in many basins; (4) reworking of sediments from adjacent basin walls; and (5) the estimated concentration (16 g/l) of the suspension much higher than that of usual current generated suspensions (Kastens and Cita, 1981). The calcarenite and calcilutite of the Peñalver Formation have characteristics such as: (1) massive lithology with no sedimentary structures suggestive of a turbidite; (2) rapid accumulation from a high density suspension; (3) the fossil assemblage suggestive of reworking from the underlying formations; and (4) wide distribution of the suspension extended to the north of the Cretaceous Cuban Arc. These characters are similar to those of a homogenite, suggesting that the Lower to Uppermost Member of the Peñalver Formation is a deep sea tsunami deposit.

Furthermore, thin conglomerate beds of the Lower Member are mainly composed of well-sorted and well-rounded pebble size mud clasts and rudist fragments whose major axes are arranged in an east–west direction parallel to the bedding. These features are similar to those of wave induced deposits, tempestites, rather than deposits induced by gravity flow such as turbidites (e.g. Einsele and Seilacher, 1991). Because a tempestite is formed by the stirring up of particles by wave action, grains are sorted to a certain size to form a sheet-like bed (Einsele and Seilacher, 1991). Assuming the Peñalver Formation was deposited more than 600 m water depth, the wave induced current which arraged pebble- to boulder-size clasts can be only explained by a tsunami. Thus, it is possible that thin conglomerate beds in the Lower Member of the Peñalver Formation may have been induced by a series of large tsunami waves followed by down-slope transport of sorted grains, such as the case for “tsunamiites”, tsunami-induced current deposits, advocated by Shiki and Yamazaki (1996). A small amount of rudist fragments in these beds may reflect back wash transport from a shallow marine platform because a tsunami wave is considered to have caused catastrophic erosion in coastal area, as is the case of study by Young and Bryant (1992).

In summary, the Peñalver Formation is considered to have been formed by a large grain flow from the platform on the Cretaceous Cuban Arc that led to the deposition of the Basal Member. The following high density suspension, which was possibly produced by tsunami waves, lead to deposition of the Lower to the Uppermost Member. Intermittent lateral supply of pebbles to boulders induced by tsunami waves was repeated more than 14 times during the early stage of settling from the dispersed suspension. After cessation of lateral supply, the settling from high density suspension continued, which formed the Middle to Uppermost Member.

7.2. Relation with the K/T impact

The depositional age of the Peñalver Formation is biostratigraphically constrained between 65.4 and 65.0 Ma. *Nannoconus* was also found in the K/T boundary “cocktail” interpreted by Bralower et al. (1998). In addition, altered vesicular glass of probable ejecta origin, and shocked quartz are found from the Basal to Lower Member and the Lower to Upper Member, respectively. These evidences strongly support that the Peñalver Formation was formed in association with the K/T impact. Although high concentration of Ir, Cr and Ni is not recognized in this study, this can be explained by either a too coarse sampling interval or the explanation that the Ir layer is present at the horizon that is not exposed in the studied area. Because the Ir anomaly usually occurs at the very top of the K/T boundary sequence around the Gulf of Mexico, and a modal grain diameter of the Ir anomaly layer is about the size of 2 μm in the Brazos River (Smit et al., 1996), it is most likely that the Ir layer is present at the boundary between the Uppermost Member and the Apolo Formation that was not observed in this study. There also is some possibility that extraterrestrial materials were diluted by a great volume of dispersed deep sea sediment.

Altered vesicular glass grains of 1–2 mm in diameter in the Basal Member and the lower part of the Lower Member, with their higher concentration toward the basal part, and the presence of a probable altered microtektite suggest deposition of the vesicular glass and tektite on the sea floor before arrival of the grain flow. On the other hand, the fact that shocked quartz grains are found in the Lower to Upper Member but absent in the Basal Member suggests that the initial grain flow arrived before deposition of the shocked quartz grains and a high density suspension was generated when the shocked quartz grains were still within the water column. The difference in arrival time to the sea floor between the vesicular glass and shocked quartz at the depositional site of the Peñalver Formation can be explained by the difference in their transportation process within the atmosphere and/or the water column. Their differentiation in the atmosphere is inferred from the facts that shocked quartz grains are abundant just above the K/T spherule-bearing layer deposited in a fluvial environment in the western interior of North America (Izett,

1990) and that shocked quartz is not found in spheroid bed and vesicular glass-bearing diamictite in Belize (Ocampo et al., 1996). These examples are consistent with the observed distributions of the impact-induced materials in the Peñalver Formation and suggest that a time lag existed between arrival of the initial grain flow and formation of the high density suspension. Together with the difference in grain composition between the Basal Member and the Lower to Upper Member, the presence of time lag further suggest that the cause of grain flow and the high density suspension should be different.

8. Conclusion

(1) The Peñalver Formation consists of normally graded calcareous clastic rocks up to 180 m in thickness around the type locality. The formation is subdivided into the Basal, Lower, Middle, Upper and Uppermost Members based on the lithology.
(2) The Basal Member consists of massive, poorly sorted, calcirudite with grain-supported fabric that contains abundant large fossils from the shallow-marine platform and occasional large intraclasts, suggesting that this member was formed by a grain flow from the carbonate platform on the Cretaceous Cuban Arc.
(3) The Lower to Uppermost Members consist of massive calcarenite and calcilutite that show upward fining. Composition of calcarenite is different from that of the Basal Member. The homogeneous appearance, coarse-tail normal grading, abundant water escape structures, and abundant reworked fossils suggest settling from high density suspension possibly caused by tsunami.
(4) Thin conglomerate beds in the Lower Member that contain well-sorted, well-rounded mud clasts and shallow marine fossils are considered to have been supplied laterally during the early stage of settling from high density suspension.
(5) The Peñalver Formation was formed in association with the K/T boundary impact event based on biostratigraphically constrained age and presence of altered vesicular glass and shocked quartz of probable impact origin.
(6) Distribution of altered vesicular glass in the Basal and lower Member and shocked quartz in the Lower to Upper Member suggest that a time lag existed between arrival of the initial grain flow and formation of high density suspension and these causes should be different.

Acknowledgements

This paper is a part of the doctoral thesis of H. Takayama under the supervision of R. Tada, and based on the joint field survey that was conducted by Japanese and Cuban scientists during 1997 and 1998. We express our special thanks to F. Fajardo Jr., (the former President of Agencia Medio Ambiente, Cuba), L. H. Arango (the former Director of Foreign Relations of Agencia de Medio Ambiente), and other Cuban government officers for their support throughout the course of this study. We also express our sincere thanks to Drs H. Gonzales, A. E. Garcia, M. Campos. R. Rojas, and other Cuban scientists for their cooperation during the field survey. We pay regard to A. Nakata (Assistant Manager at Havana, Mitsui & Co., Ltd.). The survey was supported by research funds donated to Univesity of Tokyo by NEC Corp., I. Ohkawa and M. Iizuka. We are also indebted to Mitsui & Co., Ltd. and Japan Air Line for their financial and technical support. We also acknowledge the valuable advice of Dr A. Taira of the Ocean Research Institute, University of Tokyo. We wish to thank Dr T. Shiki for his encouragement in publishing the manuscript. Thanks also go to Mr S. Yamamoto for assistance in the field work. Technical support was provided by H. Yoshida and T. Sakai at Geological Institute, University of Tokyo.

References

Albear, J.F., Iturralde-Vinent, M.A., Diaz, C., Flores, E., Sanchez-Arango, J., 1985. Contribucion a la geologia de las provincias de La Habana y Ciudad de la Habana. Editorial Cientifico-Tecnica, La Habana, 155pp.

Albertão, G.A., Martins Jr., P.P., 1996. A possible tsunami deposit at the Cretaceous–Tertiary boundary in Pernambuco, northeastern Brazil. Sediment. Geol. 104, 189–201.

Allen, J.R.L., 1982. Sedimentary Structures, Their character and Physical Basis, vol. II, Elsevier, Amsterdam (643pp).

Alvarez, L.W., Alvarez, W., Asaro, F., Michel, H.V., 1980. Extraterrestrial cause for the Cretaceous–Tertiary extinction. Science 208, 1095–1108.

Anders, E., Grevesse, N., 1985. Abundances of the elements: meteoritic and solar. Geochim. Cosmochim. Acta 53, 197–214.

Arai, S., 1990. What kind of magmas could be equilibrated with ophiolitic peridotites? In: Malpas J. et al. (Eds.), Ophiolites, Oceanic Crustal Analogues. Proc. Symp. “ROODOS 1987”, Geol. Surv. Dept., Min. Agr. Nat. Res., Nicosia, Cyprus, pp. 557–565.

Bence, A.E., Albee, A.L., 1968. Empirical correlation factors for the electron microanalysis of silicates and oxides. J. Geol. 76, 382–403.

Berggren, W.A., Kent, D.V., Swisher III, C.C., Aubry, M.-P., 1995. A revised Cenozoic geochronology and chronostratigraphy. In: Berggen, W.A., Kent, D.V. and Hardenbol, J. (Eds.), Geochronology, Time Scales and Global Stratigraphic Correlations: A Unified Temporal Framework for an Historical Geology. Spec. Publ.-Soc. Econ. Paleont. Min., vol. 54, pp. 129–212.

Bohor, B.F., 1996. A sediment gravity flow hypothesis for siliciclastic units at the K/T boundary, northeastern Mexico. In: Ryder, G., Fastovsky, D., and Gartner, S. (Eds.), Cretaceous–Tertiary Event and Other Catastrophes in Earth History. Geol. Soc. Am. Spec. Pap., vol. 307, pp. 183–195.

Bohor, B.F., Seitz, R., 1990. Cuban K/T catastrophe. Nature 344, 593.

Bourgeois, J., Hansen, T.A., Wiberg, P.L., Kauffman, E.G., 1988. A tsunami deposit at the Cretaceous–Tertiary boundary in Texas. Science 241, 567–570.

Bralower, T.J., Ituralde-Vinent, M.A., 1997. Micropaleontological dating of the collision between the North American plate and the Greater Antilles Arc in western Cuba. Palaios 12, 133–150.

Bralower, T.J., Leckie, R.M., Sliter, W.M., Thierstein, H.R., 1995. An integrated Cretaceous microfossil biostratigraphy. In: Berggren, W.A., Kent, D.V. and Hardenbol, J. (Eds.), Geochronology, Time Scales and Global Stratigraphic Correlations: A Unified Temporal Framework for an Historical Geology. Spec. Publ.-Soc. Econ. Paleont. Min., vol. 54, pp. 65–79.

Bralower, T.J., Paull, C.K., Leckie, R.M., 1998. The Cretaceous–Tertiary boundary cocktail: Chicxulub impact triggers margin collapse and extensive sedimentary gravity flows. Geology 26, 331–334.

Brönnimann, P., Rigassi, D., 1963. Contribution to geology and paleontology of the area of the city of La Habana, Cuba, and its Surroundings. Eclogae Geol. Helv. 56, 193–480.

Cita, M.B., Camerlenghi, A., Rimoldi, B., 1996. Deep-sea tsunami deposits in the eastern Mediterranean: new evidence and depositional models. Sediment. Geol. 104, 155–173.

Doehne, E., Margolis, S.V., 1990. Trace-element geochemistry and mineralogy of the Cretaceous/Tertiary boundary; Identification of extraterrestrial components. In: Sharpton, V.L. and Ward, P.D. (Eds.), Global Catastrophes in Earth History, Geol. Soc. Am. Spec. Pap., vol. 247, pp. 367–382.

Einsele, G., Seilacher, A., 1991. Distinction of tempestites and turbidites. In: Einsele, G., Ricken, W., Seilacher, A. (Eds.). Cycles and Events in Stratigraphy, Springer, Berlin, pp. 377–382.

Erba, E., Premoli Silva, I., Watkins, D.K., 1995. Cretaceous calcareous plankton biostratigraphy of Sites 872 through 879. In: Haggerty, J.A., Premoli Sliva, I., Rack, F., and Mcnutt, M.K. (Eds.), Proc. ODP, Sci. Results, 144: College Station, TX (Ocean Drilling Program), pp. 157–169.

Gibbs, R.J., Matthews, M.D., Link, D.A., 1971. The relationship between sphere size and settling velocity. J. Sediment. Petrol. 41, 7–18.

Gutiérrez, G.O. et al. (Eds.), 1989. Nuevo Atlas Nacional De Cuba. Instituto de Geografía de la Academía de Ciencias de Cuba, Havana.

Hildebrand, A.R., Penfield, G.T., Kring, D.A., Pilkington, M., Camargo, Z.A., Jacobsen, S.B., Boynton, W.V., 1991. Chicxulub crater: a possible Cretaceous/Tertiary boundary impact crater on the Yucatan Peninsula, Mexico. Geology 19, 867–871.

Hutson, F., Mann, P., Renne, P., 1998. $^{40}Ar/^{39}Ar$ dating of single muscovite grains in Jurassic siliciclastic rocks (San Cayetano Formation): constraints on the paleoposition of western Cuba. Geology 26, 83–86.

Iturralde-Vinent, M.A., 1992. A short note on the Cuban late Maastrichtian megaturbidite (an impact-derived deposit?). Earth Planet. Sci. Lett. 109, 225–228.

Iturralde-Vinent, M.A., 1994. Cuban geology: a new plate-tectonic synthesis. J. Petrol. Geol. 17, 39–70.

Izett, G.A., 1990. The Cretaceous/Tertiary boundary interval, Raton Basin, Colorado and New Mexico, and its content of shocked-metamorphosed minerals; evidence relevant to the K/T boundary impact-extinction theory. Geol. Soc. Am. Spec. Pap., vol. 249, 100pp.

Kastens, K.A., Cita, M.B., 1981. Tsunami-induced sediment transport in the abyssal Mediterranean Sea. Geol. Soc. Am. Bull. 92, 845–857.

Koeberl, C., Anderson R.R., 1996. Monson and Company: Impact structures in the United States. In: Koeberl, C., Anderson R.R. (Eds.), The Manson Impact Structure, Iowa: Anatomy of an Impact Crater. Geol. Soc. Am. Spec. Pap., vol. 302, pp. 1–29.

Kyte, F.T., Zhou, L., Lowe, D.R., 1992. Noble metal abundances in an Early Archean impact deposit. Geochim. Cosmochim. Acta 56, 1365–1372.

Leroux, H., Rocchia, R., Froget, L., Orue-Etxebarria, X., Doukhan, J.-C., Robin, E., 1995. The K/T boundary at Beloc (Haiti): compared stratigraphic distribution of the boundary markers. Earth Planet. Sci. Lett. 131, 255–268.

Lowe, D.R., 1975. Water escape structures in coarse-grained sediments. Sedimentology 22, 157–204.

Lowe, D.R., 1982. Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents. J. Sediment. Petrol. 52, 279–297.

Middleton, G.V., 1966. Experiments on density and turbidity currents. Can. J. Earth Sci. 4, 475–504.

Nakamura, Y., Kushiro, I., 1970. Compositional relations of coexisting orthopyroxene, pigeonite and augite in a tholeiitic andesite from Hakone volcano. Contribution to Mineral. Petrol. 26, 265–275.

Ocampo, A.C., Pope, K.O., Fischer, A.G., 1996. Ejecta blanket deposits of the Chicxulub crater from Albion Island, Belize. In: Ryder, G., Fastovsky, D., Gartner, S. (Eds.), Cretaceous–Tertiary Event and Other Catastrophes in Earth History, Geol. Soc. Am. Spec. Pap., vol. 307, pp. 75–88.

Palmer, R.H., 1945. Outline of the Geology of Cuba. J. Geol. 53, 1–34.

Perch-Nielsen, K., 1985. Mesozoic calcareous nannofossils. In:

Bolli, H.M., Saunders, J.B., Perch-Nielsen, K. (Eds.). Plankton Stratigraphy, Cambridge University Press, Cambridge, pp. 329–426.

Pindel, J.L., Barrett, S., 1990. Geological evolution of the Caribbean region; a plate tectonic perspective. In: Dengo, G., Case, J. (Eds.), The Caribbean region, The Geology of North America, Volume H: Geological Society of America, pp. 405–432.

Pszczolkowski, A., 1986. Megacapas del maestrichtiano en Cuba occidental y central. Bull. Pol. Acad. Sci., Earth Sci. 34, 81–93.

Rojas, R., Iturralde-Vinent, M.A., Skelton, P.W., 1995. Stratigraphy, composition and age of Cuban rudist-bearing deposits. Revista Mexicana de Ciencias Geológicas 12, 272–291.

Rosencrantz, E., 1990. Structure and tectonics of the Yucatan basin, Caribbean Sea, as determined from seismic reflection studies. Tectonics 9, 1037–1059.

Rosencrantz, E., 1994. Basement structure and tectonics in the Yucatan basin. In: Iturralde-Vinent, M.A. (Ed.), Cuban Ophiolites and Volcanic Arcs, IGCP Project 364 Special Contribution No. 1, Maiami, pp. 36–47.

Ross, M.I., Scotese, C.R., 1988. A hierarchical tectonic model of the Gulf of Mexico and Caribbean region. Tectonophysics 155, 139–168.

Shiki, T., Yamazaki, T., 1996. Tsunami-induced conglomerates in Miocene upper bathyal deposits, Chita Peninsula, central Japan. Sediment. Geol. 104, 175–188.

Shipboard, P., Shipboard Scientific Party, 1997. Explanatory notes. In: Sigurdsson, H., Leckie, P.M., Acton, G.D., et al. Proc. ODP, Init. Repts., 165 College Station, TX (Ocean Drilling Program), pp. 15–46.

Sliter, W.V., 1989. Biostratigraphic zonation for Cretaceous planktonic foraminifers examined in thin section. J. Foraminiferal Res. 19, 1–19.

Smit, J., Montanari, A., Swinburne, N.H.M., Alvarez, W., Hildebrand, A.R., Margolis, S.V., Claeys, P., Lowrie, W., Asaro, F., 1992. Tektite-bearing, deep-water clastic unit at the Cretaceous–Tertiary boundary in northeastern Mexico. Geology 20, 99–103.

Smit, J., Roep, Th. B., Alvarez, W., Claeys, P., Grajales-Nishimura, J.M., Bermudez, J., 1996. Coarse-grained, clastic sandstone complex at the K/T boundary around the Gulf of Mexico: deposition by tsunami waves induced by the Chicxulub impact? In: Ryder, G., Fastovsky, D., Gartner, S. (Eds.), Cretaceous–Tertiary Event and Other Catastrophes in Earth History. Geol. Soc. Am. Spec. Pap., vol. 307, pp. 151–182.

Sridhar, K., Jackson, M.L., Clayton, R.N., 1975. Quartz oxygen isotopic stability in relation to isolation from sediments and diversity of source. Soil Sci. Soc. Am. Proc. 39, 1209–1213.

Stauffer, P.H., 1967. Grain-flow deposits and their implications, Santa Ynez Mountains, California. J. Sediment. Petrol. 37, 487–508.

Taylor, A.T., McLennan, S.M., 1985. The Continental Crust: its Composition and Evolution, Blackwell, Oxford (312pp).

Yoshida H., Takahashi, N., 1997. Chemical behavior of major and trace elements in the Horoman mantle diapir, Hidaka belt, Hokkaido, Japan. J. Japan. Assoc. Min. Petr. Econ. Geol., vol. 92, pp. 391–409. (In Japanese with English abst.)

Young, R.W., Bryant, E.A., 1992. Catastrophic wave erosion on the southeastern coast of Australia: impact of the Lanai tsunami ca. 105 ka?. Geology 20, 199–202.

Wänke, H., Dreibus, G., Jagouts, E., 1984. Mantle chemistry and accretion history of the Earth. In: Korner, A. (Ed.). Archean Geochemistry, Springer, Berlin, pp. 1–24.