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THE RELIEF GENERATIONS ON THE ISLAND OF SKOPELOS (SPORADES)

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ABSTRACT

The presented paper is based on the mapping of Skopelos on a scale of 1: 50,000. Evidence is provided for the existence of sealed karst generations, which were formed during the Lower Cretaceous and at the turn from the Upper Cretaceous to the Palaeogene. Furthermore, the author shows that the Miocene peneplain benchland found on the island consists of various denudation systems that are of different age and that were influenced by climato-morphological processes. These denudation systems are considerably older than the piedmont benchlands of the Attic-Cycladic complex, because the geotectonic consolidation of the Pelagonian zone was finished at an earlier time. Concerning the more recent karstification of the island, the sequence of landforms consisting of pediments-glacis-karst blind valleys and poljes proves to be quite relevant. In contrast to the Southern Greek Cyclades, periglacial gelisolifluction played a significant role at all altitudes of the Magnesian Islands.

KEY WORDS: Reliefgenerationen; Piedmonttreppen; Rumpfflächen; Pedimente; Glacis; Karstblindtäler; Poljen; Paläoböden; Klimamorphologie.
relief generation; piedmont benchland; peneplain; pediment; glacis; karst blind valley; polje; palaeosols; climato-morphology.

1. INTRODUCTION

The Magnesian Islands are marked by peneplains that cut the internal structures of the Central Hellenidic Pelagonian nappes in a discordant way. The planation surfaces are developed at different altitudes. Together with the palaeosols found on them, they indicate a formation during the Miocene continental stage of the Aegean Sea. Therefore, we may also state a similarity to the piedmont benchlands of continental Greece, which were uplifted in several stages (cf. L. Sotiriadis, 1981, E. Vavliakis, 1981, A. Psilovikos, 1986). It is essential to point out, however, that the Aegean peneplain benchlands do not represent the final stage of a long-lasting development. In the course of the alpide geotectonics they were shaped within a relatively short period of time (H. Riedl, 1984) by processes of sheet erosion which were induced by a semi-humid to tropical climate. After the break-up of the Aegean continent, a border line formed south of the Magnesian Islands. Periglacial gelisolifluxion prevailed in the north and convergent periglacial processes in the south (H. Riedl, 1984).

2. GEOLOGIC KEY STRUCTURES

The central part of the island (D. Matarangas, 1992) is occupied by the **Pelagonian unit** consisting of the uppermost part of the metaclastic Skiathos series (Ladinian-Karnian) and dolomites (Norian-Rhaetian) that were formed at an early stage of diagenesis. These rocks are covered by the **mesoautochthon** which is composed of basal Fe-/Ni laterites and bauxites of Lower Cretaceous age as well as Albian transgressional conglomerates. They are succeeded by Upper Cretaceous rudist limestones and older Tertiary flysch. The

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above-mentioned mesoautochthon (E. Mposkos and A. Liati, 1991) was subject to low-grade metamorphism at 4 kbars and at a temperature of 300-330°C. The northwest of the island is dominated by the **Glossa unit** consisting of cipolino marbles, metabasalts, phyllites and greenschists derived from tholeiites with MORB type. The Glossa series (D. Matarangas, 1992) is part of the Eohellenic nappe, which was thrust over the Skiathos series and the Pelagonian dolomites between the Kimmeridge and Cenomanian (V. Jacobshagen and E. Wallbrecher, 1985). In the eastern part of the island remnants of the Eohellenic nappe and the mesoautochthon were thrust over by the Palouki unit. It consists of a low-grade, Lower Cretaceous metamorphic sequence of bedding (D. Matarangas, 1992), containing hemipelagic carbonate rocks with clastic alternate structure. They are succeeded by Upper Cretaceous rudist limestones and Lower Tertiary flysch. The Palouki unit was overthrust in Mesohellenic times (Eocene).

3. ASSOCIATIONS OF FORMS ON THE DENUDATION SURFACES

3.1. Bauxite karst and laterite karst

On Skopelos Island, the oldest generation of landforms is constituted by a palaeokarst characterised by karst troughs which are sealed by bauxites and laterites (A. Panagos and A. Liati, 1996). This palaeokarst could only develop after the Eohellenic upthrust and before the Upper Cretaceous transgression, so we may assume a Lower Cretaceous age for these features. This palaeokarst was subjected to deformation in Mesohellenic and was transported in Neohellenic times. It comprises round karren and "hohlkarren" (Blo) and 10 m broad karst troughs (Ag. Rhiginos) developed on Triassic dolomites and filled with lateritic, violet to greenish or kaolinic decomposed matadero (W. Kubiena, 1955). The metabasalts, serpentinites and schists of Eohellenic nappe remnants served as parent rocks for the development of the laterites and bauxites. Other noteworthy features are the crumpling walls of the karren and karst troughs as well as the subcutaneous sawtooth-like disintegration of the Pelagonian dolomites. In some parts we observe already collapsed hump-like karren completely mantled by decomposed matadero. Upper Cretaceous conglomerates cut all these forms of palaeokarst in a discordant way. The bauxite and laterite karst of Skopelos displays a great similarity to the sealed fossil karst at the Amvlema Pass in Middle Greece (H. Riedl, 1984), but also to the pre-Gosauian bauxite karst of the "Thermenalpen" (Eastern Alps; H. Riedl, 1973).

3.2. Preflysch karst

The subsequent generation of landforms appears in the Upper Cretaceous rudist limestones, mostly buried under Palaeogenic flysch and must have been formed at the turn from the Upper Cretaceous to the Palaeogene. In the area of the Palouki Massif (Ag. Triada) and in the Karya we observe karst cones and cockpit dolines sealed by flysch and only partly exhumed. Evidence of these karst landforms has been provided for the Gavrovo-Tripolitza zone on the Peloponnese by D. Richter and I. Mariolakos (1973, 1978). Similar results have been obtained on Kythira Island by D. Theodoropoulos (1973) and in upper Arcadia by H. Riedl (1978). These kinds of landforms are also distributed in the Parnassus-Giona zone (D. Richter and I. Mariolakos, 1978). In contrast to these results, J. Dercourt and J. Fleury (1977) deny the existence of this generation of landforms on the grounds of "couches de passage" between limestones and flysch in the Western and Central Hellenides. Such an antithesis of "transitional series" (D. Richter, 1978) and the assumed impossibility of a simultaneous subaerial karst development can be easily eliminated by considering the following aspect: In the less uplifted areas of the mesoautochthon, the marine environment may have continued for a longer period of time and transitional series between rudist limestones and flysch may have developed. This was the case on Skopelos as well as on Alonnisos (D. Matarangas, 1992), where we partly note a conformable structure of the flysch with respect to the Upper Cretaceous. Without doubt, however, sedimentation on Skopelos was stopped due to the stronger uplift of the blocks out of the Upper Cretaceous lagoon-like environment. Subsequently, subaerile karstification could set in under exogenous tropical climatic conditions. It was only the following downthrust which started the marine sedimentation of flysch, burying and fossilising the subaerile palaeokarst. We may assume that at the turn from the Upper Cretaceous to the Palae[...] [...]arine environment and

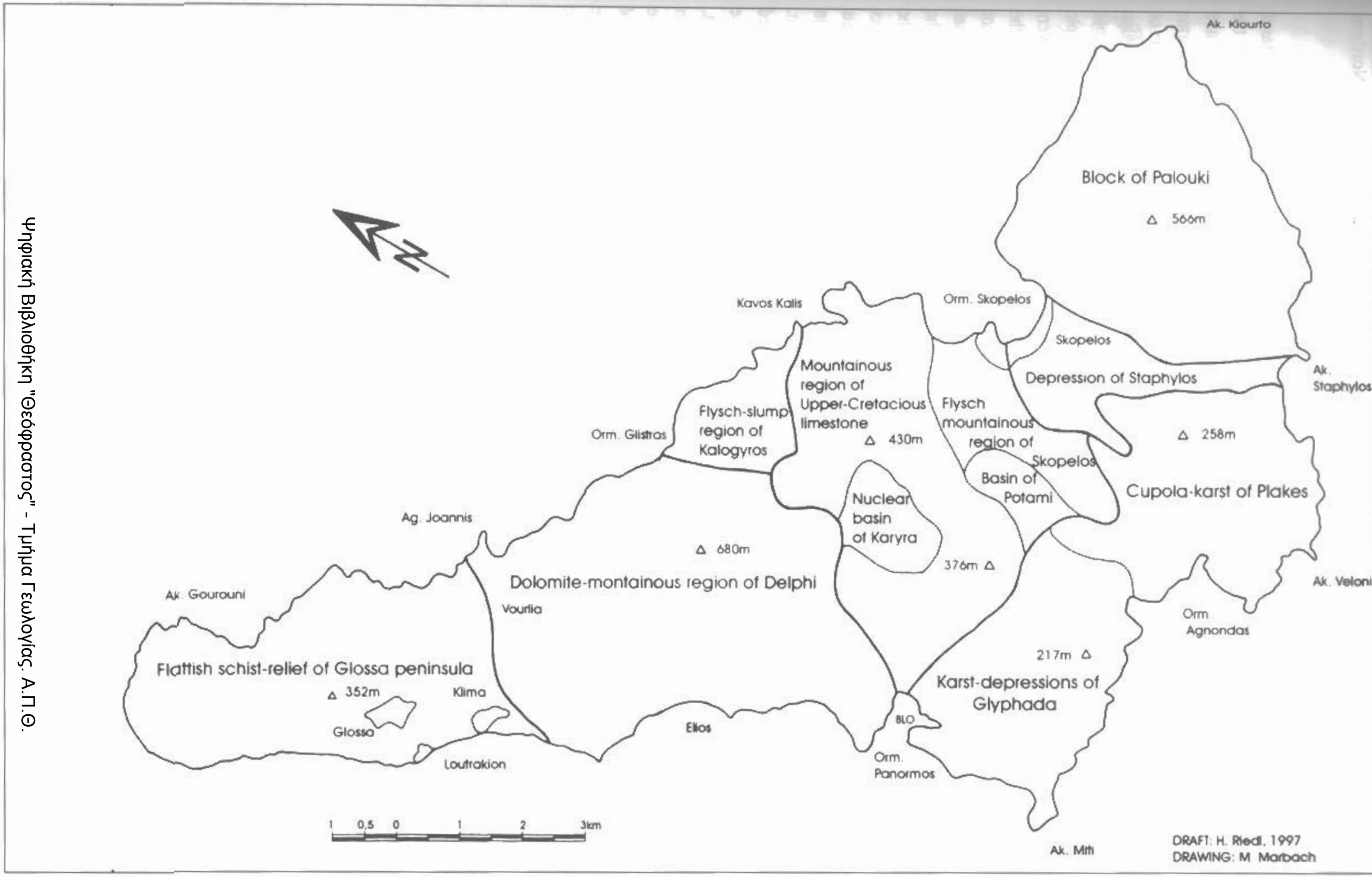

Figure 1: Geomorphological spaces of Skopelos

those shaped by subaerial denudation must have alternated on a fairly small scale on the Northern Sporades.

3.3. Lower Miocene peneplains

The oldest system of peneplains A, situated at an altitude of 550 m, dominates the Palouki Massif (Figures 1 and 2) and the southeastern Delphi ridge, formed of Pelagonian dolomites. In this part, the system rises above 600 m altitude. Lagging behind in uplift, it covers the karst blocks of Mt. Plakes and Mt. Glyphada (Figure 1) at an average height of 200 m. It may be described as an association of forms comprising solution plains, small domes, micropoljes, uvalas and bowl-shaped dolines. The thick covers of residual red soils represent another characteristic feature. These soils are loamy clays with a colour value of 10 R 4/6. The results of the heavy-mineral analysis[3] reveal that the substratum of the red soils cannot have derived from the Triassic dolomites, but only from formerly wide-spread rocks of the Eohellenic Glossa nappe (serpentinites, metabasalts, mica schists, phyllites). The erosion of the non-carbonate covers together with a simultaneous red soil weathering, the filling of the karst clefts with red soils and the development of karst plains on the Triassic basement constitute a morphodynamic unit. The high content of kaolinite of the palaeosols, measuring 15%, the content of haematite and geothite (10%) as well as that of diaspore (5%) indicate that only an ancient semi-humid to tropical climate was able to cause the formation of the denudation systems and the palaeoplastosols. With regard to the age determination of the peneplain system A, we may assume an initial subaerial land formation in the course of a Loiano effect starting with the Lower Burdigalian. This development took place during the Neohellenic tectogenesis when the Central Hellenides were thrust on the proximal parts of the Western Hellenic nappes. Already in the Aquitanian (B. Schröder, 1986), the non-marine area of sedimentation reached the Axios depression. In the Neogene basins of southern Euboea (H. Böger, 1993), the terrestrial, limnic-fluvial facies started even earlier in the Lower Burdigalian, subaerial denudation processes being evident. With regard to the climato-morphological aspect, the middle Burdigalian lignites found in the Axios-Thermaikos Basin (A. Psilovikos and G. Syrides, 1984) seem quite significant. Hence, the age of the oldest peneplain system on the Magnesian Islands differs from that of the oldest denudation systems of the Cycladic crystalline. In the first case we may assume an early Miocene age, in the latter the top surface systems have proved to be of latest Miocene age (H. Riedl, 1982) due to the young plutonism and the Upper Miocene thrust in of the Aegean nappe. The first uplift of level A must already have taken place at the climax of the Neohellenic tectogenesis during the Langhian.

3.4. Middle Miocene peneplains

Below system A we may distinguish the peneplain system B, occupying an average height of 450 m. It is located on Mt. Palouki, on the ridge of Mt. Vais and, forming a broad bench, it reaches the southeastern foot of Mt. Delphi. Furthermore, it appears as a flat surface broken up into several spurs on the northeastern slope of Mt. Delphi (Figures 1 and 2). It extends to the west slope and was lowered by 40 m due to neotectonic processes. It then turns to the Glossa Peninsula. Summing up, level B encircles nearly all sides of Mt. Palouki and Mt. Delphi. All in all, the morphological features of level B indicate a formation due to a second phase of uplift, which may be classified as an epeirogenic one involving growing and laterally expanding stages of uplift. The reconstruction of the flat surfaces shows that level B formed a wide flat valley between Mt. Delphi and Mt. Palouki, which was 7 km broad. It belonged to the type of wide open valleys with sides flaring out which were generated by sheetwash under semihumid-tropical conditions. This broad valley was situated in the centre of level A which had already been uplifted at that time. The distribution of level B as well as the development of extensive flat valley floors striking out into the air serve as clear evidence for the continental stage of landscape development. We must consider, however, that the present-day small islands had not yet emerged and that the orographic conditions were entirely different. Hence, level B displays the meagre remnants of a formerly spacious group of forms.

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[3] carried out by Em. O. Univ.-Prof. Dr. G. Frasl (Institut für Geologie und Paläontologie, Universität Salzburg)

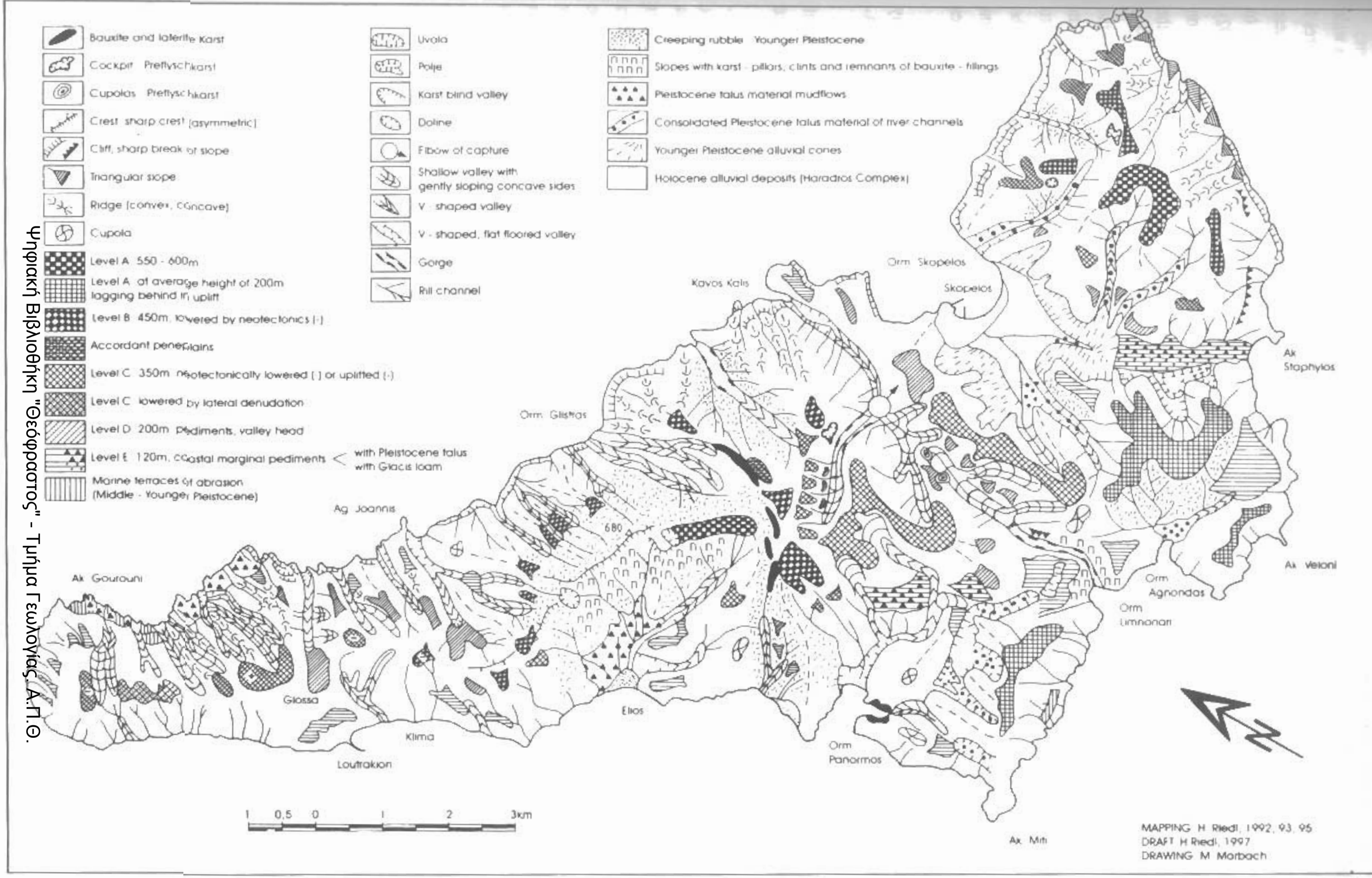

Figure 2: Geomorphological map of Skopelos

3.5. Upper Miocene peneplains

Level C is located at an average height of 350 m. Similar to both the older systems, it betrays a wide extension, transgressing all local watersheds. Level C appears dominant on the Glossa Peninsula (Figures 1 and 2) where it displays warpings with a vertical difference of 50 m. Level B was also affected by these neotectonic warpings. It is significant in both cases that the axes of the warpings strike from ENE to WSW. This fact reveals a good correspondence with D. Matarangas' lineation diagrams (1992). Accordingly, the ENE/WSW direction of the dislocation represents the youngest of all those mapped on Skopelos. This result implies a connection with the system of fractures in northern Anatolia. As the dislocations strike from ENE to WSW, they were caused by the dextral movements of the Aegean microplate (A. Peterek, 1992) in relation to the Eurasian plate. These dextral strike-slip dislocations led to the formation of the North Aegean system of troughs, which are entrenched 3 to 5 km deep. The system consists of several individual troughs, the Sporades Trough being the most important one with regard to Skopelos. The turn from Miocene to Pliocene marks a decisive period of tectonic fracturing in the development of this trough (J. Dewey and A. Sengör, 1979), even though the initial stage of its formation reaches back to the turn from Oligocene to Miocene involving the synchronous filling of terrestrial series (S. Lalechos and E. Savoyat, 1977).

It is a significant fact that neotectonic dislocations with ENE/WSW striking axes did not have any effects on levels situated below 350 m, but only on level C and higher ones. This implies that the levels A-C must be of pre-Pliocene age, whereas systems D-E turn out to be of Pliocene and post-Pliocene origin.

Level C is characterised by the formation of wide-spanned bays and passes as well as of ancient surfaces of intramontane basins. The bays are about 2 km wide which indicates the ancient predominance of deep weathering and sheet processes.

3.6. Pliocene pediments

Succeeding the uplift of level C, system D developed at an average altitude of 200 m. In comparison with the older system we note a remarkable hiatus concerning the morphological features. This level is characterised by pediments and basin-shaped valley heads. Distinct remnants of the pediments lead from Kalogyros and Nisi up to Mt. Raches, bordering the coastal plains of Skopelos. On level D, the whole slope of Mt. Palouki was modified by pedimentation, even reaching down to the depression of Staphylos. After the shaping of level D, we must assume a relatively fast uplift en bloc, as the triangular-shaped slopes on the east side of Mt. Palouki indicate. This uplift, however, did not happen continuously, but was interrupted in the following period when the coastal marginal pediments developed.

3.7. Pliocene/Pleistocene and Lower Pleistocene coastal marginal pediments

The coastal marginal pediments developed at an average altitude of 120 m may be identified as level E. The coastal marginal pediment is not only found on the smoother northern slope of the Glossa Peninsula, but it also constituted the initial surface for the present-day outline of the bays (Elios, Panormos, Limnonari). Only after the formation of the coastal marginal pediments, the tectonic break-up of the Aegean continent set in and the calas and rias we observe today were formed. Just like on the Cyclades (H. Riedl, 1982, 1983, 1984, 1986, 1995), this decisive stage of development with pedimentation extending across the present-day bays is linked to the oldest and older Pleistocene. The pediments are associated with a typical sequence of forms following a regular pattern. In the depression of Staphylos (Figures 1 and 2), the pediments, which are covered by debris mixed with loam, merge into glacis. At the counter discharge margin, formed by the slope of the karst block of Mt. Plakes, these glacis are connected with karst blind valleys. Only after the development of this synchronously modelled sequence of forms, the blind valleys were transformed into poljes and uvalas. Both types of karst depressions could only develop after the emergence of the island in post-Lower Pliocene times.

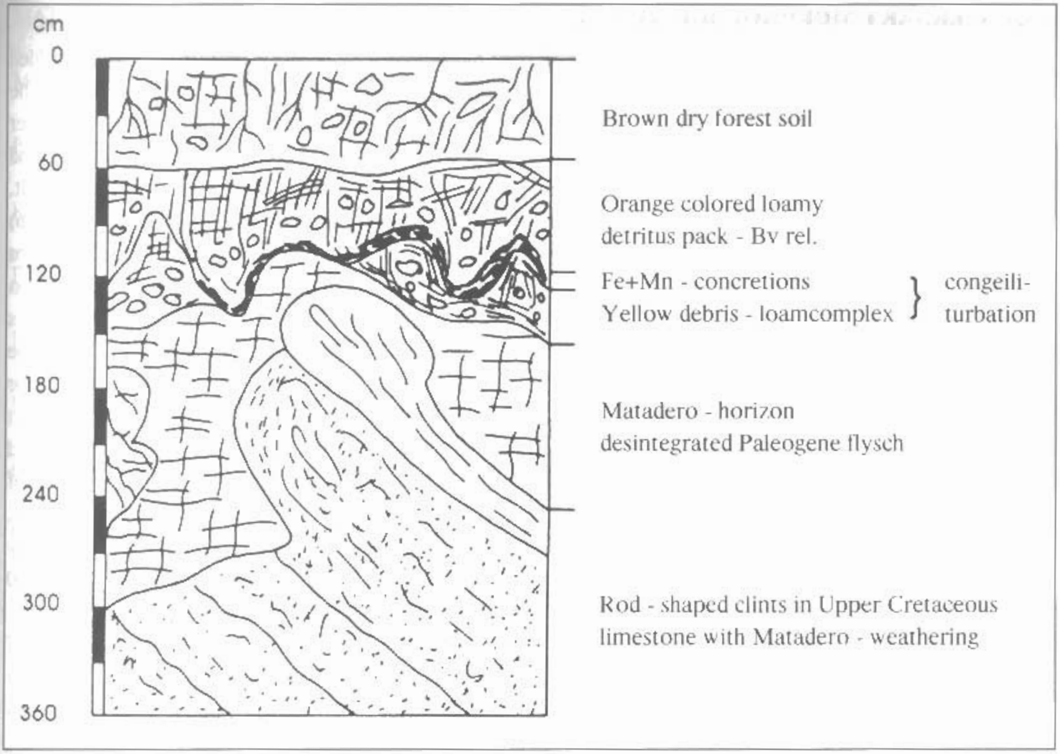

Figure 3: Soil sequence in Preflysch - Cretaceous limestone relief at the fringe of Uvala Avgheri - Delphi, 600 m.

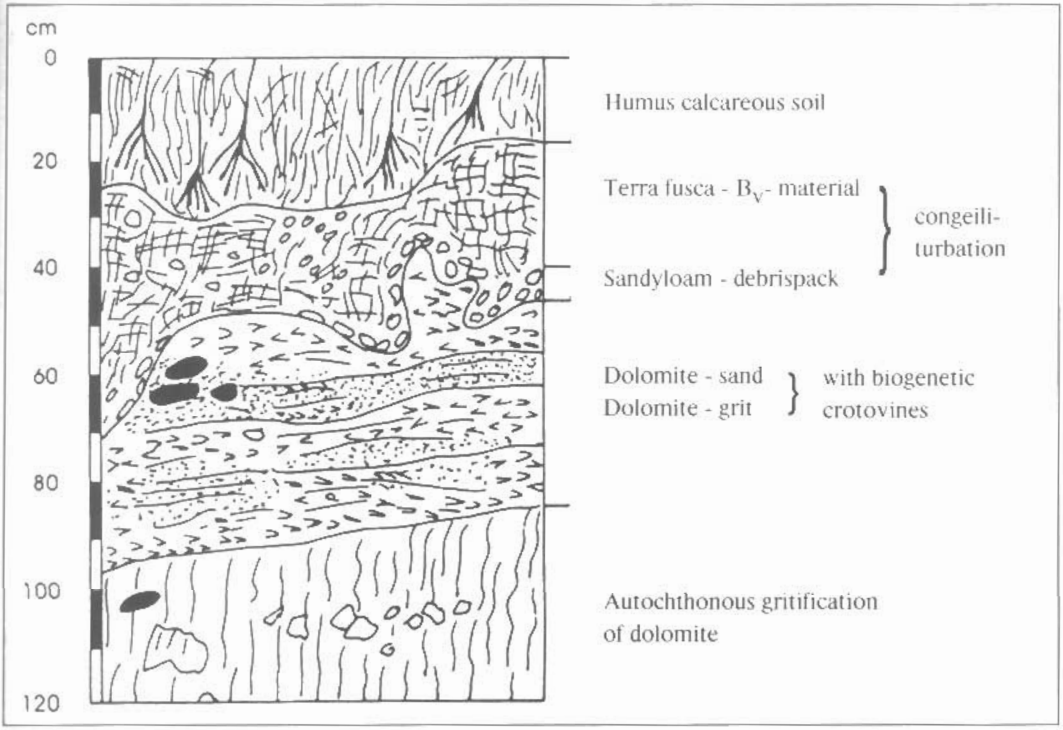

Figure 4: Alluvial cone - overprinted by congeliturbation. Bay of Agnondas.

4. QUATERNARY MORPHOLOGICAL FEATURES

The upper slopes of Mt. Delphi (Figure 3) are thickly blanketed with creeping debris covers embedded in a matrix of yellow terra fusca which fill up large dolines and uvalas in altitudes of 400 to 600 m. The cryoturbate structures and the general habit of the small-platy dolomite debris indicate younger Pleistocene periglacial-gelisolifluctional accumulations. During the glacial stages, the Miocene and Pliocene peneplains of the island were strongly moulded by periglacial processes. On the plateau of Mt. Sendoukia, for instance, we may observe that kind of "irregular karst pavement" which H. Poser (1976) has described on the Cretan Mt. Psiloriti in 2300 m altitude as a recent periglacial form. The Skopelos example, however, may only be interpreted as features resulting from glacial nivation and corrosion through melt water. If we assume that the lower border of Würmian solifluction lay 1300 m below today's solifluction belt, it would just have touched the area of the Sendoukia Plateau. At 500 m altitude, the Würmian average temperature in January ranged from -6°C to -7°C, which may be compared, for instance, with today's winter conditions at an altitude of 2000 m in the area of the East Thessalian mountain swell (H. Riedl, 1981). Cryoturbate structures (Figure 4), however, are also to be discovered below 500 m altitude extending as far down as to the sea level (for example on the coastal plain of Agnondas). In these cases we must consider that petrovariance certainly promoted this phenomenon. Nevertheless, subsequent to the Würmian solifluction zone there must have been a fringe of temporary frost activities. These conditions reveal a striking contrast to Southern Greece where the areas below 600 m were affected only by convergent periglacial phenomena (H. Riedl, 1984, 1991) without any gelifluction. Even before thick Upper Pleistocene alluvial accumulations were deposited in the V-shaped valleys of Skopelos, a deep erosion of the valleys had already taken place, which had presumably set in after the formation of the island between Günz glacial and Mindel/Riss interglacial. As these alluvial accumulations extend below today's sea level and have overridden the Würmian interstadial marine terraces (H. Riedl, 1982), this phase of deep erosion could have occurred in the Würmian or early Riss glacials. After the climax of the Würm glacial, the alluvial accumulations were strongly dissected all over Skopelos. Terraces covered with alluvial soils were encapsulated in the valleys. They are interlocked with the brown-coloured sediments of the coastal plains, which were deposited before the Hellenistic epoch and which correspond to the Haradros complex (R. Paepe, 1984).

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