Πρακτικά	6ου	Συνεδρίου	Μάϊος 1992
Δελτ. Ελλ. Γεωλ. Εταιρ.	Τομ. XXVIII/1	σελ. 165-177	Αθήνα 1993
Bull. Geol. Soc. Greece	Vol.	pag.	Athens

TECTONOMETAMORPHIC EVOLUTION OF THE GEOTECTONIC UNITS OF THE CHALKIDIKI PENINSULA

D.SAKELLARIOU

ΠΕΡΙΛΗΨΗ

Εξετάζεται η τεκτονομεταμορφική εξέλιξη των γεωτεκτονικών ενοτήτων της Χαλκιδικής και γίνεται προσπάθεια να συσχετιστούν τα διάφορα τεκτονομεταμορφικά επεισόδια, που αναγνωρίστηκαν, με συγκεκριμένα ορογενετικά γεγονότα. Η δημιουργία του τεκτονομεταμορφικού ιστού του γρανοδιορίτη της Σιθωνίας τοποθετείται στο Ηώκαινο. Η πρώτη παραμόρφωση των μολασσικών ιζημάτων του Τιθωνίου έγινε κατά την Παλαιοαλπική φάση στο Κατ. Κρητιδικό, ενώ η πρώτη παραμόρφωση και μεταμόρφωση της Περιροδοπικής ενότητας έγινε στο Ανώτ. Ιουρασικό. Η τεκτονομεταμορφική εξέλιξη της ενότητας Νέας Μάδυτου αρχίζει πιθανότατα στο Αν. Παλαιοζωικό ή στο Κατ. Μεσοζωικό (Κιμμέρια ορογένεση ;). Η εξέλιξη των ενοτήτων Βερτίσκου και Κερδυλίων κατά το Παλαιοζωικό παραμένει ως επί το πλείστον άγνωστη.

I. INTRODUCTION

During the last decades many authors tried to analyze and solve the geological structure of the Chalkidiki peninsula and gave important information about it. The most acceptable view about the geology of Chalkidiki today is the one expressed by KOCKEL et al (1977), which is the latest and most complete geological study of the area. We also relied on this paper to proceed to our present study.

The rocks occurring in the Chalkidiki peninsula belong to the following geotectonic units:

- The Stratoni granodiorite
- The Sithonia granodiorite
- The Tithonian molasse
- The Circum Rhodopian Belt
- The Arnea granite
- The Nea Madytos unit, as defined by SAKELLARIOU (1989) and SAKELLARIOU & DURR (this volume).
- The Vertiskos unit
- The Kerdilion unit

All of the above units will be examined from the scope of tectonometamor-

* Department of Geology, University of Athens, Panepistimioupoli, Zografou, 15784 Athens.

Ψηφιακή Βιβλιοθήκη "Θεόφραστος" - Τμήμα Γεωλογίας. Α.Π.Θ.

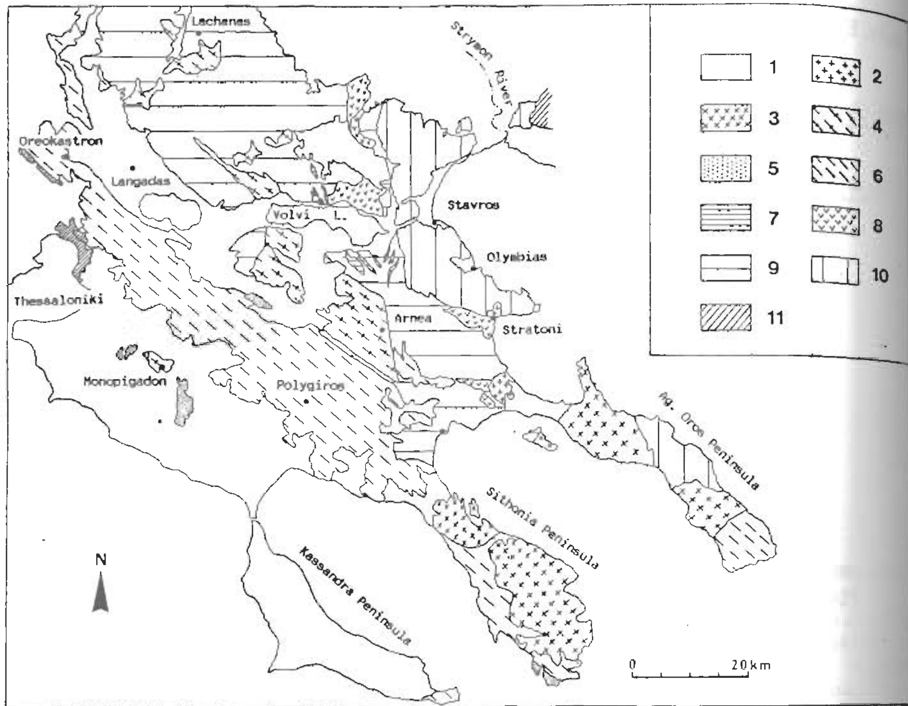

Fig. 1: Simplified geological map of the Chalkidiki Peninsula after KOCKEL et al (1977). 1. Neogene-Quarternary 2. Stratoni, Ierissos granodiorites 3. Sithonia granodiorite 4. Arnea, Lachanas, Monopigadon granites 5. Tithonian molasse 6. Circum Rhodopian Belt 7. Nea Madytos unit 8. TVG - Complex 9. Vertiskos unit 10. Kerdilion unit 11. Rhodopian massif.

phic analysis independently as well as comparatively to each other, starting from the younger to the older units, always in respect to the already existing literature.

II. TECTONOMETAMORPHIC ANALYSIS OF THE GEOTECTONIC UNITS

II.1. The Stratoni granodiorite

The Stratoni granodiorite occupies a small surfacial area to the north of the Stratoni village, intruding the gneisses and migmatites of the Kerdilion unit. NIKOLAOU (1960) characterizes it as biotite-hornblende to hornblende-biotite granodiorite with locally more dioritic parts and considers it as pre-Paleozoic. On the contrary, PAPADAKIS (1971) using the Rb/Sr radiometric method on biotites from the magmatic body determines Oligocene intrusion age of 29,6$\pm$1,4 Ma, proving the rightness of OSSWALD's (1938) view. Numerous aplitopegmatitic veins and an extensive contact aureole mark the granodiorite - gneiss boundary.

Remarkable evidence of ductile deformation are not visible in the main granodioritic body, which maintains its primary magmatic texture unaltered. That makes the Oligocene Stratoni granodiorite the oldest "geotectonic unit", which

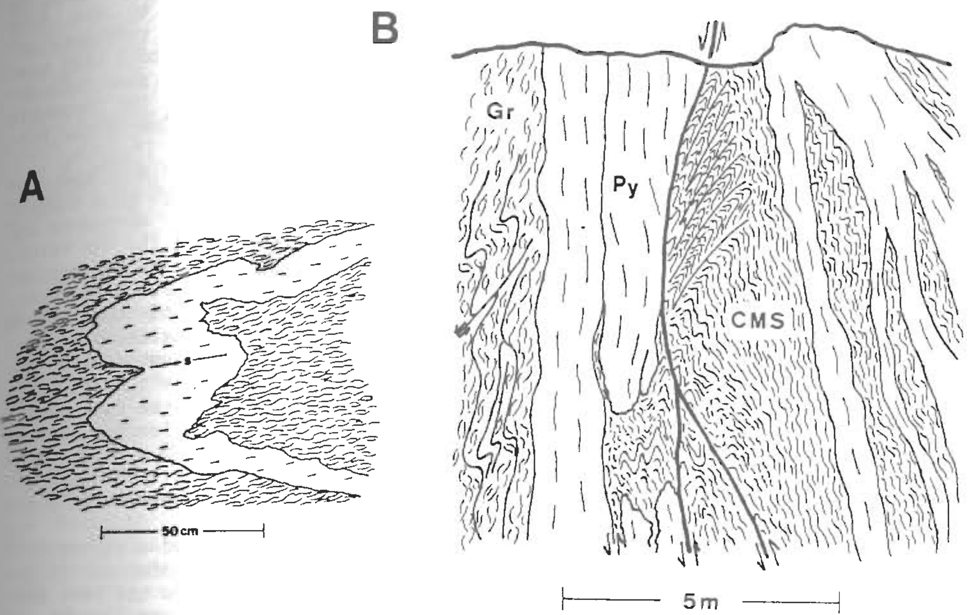

Fig. 2: See text for explanations.

has been unaffected by any tectonometamorphic event. Therefore we consider it as the temporal roof of the tectonometamorhic evolution in the Chalkidiki peninsula.

II.2. The Sithonia granodiorite

The greatest part of the Sithonia peninsula is built by the homonymous two mica to biotite granodiorite (SOLDATOS et al 1977, KOCKEL et al 1977). It intrudes the metamorphic rocks of the Circum Rhodopian Belt, especially the Jurassic Svoula flysch and the Chortiatis magmatic suite. An extended contact aureole characterized by numerous aplitopegmatitic veins and contact metamorphism phenomena has been certified along the granodiorite boundary. KOCKEL et al (1977) estimate contact metamorphism conditions to be of the albite-epidote and hornblende-hornfels facies.

DE WET et al (1989) using the K/Ar and Rb/Sr radiometric methods on micas from the Sithonia granodiorite estimate intrusion ages of 50,0$\pm$1 and 49,4$\pm$5,9 Ma. Similar intrusion ages were obtained earlier by KONDOPOULOU @ LAUER (1984), VERGELY (1984) and CHRISTOFIDIS et al (1986). Equivalent to the Sithonia granodiorite is the two mica Ouranoupolis granite, which yielded plateau ages of 44,9$\pm$2,4 and 44,5$\pm$1,1 Ma (DE WET et al, 1989).

The Sithonia granodiorite shows a completely developed deformation structure observed throughout the whole magmatic body, which gives it the character of a granodioritic augen gneiss. This structure indicates that the Eocene granodiorite has been affected by a strong ductile deformation phase under relative high strain rate, which must have taken place simultaneously at least to the late magmatic stages of the intrusion. The sketch shown on Fig. 2a proves the rightness of it. A pegmatitic vein intruding the granodiorite itself cuts the

Ψηφιακή Βιβλιοθήκη "Θεόφραστος" - Τμήμα Γεωλογίας. Α.Π.Θ.

already developed main schistosity of the magmatic body. The pegmatite itself is folded and a new weak schistosity is formed within it, parallel to the fold axial plane. It is noteworthy that the schistosity within the vein runs parallel to the host rock schistosity. That proves that the deformation of both the host rock and the vein, although of different type, resulted from the same deformation phase which, moreover, took place partly synchronously to the magmatic activity. It is obvious that this deformation phase took place during the Eocene.

The influence of the Eocene deformation is also clearly visible in the contact area of the granodiorite to the surrounding rocks. Although this contact has not completely lost its magmatic character, important tectonic movements and relatively strong ductile deformation have taken place along it. The veins, which intrude the surrounding rocks, run almost parallel to their main schistosity and are themselves ductily deformed, showing pinch and swell structures, boudinage and a weak schistosity. Although the tectonic layering of the Middle Jurassic Svoula flysch and the Chortiatis metamagmatites is considerably older than the Eocene deformation (Chapt. II.4), it has been reactivated and has undertaken the younger tectonic movements. It is also remarkable that the main schistosity of the granodiorite in the vicinity of the boundary area runs parallel to the main tectonic layering of the Jurassic rocks. A new cleavage associated with folding, occurs locally and is restricted within a 2 - 3 m thick zone from both sides of the granodiorite - host rock boundary (Fig. 2b). This new cleavage is genetically related to the tectonic movements that have taken place along the contact.

We can conclude that the Sithonia granodiorite has suffered a strong ductile deformation under relative high strain rate, which has transformed it to augen gneiss. This deformation took place partly synchronously to the late magmatic stages during the Eocene times and ceased before Oligocene, as long as no deformation structures are visible in the 29 Ma old Stratoni granodiorite.

II.3. Tithonian molasse

Limestones, conglomerates and phyllites known as "Upper Jurassic molasse" and "Prinochori beds" (MERCIER 1973, KOCKEL et al 1977) occur along the western boundary of the Circum Rhodopian Belt. The main occurences are those of Oreokastron-Neochorouda, Monopigadon, Porto Koufo and Paliourion. These molassic sediments cover transgressively the Monopigadon granite and the already once deformed and metamorphosed rocks of the Circum Rhodopian Belt, with which are involved later in new thrusting. According to KOCKEL et al (1977), they are metamorphosed under very low grade conditions.

Our observations come from the phyllites and conglomerates of the Oreokastron-Neochorouda area. The molassic sediments show a complex deformation structure resulted from the activity of two different deformation events upon them.

A first cleavage s_1 is developed within the conglomerates running parallel to the bedding and shown by two factors: the orientation of the minerals of the matrix and the elongation of the pebbles, which may reach width-to-length ratios of more than 1:10, proving ductile type of deformation.

The phyllites lie conformably on the conglomerates and show a more complicated deformation structure, due to their higher plasticity and their lower resistance to the deformation. The bedding, which is characterized by alternations of phyllitic, sandstone and conglomeratic limestone horizons, is once isoclinally to tightly folded (B_1). A first slate cleavage running parallel to the B_1 fold axial planes is well developed within the phyllitic horizons and weaker within the sandstone and the carbonate beds. New white mica and chlorite characterize the s_1-planes and show metamorphic conditions of very low grade. The B_1 folds have a relative constant N-S to NNW-SSE axial trace and a W to WSW polarity. The same deformation phase created also large coaxial folds and associated thrusts, which control the large scale structure of the molassic sediments and

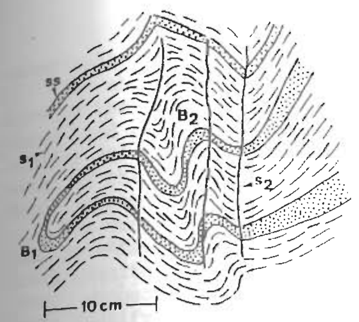

Fig. 3: Juxtaposition of two deformation phases within the molassic sediments near Neochorouda.

the underlying Circum Rhodopian Belt metamorphites.

The main deformation structure of the Tithonian molassic sediments, which was created during the first deformation phase, is rather wealky disturbed later on by a second deformation event. This younger phase created open and chevron folds B_2 usually without accompanying cleavage. The B_2 fold axes run E-W, parallel to the locally developed crenulation of the older s_1 or of the bedding. The later is often accompanied by an also weakly developed fracture to crenulation cleavage showing a S - SSW polarity (Fig. 3). Mesoscopic shear zones characterized by a top to S sence of movement resulted also during the second deformation phase.

II.4. Circum Rhodopian Belt

The Circum Rhodopian Belt has been interpreted by KOCKEL et al (1977) as the Mesozoic cover of the crystalline basement of the Chalkidiki peninsula. It is developed along the southwestern edge of the Serbomacedonian massif, which is complicated and characterized by multiple thrusting of rocks from both the basement and the cover.

The Circum Rhodopian Belt has been divided into three metasedimentary subunits of Late Permian to Middle(?) Jurassic age. The central axis of the Belt is occupied by the pelagic Melissochori-Cholomon subunit including the Triassic-Jurassic Svoula series. On both sides of it, two mainly neritic subunits are developed: the Deve Koran-Doubia to the NE and the Asprovrissi-Chortiatis subunit to the SW. The Circum Rhodopian Belt includes also the Thessaloniki-Ormylia ophiolitic complex and the Chortiatis magmatic suite.

KOCKEL et al (1977) report that the main deformation of the Belt rocks is associated with low grade metamorphism and took place during the Middle to Late Jurassic before the sedimentation of the Tithonian molasse. CHATZIDIMITRIADIS & KILIAS (1984) estimate one deformation phase in the Late Jurassic and later CHATZIDIMITRIADIS et al (1985) recognize another folding of Late Cretaceous to Paleocene age. Finally PATRAS et al (1986) report three folding phases in Middle-Late Jurassic, Early Cretaceous and Tertiary from the Circum Rhodopian Belt.

During our fieldwork we studied the metamorphites of the Circum Rhodopian Belt throughout their nearly whole expansion from the Oreokastron area to the southernmost Sithonia peninsula. We have been able to recognize three tectonometamorphic phases, whose temporal order is defined mainly with tectonic and geologic criteria (Fig. 4).

The main deformation structure of the Circum Rhodopian Belt rocks has been created during their first deformation phase in the Middle to Late Jurassic under metamorphic conditions of the greenschist facies. The main s_1 slate cleavage is developed parallel to the axial planes of the B_1 isoclinal folds of about NW - SE axial trace. Parallel to the B_1 axes runs the L_1 mineral lineation.

The second deformation phase is geometrically and kinematically identical to the first one of the Tithonian molasse and took place in the Lower Cretaceous. It created similar to isoclinal B_2 folds of N - S to NNW - SSE axial trace and new crenulation cleavage s_2 with or without new chlorite and mica. The L_2 microcrenulation runs also parallel to the B_2 axes. The second deformation phase took place under very low grade metamorphic conditions, and is responsible for the W-vergent large scale folds and thrusts of the Circum Rhodopian Belt.

Open to chevron folds B_3 with E - W axial trace and local crenulation to

Ψηφιακή Βιβλιοθήκη "Θεόφραστος" - Τμήμα Γεωλογίας. Α.Π.Θ.

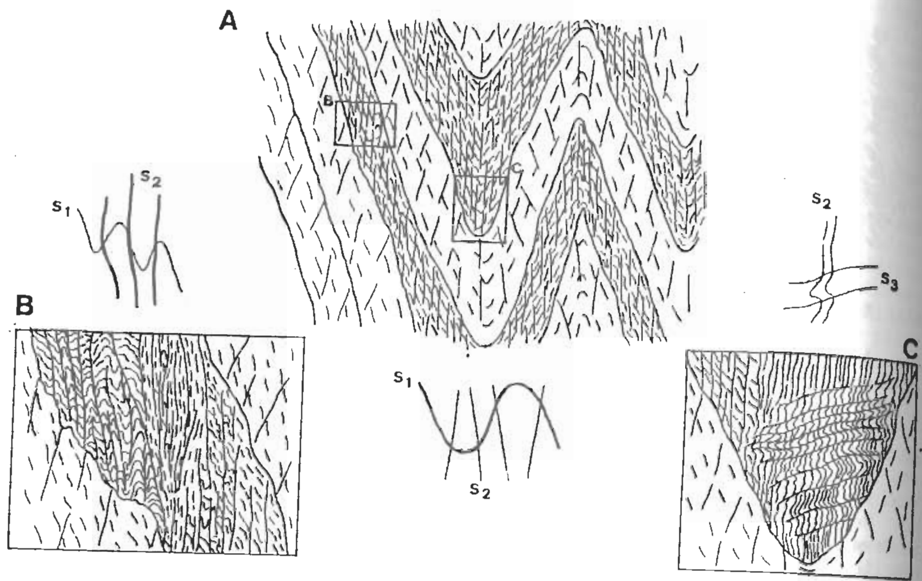

Fig. 4: Juxtaposition of three deformation phases within phyllitic - sandstone alternations of the Svoula flysch at the road Paleokastron - Polygiros.

fracture cleavage without new minerals have been created during the third deformation phase. Wide microcrenulation L_3 of E - W trace, S-vergent reverse faults, thrusting and kink bands are also expressions of the last phase, which took place under conditions of the upper tectonic stage. The similarities to the Eocene second deformation of the Tithonian molasse are obvious.

II.5. The Arnea granite

The two mica Arnea granite occupies a wide area to the south of the Volvi Lake between the Vertiskos unit to the east and the Circum Rhodopian Belt to the west. The granite is overthrusted onto the Vertiskos unit along an important mylonite zone. According to KOCKEL et al (1977), the small granitic bodies occuring within the Vertiskos unit (to the north of the Volvi Lake) and the Lachanas granite are equivalent to the main Arnea granitic body.

DE WET et al (1989) using the Rb/Sr whole rock radiometric method estimated intrusion age of the granite to be of at least 155 Ma and report that it has suffered a low grade metamorphism of greenschist facies during the Early Cretaceous. KOCKEL et al (1977) state that aplitopegmatitic veins coming from this granite intrude the Early to Middle(?) Jurassic Svoula metasediments. On the contrary, SAKELLARIOU & DuRR (in press) believe that the only veins occuring within the Svoula metasediments are young posttectonic pure quartz veins. They have moreover observed, that a narrow zone of mylonites exists between the Arnea granite and the Svoula metamorphites, which prove the tectonic character of their contact.

The Arnea granite shows a very interesting deformation structure. A very well developed schistosity s_1 accompanied by an also strong mineral lineation L_1 with almost constant N150°E trace, dominate throughout the whole exposure of the granite. Mesoscopic isoclinal to tight folds and large scale isoclinal folds are often observed. Their axes run parallel to the lineation, to which they are

genetically related.

The abovementioned tectonic features constitute the main deformation structure of the Arnea granite. It is very important to note that the tectonic features and the type of the first deformation phase of the Arnea granite are very similar and geometrically almost parallel to those of the first deformation event of the Circum Rhodopian Belt and of the third one of the Vertiskos and the Kerdilion unit, as it will be stated below.

The very low susceptibility of the rigid granitic bodies to the deformation has not favoured the creation of complete deformation structures within the Arnea granite during the weaker events of Early Cretaceous and Eocene. Nevertheless, mesoscopic shear zones of more or less ductile character as well as kink bands testify to their action upon the granite.

SAKELLARIOU (1989), relying on the way the mineral crystals of the granite were deformed during the first deformation phase, concludes that this must have taken place under medium grade metamorphic conditions. A second metamorphic event of greenschist facies, expressed mainly by extensive recrystallization, is reported also by DE WET et al (1989) and is proved to be of Early Cretaceous age.

From the abovementioned descriptions it is obvious that the Arnea granite and the Circum Rhodopian Belt have suffered the same tectonometamorphic event during Middle to Late Jurassic. We can moreover conclude that the granite is older than has been considered till now.

II.6. Nea Madytos unit

The Nea Madytos unit consists of metasediments, mainly graphitic gneisses and marbles, which occur in isoclinal fold and duplex structures within the Vertiskos unit to the north and south of the Volvi Lake. These rocks have been considered by KOCKEL et al (1977) to be equivalent to the Triassic-Jurassic Svoula series.

According to DIXON & DIMITRIADIS (1984), they have been metamorphosed under medium grade conditions. PAPADOPOULOS (1982) observes a first, possibly pre-Mesozoic amphibolite facies metamorphism, a second one of greenschist facies during the Early Cretaceous and a younger retrogression.

The nature and composition of the Nea Madytos metamorphites have favoured the creation of characteristic features and metamorphic parageneses of all four tectonometamorphic phases that affected them.

The fourth, and youngest deformation phase has created tight to open B_4 folds and microcrenulation L_4 of the main foliation with E-W axial trace, which are usually accompanied by fracture to crenulation cleavage s_4. New chlorite may develop along the s_4 planes, indicating very low grade metamorphic conditions. The geometrical and genetical similarities of this phase to the latest, Eocene tectonometamorphic event of the Circum Rhodopian Belt are obvious.

Tight to isoclinal B_3 folds of micro- to megascopic scale and parallel lineation tracing N-S have been created by the third deformation phase. The s_3 schistosity is observed locally to be parallel to the B_3 fold axial planes and is directly related to the paragenesis quartz + muscovite + Fe-rich chlorite + chloritoid observed within the graphitic gneisses. This relationship proves that the third deformation phase of the Nea Madytos unit took place under metamorphic conditions of the upper greenschist facies. This phase shows important similarities to the second Early Cretaceous deformation event of the Circum Rhodopian Belt, while PAPADOPOULOS (1982) reports also a greenschist facies thermal event in the Early Cretaceous from the same rocks to the north of Volvi Lake.

The main tectonometamorphic structure of the Nea Madytos unit has been created during the second tectonometamorphic event. The dominant s_2 schistosity is accompanied by a completely developed mineral lineation L_3 of NW-SE trace and by isoclinal folds of the same axial direction. This deformation event took place under metamorphic conditions of the lower amphibolitic facies, as can be

Ψηφιακή Βιβλιοθήκη "Θεόφραστος" - Τμήμα Γεωλογίας. Α.Π.Θ.

proved by the genetical relationship of the paragenesis quartz + muscovite + Mg-rich chlorite + garnet observed within the gneisses to the s_2 planes. The structural features of the second deformation event of the Nea Madytos unit are almost identical to those of the first one of the Arnea granite and the Circum Rhodopian Belt.

The tectonic and metamorphic features of the first tectonometamorphic event of the Nea Madytos unit are not well preserved. Isoclinal B_1 folds of assuming WNW-ESE axial trace, parallel mineral lineation L_1 and remnants of the old s_1 schistosity occur locally and are not easily recognizable. We have been able to find remnants of the first metamorphic paragenesis within the graphitic gneisses, which is consisted by quartz + muscovite + biotite + granate + staurolite and indicates medium grade metamorphic conditions.

The first tectonometamorphic event of the Nea Madytos unit predates all the events described up to now, but is identical to the second one of the Vertiskos and the Kerdilion unit. This event marks the first infolding of the Nea Madytos unit into the Vertiskos unit. From that point on these two units show identical structural and metamorphic evolution.

II.7. Vertiskos unit

The Vertiskos unit is the upper one of the Serbomacedonian massif and is considered by KOCKEL et al (1977) to be of Precambrian age. It contains various metaclastic rocks and amphibolites and is characterized by the absence of marbles. According to the same authors, it has suffered one pre-Mesozoic metamorphism of almandine-amphibolite facies and a younger diaphthoresis during the Early Cretaceous. DIXON & DIMITRIADIS (1987) distinguish three metamorphic events within the Vertiskos metamorphites to the north of Volvi Lake, the older of which may be of Variscian age.

SAKELLARIOU (1989) recognizes five tectonometamorphic events and states that any possible older events, which might have existed, are completely covered by them. The only witnesses of the first deformation phase are isolated multi-folded quartz veins, which lie within the dominant schistosity and signify the oldest recognizable s_1 foliation. No evidence, which could allow us to have any idea about the petrographic features of the first metamorphic phase, have been found. There are only some indications about the existence of an eclogitic metamorphism of unknown age within the amphibolitic rocks (DIMITRIADIS @ GODELITSAS 1991).

After the above mentioned "first" event, the tectonometamorphic evolution of the Vertiskos unit is identical to that of the Nea Madytos unit, having in mind that the oldest tectonometamorphic event of the Nea Madytos unit corresponds to the second one of the Vertiskos unit. All their tectonic features, the deformation type, the metamorphic grade and the genetical relationship between the deformation and the metamorphic phases are the same. The only difference observed is related to the composition of the corresponding parageneses, and can be explained by the different composition of the rocks of these two units.

The metamorphic parageneses of the four younger metamorphic events of the Vertiskos unit are the following:

Quartz + muscovite + biotite + garnet + staurolite + kyanite is the paragenesis of the second event and characterizes the medium amphibolite facies.

The third metamorphic event is characterized by the paragenesis quartz + muscovite + biotite + garnet + staurolite of the lower amhibolite facies.

The fourth event took place under metamorphic conditions of the upper greenschist facies and created the paragenesis quartz + muscovite + Fe-rich chlorite + chloritoid. All of the above parageneses have been observed within the metapelitic rocks of the Vertiskos unit.

The fifth metamorphic event can be recognized by the local development of new chlorite and the chloritization of older minerals. We believe that it took place under very low to low grade metamorphic conditions.

II.8. Kerdilion unit

The Kerdilion unit is the lower one of the Serbomacedonian massif and is composed mainly of biotite gneisses, amphibolites and a few marble horizons. According to KOCKEL et al (1977) it is also of Precambrian age. MANTZOS (1991) using the Rb/Sr whole rock method on the Olympias biotite gneisses estimates two different ages of 337±5 and 113±11 Ma and believes that they relate to one amphibolite facies metamorphic event in the Early Carboniferous and to a green-schist one in the Early Cretaceous respectively.

According to our observations, the two units of the Serbomacedonian massif show identical deformational evolution in relation to the number, the type and the temporal succesion of the recognizable deformation phases. The monotonous mineralogical composition of the Kerdilion metamorphites has not favoured the creation of characteristic metamorphic parageneses. The presence of hornblende within the amphibolites and the high An-component of the plagioclases proves the action of at least one amphibolite facies metamorphic event, side by side to the main third deformation phase. We have also observed the paragenesis quartz + muscovite + biotite + Fe-rich chlorite + garnet locally within the garnet-rich biotite gneises. It characterizes the upper greenschist facies and is genetically related to the s_4 planes of the fourth deformation phase. Based on these evidence we believe that the metamorphic evolution of the Kerdilion unit must have been also quite similar to that one of the Vertiskos unit.

III. DISCUSSION - CONCLUSIONS

The succession of all the recognizable tectonometamorphic events which acted upon the geotectonic units of the Chalkidiki peninsula may be summarized in Table 1.

In the authors opinion it is quite clear, that the three younger events correspond to the three alpidic orogenic events which are known from the Chalkidiki area. The youngest tectonometamorphic event acted among the other units also upon the 50 Ma old Sithonia granodiorite, but is not recognizable within the 29 Ma old Stratoni granodiorite, which is undeformed. Thus it must have taken place within the Eocene. Radiometric ages of 30 - 50 Ma obtained with the K/Ar- and Rb/Sr-methods on micas from the Kerdilion gneisses prove the existence of a thermal event in Eocene too (HARRE et al, 1968).

The fourth tectonometamorphic event may be recognized as the first structure forming event of the Tithonian molasse, but is not detectable within the Sithonia granodiorite and, therefore must be attributed to the Eshellenic phase of Early Cretaceous. Early Cretaceous radiometric ages are known from nearly the whole exposure of the Vertiskos unit, the Nea Madytos unit, the Arnea granite (HARRE et al 1968, PAPADOPOULOS 1982, DE WET et al 1989) and recently from the Kerdilion unit (MANTZOS, 1991). According to the authors mentioned above, they display the age of an extensive low grade metamorphic episode.

The third event represents the first structure forming event of the Circum Rhodopian Belt and the Arnea granite, took place in the Late Jurassic and has not affected the Tithonian molasse. This event is responsible for the creation of the dominant deformation structure and the main metamorphic parageneses of all the pre-Late Jurassic geotectonic units. Radiometric ages of Middle and Late Jurassic are also known from the Vertiskos unit to the north of the Volvi Lake (PAPADOPOULOS, 1982) and from the Arnea granite (DE WET et al, 1989).

The second tectonometamorphic event predates the third one and is presented only as remnants within the Nea Madytos, the Vertiskos and the Kerdilion units. The temporal determination of this event and its enrollment to any orogenic phase are very uncertain and are directly related to the age and the geotectonic position of the Nea Madytos unit.

If the Nea Madytos unit is equivalent to the Svoula series of the Circum Rhodopian Belt, as KOCKEL et al (1977) believed then the second tectonometamor-

Ψηφιακή Βιβλιοθήκη "Θεόφραστος" - Τμήμα Γεωλογίας. Α.Π.Θ.

TABLE 1

Phases / Units	"1st phase" ?	2nd phase ?	3rd phase Upper Jurassic	4th phase Lower Cretaceous	5th phase Eocene
Stratoni granodiorite 29 Ma	-----	-----	-----	-----	-----
Sithonia granodiorite 45 - 50 Ma	-----	-----	-----	-----	augen gneiss, s_1, B_1, tectonic movements along contact
Tithonian molasse	-----	-----	-----	s_1, B_1 (N-S), thrusting, very low grade metamorphism	s_2, B_2 (E-W), thrusting
Circum Rhodopian Belt	-----	-----	s_1, B_1 (NW-SE), L_1, low grade metamorphism	s_2, B_2 (N-S), L_2, very low grade metamorphism	s_3, B_3 (E-W), kink bands, thrusting
Arnea granite	-----	-----	s_1, B_1 (NW-SE), L_1 mylonites, medium grade metamorphism	thrusting, recrystallization, low grade metamorphism	kink banbs, thrusting
Nea Madytos unit	-----	s_1, B_1, L_1, medium grade metamorphism	s_2, B_2 (NW-SE), L_2, medium grade metamorphism	s_3, B_3 (N-S), L_3, low grade metamorphism	s_4, B_4 (E-W) L_4, very low grade metamorphism
Vertiskos unit	"s_1", Qz-veins eclogites ?	s_2, B_2, L_2, medium grade metamorphism	s_3, B_3 (NW-SE), L_3, medium grade metamorphism	s_4, B_4 (N-S), L_4, low grade metamorphism	s_5, B_5 (E-W), L_5, very low grade metamorphism
Kerdilion unit	"s_1", Qz-veins eclogites ?	s_2, B_2, L_2, medium grade metamorphism	s_3, B_3 (NW-SE), L_3, medium grade metamorphism	s_4, B_4 (N-S), L_4, low grade metamorphism	s_5, B_5 (E-W), L_5, very low grade metamorphism

Table 1: Geotectonic units versus tectonometamorphic phases. See text for explanations.

phic event must be attributed to the Late Jurassic orogenic episode. In that case, one has to explain two points: (1) why it has not affected the main body of the Circum Rhodopian Belt to the west of the Serbomacedonian massif and (2) which tectonic processes resulted in the present position of the Nea Madytos unit within the Vertiskos unit. Considering that the paleogeographic position of the Circum Rhodopian Belt was very probably to the west or southwest of the Serbomacedonian massif and that the alpidic orogenic system in that area is characterized (throughout its whole duration) by tectonic movements of westward or southwestward polarity, it seems to the author that the present position of the Nea Madytos unit cannot be easily explained in such a way.

The second possibility is that the Nea Madytos unit is independent and older than the Circum Rhodopian Belt, a view strongly supported by PAPADOPOULOS (1982) and also preffered by the author. In that case, there are two possible geotectonic interpretations of the origin and the position of the Nea Madytos unit and the age of the second tectonometamorphic event:

(1) The Nea Madytos unit is of Paleozoic age and its first structure forming event took place in the Upper Carboniferous or Lower Permian. Many pegmatites from the western Vertiskos unit yielded with the Rb/Sr whole rock method

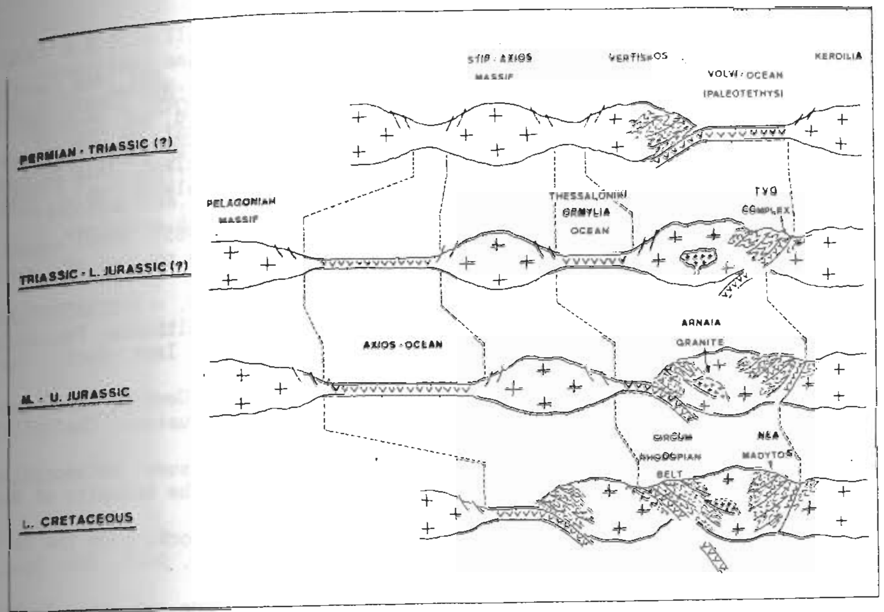

Fig. 5: Possible plate tectonic evolution of the Chalkidiki Peninsula tectonometamorphic units.

ages of 278 - 320 Ma (BORSI et al, 1965). Recently MANTZOS(1991) estimated, using the same method similar age (330 Ma) from the Kerdilion biotite gneisses near Olymbias.

(2) The Nea Madytos unit originated within the Paleotethys, the Cimmerian Ocean of SENGOR et al (1984), which should have closed before Middle Jurassic. Two radiometric ages of 191 Ma from a pegmatite from the western Vertiskos unit (HARRE et al, 1968) and 180 Ma from the Monopigadhon granite (equivalent to the Arnea granite?) (RICOU, 1965) may be considered as cooling ages from the Cimmerian orogenic event.

In the second case, one has to accept the following: (a) The Paleotethys was located between the Vertiskos unit to the south and the Kerdilion unit to the north. (b) The Vertiskos unit represents the Cimmerian Continent which drifted away from Godwana, moved to the north and leaded to the closure of Paleotethys in front of it, as well as the opening of the alpidic Tethys behind it. (c) The Kerdilion unit belonged to the Eurasian Continent. (d) The Therma - Volvi - Gomati Complex (TVG), a metamorphic and dismembered ophiolitic complex occuring along the Vertiskos - Kerdilion tectonic contact within the Serbomacedonian massif, represents an ophiolitic suture and specifically the Cimmerian ophiolitic suture. This is one of the possible interpretations of the origin of the TVG - Complex, which is not favoured but also not denied by DIXON @ DIMITRIADIS (1984) (e) The Arnea granite originated in Permian - Triassic within the magmatic arc created above the subduction zone of the Cimmerian oceanic crust.

The term "first tectonometamorphic event" includes all the possibly existed events which affected the Vertiskos and the Kerdilion units and predate the first structure forming event of the Nea Madytos unit. They are mostly not recognizable any more and their age is unknown. The evidence for high pressure eclogitic metamorphism within the Vertiskos unit (DIMITRIADIS @ GODELITSAS, 1991) may be attributed to the so called "first event".

Ψηφιακή Βιβλιοθήκη "Θεόφραστος" - Τμήμα Γεωλογίας. Α.Π.Θ.

REFERENCES

BORSI S. - FERRARA G. - MERCIER J. (1965): Determination de l' age des series metamorphicues du Massif Serbomacedonien au nord-est de Thessalonique (Grece) par le methodes Rb/Sr et K/Ar. Ann. Soc. geol. Nord, 84, 223-225.

CHATZIDIMITRIADIS E.A. - KILIAS A. (1984): Paleozoisch-mesozoische Faltungsphasen im Gebiet ostlich von Thessaloniki / Nord Griechenland. (Ein Beitrag zur Geologie der inneren Helleniden-Zone). N. Jb. Geol. Paleont. Mh., 1984: (4), 193-203.

CHATZIDIMITRIADIS E.A. - KILIAS A. - STAIKOPOULOS G. (1985): Nuovi aspetti petrologici e tettonici del masiccio Serbomacedone e delle regioni adiacenti della Grecia del nord. Boll. Soc. Geol. It., 104, 515-526.

CHRISTOFIDIS G. - D'AMICO C. - DEL MORO A. - ELEFTHERIADIS G. - KYRIAKOPOULOS C. (1986): A Rb/Sr study of the granitoids of the Sithonia Peninsula (Northern Greece). 6th Int. Conf. Geochron. Cosmochron. Isot. Geol., Cambridge.

DE WET A.P. - MILLER J.A. - BICKLE M.J. -CHAPMAN H.J. (1989): Geology and geochronology of the Arnea, Sithonia and Ouranopolis intrusions, Chalkidiki Peninsula, northern Greece. Tectonoph., 161, 65-79.

DIMITRIADIS S. - GODELITSAS A. (1991): Evidence for high pressure metamorphism in the Vertiskos group of the Serbomacedonian massif: the eclogite of Nea Roda, Chalkidiki. Bull. Geol. Soc. Greece, XXV/2, 67-80.

DIXON J.E. - DIMITRIADIS S. (1984): Metamorphosed ophiolitic rocks from the Serbo-Macedonian Massif, near Lake Volvi, North-east Greece. Geol. Soc. Spec. Publ., 17, 603-618.

DIXON J.E. - DIMITRIADIS S. (1987): The metamorphic evolution of the Serbo-Macedonian massif in Greece. Terra cognita, 7, (2-3), p. 106.

HARRE W. - KOCKEL F. - KREUZER H. - LENZ H. - MUELLER P. - WALTHER H.W. (1968): uber Rejuvenationen im Serbo-Mazedonischen Massiv (Deutung radiometrischer Altersbestimmungen). Acad. Prague, 23rd Int. Geol. Congr., 6, 223-236.

KOCKEL F. - MOLLAT H. - WALTHER H.W. (1977): Erlauterungen zur Geologischen Karte der Chalkidiki und angrenzeder Gebiete 1:100.000 (Nord Griechenland). BGR, 119 p., Hannover.

KONDOPOULOU D. - LAUER J.P. (1984): Paleomagnetic data from Tertiary units of the north Aegean zone. Geol. Soc. London, Spec. publ., 17, 681-686.

MANTZOS L. (1991): Rb/Sr whole rock geochronology of gneisses from Olymbias, Chalkidiki. Bull. Geol. Soc. Greece, XXV/1, 147-162.

MERCIER J. (1973): Etude geologique des zones internes des Hellenides en Macedoine centrale (Grece). Ann. Geol. Pays Hell., 20, 792 p.

NIKOLAOU M. (1960): L' intrusion granitique dans le region de Stratoni - Olymbiade et sa relation avec la metallogenese. Ann. Geol. Pays Hell., 11, 214-265.

OSSWALD K. (1938): Geologische Geschichte von Griechisch-Nordmazedonien. Denkschr. geol. Landesanst. Griechenland, 3, 141 p.

PAPADAKIS A. (1971): On the age of the granitic intrusion near Stratonion, Chalkidiki (Greece). Ann. Geol. Pays Hell., 23, 297-300.

PAPADOPOULOS C. (1982): Geologie des Serbo-Mazedonischen Massivs noerdlich des Volvi-Sees. Dissertation, 159 p., Wien.

PATRAS D. - KILIAS A. - CHATZIDIMITRIADIS E. - MOUNTRAKIS D. (1988): Study of the deformation phases of the Internal Hellenides in Nothern Greece. Bull. Geol. Soc. Greece, XX/1, 139-158.

RICOU L.E. (1965): Contribution a l' etude geologique de la bordure sud-ouest du Massif Serbo-Macedonien aux environs de Salonique. These, 3e cycle, Univ. Paris, 121 p.

SAKELLARIOU D. (1989): Geologie des Serbomazedonischen Massivs in der nordostlichen Chalkidiki, N-Griechenland - Deformation und Metamorphose. Diss., Univ. Mainz, Geol. Monogr. 2, 177 p, Athens.

SAKELLARIOU D. - DuRR St. (in print): Geological structure of the Serbomacedonian massif in NE Chalkidiki Peninsula. Bull. Geol. Soc. Greece.

SENGOR A.M.C. - YILMAZ Y. - SUNGURLU O. (1984): Tectonics of the Mediterranean Cimmerides: nature and evolution of the western termination of Paleo-Tethys. Geol. Soc. London, Spec. Publ. 17, 77-112.

SOLDATOS K. - SAPOUNTZIS E. - CHRISTOFIDIS G. - ELEFTHERIADIS G. (1976): Contribution to the study of the Sithonia plutonic complex I. (N. Greece). Ann. Geol. Pays Hell., 28, 62-98.

VERGELY P. (1984): Tectonique des ophiolites dans les Hellenides Internes (deformations, metamorphismes, et phenomees sedimentaries). Consequences sur l' evolution des regions Tethysiennes Occidentales. Thesis, Orsay.

Ψηφιακή Βιβλιοθήκη "Θεόφραστος" - Τμήμα Γεωλογίας. Α.Π.Θ.