THE ALPINE EVOLUTION OF THE CYCLADES.

AN INTRODUCTION AND EXCURSION GUIDE TO THE GEOLOGY OF THE ISLANDS OF SIFNOS AND TINOS

Dov Avigad

Alan Matthews

Zvi Garfunkel

Department of Geology

Institute of Earth Sciences

The Hebrew University of Jerusalem

Jerusalem 91904, I S R A E L

LIST OF CONTENTS

ACKNOWLEDGEMENTS

This guide was prepared in connection with the excursion organized by the Geological Society of Israel to the Cyclades in October 1988. We would like to thank Dr. Y. Arkin of the Geological Survey of Israel for his help in coordinating the excursion.

The work of a number of people has contributed to the preparation of this guide, and many have given advice and assistance which has considerably aided our own work. We would particularly like to thank Manfred Schliestedt, Bernard Evans and Michael Broecker.

Preparation of this guide forms part of the project 'Geodynamic Evolution of the Blueschist and Greenschist Facies Metamorphism, Cyclades, Greece', funded by a grant from the United States-Israel Binational Science Foundation (BSF), Jerusalem, Israel. We thank the director of the I.G.M.E. in Athens for permitting the research in the Cyclades

1. A BRIEF INTRODUCTION TO THE GEOLOGY OF THE CYCLADES

The Cycladic massif in the central Aegean Sea forms part of an Alpine orogenic belt stretching from continental Greece to the Turkish mainland (Fig 1.1). Its geodynamic evolution is dominated by processes of collision, subduction and uplift of Alpine age, which have resulted in several stages of metamophism, deformation and magmatism.

The Cycladic massif (Fig. 1.2) is generally considered to be composed of at least two tectonic units (Durr et al., 1978). The dominant unit is the lower high pressure metamorphic unit which consists of metamorphosed carbonates, clastic sediments and acid and basic volcanics. This sequence was apparently deposited on an older Hercynian basement which outcrops on the islands of Ios and Naxos (Henjes-Kunst and Kreuzer 1982, Andriessen et al. 1987). The upper tectonic unit, which consists of Permian and Triassic sediments, low-P crystalline metamorphics and ophioltic material, appears to have been thrust onto the lower unit at around the Miocene - Pliocene boundary (Altherr and Seidel, 1977).

The high pressure unit of the Cyclades is dominated by the formation of Blueschist and Eclogite facies rocks which are now best preserved in the islands of Sifnos, Syros and Tinos (Fig. 1.2). Rb/Sr and K-Ar age determinations of micas indicate that this subduction related metamorphism occurred at around 40 - 45 Ma (Altherr et al., 1979; Andriessen et al., 1979). Temperature and pressure conditions of the high pressure metamorphism have been estimated at 450 - 500°C and 15 kbars (Schliestedt et al., 1987).

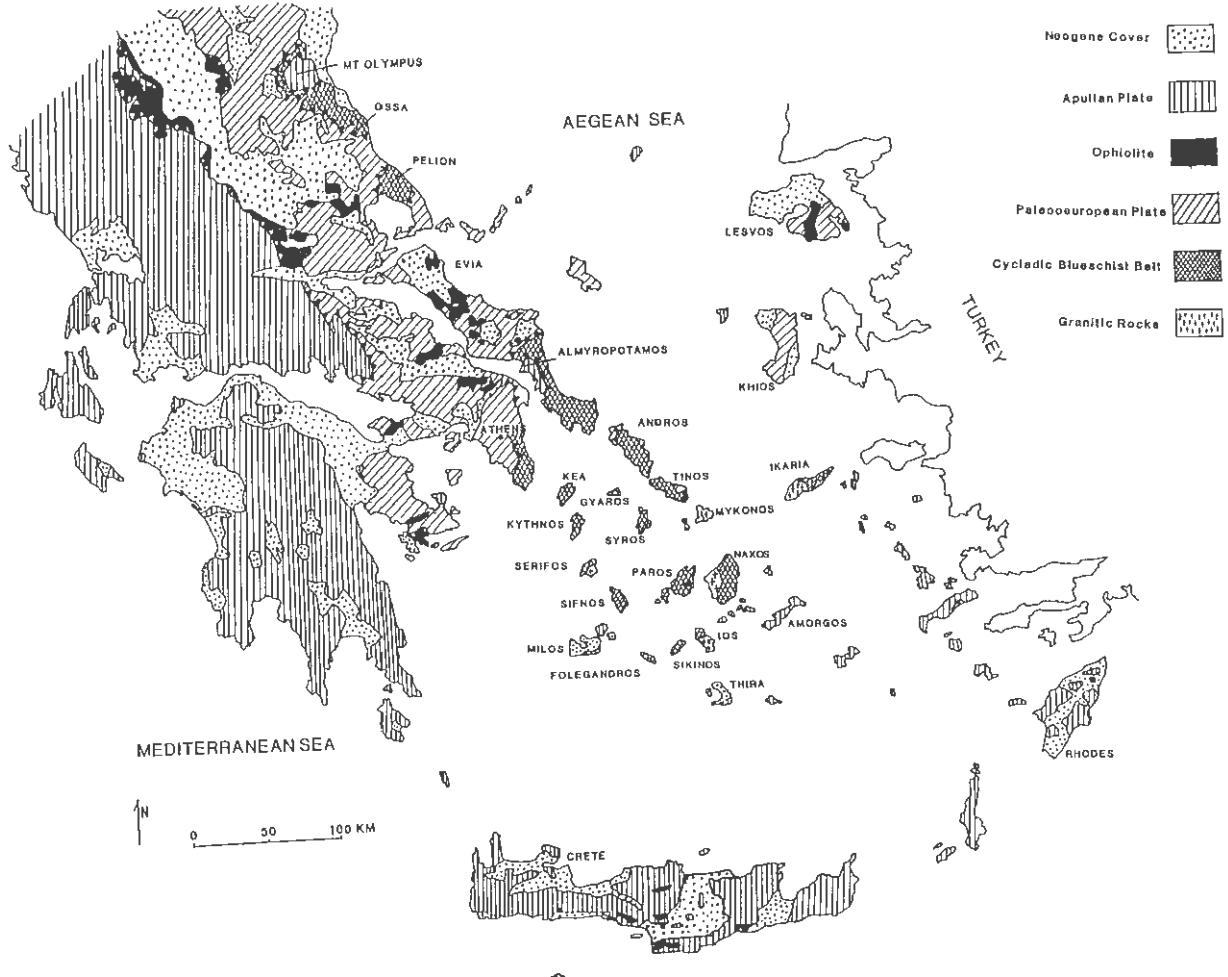


Fig. 1.1 Geologic setting of the Cycladic blueschist belt, Greece(from Blake et al, 1981)

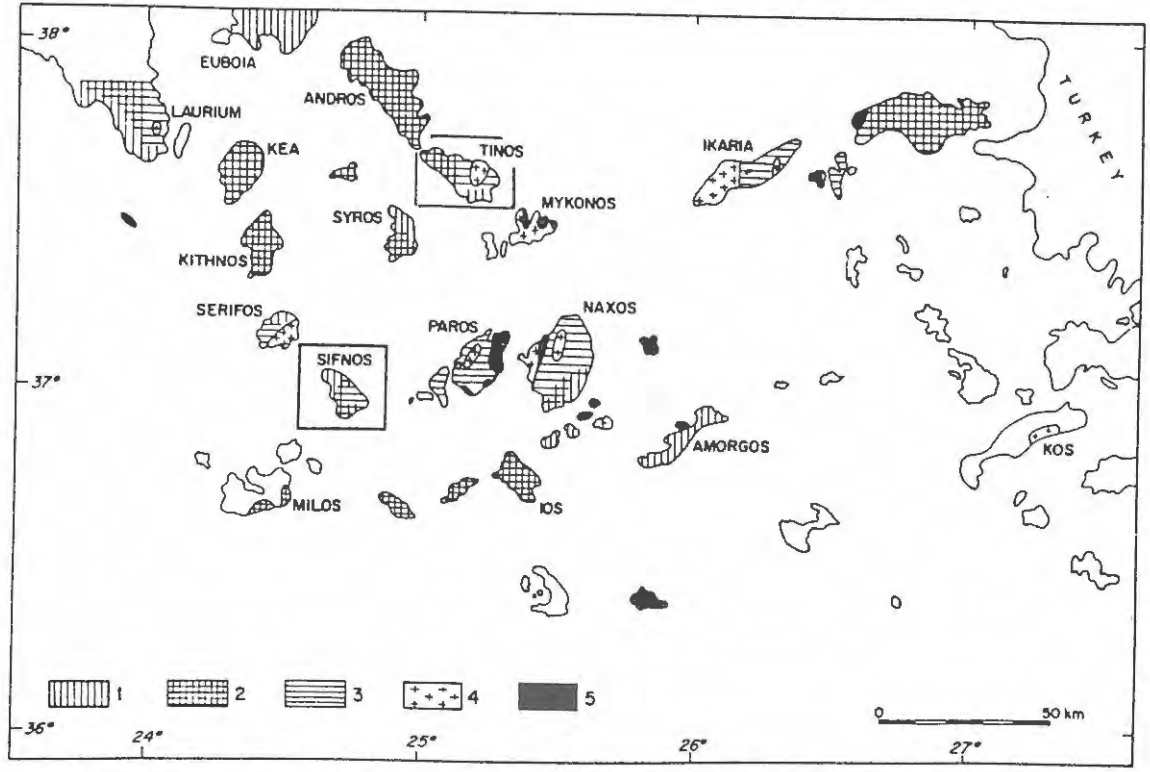


Fig. 1.2. Metamorphic rocks of the Attic-Cycladic complex. Lower (Basal) unit, 1–4. *1* = Eocene high-pressure metamorphic rocks; *2* = Eocene high-pressure metamorphic rocks overprinted by the Oligocene/Miocene low/medium-pressure metamorphism; *3* = Oligocene/Miocene low/medium-pressure metamorphic rocks; *4* = Oligocene/Miocene granitoids; *5* = undifferentiated rocks of the upper tectonic unit. Geology is generalized after Altherr et al. (1979, 1982)

The metamorphism is thus believed to result from tectonic burial of surface derived rocks (sediments and volcanics) to depths of 50 km or more.

During the late Oligocene to early Miocene, a major metamorphic overprint occurred throughout the massif and both partially and completely effaced the assemblages of the Eocene high pressure metamorphism. K-Ar ages on hornblendes and K-Ar and Rb/Sr ages on micas are in the range of 20-25 Ma (Andriessen et al. 1979; Altherr et al., 1979, 1982).

The Oligocene - Miocene metamorphic phase appears to be related to regional scale movement of fluids, whose provenance still remains an enigma. Greenschist facies conditions were regionally typical of this metamorphic phase and temperatures and pressures of 450°C and 5-7 kbars prevailed (Schliestedt and Matthews, 1987). Post Eocene uplift of the blueschist rocks was thus essentially isothermal. Locally, as on Naxos, higher temperature conditions occurred, culminating in the formation of amphibolites and migmatites at T >600°C and P= 5 to 7 kbar (Jansen and Schulling, 1976).

Approximate Pressure-Temperature-time (P-T-t) paths on Sifnos and Naxos are shown in Fig. 1.3.

The Oligocene - Miocene Greenschist overprint was followed in the Early Miocene by extensive emplacement of granitoid rocks whose intrusion was accompanied by local contact metamorphism (Altherr et al., 1976; Baltazis 1981; Salemink 1985). Radiometric Age measurements resulted in the following assessment (Altherr et al. 1988, and references cited therein) : 18 Ma (Ikaria and Tinos), 15

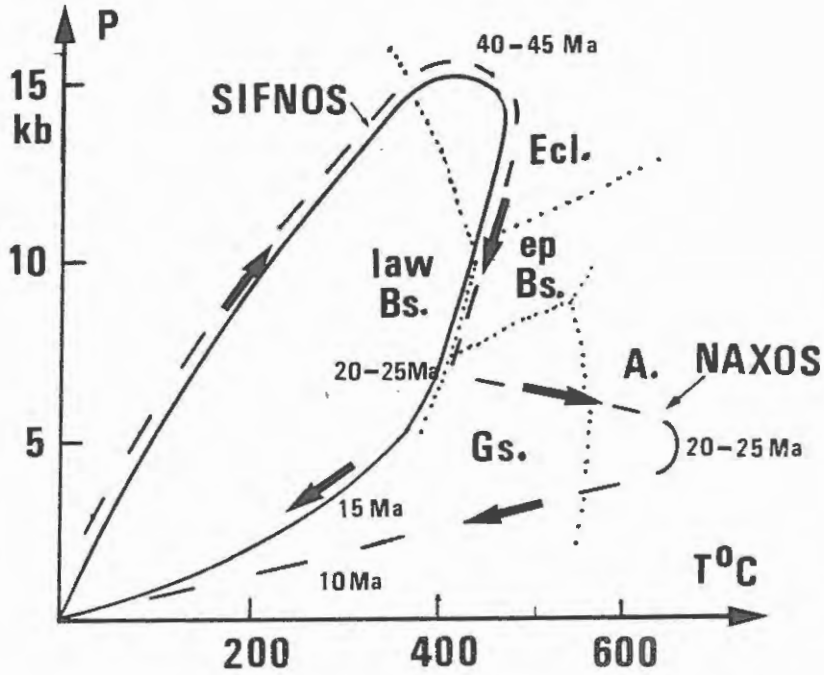


Fig. 1.3 Schematic P-T-t path showing the metamorphic evolution as examplified by the islands of Sifnos and Naxos.Numbers on curves refer to ages in Ma. Very approximate boundaries separate relevant metamorphic facies are also shown in dotted lines. Ecl= Eclogite, law Bs= (lawsonite) Blueschist, ep Bs=(epidote) Blueschist, A= Amphibolite, Gs= Greenschist.

Ma (Myconos and Delos), 12 Ma (Naxos, Kos and Samos) and 10 Ma (Laurion and Serifos). The granitoids show a regional trend from Granodiorites in the south west (e.g. Serifos, Laurium) to Granites in the center (Naxos, Mykonos, Tinos, Ikaria) to Monzonites in the south east (Samos and Kos). This regional trend and the systematic variation of initial $^{87}Sr/^{86}Sr$ ratios, $\delta 180$ and 1/Sr lead Altherr et al., (1988) to propose that the granitoids were generated by a mechanism of assimilation of older continental material by mantle derived arc magmas with combined fractionation.

Altherr et al. (1982) consider that the granites and the greenschist overprint are part of a high temperature zone which forms a paired metamorphic belt with the Oligocene Miocene high pressure metamorphic rocks of Crete and the Pelopponesus (Seidel et al., 1982).

The Cycladic massif is one of the largest units of the Hellenides but its structural position within the orogenic framework is not very clear. Blueschists, considered to be the northward extension of the Cycladic massif, also occur on mainland Greece (Fig. 1.1). They form a distinct tectonic unit within the nappe pile which comprises the Hellenides. The high pressure rocks are overlain by the Pelagonian nappe which consists of reworked Hercynian basement and its Mesozoic cover, and overlie, with a tectonic contact, Eocene flysch (Godfriaux, 1977). Although this nappe structure cannot be readily extended into the Aegean region, recent discoveries from the Island of Tinos (Avigad and Garfunkel, 1988) support current tectonic models (Papanikolaou, 1984; Bonneau 1984) and indicate that the blueschist unit of the Cyclades form an

allochtonous thrust sheet. This thrust sheet was emplaced onto a low grade carbonate series later than the Eocene and was displaced at least 50 km southwards.

The collision processes involved in the Cycladic blueschist metamorphism have been attributed to the northward or north-eastward directed underthrusting of the Apulian microplate beneath the Eurasian plate (Biju-Duval et al., 1977). A detailed geological evolution of the Cyclades within the tectonic framework of the Hellenides as outlined by Altherr and Seidel (1977) is illustrated in Fig. 1.4.

The present form of the Aegean area has been shaped by a process of fragmentation which started in the middle to late Miocene (Angelier, 1978) and resulted in the formation of the Aegean sea above a thin continental crust (Makris, 1978). The extensional tectonics affecting the Aegean is related to the underthrusting of the African continent beneath the Aegean plate. In the late Pliocene (Fytikas et al., 1984) a calc-alkaline volcanic arc evolved from Aegina in the west through Milos and Santorini to Nisiros in the east. This volcanic arc is situated above an inferred Benioff zone (Le Pichon and Anglier, 1979) and is still active today.

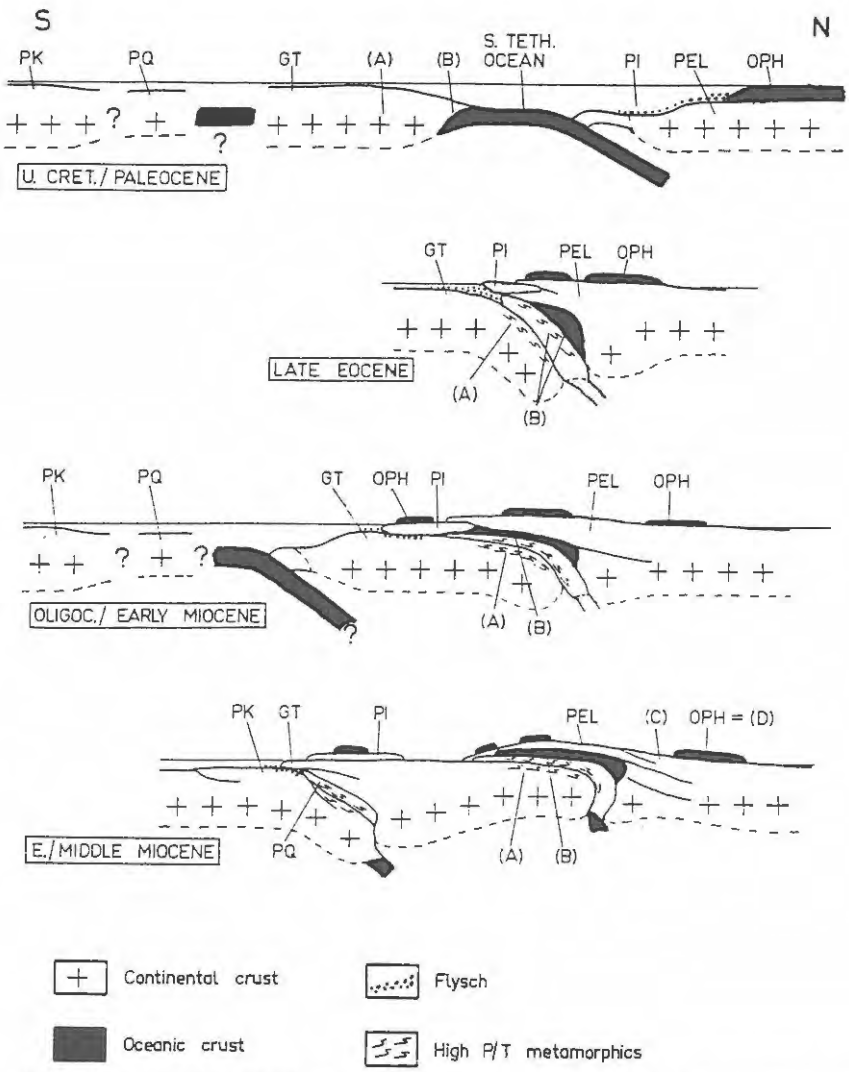


Fig. 1.4 Plate tectonics model for the Tertiary evolution of the Southern Aegean. OPH = Ophiolite nappe, PEL = Pelagonian, PI = Pindos, PQ = Phyllite-Quartzite series, PK = Plattenkalk series, GT = Gavrovo-Tripolitza, S.TETH.OCEAN = Southern Tethyan ocean. (from Altherr & Seidel, 1977)

2. GEOLOGICAL EVOLUTION OF SIFNOS

2.1 Introduction and Stratigraphy

The Cycladic island of Sifnos is located in the western central part of the Aegean sea, about 150 km SE from Athens (Figs. 1.2; 2.1). The island is about 15 km long and has a maximum width of about 7 km. The highest topographic summit is Mt. Profitis Elias, rising 700 m above sea level.

Sifnos is bounded and transected by steep cliffs which formed as a result of extensive faulting associated with the development of the Aegean Sea. This Late Cenozoic faulting shaped the present structure of the island, which is dominated by a series of tilted blocks. This structure can be readily appreciated in the N-S cross section given in Fig. 2.2.

The island consists of sequences of metamorphic rocks, which form part of the basal tectonic unit of the Cycladic blueschist belt. The stratigraphic sequence is ~2 km thick and can be divided into 4 lithological subunits. The four subunits are (Fig. 2.1): I. A lower greenschist unit in which relics of high pressure rocks occur locally. The unit is dominated in its lower parts by metapelites and by basic metavolcanics in its upper parts. Marble intercalations a few tens of meters thick are also prominent features of this unit, with rare layers of quartzites and meta-acidic rocks. The base of this unit is not exposed and consequently the underlying substratum is not known. II. The lower greenschist unit is overlain by a 700m thick Main Marble sequence. The sequence is mainly composed of calcite marbles, but a

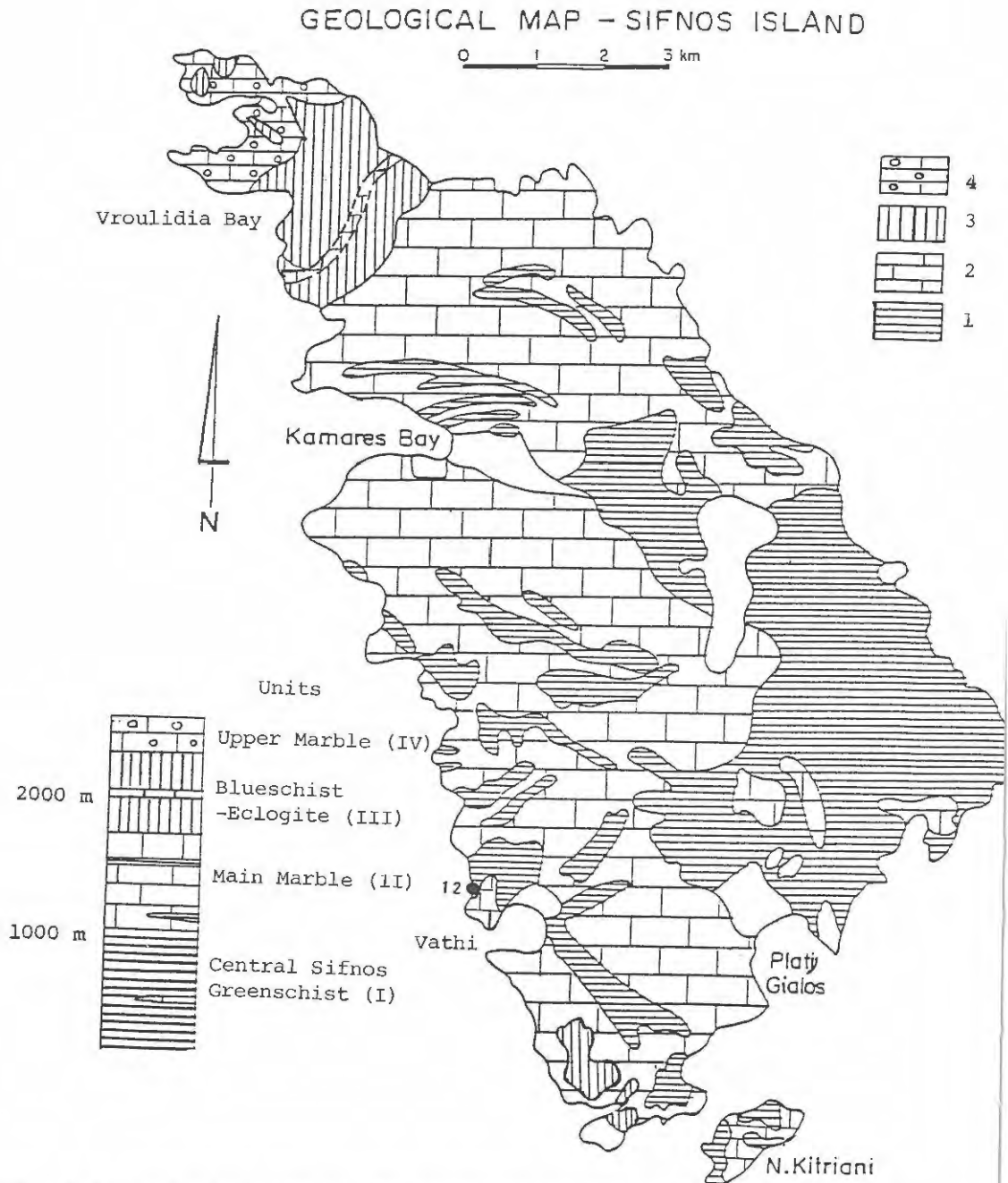


Fig. 2.1 Geological map of the island of Sifnos modified after Davis (1966)

1) mainly greenschist facies rocks 2) main marble 3) eclogite facies rocks 4) upper marble

N - S STRUCTURAL CROSS SECTION OF THE ISLAND OF SIFNOS

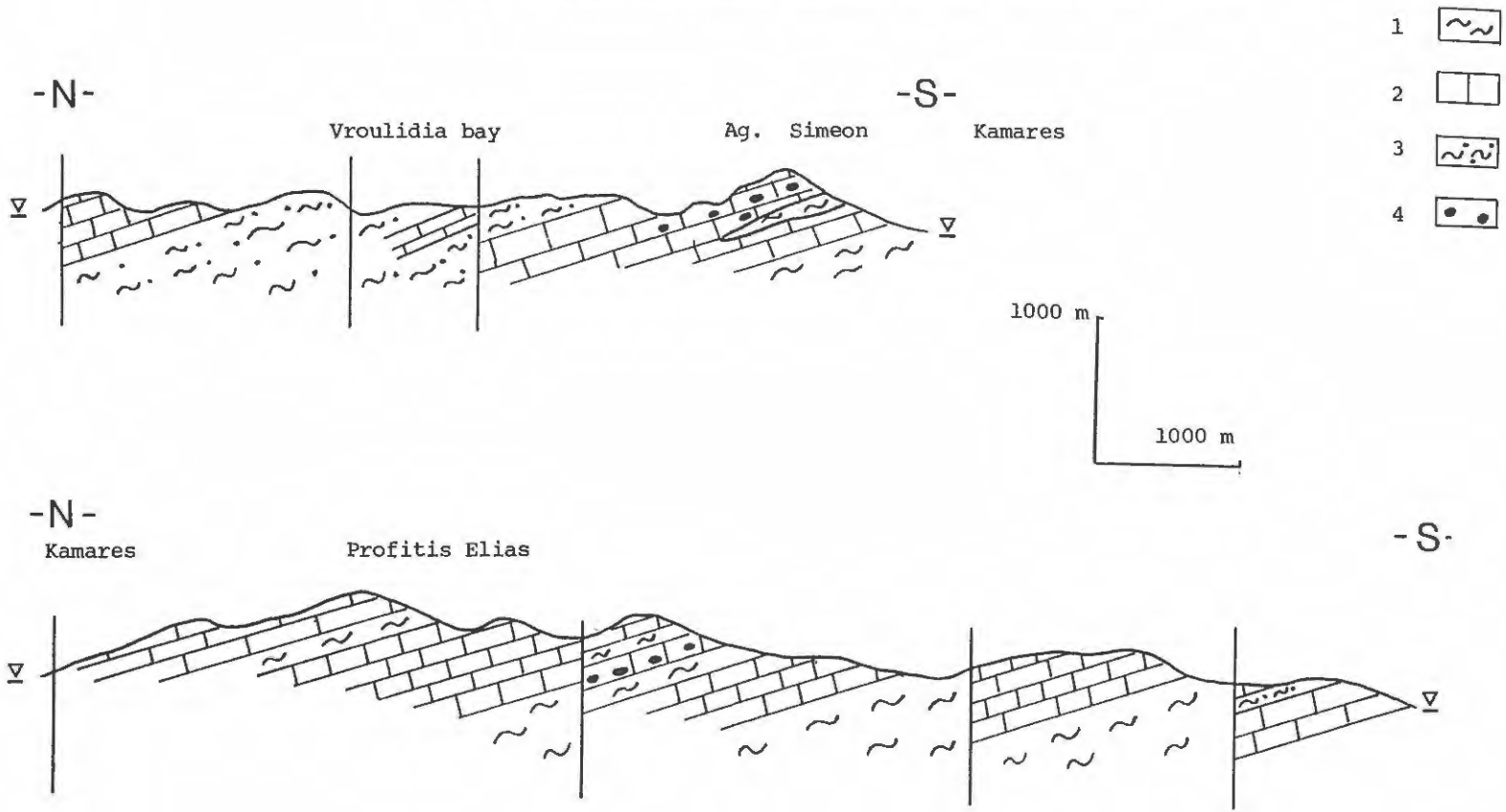


Fig.2.2 N - S cross section, Sifnos, 1) greenschists 2) marble 3) blueschists-eclogites 4)conglomerates

few distinct thick horizons of dolomite marbles occur in the lower half of the section. Locally, the marble grades into calc-schists and in places conglomerate-type structures appear. Metavolcanic rocks occur as layers within the section, with the thickest horizon being about 100 m thick. These metavolcanic intercalations are composed of greenschists, and partly transformed blueschists and eclogites. III. Concordantly overlying the Main Marble unit is the high pressure Blueschist and Eclogite sequence. The sequence is about 800 m thick and almost exclusively consists of well preserved blueschists, eclogites, jadeite gneisses and metasediments. These high-pressure eclogite facies metamorphic rocks are characterized by a variety of rock types in which the dominant feature is the interlayering of acid and basic metavolcanics. Layering on the scale of tens of cm indicates that some of the rocks had a pyroclastic origin. A marble layer and horizons of metapelites occur within the section, together with a few layers of quartzites and metacherts and a small outcrop of the iron amphibole, deerite ($Fe_6(FeAl)_3Si_6O_{20}(OH)_5$)-bearing meta-ironstone. IV. The uppermost unit of the island, concordantly overlying the Blueschist and Eclogite unit, is the 300mthick Upper Marble Unit, consisting of sequences of calcite and dolomite marbles. Near the top of this sequence the marbles are composed of calcite pebbles, presumably of clastic origin. Interlayered with the marble is a several meter thick coarse grained quartzite horizon of sandy origin. Small remnant outcrops of blueschist metavolcanics and metasediments are preserved above the marble. Generally, they have a less well preserved high-pressure mineralogy compared to the underlying Blueschist and Eclogite sequence.

2.2 Metamorphic Evolution

From the perspective of metamorphic evolution, the stratigraphic units on Sifnos can be subdivided into two categories. The Upper Marble and Blueschist-Eclogite units in the north of the island are dominated by rocks of the high pressure metamorphic facies. Schistose rocks within the Main marble and the underlying Central Sifnos Greenschist unit mainly contain assemblages showing transformation from high pressure to greenschist facies conditions. The mineral assemblages resulting from the combination of the Eocene high-pressure metamorphism and its subsequent Oligocene Miocene overprint are a function of pressure, temperature, bulk chemical composition and the composition of the fluid phase. Bearing in mind the need to define these parameters, we will examine the metamorphic evolution on Sifnos from three aspects: a) the petrology of the high-pressure metamorphic rocks of Northern Sifnos; b) petrology and petrography of the Greenschist unit of Central Sifnos; c) petrology of the Eclogite-Blueschist-Greenschist transformation and d) evidence for fluid infiltration on Sifnos.

2.2.a Petrology of the high-pressure metamorphic rocks of Northern Sifnos.

A variety of metamorphic rock-types occur including blueschists, mica and glaucophane-bearing eclogites and actinolite-bearing metabasites, jadeite gneisses and deerite bearing meta-ironstones. Metamorphic assemblages of the various rock-types are presented in Table 1. Minerals considered to be typical of high-pressure metamorphism, glaucophane, omphacite, jadeite, the

TABLE 1. Mineral Assemblages of the Eocene High Pressure Metamorphic Rocks of Sifnos (After Schliestedt *et al.*, 1987).

Metabasites	
Eclogites (± Qtz, Sph)	Om-Ph-Ga-Gl-Ep Om-Ga-Ep
Blueschists (± Qtz, Ru, Ph)	Gl-Ep-Ga-Om-Pg Gl-Ep-Ga-Ctd-Pg
Act-bearing metabasites (± Qtz, Sph)	Om-Act-Ga-Gl-Ep-Ph Act-Chl-Ep±Gl±Ph
Meta-acidites	
Jadeite gneisses (± Ru, Ph, Ep)	Qtz-Jad-Gl-Ga-Pg±Ctd
Metasediments	
Marbles	Cc±Qtz±Ph±Ep±Gl Cc-Dol-Qtz-Ph
Quartzites	Qtz-Ga; Qtz-Gl; Qtz-Om-Ep Qtz-Dee-Mt-Ga-Aeg-Cro-Cum-Act
Metapelites	Qtz-Ph-Ga-Gl-Ep±Ru±Sph±Cc

Abbreviations: Ab - albite; Act - actinolite; Aeg - aegirine-augite; Barr - barroisite; Car - carpholite; Cc - calcite; Clz - clinozoisite; Cm - chloromelanite; Cor - corundum; Cro - crossite; Cum - cummingtonite; Ctd - chloritoid; Dee - deerite; Dia - diaspore; Dol - dolomite; Ep - epidote; Ga - garnet ; Gl - glaucophane; Jad - jadeite; Ky - kyanite; Law - lawsonite; Mt - magnetite; Mu - muscovite; Om - omphacite; Pg - paragonite; Ph - phengite; ps - pseudomorphs; Qtz - quartz; Ru - rutile; Sil - sillimanite; Sp - spinel; Sph - sphene; St - staurolite.

rare amphibole deerite and the 3T polytype of phengite rich in the celadonite component occur, but lawsonite is notably absent. The occurrence, instead, of pseudomorphs after lawsonite, composed of paragonite + clinozoisite + quartz, as inclusions within the garnets of jadeite gneisses, and the ubiquitous presence of glaucophane-epidote in many rock-types, point to metamorphic conditions typical of the higher temperature parts of the blueschist-facies. However, the presence of eclogites indicate that eclogite facies conditions are attained. Schliestedt (1986) has demonstrated that eclogites, blueschists and actinolite-bearing metabasites represent basic volcanic rocks of different chemical composition, which have recrystallized under the same metamorphic conditions.

There has been some debate in the literature concerning the definition of the blueschist facies. We adopt here the comprehensive definition (Evans and Brown, 1987) which includes assemblages containing both glaucophane + lawsonite and/or glaucophane + epidote. The latter is suitable for Sifnos rocks; however, Evans and Brown (1987) further suggest that where the blueschists are interlayered with omphacite garnet rocks (eclogites), the conditions are better defined as eclogite facies. This eclogite facies definition is most appropriate for the metamorphic evolution on Sifnos. The transformation from high-pressure-eclogite facies to greenschist facies involves an intermediate phase of blueschist facies equilibration in which glaucophane-epidote albite assemblages develop as an overprint of the original high pressure assemblages.

One of the remarkable features of the high-pressure assemblages is their fine state of preservation. Retrogression is generally minor and mainly expresses itself by the presence of calcite instead of aragonite (the stable polymorph of $CaCO_3$ at eclogite facies conditions) and the breakdown of jadeite+quartz to albite. This breakdown frequently went to completion with large porphyroblasts of albite replacing jadeite, but sometimes only reaction rims of albite and aegirine containing about 10 mol% jadeite developed between jadeite and quartz (Okrusch et al., 1978). In metabasic and metasedimentary rocks partial replacement of garnet by chlorite is observed and albite-chlorite forms as the breakdown products of omphacite + paragonite in eclogites of glaucophane in blueschists.

Estimates of the P-T conditions for the high pressure metamorphism are summarized in Fig. 2.3. Temperature ranges of 440-500°C are deduced by oxygen isotope thermometry, Mg/Fe exchange thermometry between garnet and omphacite in eclogites and solvus thermometry based on the $MgCO_3$ content of calcite coexisting with dolomite (Schliestedt 1980, 1986; Matthews and Schliestedt, 1984). A particularly reliable temperature estimate is 470±25°C deduced by Schliestedt (1986) from well constrained K_D values in the range 21-27 (Fig. 2.3) for garnet-omphacite Mg/Fe exchange. Further support for these temperatures comes from the presence of paragonite-clinozoisite-quartz pseudomorphs after lawsonite in jadeite gneises (i.e. conditions above curve 4 in Fig. 2.3) and deerite in meta ironstones (curve 2). Lower and upper limits of pressure are given by the occurrence of almost-pure jadeite + quartz and paragonite instead of omphacite + kyanite (curves 1 and 3

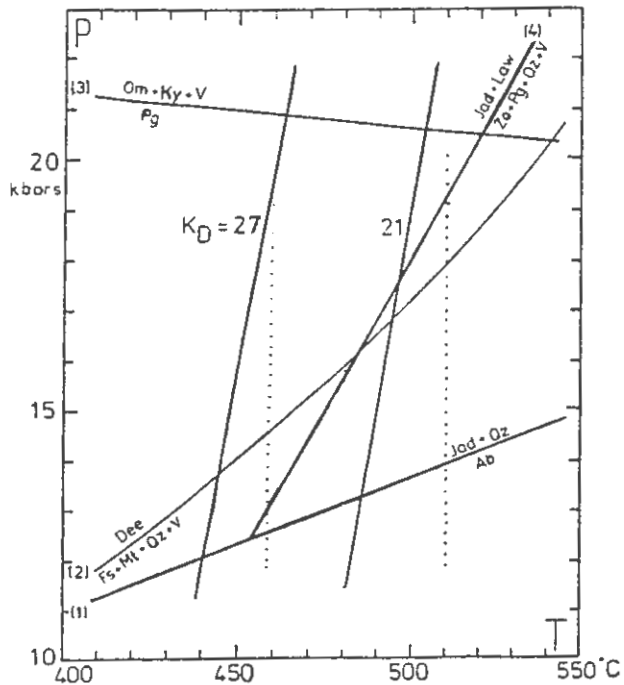


Figure 2.3 P-T diagram showing estimates of the physical conditions for the high-pressure metamorphism on Sifnos. K_D values are for Mg/Fe exchange between garnet and omphacite. Dotted lines give ranges of oxygen isotope temperatures revised after Matthews (1988). Source of diagram: Schliestedt et al., 1987.

respectively). These pressure limits of 12-20 kbar are further constrained by the presence of the assemblage deerite-magnetite-crossite for which Evans (1986) calculated a pressure range of 14-18 kbars, corresponding to a depth of burial of ~50-65 km at a nominal rock density of 2.75.

The occurrence of pseudomorphs after lawsonite and H2O-rich fluid inclusions point to a very low mole fraction of CO2 in the fluid phase (Schliestedt and Matthews, 1987).

Trace element abundances (Schliestedt, 1980) are consistent with the parent rocks being tholeitic basalts for the blueschists and calc-alkaline basalts with shoshonitic affinities for other metabasic rocks. $\delta 180$ analyses of metabasic rocks are generally in the range 9.5 to 11.5o/oo. Comparing this to the range of 5.5 to 7.5o/oo characteristic of by fresh basalts, suggests that the basic rocks undergone low-temperature sea-water alteration process (spilitization) which raised $\delta 180$ to the present values, prior to subduction and burial. High sodium contents observed in jadeite gneisses can also be interpreted as the result of low temperature sea-water alteration of parent rhyolites (Schliestedt and Okrusch, 1988).

2.2.b Petrology and petrography of the Greenschist unit of Central Sifnos.

In the two lowest stratigraphic units on Sifnos, we have noted that schistose rocks generally reveal evidence of the transition from higher to lower pressure conditions as a result of Oligocene-Miocene overprint. The process of transformation

ultimately results in the formation of classical greenschist facies assemblages: albite-epidote chlorite-actinolite and albite-epidote-calcite-actinolite assemblages in metabasics and albite-quartz-muscovite chlorite in metasediments.

Relict blueschist minerals and mica-rich eclogites layers are found within the Central Sifnos Greenschist unit. Relict glaucophane frequently occurs within albite and epidote crystals. It is, however, more important to note that most of the glaucophane in the Greenschist unit is not a relict of the Eocene high-pressure metamorphic phase, but has formed during uplift as fresh porphyroblastic minerals. Texturally, these porphyroblastic glaucophanes are in stable coexistence with epidote and new albite grains. Rocks with such parageneses represent an intermediate stage of epidote blueschist facies equilibration during the transformation from the eclogite facies conditions of the Eocene metamorphic phase.

The later transformation to greenschist facies rocks is evidenced by the chloritization of the glaucophane.

A few partly retrogressed eclogites are interlayered within and grade into greenschists and glaucophane-albite blueschists near the town of Apollonia. Texturally these rocks also show evidence of the transformation to lower pressure epidote blueschist facies conditions with omphacite+paragonite breaking down to give glaucophane+albite. The glaucophanes in these eclogites then show evidence of further breakdown, either with the development of calcic amphibole rims or chloritization. The eclogites are relics of an earlier high pressure metamorphic phase.

Conditions of final equilibration in the greenschist facies are a little more difficult to define. Isotopic temperatures range from 400 -480°C and consequently are quite similar to those of the high pressure metamorphism. From an analysis of phase equilibria on Sifnos and other islands, Schliestedt et al. (1987) concluded that temperatures of 400- 450°C and pressures of 5-7 kbar would be the most reasonable estimate for the regional greenschist facies metamorphism. Aegirine rims on jadeites, containing about 10 mol% jadeite would also be consistent with final equilibration at about 5 kbar.

2.2.c Petrology of the Eclogite-Blueschist-Greenschist Transformation

The various mineral parageneses observed in metabasic rocks during the transformations from eclogite facies conditions through blueschist to greenschist, allow the definition of a representative chemographic sequence. The sequence deduced by Schliestedt and Matthews (1987) is given in 2.4 and consists of four chemographic triangles with components A (=Al+Fe3+) - NA (=Na(Al+Fe3+)) and CFM (= Ca(Mg+Fe2+)). The mineral phases are projected into the triangles from epidote, quartz, H2O and CO2. Triangles a and d represent coexisting mineral assemblages for the eclogite and greenschist facies conditions, respectively. The intermediate epidote-blueschist facies assemblages are represented in triangles b and c. The sequence of reactions compatible with the diagrams is given in the caption to the figure commencing with reactions involving the breakdown of omphacite (such as the reaction of omphacite + paragonite) followed by the breakdown of glaucophane

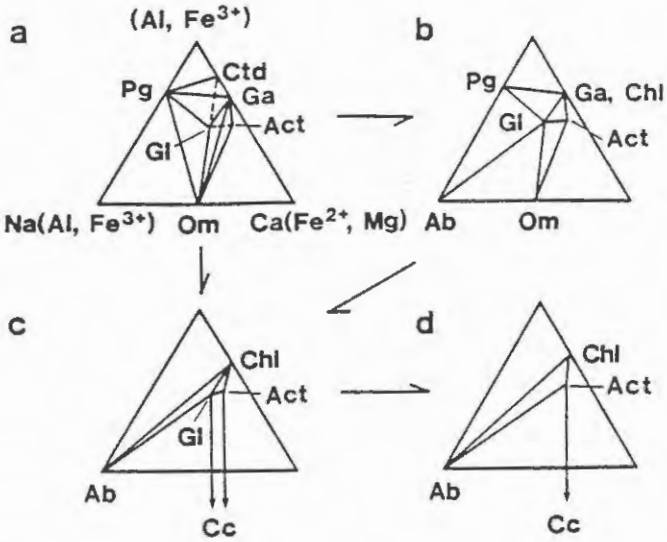


Figure 2.4 Chemographic diagrams illustrating the breakdown from high pressure eclogite facies conditions (Triangle A) through intermediate blueschist facies (B & C) to greenschist facies assemblages (D). Reactions are: A ---> B: om + pg = abl + gl + ep + H_2O; gl + ctd + H_2O = ab + chl + ep ; B --> C: gl + pg + H_2O = ab + chl + q; om + H_2O + CO_2 = ab + chl + cc + ep; ga + H_2O = chl + ep + q; C ---> D: gl + ep + H_2O = ab + chl + act. Mineral abbreviations as in Table 1. From Schliestedt and Matthews (1987).

bearing assemblages to give greenschists.

Calculated P-T-XCO2 conditions for the equilibration of the intermediate blueschist facies assemblages are plotted in Fig. 2.5 a and b. The calculations (and the previous PT estimates based on the jadeite composition in pyroxenes) were made using the Thermocalc thermodynamic dataset of Holland and Powell (1985) and Powell and Holland (1988) with added data for glaucophane (Holland 1988). In Fig. 2.5 we see four equilibrium curves for hydration reactions involving the equilibration of omphacite and glaucophane bearing assemblages at XCO2=0. The reactions and the transitions they are equivalent to in the chemographic sequence are:

1. 2di+11gl+8H2O=28abl+6clin+1tr (a -> c)
2. 1jd+5gl+3pa+4H2O=14abl+3clin (a -> c)
3. 5gl+3pa+4H2O=13abl+3clin+1q (b -> c)
4. 25gl+6cz+7q=50abl+3clin+6tr (c -> d)

As is evident from the diagram, the reactions which are calculated using actual chemical compositions of Sifnos minerals, are stable at similar P-T conditions. At temperatures of about 450-500°C, which would be appropriate for the reaction conditions on Sifnos, pressures of around 8 to 10 kbar appear reasonable.

Fig. 2.5b shows a number of equilibria involving H2O and CO2 on a diagram of P vs XCO2 at 450°C. Of these perhaps the most revealing is the greenschist facies reaction 7: (stoichiometries of 5, 6, see caption to Fig. 2.5).

7. 3tr+2c+8H2O+10CO2=3clin+10cc+21 q

Since mineral assemblages on both sides of this equation exist in Sifnos greenschists, it appears that there is a slight increase in

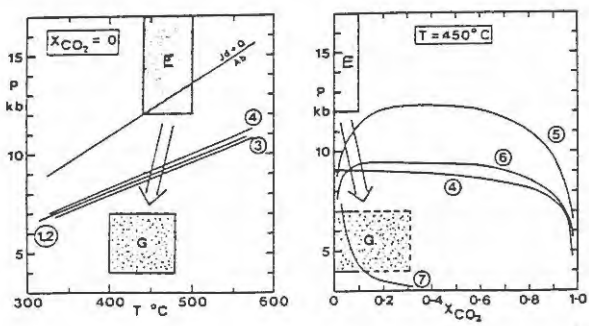


Figure 2.5 P-T at XCO_2 = 0 and P-XCO_2 at 450°C (XCO_2 = mole fraction of CO_2 in the fluid phase) showing the transformation from eclogite facies (E) to final greenschist facies conditions. The equilibrium reactions 1 to 7 are calculated using the thermodynamic data set of Holland and Powell (1985) and data for glaucophane (Holland, 1988). Reactions 1-4 and 7 are detailed in the text;

5: gl + 2 cz + 6 H_2O + 4 CO_2 = 10 abl + 3 clin + 4 cc + 7q

6: 15 gl + 4 cz + 10 H_2O + CO_2 = 30 abl + 6 clin + 3 tr +2cc.

Note that for hydration reactions like 1 to 4, changes in XCO_2 would not much affect equilibrium T conditions, whereas the occurrence of hydration-carbonation reactions 5-7 could be significantly influenced by changes in XCO_2.

Notation: di = diopside; pa = paragonite; abl = low albite; clin = clinochore; tr = tremolite; cz = clinozoisite; jd = jadeite; cc = calcite; q = quartz. Diagram is from Figure 6 of Schliestedt and Matthews (1987) where details of compositions and activities can be found.

XCO2 of the vapour phase during the transformation from high-pressure to greenschist facies metamorphic conditions.

The P-T conditions deduced for the eclogite facies, greenschist facies and epidote-blueschist facies equilibrations present a clear picture of the P-T path the uplift history of the Sifnos units. Following an isotopic and petrological equilibration at 40-45 Ma during tectonic burial to depths of around 50 km or more and at temperatures of 450-500°C, the rocks were virtually isothermally uplifted to pressures of 10 through 5 kbar where blueschist and greenschist facies assemblages developed 20 Ma later. This type of near isothermal uplift path is consistent with the general form of the calculations of England and Thompson (1984) for pressure-temperature-time paths in continental crust thickened by underthrusting and exhumed by erosion. However, there are still a number of problems with the P-T time path of Sifnos, such as the extreme rapidity of erosion implied by a 20 my timespan and the question of where all this eroded material has disappeared to, which suggest that other, tectonic, factors may also have been involved in the uplift process (see introductory section 1).

Another point which should be appreciated from the reactions of Fig. 2.5 is that blueschists and greenschists sensu stricto may coexist. The classical blueschist greenschist reaction is reaction 4 in Figure 2.5. It is a continuous or sliding reaction, whose P-T conditions depend on the bulk rock chemical compositions; a small difference in the chemical compositions of rocks at a certain P-T value could mean that in one rock glaucophane-epidote assemblages are stable while in another the greenschist minerals: albite

chlorite-actinolite are completely formed. Blueschists and greenschists can therefore, under a certain range of P-T conditions, coexist as, say, compositionally different interlayered rocks, in the same way as at higher pressures blueschists and eclogites coexist on Sifnos. And, as in the latter case, where conditions are discribed as eclogite facies, so too where blueschists and greenschists are interlayered, the conditions should be described by the higher grade facies assemblage - i.e. as epidote blueschist facies.

2.2.d Evidence for fluid infiltration on Sifnos

The petrographically observed reactions defining the transformation from eclogite to blueschist to greenschist facies conditions are, with very few exceptions, reactions of hydration and carbonation. On average, the H2O content of basic rocks is 1.2 wt.%, higher in greenschist facies rocks compared to eclogite facies rocks (Schliestedt and Matthews, 1987). Hydration requires fluids and the question arises from where do these fluids derive? There is evidence that externally derived fluids may have infiltrated into the schistose rocks within the Main Marble and the Central Sifnos Greenschist Unit (but not the eclogite facies rocks of northern Sifnos) producing the hydration reactions we have described previously. The evidence is mainly from oxygen isotopes and we will summarize it briefly.

The critical aspect of the oxygen isotope data on minerals and whole rocks on Sifnos is that whenever chemically equivalent high pressure eclogite facies and blueschist/greenschist facies rocks are

compared, the $\delta 180$ of the latter are always higher. This can be seen in Fig. 2.6, which plots $\delta 180$ compositions of whole rocks from the high pressure metamorphic rocks at Vroulidia Bay (station 3, excursion guide), greenschist facies rocks within the main marble at Kamares Bay (station 9, excursion guide) and epidote-glaucophane, blueschists and greenschists from the Central Sifnos. The $\delta 180$ compositions are plotted against an index of chemical composition which is necessary for comparing rocks with different chemical compositions (loosely, the more silica rich a rock the more it will concentrate 180 in a series of rocks formed under the same conditions - see Schliestedt and Matthews, 1987, for an explanation). The diagonal lines in the figure show calculated $\delta 180$ compositions of waters in equilibrium with rocks of different chemical composition at an assumed temperature of 450°C. Since a series of interlayered rock in isotopic equilibrium with one another will also be in equilibrium with same fluid, a series of rocks in equilibrium should give a trend parallel to one of these lines, the slope of which depending on temperature. It can be seen that given 450°C as a reasonable temperature the expected trends are observed for the interlayered rocks at Vroulidia and Kamares Bays and less so by the Central Sifnos rocks, which are more randomly sampled.

The Kamares Bay and Central Sifnos samples are enriched in 180 relative to the Vroulidia high pressure samples. If the former were originally compositionally similar to the Vroulidia Bay samples, then evidently the Oligocene-Miocene metamorphic overprint resulted in an increase in $\delta 180$. This could only come about by the introduction of an 180-rich component; a fluid enriched in 180.

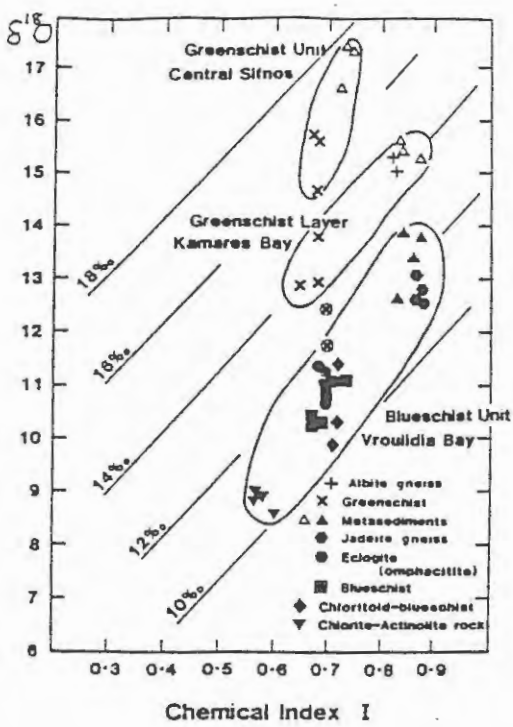


Figure 2.6 Diagram of $\delta^{18}O$ against the Garlick (1966) chemical index I from (a) high pressure metamorphic rocks of the Blueschist/Eclogite Unit at Vroulidia-Bay. (b) greenschist facies layers within the Main Marble Unit at Kamares Bay and (c) the Central Sifnos Greenschist Unit. The diagonal lines show isotopic compositions of waters in equilibrium with whole rock samples at an assumed representative temperature of 450°C. Samples indicated by the ⊗ symbol are relict eclogites within the greenschist layers at Kamares Bay. Diagram after Schliestedt and Matthews (1987).

This process is exemplified by the analyses of two eclogite relics within the Kamares Bay layer, which are significantly lower in $\delta^{18}O$ than greenschists in the section, and thus still preserve the lower $\delta^{18}O$ of the high-pressure metamorphism.

There is little problem in enriching a fluid in 180 in the Sifnos section. The calcites of the marble units have $\delta^{18}O$ compositions in the range 24-28‰ and $\delta^{13}C$ from 0 to +4‰. In other words, they still retain the isotopic compositions they had as marine sedimentary precursors. Any fluid which passes through these marbles at temperatures of $\sim 450^{\circ}C$ would be enriched in 180. The isotopic results thus lead to the concept that the 180 enrichment on Sifnos was the result of the infiltration of 180 enriched fluids, whose isotopic compositions were controlled by exchange with Marble units. With this model, one can calculate that the open-system minimum Water/Rock ratios for an infiltrating fluid would be approximately 10 and 20 wt% respectively for the Kamares Bay layer and Central Sifnos samples. A corollary of this argument is that the high-pressure metamorphic rocks of Northern Sifnos, which were not affected by the overprint, also did not see the fluids that brought about the hydration reactions and isotopic enrichment.

The question arises as to the provenance of these fluids. Metamorphic dehydration fluids derived from deeper levels below Sifnos, fluids from a crystallizing granite at depth, mixing of previously immiscible fluids in carbonate and schistose rocks are all potential sources.

2.3 Structure and Tectonics

The study of the kinematics of subduction and uplift processes during orogenic evolution mainly relies on study of the geometrical relations and spatial distributions among preserved structural features. The tectonic structures observed in the Eocene high pressure rocks of Sifnos were developed during their burial and metamorphism in the collision events. It follows that the metamorphic structures in the high pressure sequence may provide important insights into style of deformation and characteristic strain features developed in the underthrust plates of convergent plate margins. On the other hand, post metamorphic structures or structures developed at lower pressures may supply evidence of the nature of the uplift and exhumation processes of such blueschist terrains.

Syn-metamorphic strain in the subducted rock mass of Sifnos resulted in folding and the development of a penetrative layer parallel cleavage. Thrusting related to subduction was not observed on Sifnos but its role as a thickening mechanism in other segments of the high pressure belt has been demonstrated on a number of islands (e.g. Naxos, Ios, Syros, Andros and possibly, Tinos).

The penetrative cleavage on Sifnos is developed in all rock types and is microscopically defined by synkinematic growth and recrystallization of the high pressure minerals. The cleavage planes usually trend to the north, and dip as much as 40°. The precursor layering was rotated, flattened and transposed parallel to the cleavage. Hinges of folds to which the cleavage is axial

planar, were not found although repetition of layering is locally observed. This indicates that the cleavage was formed as a result of a strong isoclinal folding (D0)and that the hinges were flattened out. The scale of this early phase is not yet clear. Macro (island) scale stratigraphic inversion does not occur. Flattened calcite pebbles found within the cleavage planes resemble pancakes indicating that flattening deformation played a role in the formation of the fabric.

The earliest observable fold axes were termed D1 and they clearly deform the main cleavage. D1 is mostly found in the northern part of Sifnos in the upper marble and the quartzites. The fold axes trend northeasterly; the fold hinges are on a meter-scale and are very tight to isoclinal. Most D1 folds are overturned to the SE but a few examples showing the opposite sense of vergence exist. Recrystallization of the high pressure minerals, glaucophane and omphacite, is observed to occur in the fold hinges. Locally glaucophane forms a mineral lineation parallel to the fold axes, whilst large glaucophane porphyroblasts also developed randomly on fold axes. This implies that the D1 phase affected the rocks at high pressures.

The glaucophane lineation is frequently parallel to the NE trending fold axes or parallel to associated microcrenulations. But on the limbs it may be transverse to or oblique to the axes. Thus the mineral lineation cannot be used as a clear kinematic indicator. No evidence has been found for the rotation of the fold axes into a parallel alignment with the mineral lineation and there are also no indications that the folds form part of sheath fold system.

The microstructural characteristics of the D1 folding phase show that it was produced during the collision event, while the rocks were undergoing high pressure metamorphism. Unless the orientation of the folds was modified in later stages of the tectonic history, the direction of the fold axes suggest that the high pressure rocks were compressed and metamorphosed between NW to SE converging blocks (Avigad 1987). Since folds of similar style, trend and relationship to metamorphic history occur throughout the Cyclades, we refer to D1 as the "Cycladic trend".

Most workers have regarded the subduction and collision processes of the Cycladic rocks as a part of the Alpine evolution within the Hellenides. The kinematics of the Hellenides is constrained by studies of nappe movements and the recognition of tectonic windows underneath allochtonous basement nappes on the mainland. Convergence and nappe propagation took place, between the internal Rhodope massif in the NE and the external Apulia subplate in the SW. The NW-SE compression direction deduced from the D1 structures on Sifnos (and from other islands in the massif) is normal to the direction of the Hellenides and parallel to the strike of the present mountain belt. This would suggest that the collision event forming the Cycladic rocks did not occur within the tectonic framework of the Hellenides. This interpretation was generally rejected and it has been preferred instead not to assign any kinematic significance to the D1 folds. The direction of convergence in the Cyclades has been deduced purely from the orientation of the glaucophane lineation. The lineation is considered to be the direction of stretching and is presumably

oriented perpendicular to the strike of the convergent boundaries. The NE trend of the lineation in the Cyclades would then suggest compression between NE and SW, in agreement with the observed movement in the Hellenides.

A second folding phase, D2, affected the rocks on Sifnos. The folds are isoclinal and their axes trend 020°. Hinges, from tens of meters and up to a hundred meters in size, are observed. A crenulation cleavage is developed, with marked recrystalliation and polygonization of the high pressure minerals. D2 is well developed in the high pressure sequence of northern Sifnos, in the main marble and also in the greenschist unit. Glaucophane porphyroblasts, previously noted to overgrow D1, also postdate D2. A glaucophane lineation (L2) trending 010-030° is locally well developed on cleavage planes and was formed during D2 phase. Considerable thickening accompanied the D2 phase, since parts of the stratigraphic sequence are clearly duplicated. Large albite porphyroblasts, which grew during the greenschist overprint, postdate D2. It should be noted that the D2 phase differs from D1 in its trend but relates in similar way to the metamorphic history. In the outcrop of Kamares bay a D1 hinge is observed refolded by D2 but both phases could be a result of a single continuous deformation.

A subsequent folding phase, D3, can be recognized in Sifnos, particularly in the center of the island in the area from Kamares to Platygialos. D3 axes trend NW, and the folds are recumbent and overturned to the NE, having axial planes which dip shallowly to the SW. Hinges are observed with wavelength of tens of meters. The

fabric formed is weak, and only a local fracture cleavage is developed around the fold hinges. Glaucophane has recrystallized parallel to but overprints D3 hinges. This glaucophane belongs to the intermediate phase of blueschist facies metamorphism accompanying the transformation from eclogite to greenschist facies conditions during uplift. The mineral lineation L2 is refolded by D3 folds in several localities. Albite poikiloblasts which were formed during the decompression are mostly unaffected by D3, but are sometimes slightly rotated or partly polygonized along fracture cleavage planes. It can therefore be concluded that this NW phase affected the rocks at pressures of the blueschist facies, close to the conditions of the greenschist facies. As Fig. 2.5 indicates, pressures of around 8 kbars are appropriate for this stage. D3 appears to have resulted from one of the tectonic processes which facilitated the rapid uplift of the high pressure rocks. The hinges of D3 observed in the center of the island are overturned to the NE in the lower part of the section (greenschist unit) and to the SW in the main marble on top. The D3 trend on Sifnos is parallel to the major Tertiary fold axes in the Hellenides and thus we refer to it as the "Hellenic trend".

The greenschist facies overprint was not accompanied or followed by any ductile deformation. The overprint is dominated by static and mimetic growth of the greenschist mineralogy. The three observable phases of deformation were thus confined to a period of aproximately 20 m.y., between the middle Eocene high pressure metamorphism and the late Oligocene-early Miocene greenschist overprint. Plastic flow thus occurred in the pressure range of 15

kb to 8kb. The fact that the ductile deformation did not take place while the rocks were at shallower levels in the crust (e.g. at 20 km during the greenschist phase) suggests that 20-25 km could be approximately the level of the brittle-ductile transition in the Cyclades, and that deformation at shallower depth occurred by brittle fracturing and movement along faults. The brittle-ductile transition is normally considered to occur at depth of 10-15 km. The depression of the brittle-ductile transition to a greater depth is consistent with the low geothermal gradients of the regional metamorphic evolution.

2.4 EXCURSION GUIDE FOR SIFNOS

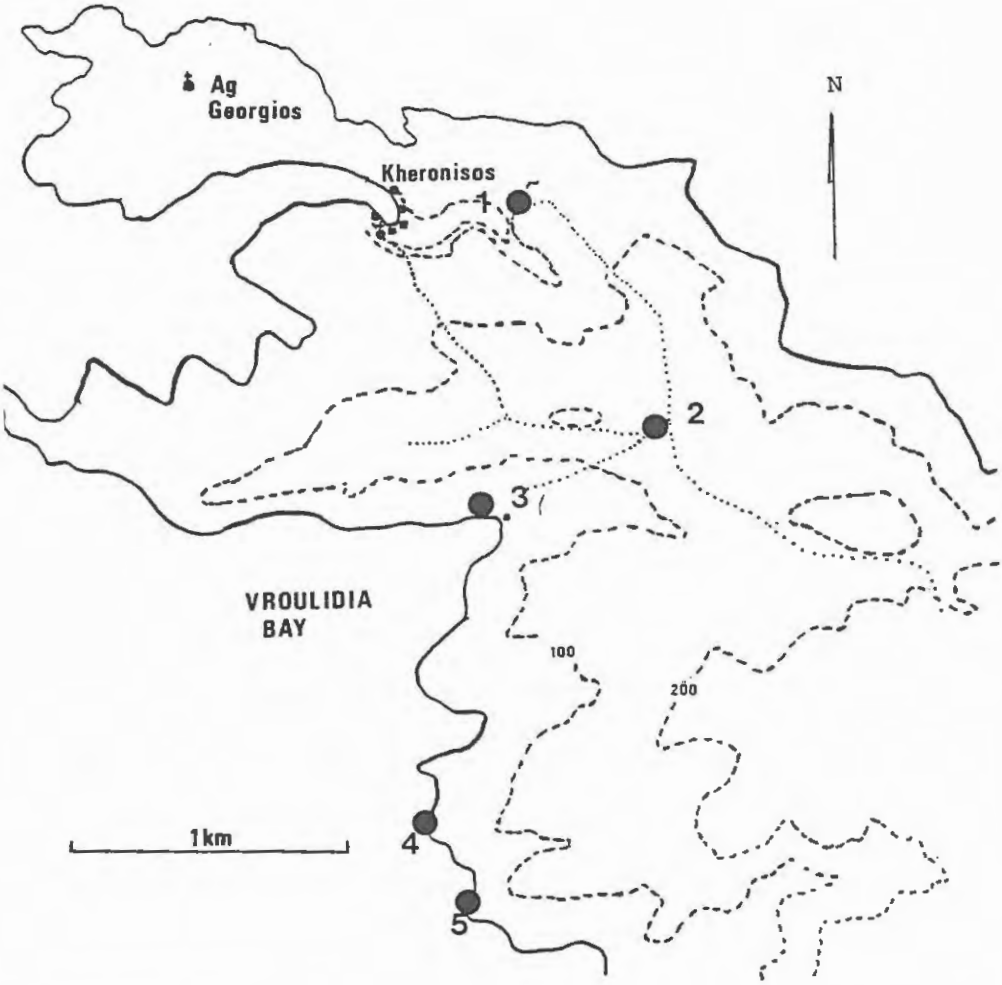
DAY 1: THE HIGH PRESSURE ROCKS OF NORTHERN SIFNOS: STRATIGRAPHY, MINERALOGY AND STRUCTURES. LOCATION:MAP 1. FROM THE VILLAGE OF KHERONISOS TO VROULIDIA BAY.

Way to station 1. The track starts at the village of Kheronisos. Walk south into the main valley just behind the village. Go past a dolomite block on your left and climb the eastern marble wall of the wadi, which is several tens of meters high. Continue walking south on a goat path to station 1.

Station 1. The upper marble section and interlayered quartzite.

The uppermost part of the stratigraphic section of Sifnos is made of a 300 meter-thick marble unit containing intercalations of micaschists. The carbonate rocks include particularly well layered gray and white calcite marbles. These marbles form the northern wall of Kheronisos Bay and mark the top of the carbonate section.

Carbonate fragments of clastic origin are abundant in the gray layers near the church of Ag. Georgios. Beneath these layers, forming the rocks on which Kheronisos village is built and most of the southern part of the bay, is a sequence of brown dolomites tens of meters thick. Below this dolomite, the lowest part of the marble section is composed of blue-gray and white calcite marble containing numerous horizons of brownish dolomite. A few quartzite horizons occur within the carbonates and Station 1 is located on one of them. The quartzite here is about 3 meters thick and can be traced along



LOCATION MAP 1

strike to the western side of the island. The quartzite is impure, containing various amounts of white mica, and is most probably a metasandstone. The quartzite is folded by the first folding phase, D1. The fold is defined by the more quartz rich layers occuring within an impure quartzitic matrix. Folding is enhanced at the contact between the quartzites and the overlying marble as a result of the differences in competence. The fold hinges observed at this stop are tens of centimeters in size. The axes trend 060°, and the folds are tight to isoclinal, assymetric, and overturned to the SE. The hinges were thickened by plastic flow as is evident by a preferred crystallographic orientation of the quartz. The axial plane is subhorizontal and almost parallel to the schistosity. The schistosity in the area is parallel to the bedding planes and refolded by D1. Microstructural investigations of D1 fold hinges from other localities in Sifnos show that high pressure minerals, including omphacite, grew during and subsequent to D1. Thus both cleavage formation (in which high pressure minerals crystallized) and D1 occurred under high pressure metamorphic conditions and it is not considered that any metamorphic or time discontinuity separates the two phenomena.

Way to Station 2. Walk South along the marble cliff then follow the stone wall to the intersection of path above Vroulidia Bay. Station 2 is located in the north part of the saddle above Vroulidia Bay near a waterhole. On the way we cross the contact between the marbles and the underlying gneisses and schists of the blueschist-eclogite unit. The sequence of the blueschists, eclogites and acid gneisses is layered on a scale of centimeters to

several meters. A glaucophane lineation trending NE is observable, particularly in acid gneisses.

Station 2. Jadeite-bearing acid gneisses.

At this station we observe a thick horizon of light coloured acid gneiss interlayered with blueshists and eclogites. The acid gneiss contains quartz, jadeite, garnet, glaucophane, epidote, phengite and albite. The jadeite which is well preserved in this outcrop, is almost an end member composition.

In thin section it is observed that jadeite and quartz were recrystallized during the formation of the (S1) schistosity and, at pressures greater than 12 kb, developed during the subduction related deformation. Albite grew later at the expense of jadeite along its contacts with quartz, indicating that jad+q=ab equilibrium curve was crossed from the high pressure to the low pressure side, and that albite was generated during uplift.

Way to Station 3. Enter through the gate in the SW stone wall and descend 250 m along a goat path to Vroulidia Bay. Just before reaching the water front and about 30 m north of the deserted fishermans house is a series of interlayered eclogites, metasediments and acid gneisses. This series forms part of a sequence of interlayered rock types occuring on the northern edge of Vroulidia Bay. The sequence is illustrated in Fig.2.7, where it is subdivided into 12 rock units, of which unit 12 is the one mentioned above. To get to the remaining units, climb up approximately 15 meters from the water level (the rocks here are blueschists of unit 11), walk above the cliff (eclogite unit 12) for about 50 meters,

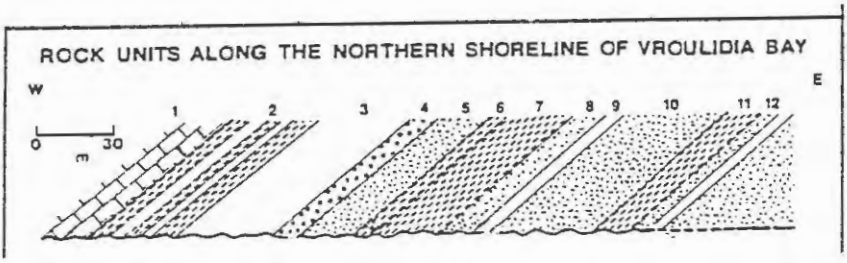


Figure 2.7 Interlayered eclogite facies rock units along the north shore and immediate hinterland of Vroulidia Bay (station 3). 1. Upper Marble; 2. Interlayered Blueschists and Jadeite Gneisses; 3. Jadeite Gneiss; 4. Micaschist with Calcite Lenses; 5. Coarse-grained Eclogite; 6. Chloritoid-bearing Blueschist; 7. Blueschists with Minor Eclogites; 8. Chlorite Actinolite Schist; 9. Garnet Micaschist;10. Eclogites;11. Blueschist; 12. Interlayered Eclogites and Metasediments. Diagram after Schliestedt and Matthews (1987).

then descend once again down to the water front along the northern side of the bay. This bring us down to the area of the contact between the units 10 and the garnet micaschist of unit 9. Continuing (clambering) along the water front we pass successively through interlayered units 9 to 1.

Station 3. A cross section of the high-pressure metamorphic rocks on the Northern side of Vroulidia Bay.

The various rock types presented in the schematic cross section are:

12 - interlayered eclogites and metasediments. The eclogites are greenish rocks, mainly composed of omphacite, with garnet, phengite, epidote, glaucophane and quartz. The metasediment mineralogy includes phengite, jadeite-acmite, garnet, epidote, glaucophane, calcite and rutile, and is often crenulated. Albite is present as late porphyroblasts overgrowing the microfolds.

11 - blueschists. Metabasic rocks whose mineral assemblage consists of glaucophane, epidote, garnet, phengite, rutile, sphene and magnetite. As we noted earlier, although these rocks are typical high temperature epidote blueschists, conditions of metamorphism are best described as eclogite facies.

10 - eclogites. Garnet-omphacite thermometry of eclogites from Vroulidia Bay gave average temperatures of 470°C . The presence of mica in the eclogites also indicates that they were formed in the presence of a aqueous metamorphic fluid.

9 - garnet micashist. The micas are mainly phengites. Some are uniaxial, belonging to the 3T high pressure structural polymorph.

8 - chlorite actinolite schist. This is an unusual rock type with a unique mineralogy which is not very typical of the high pressure metamorphic facies. Both actinolite and chlorite in this rock are primary high pressure minerals showing typical early microstructures. Chemically the rock is almost ultrabasic, being unusually rich in Mg.

7 - blueschist with minor eclogite intercalations.

6 - chloritoid bearing blueschist. Chloritoid is not particularly abundant in the blueschists of Sifnos. Here it is restricted to a specific horizon where it overgrows garnet porphyroblasts.

5 - coarse grained eclogites.

4 - micaschist with calcite lenses. White calcite lenses are embedded in a micaceous matrix. They may have been derived from earlier clasts or from the boudinage of a continuous layer. The lenses are flattened within the schistosity and frequently resemble a pancake with no preferred elongation direcion. (This rock unit is a particularly good marker if you get lost in the section).

3 - jadeite gneiss. Most of the jadeite in these gneisses has retrogressed to albite during the uplift.

2 - interlayered blueschist and jadeite gneisses. The uppermost part of the section is made of alternating layers of blueschist and jadeite (now mostly albite) gneisses. The layers are a few tens of centimeters thick and it is possible that the protoliths were tuffs or volcanoclastics.

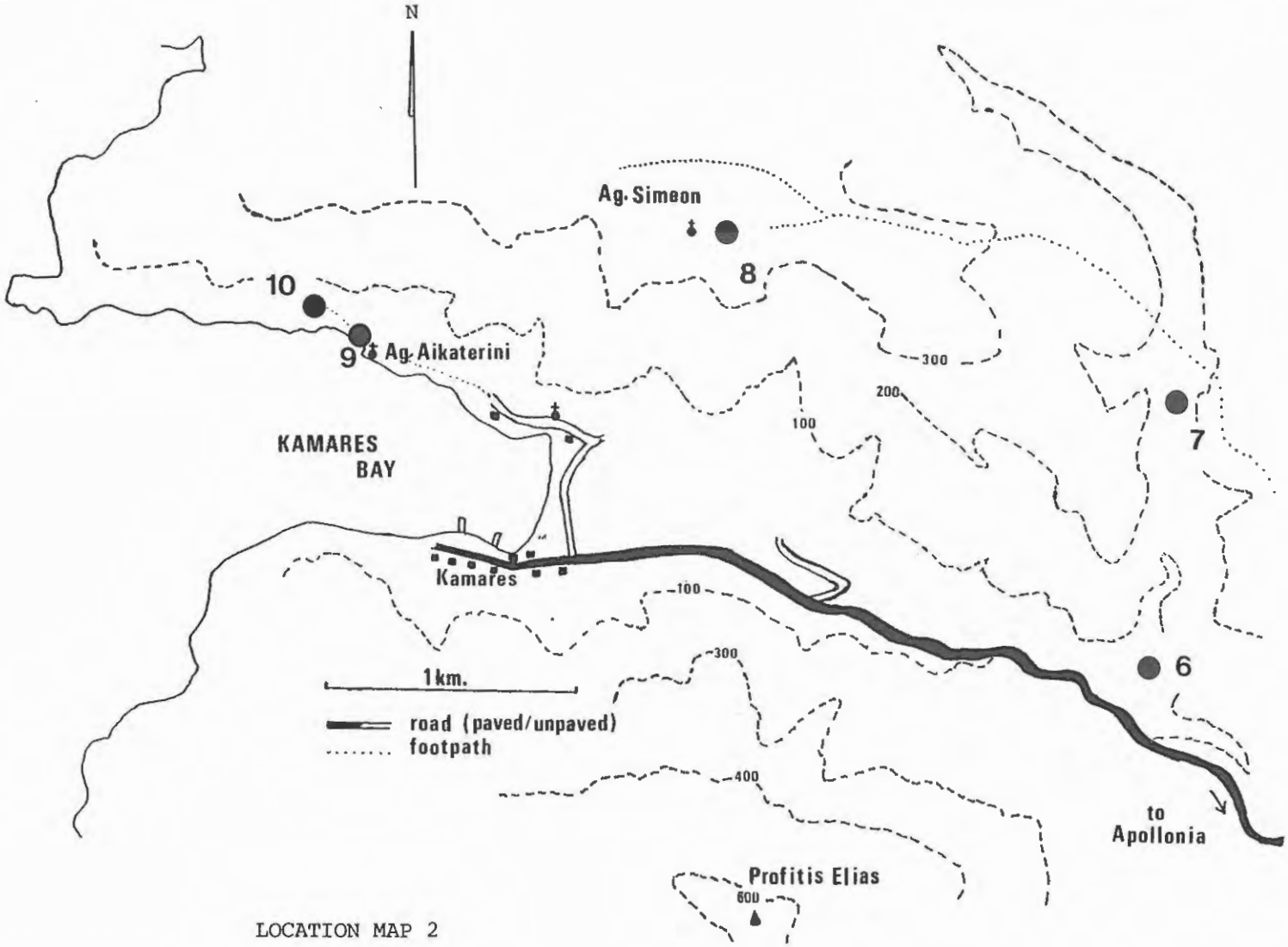
1 - upper marble.

The Vroulidia Bay section thus seen to contain a wide range of lithologically and chemically different rocks. Particularly important is the interlayering of acid and basic rocks and the occurrence in the sequence of different metabasic rock types. All the different high pressure mineralogies of these rocks were formed under similar P-T conditions, as Schliestedt (1986) demonstrated.

Following the visit to Vroulidia Bay we return to Kheronisos. Climb back up to the jadeite gneiss and at station 2 take the path to the west, following the western ridge down to the village.

On the return back to Kamares by boat, D2 isoclinal fold hinges in blueschists along the coastal cliffs may be observed at stations 4 and 5 (location map 1).

DAY 2 THE GREENSCHIST UNIT OF CENTRAL SIFNOS. STRATIGRAPHY AND STRUCTURES IN THE MAIN MARBLE HORIZON. PETROLOGY OF BLUESCHIST GREENSCHIST TRANSFORMATION. LOCATION: MAP 2. FROM THE MAIN KAMARES VALLEY INTO THE WADI TO AG. SIMEON, BACK TO KAMARES AND ALONG THE NORTHERN SIDE OF KAMARES BAY.



LOCATION MAP 2

Walk up the main valley from Kamares up till the entrance of the wadi shown in the location map 2.

Station 6. The lower parts of the wadi to Ag. Simeon.

Rocks exposed on the sides of Kamares Valley and at the entrance to the wadi have greenschist facies mineralogies. At the station, the rocks are composed of albite, muscovite, chlorite, epidote and quartz. No high pressure minerals are found. The rock contains abundant albite xenoblasts, which are undeformed and post kinematic. The schistosity is defined by the preferred orientation of the mica.

A few tens of meters into the wadi, a fold hinge trending 320° are exposed. The axial plane dips 30° to the SW, and sets of new fracture cleavages are developed parallel to it. The second cleavage is well developed in rocks rich in mica. The albite blasts were not polygonized during the formation of the second cleavage but have been sometimes slightly rotated. The D3 fold observed here is a part of a larger hinge zone that extends to Apollonia and to the village of Platygialos at the SE side of the island. This hinge zone is the major fold structure parallel to the Hellenic trend, observable in central Sifnos.

About a hundred meters into the wadi several tens of meters of a metasedimentary micaschist sequence is exposed. The rocks are well cleaved and in the past were quarried by the Sifniots for paving stones.

The rocks consist of muscovite, chlorite, albite, calcite, quartz and graphite. No traces of an original high pressure mineralogy are found. The mineral assemblage suggests that the rock was derived from the metamorphism of a marl; the graphite, derived from former organic matter, is commonly found in rapidly deposited marls.

Further on up the wadi we come to a 2 meter thick marble layer. At the base of this layer, at the contact with underlying schists, fresh glaucophane is preserved, with a mineral lineation oriented 030° This glaucophane is the first indication of high pressure metamorphism in this section of the greenschist unit.

Immediately above this marble is a massive gray rock containing relict glaucophane within an albite iron-ore symplectite. The glaucophane is almost completely replaced by chlorite, biotite and actinolite.

Glaucophane relics become much more abundant in the upper parts of the wadi section, with a series of layered rocks forming the rest of the sequence. Most rocks are greenschists, consisting of albite, chlorite and varying amounts of actinolite, epidote, muscovite, quartz, calcite, sphene and hematite. Fine scale layering and kinking is frequently observed, including abundant quartz and calcite laminae. Variegated layering of the type found in Vroulidia Bay is not observed. Garnet pseudomorphs, now replaced by chlorite, are the remnants of the former high pressure mineralogy in albite rich rocks.

Station 7. Glaucophane bearing rocks in the upper parts of the Greenschist unit. Wadi to Ag. Simeon, near contact with Main marble.

As we come out of the wadi into a wide stony river bed, we reach the contact with the main marble unit. Beneath the main marble, on the western side of the wadi, a 2 meter thick glaucophane-bearing layer is exposed. Glaucophane is fresh, up to 2 cm in length and it forms a mineral lineation oriented 020 . The rocks also contain albite, calcite and epidote. Microstructures indicate that glaucophane is stable with the albite and epidote. This is the first example we have seen of blueschist facies rocks intermediate between eclogite facies and greenschist facies assemblages.

Just below the glaucophane bearing rock, a layer consisting of small gray concretions made of albite and ore symplectites, without any relict glaucophane, can be seen. The concretions are flattened and the schistosity wraps around it. The albite within the concretions is strain free; thus flattening occurred prior to the growth of the albite. It is probable that the concretions were previously more competent eclogite layers, which were boudinaged during the high pressure deformation. The eclogite boudains later retrogressed to albite blasts and iron symplectites which now form the concretions.

In the basic rocks of station 7, actinolite occurs with calcite and chlorite. Sphene is also abundant. Both features indicate that XCO2 in the fluids associated with the greenschist metamorphism was

low, possibly less than 0.1 (Fig. 2.5b). Way to station 8. Following the river bed we enter the main marble. At the sign to Agios Simeon, on the left hand side of the way, we follow the path to the west to the top of the ridge above Kamares Bay.

Station 8. General view of the large scale stratigraphy and block faulting, Agios Simeon.

The general stratigraphy of the island can be seen from this point. The sequence as a whole is tilted to the north. In the north the blueschists and eclogite sequence is exposed. Beneath is the main marble sequence on which Ag. Simeon is located. Within the main marble thick (up to 100 m) intercalations of schists can be seen. The greenschist unit is well exposed to the east and south east, below the marble. To the south, immediately across Kamares Bay, the island is bisected by a major normal fault. The fault extends from Kamares up Kamares valley to Apollonia and a prominent breccia zone is associated with it. The faulting is attributed to the post Miocene extension tectonics of the Aegean region. The roof of the main marble reappears in the hanging block south of Kamares valley, and is inclined to the north.

Way to Station 9. Either climb down one of the steep valleys to Kamares or take the unpaved road that leads from Ag. Simeon down to the beach. From the beach continue west along the northern side of the bay to the church of Ag. Aikaterini (Map 2).

On the way to Ag. Aikaterini, at the NE corner of Kamares Bay thick dolomitic marbles can be seen just above the track. These marbles gives the typical marine sedimentary carbonate analyses found in the Main Marble (dolomites $\delta 180 \sim 29o/oo$, $\delta 13C$ 4o/oo; calcites $\delta 180 \sim 28o/oo$, $\delta 13C \sim 3.5o/oo$). Further towards Ag. Aikaterini, we pass two schist intercalations. The schist intercalation whose $\delta 180$ compositions were previously discussed in Fig. 2.6 occurs behind the church.

Station 9. Greenschist layer, Ag. Aikaterini.

The rocks here consist of acid and basic types, interlayered and very similar in appearance to rocks from Vroulidia Bay (e.g. unit 12) although they are now in greenschist facies. The section behind the church is illustrated in Fig. 2.8. The lower parts of the section can be seen by going down the steps at the back of the church to the water front. Here greenschists and micaschists crop out. Above these are the interlayered acid-basic rocks. Climbing approximately 30 m above the top of the steps we find the interlayered acid-basic rock types and the relict eclogite lenses. The acid rocks are now albite gneisses and the basic rocks, typical greenschists containing albite-epidote chlorite-actinolite. All of these rocks (aside from the eclogites) have a higher $\delta 180$ composition than chemically equivalent rocks from Vroulidia Bay (e.g. unit 12). A comparison of the $\delta 180$ compositions of the Kamares Bay layers with unit 12 of Vroulidia is shown below the section in Fig. 2.9. This is evidence for the infiltration of 180 enriched fluids.

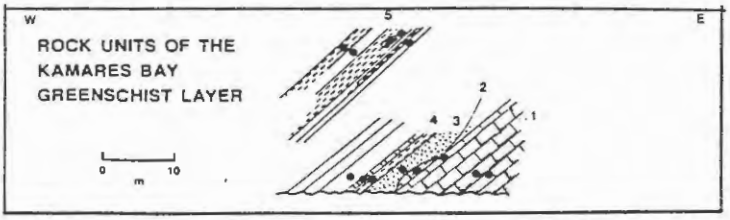


Figure 2.8 Interlayered units of greenschist facies rocks within the Main Marble at Kamares Bay (Ag Eikaterini). 1 = Main Marble; 2. Marble at contact;3.micaschist; 4 and 5 interlayered sequences of greenschists containing relict eclogite lenses, and micaschists and albite gneisses. Diagram after Schliestedt and Matthews (1987).

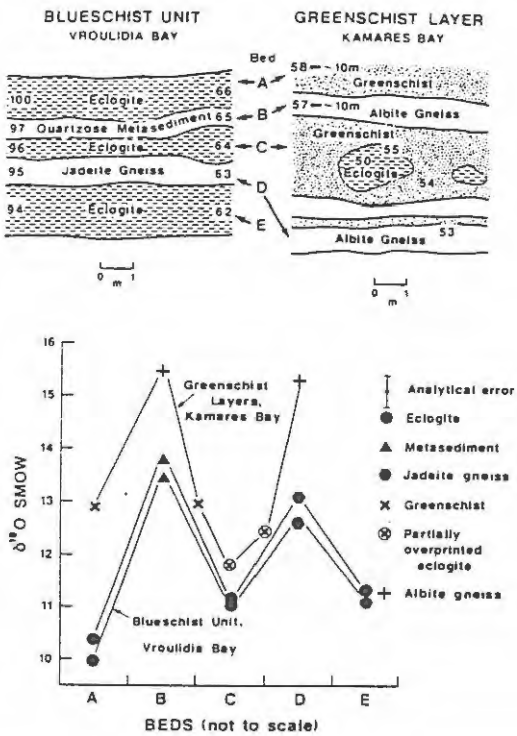


Figure 2.9 Comparison of the isotopic compositions of interlayered basic and acid rocks from the high pressure sequence in Vroulidia Bay (unit 12, Figure 2.7) with the Kamares Bay layers in Figure 2.8 above. From Schliestedt and Matthews (1987).

Way to station 10 Climb above the schist intercalation on a narrow goat path and walk west. Cross the upper contact of the schist intercalation with the marble. Continue about 50 m along the path up to a water hole, located in a gully at a height of 40 m.

Station 10. - Subduction related megastructures within the main marble.

Along the marble cliffs of Kamares Bay the complexity of the structures associated with the subduction process can be examined. Large scale folded structures can be traced by following individual marker horizons. The station is located on a large D1 hinge zone mainly defined by folded brownish dolomite and white marble. The fold is isoclinal, the axis trends 020°-040° and the fold closes to the west. The fold can be traced from the first calc-schist horizon immediately overlying the large schist intercalation of station 9. This calcschist horizon forms the outer envelope of the structure. Tracing the dolomite uphill to the east for several hundreds of meters, another hinge zone can be observed. Large scale superposed folding relations are illustrated in Fig. 2.10. Above the dolomite and the white marble, further along the path towards west, more of the marble sequence is exposed. Immediately above the calc-shist envelope lies a blue gray marble horizon. If it is traced towards the water line a large D3 hinge zone is observed trending 320°. The remainder of the cliff to the west and close to the water is made of blue marble that is repeatedly folded. These folds, which are on average tens of meters in size, are probably only parasitic folds on the limb of a much larger structure, the closure of which lies in the sea. Several layers above the marble have a conglomeratic

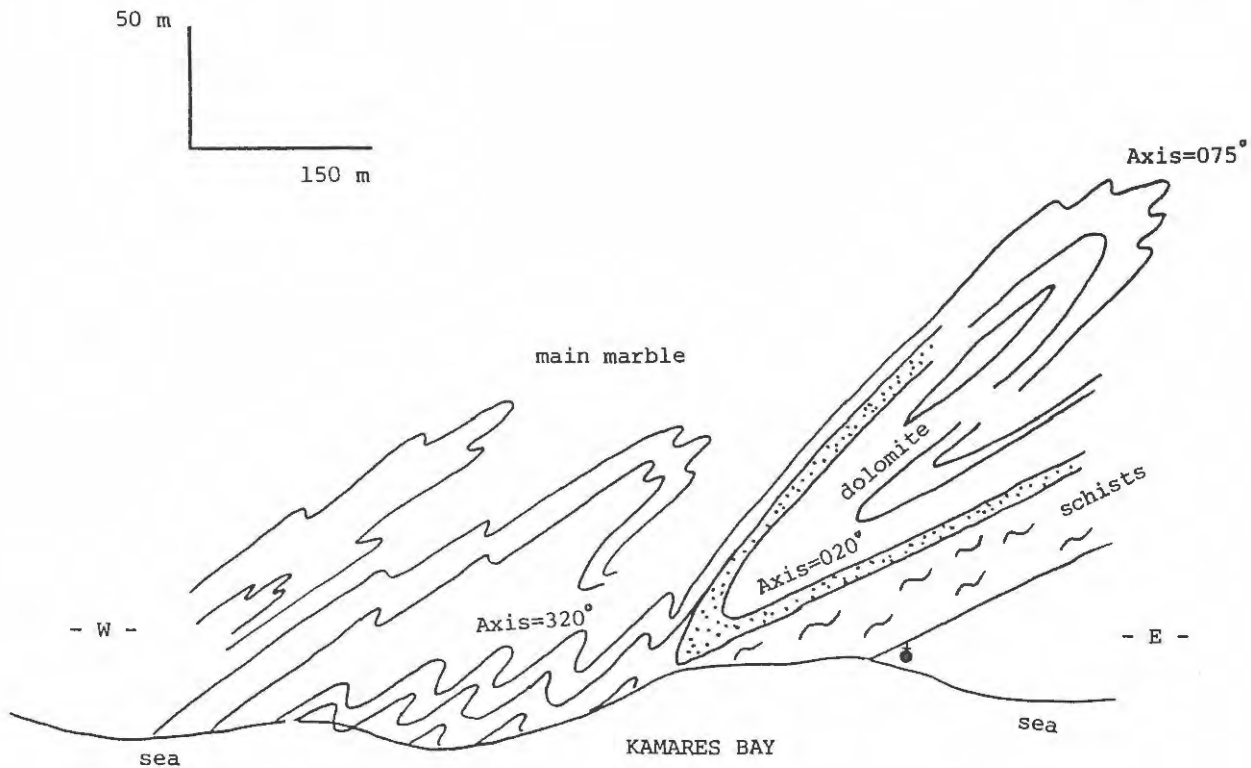


Fig. 2.10 Fold structures in the main marble, northern side of Kamares bay. Tracing of individual marble horizons. Calc-schist - dotted.

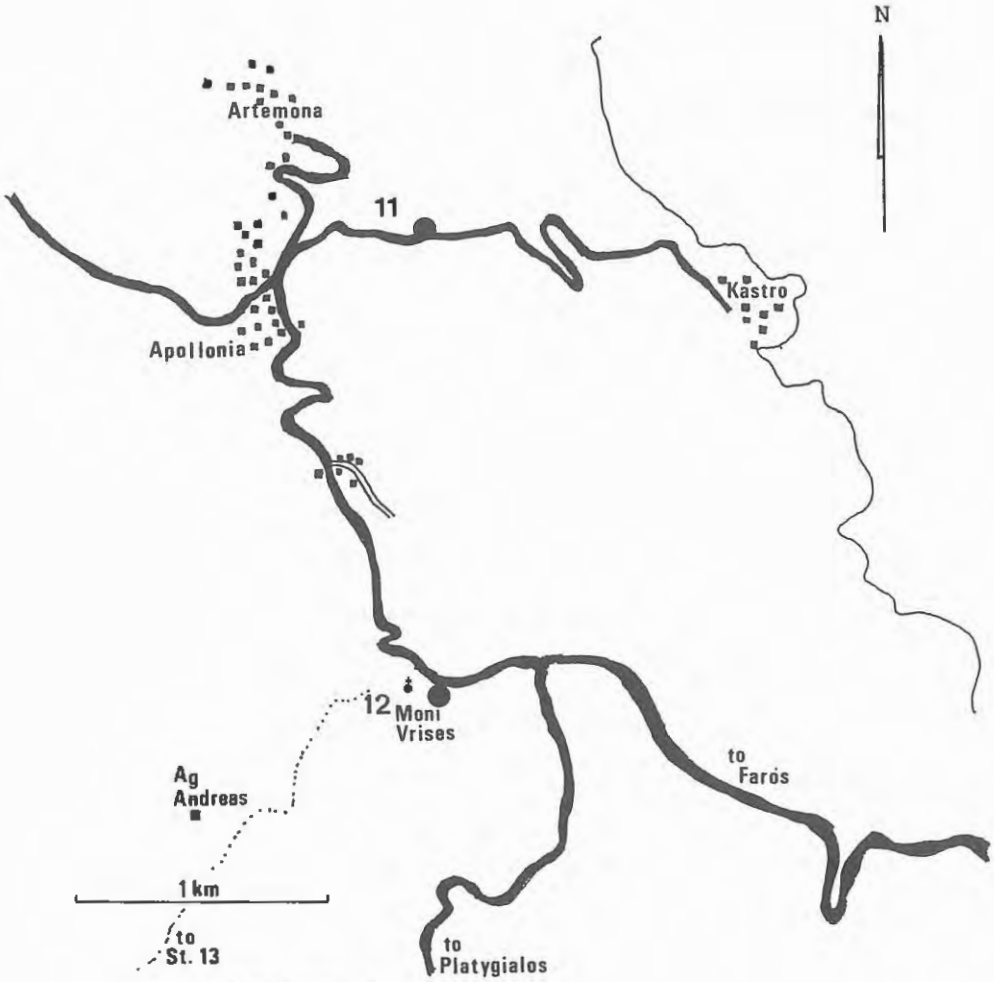
structure. They consist of mafic pebbles in an impure marble. Some of the mafic material is boudinaged and deformed. The mafic rocks contain undeformed albite porphyroblasts. Fresh glaucophane forms both in the necks of the boudins and as a randomly oriented mineral overprint. This indicates that the D3 folding occurred while the rocks were still at relatively high pressures and prior to the final greenschist overprint.

DAY 3 GREENSCHISTS, BLUESCHISTS AND ECLOGITES REMNANTS IN THE CENTRAL SIFNOS UNIT. PETROLOGY AND STRUCTURE. FAULTED BLUESCHIST BLOCKS OF S-W SIFNOS. LOCATION- MAP 3. ROADS FROM APOLLONIA TO KASTRO, AND FROM APOLLONIA TO MONI VRISES. AREA OF VATHI.

Way to station 11. From the main intersection in Apollonia walk 300 m on the road to village of Kastro. Mafic schists are exposed along the road and many contain relict omphacite. Garnet-omphacite bearing rocks occur sporadically along the road as thin grassy-colored mica rich layers. Locally, reddish garnet is preserved and often glaucophane is preserved next to it within a layered greenschist matrix.

Station 11. Eclogite micaschist remnants in the greenschists of central Sifnos.

The station is located in the northern side of the road in a curve before a church. At this outcrop, centimeter-scale layering involves albite greenschists, garnet glaucophane schists and omphacite- garnet mica schists. Various stages of the transformation from the high pressure omphacite-garnet rocks to greenschists can be observed. Albite blasts have crystallized at



LOCATION MAP 3

the expense of clinopyroxene and mica, and frequently mantle garnets.

Texturally, the eclogite schists contain a crenulation cleavage which is defined by microfolding and recrystallization of fine grained omphacite and phengite. The cores of omphacite grains have a jadeite content of about 30% (indicating a minimum P of 9-10 kbar) which decreases to less then 10% in the rims.

Way to station 12. Return to Apollonia then take the road to Platygialos. Station 12 is located on the section of the road that begins from Moni Vrises and continues a few hundred meters to the East until the junction of the road to Faros.

Station 12. Blueschist and greenschist rocks, and the D3 "Hellenic trend".

The rocks in the section exposed along the road belong to the upper parts of the greenschist unit of central Sifnos. Further to the south they are overlain by the main marble. In the outcrop, just beneath the church of Moni Vrisis, glaucophane is abundant in both micaceous and more basic rock types. The glaucophane occurs on schistosity planes, appears fresh and is randomely oriented. In adjacent layers traces of it can be found as pseudomorphs made of chlorite.

Across the valley to the east, massive layers of rocks consisting mainly of albite, epidote, glaucophane, calcite, sphene and hematite can be seen. Texturally, the glaucophane in all rocks of this area shows rational contacts with albite. The glaucophane

is thus not a remnant phase of the older high pressure metamorphism, but grew with albite in the blueschist facies. The blueschist facies assemblages of central Sifnos represent a stage of the Alpine evolution in which the uplifted eclogite-facies rocks re-equilibrated to form a new stable paragenessis at lower pressure epidote blueschist conditions. These rocks later reacted to form greenschists through the breakdown of glaucophane to chlorite and actinolite.

At this outcrop, open, tight and isoclinal D3 folds can be seen. They trend 320° and belong to the "Hellenic trend". The folds are part of a large hinge zone that stretches to Kamares Bay in the west and to Platgyialos in the SE. The fold is overturned to the north and there is little fabric developement associated with it. Locally glaucophane is aligned parallel to the fold axes but more often it postdates the folding and is randomely oriented.

Up the hill south of Moni Vrises, a superposed folding structure is observed. A D1 isoclinal fold is refolded by 320 trending fold of the " Hellenic trend" . Nowhere is the "Hellenic trend" itself observed to be folded. It is the last folding phase affecting the rocks, and as we have formerly shown, occurs during the transitionary blueschist facies equilibration.

Way to Station 13 (optional if time permits). Walk south into the wadi and uphill from Moni Vrises towards the prehistoric settlement at Ag. Andreas. Walk around the mountain of Platia Rakhi to Ag. Nikolao and downhill toward Vathi (6 km distance). On the way cross the main marble section and an intercalation of

partially replaced blueschists in the marble.*

Station 13. Faulted blueschist blocks (location, see Fig. 2.1).

Steep faults usually shape the borders of the island, forming cliffs that rise straight out from the sea. At station 13, a blueschist block is faulted against the lower part of the main marble. The blueschist rocks in the outcrop consist of alternating layers of basic rocks and acid gneisses. The blueschists are fresh, with almost no sign of retrogression. The stratigraphy and the preservation of the high pressure mineralogy are reminiscent of the sequence at Vroulidia Bay. We suggest that the faulted block was derived from the northern high pressure sequence, and estimate a 1000 m throw on the fault.

A fault bounds the high pressure sequence that crops out in the South of the island, and presumably a similar interpretation can also be applied to this.

* Alternatively take a boat from Kamares to Vathi

3. GEOLOGICAL EVOLUTION OF TINOS

3.1 Introduction and stratigraphy

The island of Tinos is situated approximately 75 km to the NNE of Sifnos (Fig. 1.2). The island forms a 30 km NW-SE striking elongated ridge and with an area of ~200 km^2 is the third largest island of the Cycladic group. Mt. Chiknias, the highest topographic peak of Tinos, reaches 730 meters. The island was mapped by Melidonis (1980) who recognized the following stratigraphic units: Main lower blueschist unit, an upper unit, an intrusive granodiorite, minor volcanic rocks and small outcrops of Miocene-Pliocene transgressive sediments (Fig. 3.1). The metamorphic rocks of the lower blueschist unit contain glaucophane-garnet-epidote-phengite bearing high pressure assemblages and mainly occur in the western part of the island. As on Sifnos, these high pressure assemblages were affected by the early Miocene greenschist facies overprint. The upper unit tectonically overlies the blueschist rocks of the lower unit. It consists of a variety of rock types, including phyllitic greenschists, banded amphibolites, ultramafics, ophicalcites and metagabbros. The effects of high pressure metamorphism are not seen in the upper unit and preliminary cooling ages on the amphibolites give 70 Ma. (Patzak 1988 and ages obtained on hornblende separates in the laboratory of Geochronology, Geological Survey, Jerusalem).

Regionally, the lower unit can be correlated with the basal unit of the Cycladic massif, whereas the upper unit correlates with other remnants of low pressure metamorphic rocks which overlie the

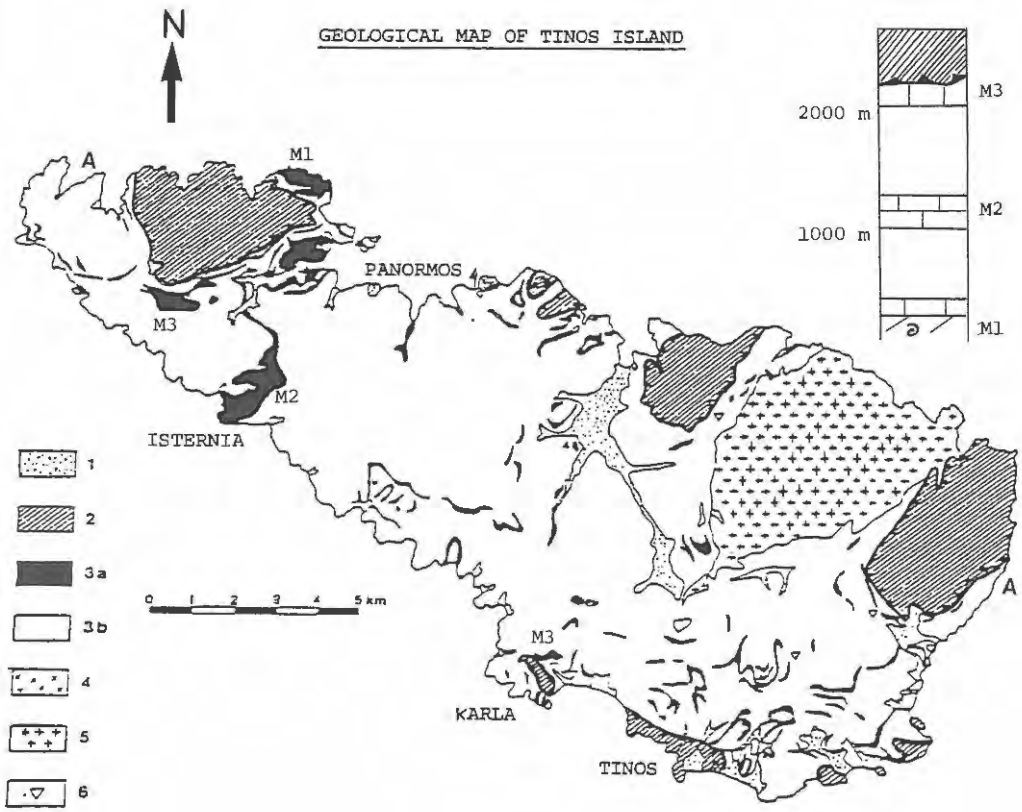


Fig. 3.1 Geological map of Tinos island modified after Melidonis (1980). 1) Quaternary; 2) Upper unit; 3a) marble of the blueschist unit; 3b) schists and gneisses of the blueschist unit; 4) diorite; 5) granite; 6) volcanics

blueschists throughout the Cyclades. The age of the intrusive granodiorite of Tinos has been estimated at 18 Ma and belongs to the phase of regional batholithic granitoid activity that affected the Cyclades throughout the Miocene (Altherr et al. 1982, 1988).

Melidonis (1980) mapped late Miocene transgressive sediments on top of the metamorphic rocks of Tinos. The sediments were deposited in the proto Aegean sea and date the final exposure of the metamorphic rocks.

The present geographic and topographic features of Tinos island were shaped by normal fault tectonics of the Aegean region which have occurred since the late Miocene. The island as a whole forms a block which is tilted to the NE and the structure continues along strike to the island of Andros (Papanikolau 1978). Several NW trending fault escarpments can be traced, and two of them are accompanied by parallel dikes. The dikes do not form fissures for any exposed extrusives, and were intruded prior to the late Miocene sedimentation whilst the rocks were still at depth. A major NNE trending fault scar transects the island, with the southern part of the island forming the downthrown block.

Stratigraphy of the lower blueschist unit:

The lower unit section, as defined by Melidonis, is up to 2 km thick and includes three distinct marble horizons. The lowest marble horizon (M1) forms the base of the section, whereas the third marble horizon (M3) occurs near the top (Fig 3.1). The occurrence of M1 marble is restricted to the northern part of the island at Panormos Bay, and to a small outcrop in the bay of Vathi. M1

consists of 150 meters of massive dolomite capped by a well layered quartzite bearing calcitic marble.

Schists and gneisses of the lower unit overlie the M1 marble. They comprise pelitic and quartzo-feldspathic sediments and basic and acid metavolcanics of a probably volcanoclastic origin. Glaucophane bearing blueschists and a unique horizon of meta-ironstones occur near the bottom of the second marble layer M2 in the middle of the section (the region of Isternia, station 13).

M2 is 150 meters thick and consists of well layered and recrystallized calcitic marbles. Several horizons of schists are intercalated within the marble. Above M2, metasediments and metavolcanics reappear in the section. Close to the contact with M3 in the northern part of the island, Broecker and Okrusch (1987) hav distinguished a metaconglomerate consisting of carbonate and some metavolcanic pebbles. M3 is thick and well layered in the north of the island but occurs as thin interlayers in the SW. Both above and below M3, glaucophane bearing blueschist rocks occur. In the area of Karla near Tinos town, thin M3 marbles are interlayered with blueschists, eclogites, quartzite and metacidites.

The high pressure metamorphic rocks on Tinos are generally less well preserved than those on Sifnos. The high pressure rocks of Tinos also do not occupy a restricted position within the section. The large mass of eclogites is confined to the upper part of the section in the area of Karla, in a similar fashion to the occurrence of eclogites on Syros and Sifnos.

3.2 Metamorphic evolution

3.2.a High pressure assemblages of the lower unit in the Karla area:

The Eocene high pressure metamorphic phase is best represented in the area of Karla (Fig. 3.2) where a representative variety of rock types are found. The mineral assemblages are very similar to those detailed in Table 1 for Sifnos, i.e. blueschists: glaucophane, epidote, garnet, phengite, rutile, sphene and opaques. Eclogites: omphacite, garnet, phengite, epidote, glaucophane, quartz, rutile, sphene (albite). Jadeite gneisses: jadeitic pyroxene with up to 82 mol% jadeite, garnet, mica, quartz + rutile, epidote, zircon (albite).

The blueschists and eclogites frequently occur in close contact with each other. The jadeite gneisses however, occur as large boudins or knockers some tens of meters in size. Pseudomorphs after lawsonite were found in mafic rocks, and consist of fine grained zoisite, paragonite and quartz. The texture indicates that conditions close to those of reaction 4 in Fig 2.3 were exceeded. Garnet-pyroxene thermometry on eclogites yields temperatures of 400 -500°C. At temperatures in the range 450-500°C, the ca. 80 mole% jadeite content of the cpx in meta-acidic rocks indicates minimum pressures of 11 - 12 kbars. Clearly P-T conditions in the eclogite facies rocks of Karla area were very similar to those in northern Sifnos.

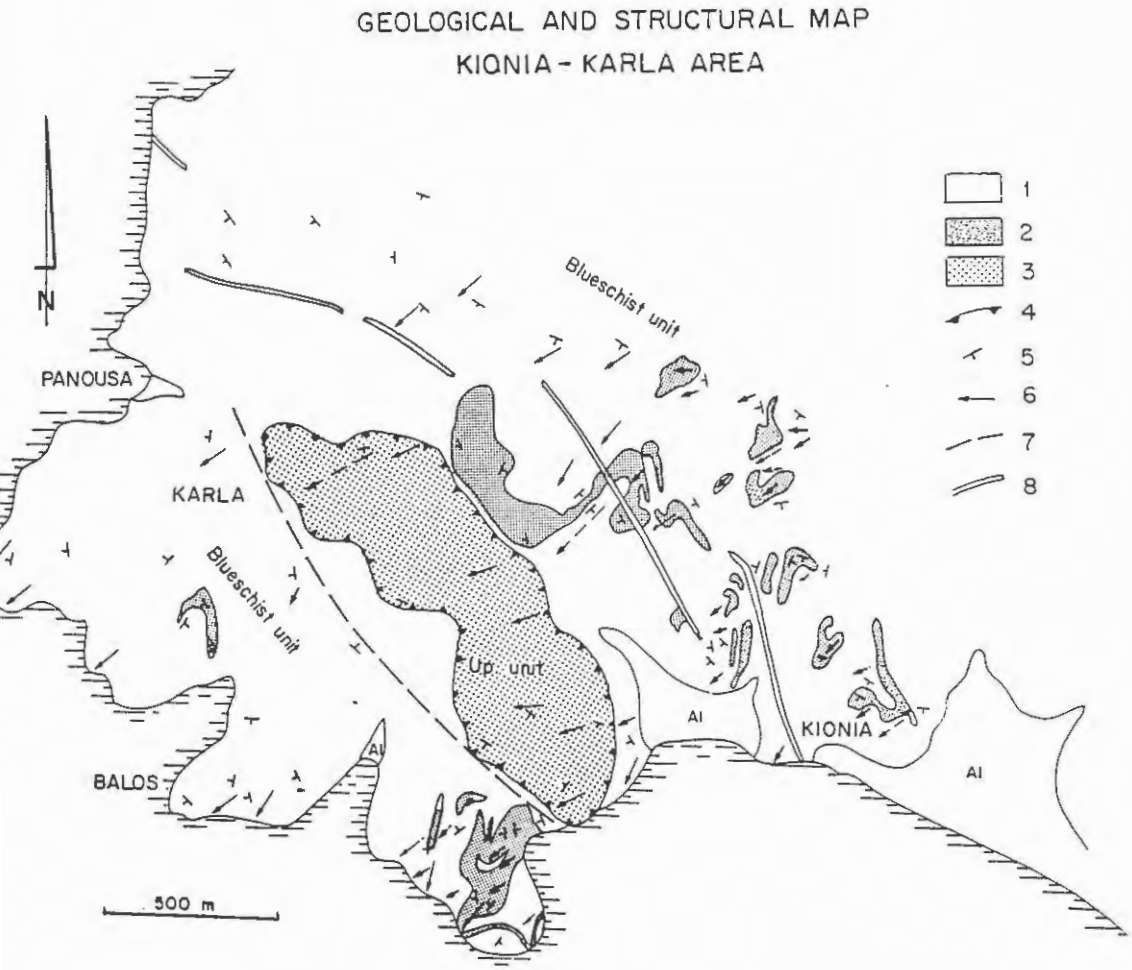


Fig. 3.2 Geological and structural map of Kionia - Karla area, Tinos. 1) lower blueschist unit; 2) marble intercalation within the blueschist unit; 3) Upper unit; 4) low angle tectonic contact; 5) dip of cleavage; 6) fold axis; 7) fault; 8) volcanic dike

3.2.b The transformation from high pressure to greenschist facies conditions

The high pressure metamorphic rocks of Tinos underwent an extensive transformation to greenschist during uplift, and as a result most of the rocks of the lower unit have a greenschist facies mineralogy.

The reactions occuring during the transformation to greenschist facies conditions are similar to those described for Sifnos. In eclogites, omphacite and paragonite react to form albite. In blueschists, glaucophane is often rimmed or replaced by green calcic amphibole, with or without secondary chlorite and albite. Albite poikiloblasts crystallized at the expense of glaucophane and paragonite and often contain remnants of clinopyroxene. Biotite and chlorite often replace garnet, and rutile is frequently rimmed by sphene.

An intermediate phase of epidote blueschist facies equilibration also appears to have occured on Tinos. Large idioblastic crossites coexist with albite blasts, epidote and calcite. Relicts of clinopyroxene occur within the albite, suggesting that the albite and glaucophane were formed as the result of the breakdown of omphacite bearing assemblages. The epidote-glaucophane-albite parageneses were then later transformed into greenschists.

Insufficient mineral chemistry data is available for the P-T path on Tinos to be properly constrained, but it appears to be very similar to that experienced by the Sifnos units. Our stable isotope geochemistry is still in early stages , but the initial results support the correlation of Sifnos and Tinos P-T histories. The samples from Karla area have $\delta 180$ compositions identical to those observed from Vroulidia Bay, northern Sifnos, whilst the lower parts of the lower unit in Tinos bear many similarities to the greenschist layers of Kamares Bay and the rocks of the central Sifnos unit.

3.2.c Metamorphic assemblages of the upper unit:

The upper unit is comprised of rocks which exhibit a variety of metamorphic grades. The phyllites consist of quartz, muscovite, epidote and plagioclase, and were probably metamorphosed in the greenschist facies. The amphibolites consist of green hornblende with plagioclase, epidote, biotite, rutile and sphene. In the ultramafic rocks primary igneous minerals are often preserved; chromite in ophicalcites, and clinopyroxene and olivine in the ultrabasic rocks of the Chiknias mountain.

No evidence for high pressure metamorphism of the type observed in the lower unit was recognized in the Upper unit and clearly a marked metamorphic discontinuity separates the two units.

3.3 Structure and tectonics

The discussion on the tectonics of Tinos can conveniently be divided into three parts: a) Metamorphic related structures in the main lower blueschist unit. b) Structures of the upper unit and the

intrusive granite. c) Large scale nappe tectonics.

3.3.a Structures in the lower blueschist unit.

Several deformation phases can be recognized within the lower metamorphic unit. A first deformation phase, D1, produced a layer-parallel cleavage, whose formation was associated with the crystallization of the high pressure minerals. The axes of D1 trend to the NE, and the hinges are tight to isoclinal, with almost flat axial planes. The hinges range in size from microcrenulations to tens of meters and are overturned to the SE. In the area of Isternia, this folding phase is seen particularly well in meta-ironstones. The orientation of D1 in the Karla area, where high pressure assemblages are well preserved, is shown on the structural map of Fig. 3.2. A mineral lineation, characterized by alignment of glaucophanes, is often developed parallel to the axes of microfolds. Both in style and trend, D1 can be correlated with the NE trending D1 phase of Sifnos. Both are associated with the high pressure metamorphism, and hence formed the subduction related structures. Similar subduction related NE trending D1 structures are also observed on Syros (Ridley 1984) and thus a constant orientation is maintained over a distance of more then 80 km.

An additional folding phase, D2, can be traced from the road cut above Isternia towards the ridge just above the village of Kardiani, and can be recognized also near Tinos beach (Kionia). The folds trend NW and have hinges several tens of meters in size. A shallow fracture cleavage is developed along the axial planes . This fabric is local and nonpenetrative. Albite porphyroblasts are

sometimes rotated along the new cleavage planes but are mostly undeformed. Near Kardiani, newly formed crossite grains overgrow the hinges of D2 micro-crenulations; the albite in these rocks is undeformed. The microstructural evidence suggests that the D2 phase affected the rocks prior to the growth of porphyroblastic albite, and at pressures sufficiently high to enable the crystallization of crossite. A similar trend, style and relation to metamorphism is observed in the NW trending D3(!) phase in Sifnos. As we have discussed earlier this NW trending phase is termed the "Hellenic Trend", in view of its parallelism to the major fold axes in the Greek mainland.

An additional folding phase, D3, can be observed within the lower unit of Tinos. Its axes trend 020°,and the hinge zone is well developed from Karla area over to the village of Kalloni on the other side of the island. The hinges are on a scale of hundreds of meters, and above the road in the area of Linaria are overturned to the east. Strong kinking of micas is observed in the hinge zones, and albite in metapelites is deformed. In metabasic rocks, actinolite and chlorite are repeatedly folded, although albite is strain free. In the area of Karla, superposition relations are observed with the D1 axes refolded by the 020° phase.

D3 in Tinos cannot be correlated to the 020° trend on Sifnos. In Sifnos the 020° trend (D2 in the terminology of Sifnos) is associated with blueschist metamorphism, whilst on Tinos the trend is associated with deformation of greenschist minerals. We suggest that D3 on Tinos should be correlated instead to the 015° folding phase which is known to affect the rocks of Naxos in the late

Oligocene and early Miocene. The structural dome of Naxos was formed during the Barrovian overprint and the amphibolite facies minerals are frequently deformed. An older phase of deformation trending 015° is recognized in Naxos and is related to the postulated early high pressure metamorphic facies. We suggest that the D2 of Sifnos is comparable to it.

3.3.b Structures within the upper unit and Miocene granite:

Ductile deformation structures are mainly developed within the phyllitic greenschists of the upper unit. The cleavage is observed to be subhorizontal and parallel to layering. In fine grained slates it is a crenulation cleavage. Tight to isoclinal folds can be observed in the upper unit. In the rock of the Upper unit in the area of they trend NE, identical to structures in the underlying blueschists. The upper unit as a whole is preserved in synformal pockets which were formed by gentle folding of the contact between the upper and lower units subsequent to emplacement. The gentle folds involving the contact are upright and trend NE. The NE trending isoclinal folding thus terminated prior to the emplacement of the Upper unit onto the lower unit. The internal NE trending isoclinal folds observed in each unit must have formed before the two units were juxtaposed.

The Miocene intrusive granite is relatively undeformed but locally in the area of Livados Bay it is mylonitized and shows developement of a NE trending fabric. The granite is exposed in the core of an upright antiform which was formed subsequent to the emplacement of the upper unit.

NE trending structures have thus been periodically recorded since the upper Cretaceous. They occur in the Eocene metamorphic structures (D1) of the subducted high pressure units, and postdate the intrusion of the granite in the middle Miocene.

3.3.c Large scale nappe tectonics:

The lowest marble horison (M1) in Tinos was considered by Melidonis as the base of the section. Recent investigations indicate that the lower carbonate sequence (M1) is not an integral part of the blueschist sequence but forms a different tectonic unit. This is based on detailed observations on the metamorphism and the structure of the lower carbonate section and comparison to the overlying blueschist unit (Avigad and Garfunkel 1988). The structure of Tinos island is shown in Fig. 3.3a.

The structurally lowest part of the section is exposed in the north of the island around Panormos Bay, in the core of an upright antiform (Fig. 3.3b, 3.4). The base of the section consists of more than 100 m thick massive dolomites. The dolomites contain Upper Triassic fossils including algae and fragments of corals, as observed by Melidonis (1980). Fossiliferous banks were also found by the authors at a different locality where coral fragments, algae and fragments of echinoids preserve their original shape and fabric. The dolomite section ends with a 2 m thick layer of quartzites and micaceous phyllites. The phyllites consist of muscovite, quartz, albite and opaques. The schistosity is defined by the preferred orientation of micas in a subpolygonal quartzo felspathic matrix, and was locally refolded. An axial planar crenulation cleavage was

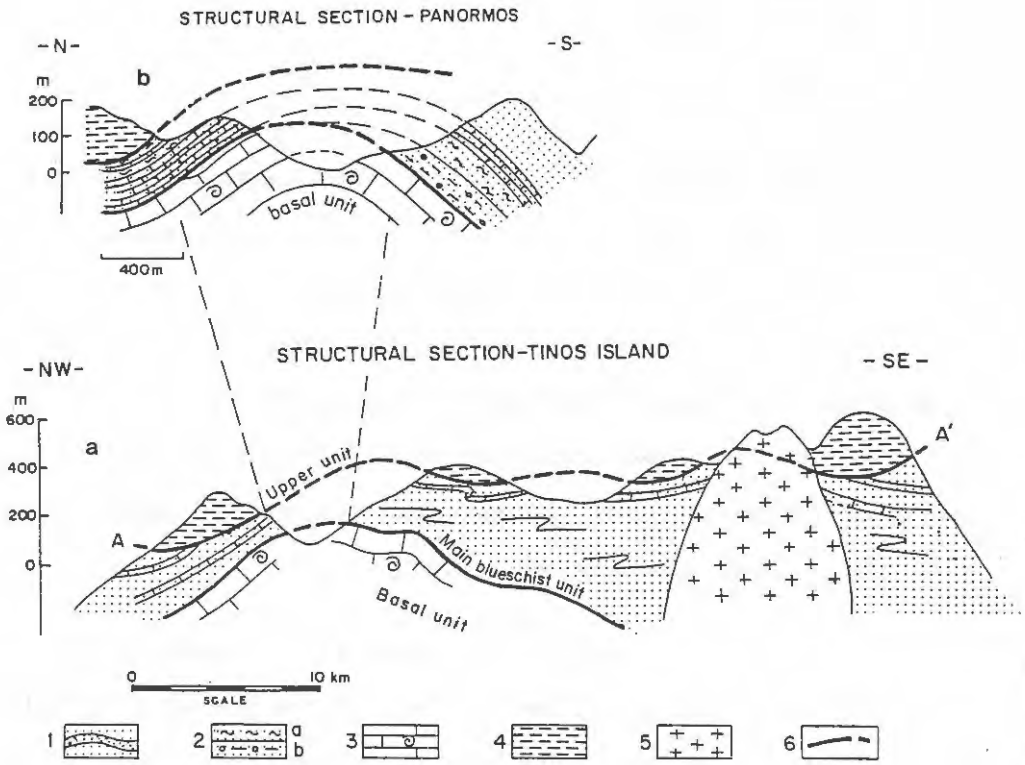


Fig. 3.3 a) structural cross section of the island of Tinos (A - A') showing the three units of the Cycladic massif in Tinos island : the Basal carbonate unit, the overlying Blueschist unit and the Upper unit at the top. b) Enlarged structural section of the tectonic window around Panormos bay. In the north the lower tectonic contact is placed above the dolomite section and below the calcite marble, although a structural discordancy will still exist if the contact is placed above the marble.
Legend: 1) Main Blueschist unit with marble intercalations. 2) a) albite schist. b) calcschist. 3) Basal carbonate unit (M1), fossiliferous dolomites. 4) Upper low-P unit. 5) Miocene granite. 6) Tectonic contact.

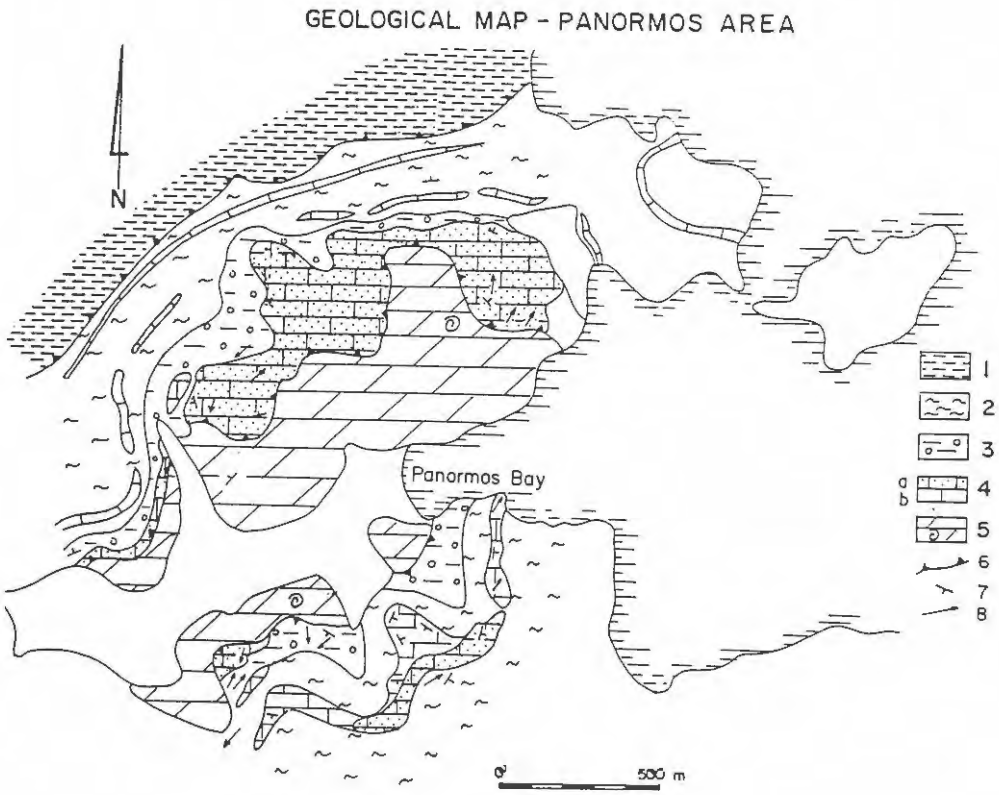


Fig. 3.4 Geological map of Panormos area, Tinos. 1) Upper unit 2) schists and gneisses of the blueschist unit 3) calcschist 4a) calcite marble with quartzite 4b) calcite marble 5) fossiliferows dolomite

produced in secondary fold hinges.

In the north of the dome structure, the massive dolomite and the phyllites are overlain by a 50 m thick calcite marble. This marble is well layered and contains abundant quartzitic horizons. Above the calcite marble, relict glaucophane occurs within albite-actinolite schists of the blueschist unit. In the south, a wedge of calcschists and albite schist rests directly on the dolomite. A quartzite bearing calcitic marble reappears on top of these schists. The relations thus indicate a structural discordancy at the top of the dolomite-phyllite section (Fig. 3.3b).

A comparison of the blueschists and basal dolomites units reveals important metamorphic and microstructural discontinuities. The blueschists were metamorphosed and deformed at high pressures and at temperatures of nearly 500°C, whereas the underlying rocks were affected only by low grade metamorphism in the quartz-muscovite subfacies. Glaucophane is abundant in the blueschist unit above the contact but is absent in the basal section. The marbles in the blueschist unit consist of a polygonal mosaic and sedimentary features have been obliterated, but the carbonates of the basal section contain undeformed fossils. The repeated folding and syn-kinematic crystallization of the phyllites contrasts with the static mineral growth of greenschists within the main blueschist unit. These differences indicate that the basal dolomite and associated phyllites underwent a different tectonometamorphic history than the overlying blueschists. Therefore the basal section should be separated from the rest of the overlying blueschists. The contact between the two units is interpreted to be tectonic and is

marked by the structural discordance observed at the top of the basal dolomites.

The blueschist unit of Tinos is thus allochtonous and occupies a median position within a nappe pile. The contact between the low grade carbonate and the blueschist unit juxtaposes deeply buried rocks over shallower rocks and is thus a thrust fault.

Above the main blueschist unit lie remnants of the Upper allochtonous unit (Figs.3.1, 3.3). A rather flat but undulating tectonic contact is well exposed at its base. The contact is defined by a zone of several meters of brecciation, above which lie fractured serpentinite and ophicalcite bodies. Mylonitization and features indicating plastic flow are absent. The marked metamorphic and age discontinuities have been noted previously.

The contact between the blueschists and the overlying Upper unit places rocks metamorphosed at mid crustal levels on top of rocks metamorphosed at great depths. A thick crustal section is thus apparently missing between the two units. Hence, the upper contact operated as a normal fault.

The blueschist unit of Tinos is thus sandwiched between a thrust fault at its base and an almost flat normal fault at its top. Both contacts evolved later than the Eocene, as the basal carbonate unit and the Upper unit were not affected by the high pressure metamorphism. The Upper unit was also unaffected by the greenschist overprint in the late Oligocene since it yields upper Cretaceous ages. On the other, hand investigation near the contact with the Miocene granite indicate that the emplacement of the Upper unit predated the intrusion of the granite 18 Ma ago.

3.4 EXCURSION GUIDE

DAY 1 PETROLOGY AND STRUCTURE OF THE HIGH PRESSURE METAMORPHIC ROCKS OF THE KARLA-KIONIA AREA . REMNANTS OF THE UPPER UNIT. LOCATION MAP 4. (stations 1 - 7) LOCATION MAP 5. (stations 8, 9)

Take the paved trail just behind Tinos Beach hotel at Kionia and climb uphill.

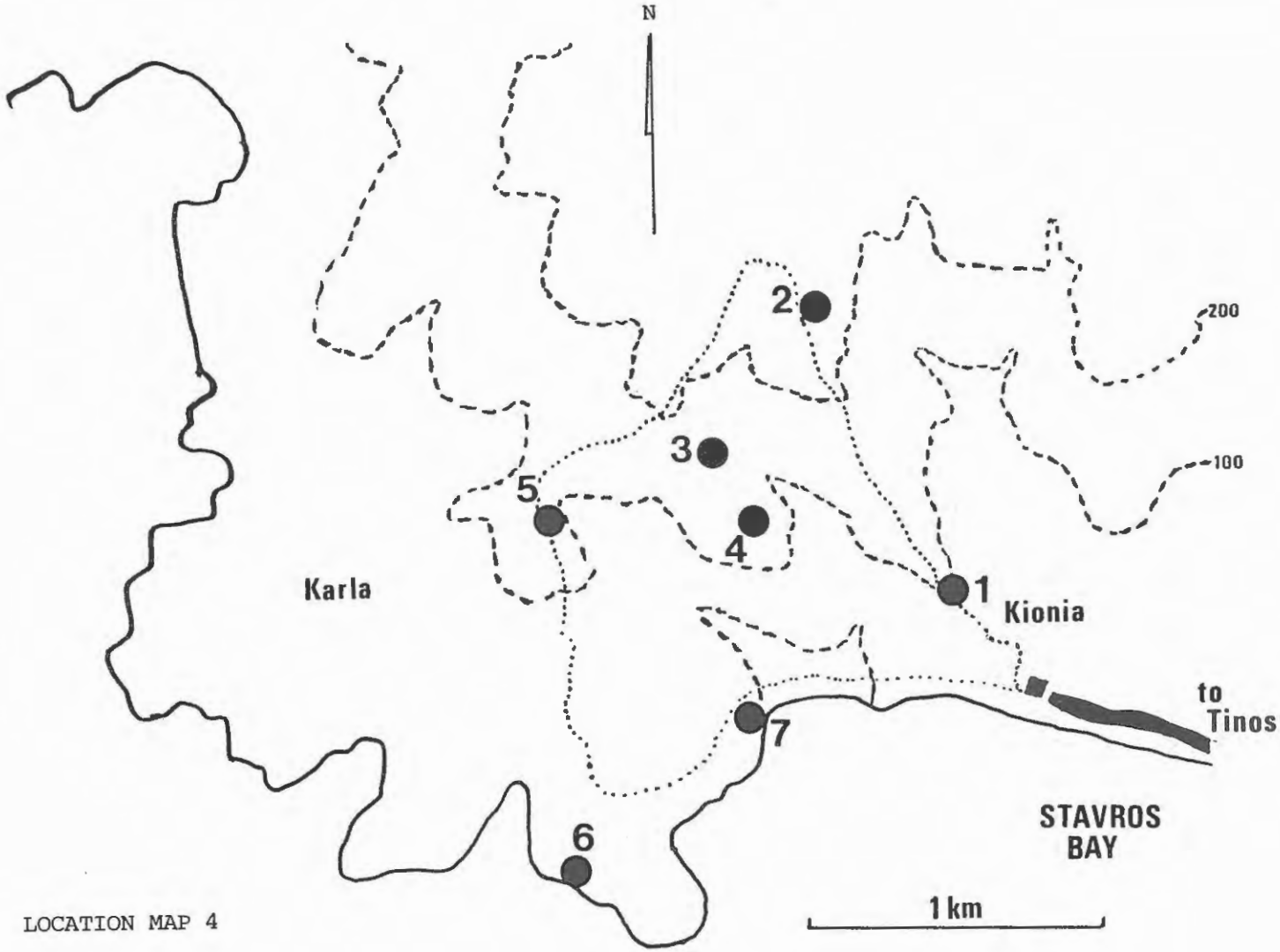
Station 1. M3 marble layer

Marble layers several meters in thickness are intercalated within various schists. These marbles belong to the M3 horizon which marks the uppermost part of the lower blueschist unit. The layers are often folded by NE folds. The fold hinges are tight or isoclinal and are mostly overturned to the SE. The folds belong to the first deformation phase D1.

Climb uphill to station 2.

Station 2. Blueschists and eclogites.

On the right side of the trail behind a stone wall a massive block several tens of meters in size can be seen. The block contains basic rocks which preserve a high pressure metamorphic mineralogy. Eclogites and blueschists occur in layers a few centimeters to several meters thick. Their mineralogy was described by Broecker and Okrusch (1987). Eclogites consist of omphacite, garnet, epidote, phengite and glaucophane. Blueschists consist of glaucophane, epidote, garnet, phengite-paragonite. No signifcant chemical differences between the two rock types are found and



LOCATION MAP 4

Broecker and Okrusch have suggested, on textural grounds, that the eclogites were overprinted by the blueschists. Preliminary garnet-pyroxene thermometry on the eclogites suggest temperatures of 400 -500°C.

An acid gneiss occurs a few meters below the eclogites. Its mineral assemblage contains albite, quartz, biotite and garnet porphyroblasts. Lower in this section greenschist mineralogy becomes dominant in basic rocks and no traces of the high pressure mineralogy can be found.

Way to station 3. Continue to walk along the trail down into the valley and cross to the next ridge. On the way a distinct horizon of fine-grained quartzite with a strongly defined NE trending fold axis and a parallel stretching lineation can be observed.

Station 3. Blueschists, eclogites and greenschists.

The outcrop here consists of a 2 meter block located near the crest of the ridge. Here interlayering of blueschists, eclogites and greenschists can be studied. The block is mostly massive, consists of fine grained glaucophane, with some epidote, garnet and white mica. Within the glaucophane rich block, eclogite layers from a few millimeters to several centimeters thick are found. The layers are frequently boudinaged, and sometimes occur as isolated lenses. In thin section, the transformation of omphacite to glaucophane can be observed. The block as a whole is surrounded by a greenschist facies matrix. Attached to its top is a mafic greenschist layer composed of albite, chlorite and actinolite. The

albite is poikiloblastic and undeformed, indicating that a deformation phase must separate the nucleation of the albite from the formation of blueschists and associated eclogitic rocks.

Way to station 4: Walk along the ridge crest towards the sea. On the way, a ten meters thick, vertical-walled volcanic dike can be observed to cut the metamorphic rocks. The dike is andesitic, trending NW and is a marker for NE-SW extensional processes which occurred in the late Miocene. Station 4 is located near the dike close to its eastern wall, about 100 m down hill.

Station 4. Fresh jadeite gneisses.

The jadeite gneisses occur as large blocks, a few meters in size, and could either be mega-boudins or knockers.

The rock is hard and massive and has a greenish color. The matrix is made of blocky sodic clinopyroxenes and garnet porpyroblasts. The pyroxene includes two compositional varieties: An almost pure jadeitic pyroxene with 82 mol% jadeite component; and a more acmitic jadeite with only 65 mol% jadeite. The rock also contains some quartz, phengite, rutile and sphene. Minimum pressures estimated from the jadeite 82 mol% composition at temperatures of 450 -500°C, are around 13 Kbars.

Way to station 5. Walk uphill back to the trail and turn west. On the way, cross again the volcanic dike and a quartzite. Cross the gully to the next ridge to the north west.

Station 5. Tectonic contact between the blueschist unit and the Upper unit.

Along the trail, near a small cement bridge, the contact with the Upper metamorphic unit is exposed. The contact is defined by a few meter thick zone of brecciation. The rocks of the upper unit in this area consist of fine grained phyllites. There are no signs of mylonitization along the contact and the emplacement evidently occurred in a brittle regime. As we have discussed in section 3.3, this rather unspectacular contact is connected with a large displacement.

Way to station 6. Cross the ridge and the section of the upper unit to the west and walk along the path to the beach.

Station 6. Superposed folding, the "Cycladic" and "Naxotian" trends.

The cliffs that rise from the sea expose folds. At this station an upright fold hinge is exposed at the cliff face (be careful!). The axes trend 020° and form the continuation of a 020° trending D3 deformation structure that extends as far north as the village of Kalloni. Folded pelitic samples taken from the hinge show strong deformation of albite blasts. This indicates that the deformation took place after the formation of the albites which developed during the greenschist overprint. Thus the 020° trend affected the rocks more recently than 24 Ma. ago. A fold generation with similar relation to metamorphism, of the same age and with a similar trend, was recognized on Naxos (Jansen 1976). The trend is therefore termed the "Naxotian trend".

On the western limbs of the fold at this station, just few meters from the hinge, small microcrenulations occur. The crenulations trend 060° and are folded by the Naxotian trend. These crenulations are related to the isoclinal folds of the same trend that can be seen on the way back along the beach. The 060° trending folds are part of the D1 "Cycladic" folding phase.

Way to station 7. Walk south along the beach. The next station is located at the NW tip of the sandy beach west of Kionia.

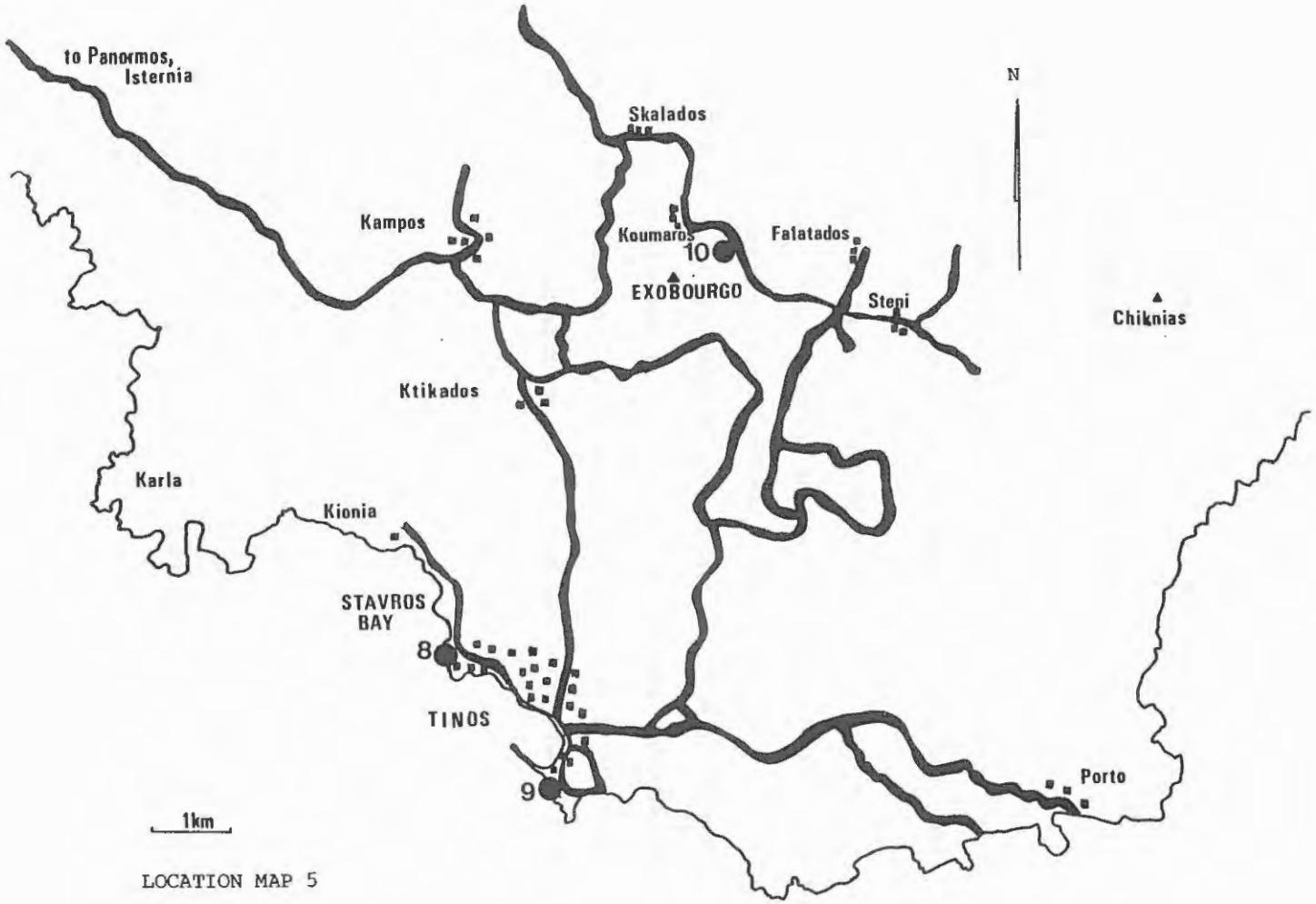
Station 7. Lawsonite pseudomorphs.

On the cliffs bordering the beach, white pseudomorphs after lawsonite are widespread. The pseudomorphs are up to 2 centimeters in length and are embedded in a mafic matrix. The pseudomorphs consist of zoisite, paragonite and albite. These pseudomorphs indicate that conditions in the higher temperature parts of the blueschist facies were attained during progressive metamorphism. The matrix consists of chlorite, albite, actinolite and epidote.

Way to station 8. Walk along the beach to the Tinos beach hotel. Continue to walk along the road towards Tinos town. (location map 5).

Station 8: Metagabbro of the Upper unit.

In the area of Stavros a large block of metagabbro is exposed. The coarse grained metagabbro consists of zoned hornblende and plagioclase and lacks deformation structures. The block is underlain by phyllites. Across the road, at the base of this Upper unit slice, serpentinite can be seen.



LOCATION MAP 5

Station 9. The amphibolite of the Upper unit.

Near the football ground in Tinos town, a thick block of amphibolite belonging to the Upper unit is exposed. The base of this block and the contact with the underlying blueschist unit cannot be seen, but traces of the upper unit can be seen in Tinos town, and the contact is fully exposed in a road cut near the windmills above the town, where it is marked by serpentinites. The amphibolites are layered on a scale of a centimeters to meters. K-Ar geochronological studies (Patzak 1988, and ages obtained in the laboratory of geochronology of the Geological Survey, Jerusalem) yielded Upper Cretaceous ages.

DAY 2. THE MIOCENE GRANODIORITE. THE BLUESCHISTS AND GREENSCHISTS NEAR M2 AND M3 MARBLES. THE UPPER UNIT IN MARLAS. THE TECTONIC WINDOW OF PANORMOS BAY. LOCATION MAP 5. (station 10) LOCATION MAP 6. (stations 11 - 15)

Drive to Exobourgo above Tinos harbour.

Station 10. The granodiorite and its contact aureole.

The contact between the granodiorite and its country rock is exposed near the top of the mountain and the station is located in various points along the road that descends from the mountain to the village of Koumaros.

The contact aureole is defined by the development of plagioclase hornblende fels in mafic schists. In places calcsilicates and other skarn minerals are developed. Down the road a swarm of pegmatitic granitic veins occur. The granite consists of

quartz, plagioclase, alkali feldspar, biotite and hornblende and is more eccurately classified as monzogranite (Altherr et. al., 1982). Altherr et al. (1988) assigned an age of 18 Ma to it.

Way to station 11 (location map 6). Drive along the main E-W road to the junction to Isternia Bay. Walk down the road towards Isternia Bay. On the way blueschists and greenschists can be seen.

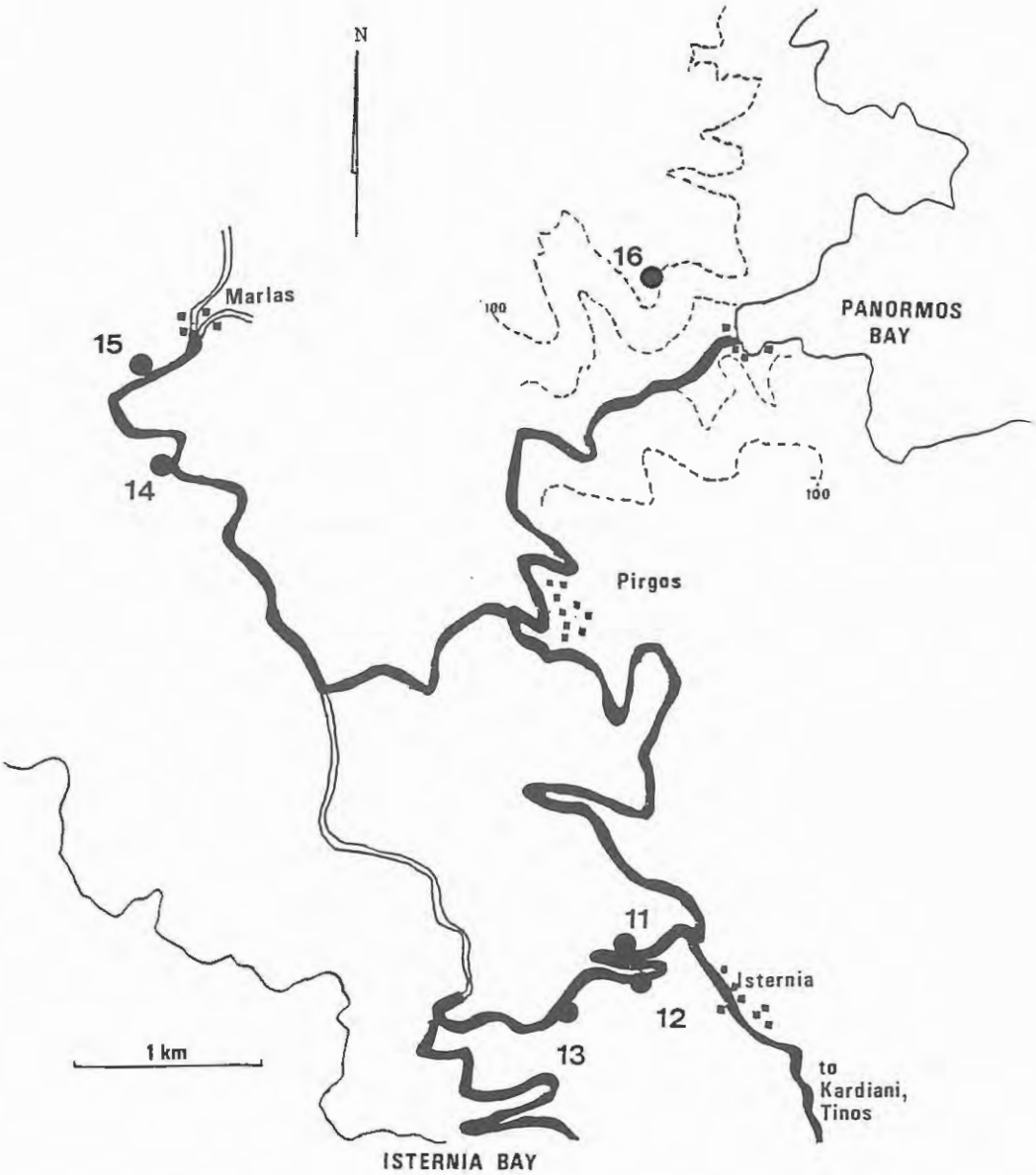
Station 11. Meta-ironstone and the Cycladic trend.

A block of meta-ironstone occurs within the section near the first road bend. The rock is made up of fine layers of quartz, which alternate with layers consisting of magnetite, glaucophane, garnet and epidote. The layers are isoclinally folded. The axes trend 050° and the folds are asymmetric and overturned to the SE. Thickening of quartz layers in the hinges can be observed. Glaucophane mineral lineation is developed along the fold axes. Glaucophane may be oriented at an angle oblique to the axes of the fold limbs. This folding phase belongs to the D1 "Cycladic trend". It is clear from the crystallization of glaucophane parallel to the fold axes that the deformation occurred during the high pressure metamorphic events.

Walk along the road down the hill.

Station 12. Folds of the the "Hellenic trend".

In semipelitic rocks, the developement of a new cleavage overprinting S1 can be observed. S1 here is strongly crenulated and a new crenulation cleavage is formed. The crenulation cleavage is subhorizontal and forms an axial plane to 320 trending folds.



LOCATION MAP 6

Under the microscope, the micas are crenulated and deformed, but albite is almost unaffected and only slightly rotated parallel to the new cleavage. Near the village of Kardiani, crossite coexisting with albite grains has been observed to overgrow these microcrenulations. Thus the folds predate the transformation to greenschist facies, and perhaps even precede a phase of epidote-blueschist facies equilibration represented by the crossite-albite assemblages. This phase is similar in trend and relation to metamorphic history to the 320° "Hellenic trend" in Sifnos.

Way to station 13. Continue to walk along the road towards the contact of the marble (M2). Station 13 is located several tens of meters before the road enters the marble.

Station 13. Blueschist-greenschist transformation.

Layers of blueschists and greenschists, several centimeters in thickness, alternate along the road cut at station 13. Here the nature of the blueschist to greenschist transformation is well established in thin section. A transition occurs from a glaucophane, epidote, garnet and phengite-bearing rock, to chlorite, albite, actinolite and epidote bearing greenschists, with all gradational stages being observable. Texturally the schistose fabric of the blueschists is replaced by greenschist minerals in which the fabric is dominantly static and undeformed. Microstructurally the transformation can be seen to occur on a millimetric scale. Blueschists and partially transformed rocks are dissected by a net of anastamosing micro shear zones, but in the

greenschists no traces of these shear zones remain. In partially transformed rocks albite and chlorite preferentially nucleate along the small shear zones. The shear zones probably served as channels for the metamorphic fluids necessary for the hydration reactions of the blueschist- greenschist transformation. Growth of the new minerals probably affected the rocks in two ways: filling up the micro channels and inhibiting further fluid flow; erasing the fabric of the microshears.

Way to station 14. Walk or drive around M2 on the unpaved road towards the village of Marlas. Metasediments form most of the section along the way. Get to the main road and walk towards the village of Marlas.

Station 14. Contact of the Upper unit with the lower blueschist unit.

In a small gully on the left of the road near Marlas, the contact with the upper unit is exposed. Phyllites of the upper unit are kinked and deformed. Slip fiber serpentine and talc occur along the faulted zone.

Station 15. A block of ophicalcites within the basic phyllitic matrix of the Upper unit.

In the quarry near the village of Marlas, a large block of ophicalcite more than 100 meters in size is exposed. The ophicalcite is made up of serpentinite and calcite. Accessory minerals include chromite mantled by magnetite, chlorite, talc and tremolite. The ophicalcite is a sedimentary or tectonic breccia

which includes the erosional products of an ultramafic ophiolite suite. On the right side of the quarry, the contact of the block with the surrounding basic phyllites is well exposed. The contact is subvertical, discordant with the layering in the surrounding phyllites, and some mineralization occurs along the contact. The ophicalcite block extends to the other side of the road and down into the valley. It was extensively quarried as the "green marble" of Tinos.

Way to station 15. Walk along the road to the east on the flanks of the valley towards the bay of Panormos. On the way cross the contact of the upper unit with the underlying blueschist. Walk to the lower marble (M1) near Panormos. Descend through a gray calcite marble containing numerous quartzite horizons, to the massive dolomite section. Station 15 is located at the contact between the calcite marble and the dolomite.

Station 16. Phyllite in M1.

The rocks in the station consist of a 2 meter thick layer of micaceous phyllites intercalated within fine quartzitic layers. The phyllites are of low metamorphism grade and high pressure minerals are totally absent. A thick gray calcite marble with abundant quartzite layers outcrop above the phyllite. A massive dolomitic sequence forms the base of the section. The dolomite is poorly recrystallized and contains fossils. It was suggested in the section 3.3 that the phyllites are in metamorphic and structural discordancy with the overlying blueschist rocks. The gray calcite marbles belong to the overlying blueschist unit, whereas the

phyllites and the fossiliferous dolomite constitute a different low grade tectonic unit.

4. REFERENCES

Altherr, R., Keller, J. and Kott, K., Der jungtertiare monzonit von Kos und sein Kontakthof (Aegaeis, Griechenland). Bull. Soc. geol. France, 18:7 (1976) pp. 403-412.

Altherr, R. and Seidel, E., Speculations on the geodynamic evolution of the Attic-Cycladic crystalline complex during Alpidic times. In D. Kallergis (ed.), Proceedings of the VI Colloquium on the Geology of the Aegean Region, Inst. Geological and Mining Research, Athens, Greece, (1977) pp. 347-352.

Altherr, R., Schliestedt, M., Okrusch, M., Seidel, E., Kreuzer, H., Harre, W., lenze, H., Wendt, I. and Wagner, G.A., Geochronology of high pressure rocks on sifnos (Cyclades, Greece). Contrib. Mineral. Petrol., 70 (1979) pp. 245-255.

Altherr, R., Kreuzer, H., Wendt, I., Lenz, H., Wagner, G.A., Keller, J., Harre, W. and Hohndorf, A., A Late Oligocene/Early Miocene high temperature belt in the Attic-cyclades crystalline complex (SE Pelagonian, Greece). Geol. Jb., E23 (1982) pp. 97-164.

Altherr, R., Henjes-Kunst, F., Matthews, A., Friedrichsen, H. and Hansen, B.T., O-Sr isotopic variations in Miocene granitoids from the Aegean: Evidence for an origin by combined assimilation and fractional crystallization. Contrib. Mineral. Petrol., (1988) in press.

Andriessen, P.A.M., Boelrijk, N.A.I.M., Hebeda, E.H., Priem, H.N.A., Verdurmen, E.A.Th. and Verschure, R.H., Dating of the events of metamorphism and granitic magmatism in the Alpine Orogen of Naxos (Cyclades, Greece). Contrib. Mineral. Petrol., 69 (1979) pp. 215-225.

Andriessen, P.A.M., Banga, G. and Helbeda, E.H., Isotopoc age study of pre-Alpine rocks in the basal units on Naxos, Sikinos and Ios, Greek Cyclades. Geol. Mijnbouw, 66 (1987) pp. 3-14.

Angelier, J., Tectonic evolution of the Hellenic Arc since the Late Miocene. Tectonophysics, 49 (1978) pp. 23-36.

Avigad, D., The structural history associated with the Eocene blueschist metamorphism on Sifnos (Cyclades) Greece. Terra Cognita, 7:2-3 (1987) p. 169.

Avigad, D. and Garfunkel, Z. Low angle faults underneath and above a blueschist belt - Tinos Island, Cyclades, Greece. (1988) Submitted to Nature.

Baltatzis, E., Contact metamorphism of a calc-silicate hornfels from Plaka area, Laurium, Greece. N. Jb. Miner. Mh., 1981 (1981) pp. 481-488.

Blake Jr., M.C., Bonneau, M., Geyssant, J., Kienast, J.R., Lepvrier, C., Maluski, H. and Papanikolaou, D., A geologic reconnaissance of the Cycladic blueschist belt, Greece. Geol. Soc. Am. Bull., 92:I (1981) pp. 247-254.

Biju-Duval, B., Dercourt, J. and Le-Pichon, X., From the Tethys ocean to the Mediterranean seas: A plate tectonic model of the evolution of the Western Alpine system. In: Structural History of the Mediterranean Basins, Split, (1976) pp. 143-164.

Bonneau, M., Correlation of the Hellenide nappes in the south-east

Aegean and their tectonic reconstruction. In J.E. Dixon and A.H.F. Robertson (eds), The Geological Evolution of the Eastern Mediterranean, Blackwell Scientific Publications, London, (1984) pp. 517-527.

Broecker, M. and Okrusch, M., Geochemistry of metabasites and associated metasediments from Tinos Island, Cyclades. Terra Cognita, 7:2-3 (1987) p. 196.

Davis, E.N., Der geologische Bau der Insel Siphnos. Geological and Geophysical Research, Athens, 1966.

Durr, S., Altherr, R., Keller, J., Okrusch, M. and Seidel, E., The median Aegean crystalline belt: Stratigraphy, stucture, metamorphism, magmatism. In: H. Cloos, D. Roeder, and K. Schmidt (eds), Mediterranean Orogens, Inter-Union Commission on Geodynamics, Scientific Rep., (1978) pp. 455-477.

England, P.C., and Thompson, A.B., Pressure-temperature-time paths of regional metamorphism, I, heat transfer during the evolution of regions of thickened continental crust. J. Petrol., 25, (1984) pp. 894-928.

Evans, B.W., Reactions among sodic, calcic, and ferromagnesian amphiboles, sodic pyroxene, and deerite in high-pressure metamorphosed ironstone, Siphnos, Greece. Am. Mineral., 71 (1986) pp. 1118-1125.

Evans, B. W. and Brown, E.H., Reply to Comment on Blueschists and Eclogites. Geology, (1987) pp. 773-775.

Fytikas, M., Innocenti, F., Manetti, P., Mazzouli, R., Peccerillo, A. and Villari, L., Tertiary to Quaternary evlution of volcanism in the Aegean region. In J.E. Dixon and A.H.F. Robertson (eds.), The Geological Evolution of the Eastern Mediterranean, Blackwell Scientific Publications, London, (1984), pp. 687-700.

Garlick, G.D., Oxygen isotope fractionation in igneous rocks. Earth Planet. Sci. Lett., 1, (1966) pp. 361-368.

Godfriaux, I., L'Olympe. Bull. Soc. geol. France, 19:1 (1977) pp. 45-49.

Henjes-Kunst, F. and Kreuzer, H. Isotopic dating of Pre- Alpidic rocks from the Island of Ios (Cyclades, Greece). Contrib. Mineral. Petrol., 80 (1982) pp. 245-253.

Holland, T.J.B., Preliminary phase relations invoving glaucophane and applications to high presure petrology: New heat capacity and thermodynamic data. Contrib. Mineral. Petrol., 99 (1988) pp. 134-142.

Holland, T.J.B. and Powell, R., An internally consistent thermodynamic dataset with uncertainties and correlations. 2. Data and results. J. Metam. Geol., 3 (1985) pp. 343-370.

Jansen, J.B.H. and Schuiling, R.D., Metamorphism on Naxos: Petrology and geothermal gradients. Am. J. Sci., 276 (1976) pp. 1225-1253.

Jansen, J.B.H., The Geology of naxos. Geological and Geophysical research 1, I.G.M.R., Athens, (1977) 100 p.

Le Pichon, X. and Angelier, J., The Hellenic arc and trench system: a key to the neotectonic evolution of the Eastern Mediterranean area. Tectonophysics, 60 (1979) pp. 1-42.

Makris, J., The crust and upper mantle of the Aegean region from

deep seismic soundings. Tectonophysics, 46 (1978) pp. 269-284.

Matthews, A., Oxygen isotope thermometry in regional metamorphism. International Congr. Geochem. Cosmochem., Paris, (1988).

Matthews, A. and Schliestedt, M., Evolution of the blueschist and greenschist facies rocks of Sifnos, Cyclades, Greece. Contrib. Mineral. Petrol.,88 (1984) pp. 150-163.

Melidonis, N.G., The Geology of Greece: The geological structure and mineral deposits of Tinos island (Cyclades - Greece) (A Preliminary Study). Inst. Geol. Mineral. Exploration, Athens, 1980.

Okrusch, M., Seidel, E. and Davis, E.N., The assemblage jadeite-quartz in the glaucophane rocks of Sifnos (Cyclades Archipelago, Greece). N. Jb. Miner. Abh., 132:3 (1978) pp. 284-308.

Papanikolaou, D.J., Contribution to the geology of Aegean Sea: The Island of Andros. Ann. Geol. Pays Helleniques, 29 (1978) pp. 477-553.

Papanikolaou, D.J., The three metamorphic belts of the Hellenides: a review and a kinematic interpretation. In J.E. Dixon and A.H.F. Robertson (eds.), The Geological Evolution of the Eastern Mediterranean, Blackwell Scientific Publications, London, (1984), pp. 551-561.

Patzak, M., Der Amphibolit-Gneis-Koerper von Akrotiri, Insel Tinos (Greichenland). Diplomarbeiten fur das Studienfach Mineralogie, Universitat Wurzburg, 1988.

Powell, R. and Holland, T.J.B., An internally consistent dataset with uncertainties and correlations. 3. Application to geobarometry, worked examples and a computer program., J. Metam. Geol., (1988).

Ridley, J., Evidence of a temperature-dependent 'blueschist' to 'eclogite' transformation in high-pressure metamorphism of metabasic rocks. J. Petrol., 25:4 (1984) pp. 852-870.

Salemink, J., Skarn and ore formation at Serifos, Greece, as a consequence of granodiorite intrusion., Geologica Ultrajectina, 40 (1985) pp. 1-232.

Schliestedt, M., Phasengleichgewichte in Hochdruckgesteinen von Sifnos, Griechenland. Dr.rer.nat. thesis, Univ. Braunschweig, F.R.G., (1980).

Schliestedt, M., Eclogite-blueschist relationships as evidenced by mineral equilibria in the high-pressure metabasic rocks of Sifnos (Cycladic Islands), Greece. J. Petrol., 27:6 (1986) pp. 1437-1459.

Schliestedt, M. and Okrusch, M., Meta-acidites and silicic metasediments related to eclogites and glaucophanites in Northern Sifnos. In D.C. Smith (ed.), Developements in petrology: Eclogites and Eclogite-facies Rocks, Elsevier, Amsterdam, (1988).

Schliestedt, M., Altherr, R. and Matthews, A., Evolution of the Cyclades crystalline complex: Petrology, Isotope geochemistry and geochronology. In H.C. Helgeson (ed.), Chemical Transport in Metasomatic Processes, Reidel Publishing Company, (1987) pp. 389-428.

Schliestedt, M., and Matthews, A., Transformation of blueschist to greenschist facies rocks as a consequence of fluid

infiltration, Sifnos (Cyclades), Greece. Contrib. Mineral. Petrol., 97 (1987) pp 237-250.
Seidel, E., Kreuzer, H. and Harre, W., A Late Oligocene- Early Miocene high pressure belt in the External Hellenides. Geol. Jahrb., E23 (1982) pp. 165-206.

רשימות