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METAMORPHISM ON NAXOS: PETROLOGY AND GEOTHERMAL GRADIENTS

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ABSTRACT. The regional metamorphism of metasediments and related rocks on Naxos, Greece, is in general of the kyanite–sillimanite type, although locally the three Al-silicates have been found in the same rock. From a study of the metamorphic parageneses in pelitic schists, marbles, meta-volcanic rocks, metamorphosed ultramafic rocks, and meta-bauxites, it can be inferred that the metamorphism took place within a gradually increasing temperature range from about 400°C in southeast Naxos to 700°C in the central part of the metamorphic complex under a total pressure of around 5 to 7 kb. The geothermal gradients responsible for the regional metamorphism in the Cycladic Massif were significantly different over horizontal distances of 10 to 20 km; 30°C per km in the western part of the metamorphic complex on Naxos, 22°C per km in southeast Naxos, and less than 15°C per km on the islands of Ios, Siphnos, and Syros.

INTRODUCTION

The island of Naxos, Cyclades (see fig. 1), Greece, is roughly oval-shaped and measures about 20 by 35 km. Geologically, it consists mainly of a domed metamorphic complex and a pluton, granodioritic in composition, which occupies the western part of the island (fig. 2). Naxos forms part of the Attic-Cycladic Massif, of probable Alpine age. Age determinations now in progress indicate a Late Tertiary age for the granodiorite, which is intrusive into the metamorphic complex. Rb-Sr mica ages are young, ranging from 6 to 13 m.y. for muscovites from the high-grade migmatites and gneisses to 40 m.y. for muscovites from the low-grade micaschists of the metamorphic complex. The biotite of the granodiorite shows a mineral age of about 12 m.y. (Priem and others, 1969; Andriessen, personal commun., 1975). Fossiliferous unmetamorphosed Permian limestones have been found on Naxos (Marks and Schuiling, 1965) and on Mykonos (Papastamatiou, 1963). On another island of the massif, Antiparos, the Permian seems to be slightly metamorphosed (Anastopoulos, 1963). According to our present interpretation, the Permian in the central part of the Cyclades belongs to an allochthonous tectonic unit emplaced during the Miocene (for Naxos, Jansen, 1973).

The contact relations of the granodiorite show unambiguously that it is intrusive into the metamorphic sequence. It has developed an aplitic, tourmaline-rich border facies near the contact with the metamorphic rocks, and veins of aplite and granodiorite cut through the country rocks. The latter show evidence of contact-metasomatism, as exemplified by skarn formation and minor scheelite mineralization, and contact-metamorphism of the andalusite–sillimanite type (fig. 3).

The main part of the island consists of a metamorphic complex, the core of which is formed by a migmatitic gneiss dome. This metamorphic complex consists mainly of an alternation of calcitic and dolomitic marbles with mica schists. The carbonate rocks are predominant in the

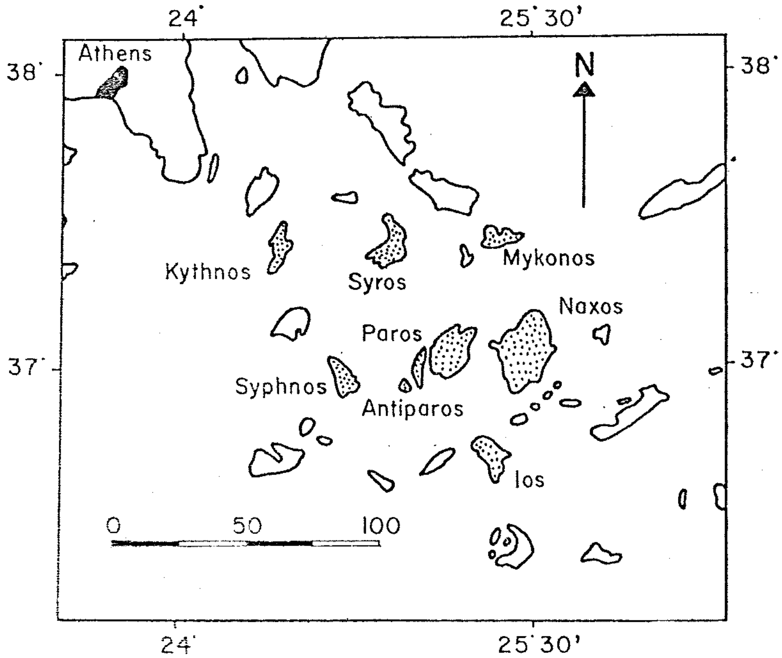


Fig. 1. Location of the Cyclades, Greece. The dotted islands are mentioned in the text.

upper part of the sequence, whereas in the stratigraphically lower part mica schists, gneisses, and amphibolites predominate. There is also a marked variation in lithology along strike. Minor, but distinctive rock types are formed by metamorphosed ultramafic rocks and by metamorphic bauxites which occur in most of the marble beds throughout the whole metamorphic series (fig. 2). This rather special lithology lends support to the thesis that the whole metamorphic series is part of one single tectonic and stratigraphic unit and is not composed of different unrelated tectonic units superimposed on each other. However, the two different, discontinuous horizons of ultramafic bodies may also be considered as remnants of ophiolites emplaced along premetamorphic thrust-planes. Although folding is fairly common, there is a general lack of large scale isoclinal folding, apart from some flow-folds in the high-grade marbles. Isoclinal folding with limbs up to a few meters are found through the whole sequence.

There is a very pronounced north 15° east lineation in the metamorphic rocks, which has also been noted in other metamorphic massifs in the Eastern Mediterranean, such as in the Rhodope Massif, the Kaz-Dag complex, and the Menderes Massif (Schuiling, 1962a). This direction of lineations and fold-axes has been interpreted as evidence of Hercynian events in these massifs (Trikkalinos, 1947; Schuiling, 1962a). In the light of the fact that updoming with a north 15° east axis and

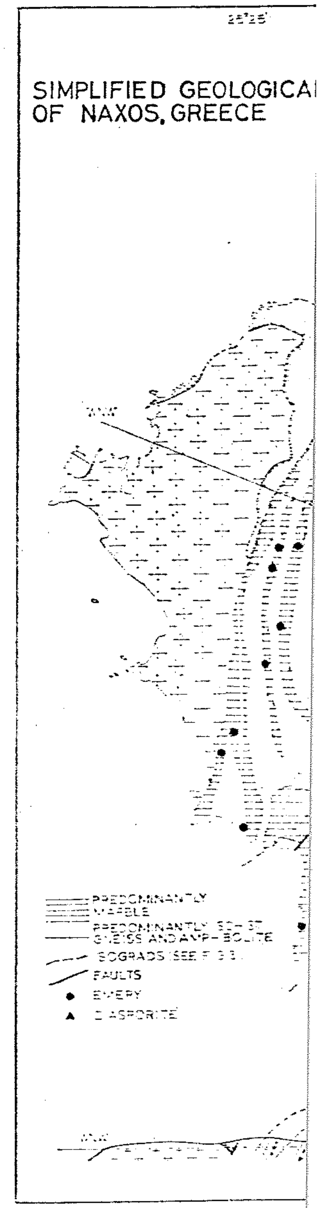
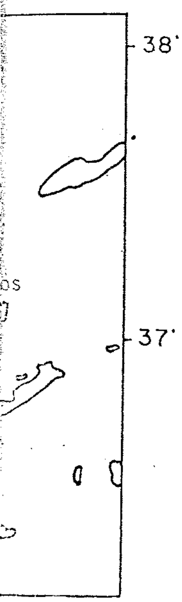


Fig. 2. Simplified geological
west-northwest east-southeast. N
through the whole metamorphic
outlines of the lithologic units.



are mentioned in the

ically lower part
There is also a
t distinctive rock
and by metamor-
s throughout the
l lithology lends
es is part of one
posed of different
However, the two
may also be con-
tamorphic thrust-
general lack of
n the high-grade
neters are found

ion in the meta-
morphic massifs
Massif, the Kaz-
62a). This direc-
as evidence of
chuiling, 1962a).
5° east axis and

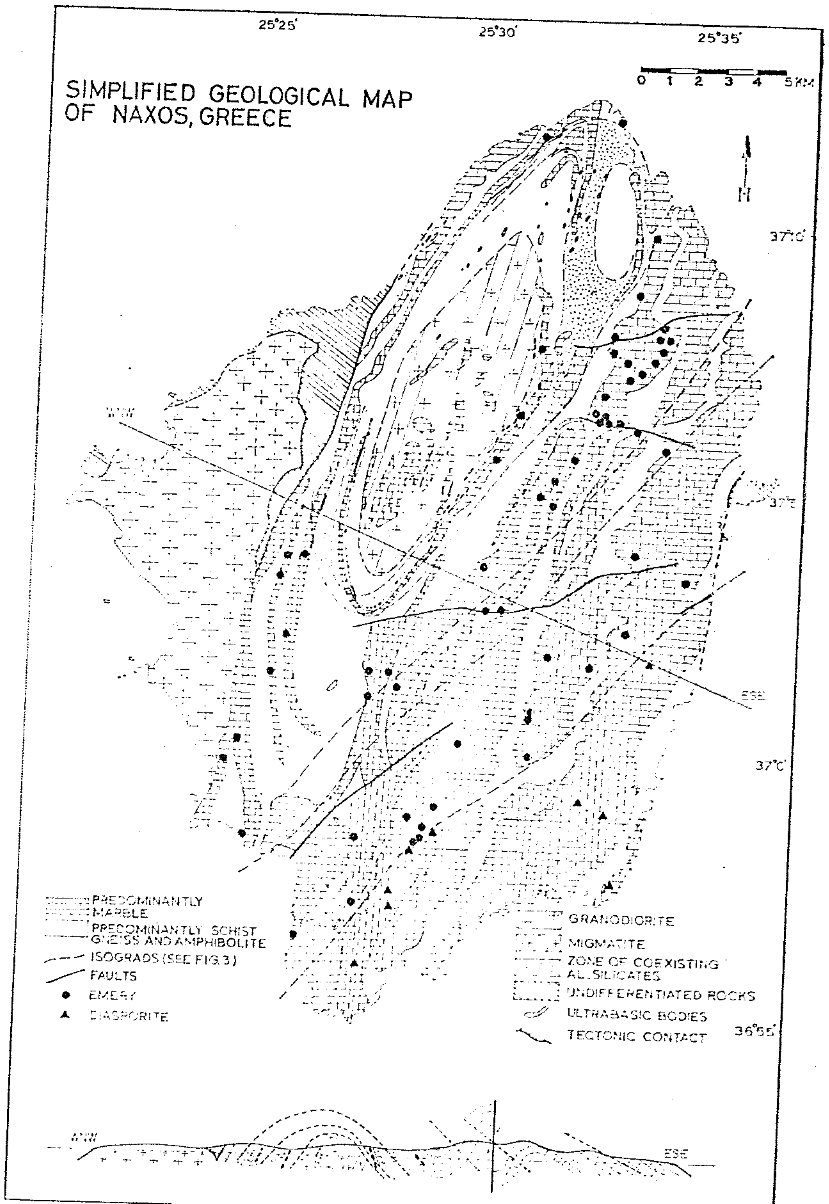


Fig. 2. Simplified geological map of Naxos, Greece, with schematic cross section west-northwest east-southeast. Note the distribution of the meta-bauxite occurrences through the whole metamorphic complex and the fact that the isograds intersect the outlines of the lithologic units.

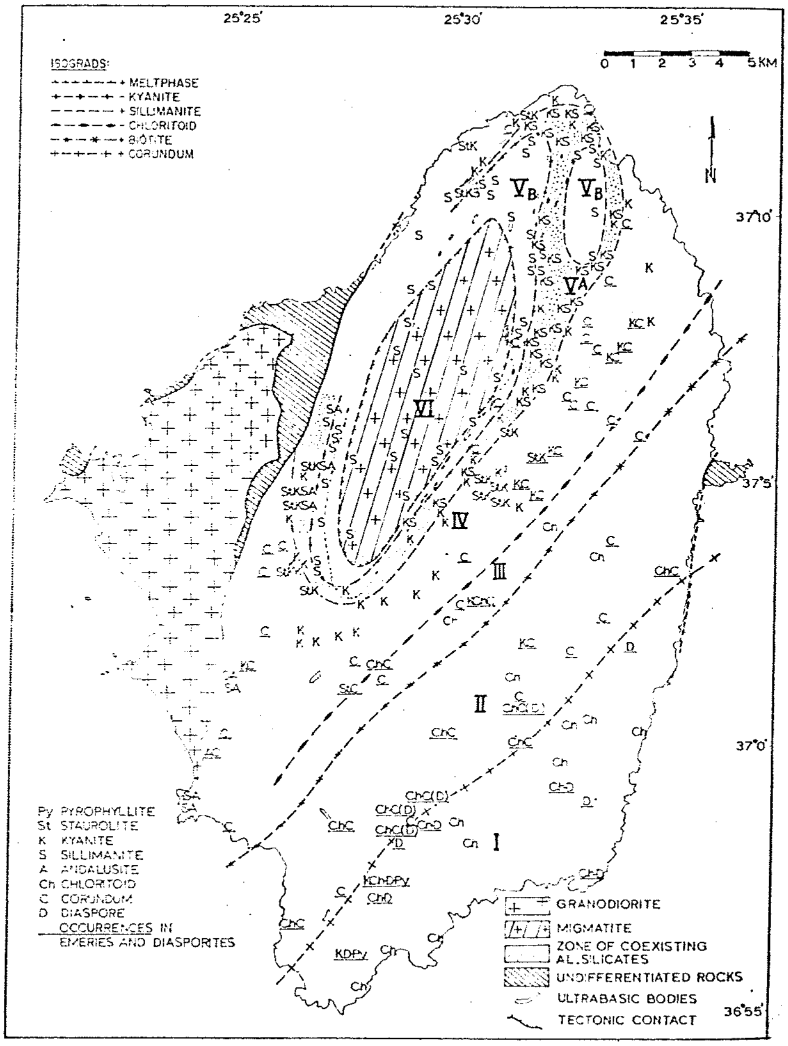


Fig. 3. Petrological map of Naxos, Greece, showing the distribution of several aluminous minerals. Roman numbers indicate metamorphic zones: I = diaspore zone, II = chlorite-sericite zone, III = biotite-chloritoid zone, IV = kyanite zone, V_A = kyanite-sillimanite transition zone, V_B = sillimanite zone, and VI = migmatic zone. In order of increasing grade of metamorphism toward the center of the dome the isograds are respectively the + corundum, + biotite, — chloritoid, + sillimanite, — kyanite, and + meltphase.

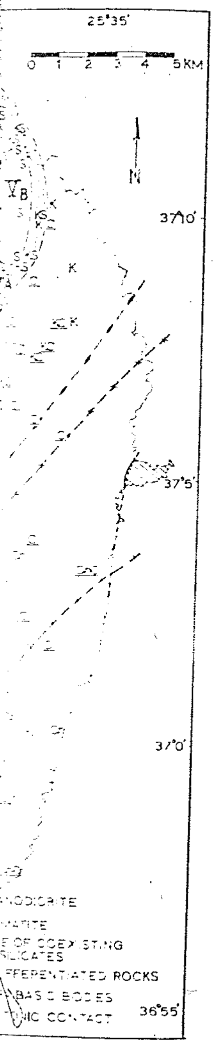
metamorphic crystallizati
must be abandoned, at l
cerned.

It seems unlikely th
within the metamorphic
was rejuvenated during
the map of figure 2, cert
does also a discontinuous
It is likely, therefore, tha
melted equivalent of the

The metamorphic gr
decreasing distance to the

Zones and Isograds / Minerals
Pelitic compositions
Albite
Oligoclase-Andesine
Paragonite
Biotite
Muscovite
Chlorite
Epidote, zoisite
Kyanite
Sillimanite
Andalusite (west)?)
Staurolite (west)?)
Staurolite
Chloritoid
Garnet
K-feldspar
Carbonate-rich compositions
Albite
Plagioclase > 15% An.
Muscovite
Phlogopite
Chlorite
Epidote-zoisite
Talc
Tremolite
Diopside
Scapolite
Grossularite
Vesuvianite
Hematite
Magnetite

Fig. 4. Range of occurren
carbonite-rich compositions in
parts of the line indicate minor
side of the migmatite dome, an
grade of metamorphism.



distribution of several
nes: I = diaspore zone,
= kyanite zone, V_A =
I VI = migmatic zone.
enter of the dome the
loritoid, + sillimanite,

metamorphic crystallization seem to be Alpine events, this interpretation must be abandoned, at least as far as the Attic-Cycladic Massif is concerned.

It seems unlikely that the migmatite dome, as it manifests itself within the metamorphic complex, is part of an earlier basement, that was rejuvenated during regional metamorphism. As can be seen from the map of figure 2, certain marble beds enter the migmatite dome, as does also a discontinuous horizon of metamorphosed ultramafic bodies. It is likely, therefore, that the migmatite dome is a high-grade, partially melted equivalent of the mica schists and gneisses that lie on top of it.

METAMORPHIC ZONES

The metamorphic grade of the rocks increases in a regular way with decreasing distance to the central migmatic dome. The mapped isograds

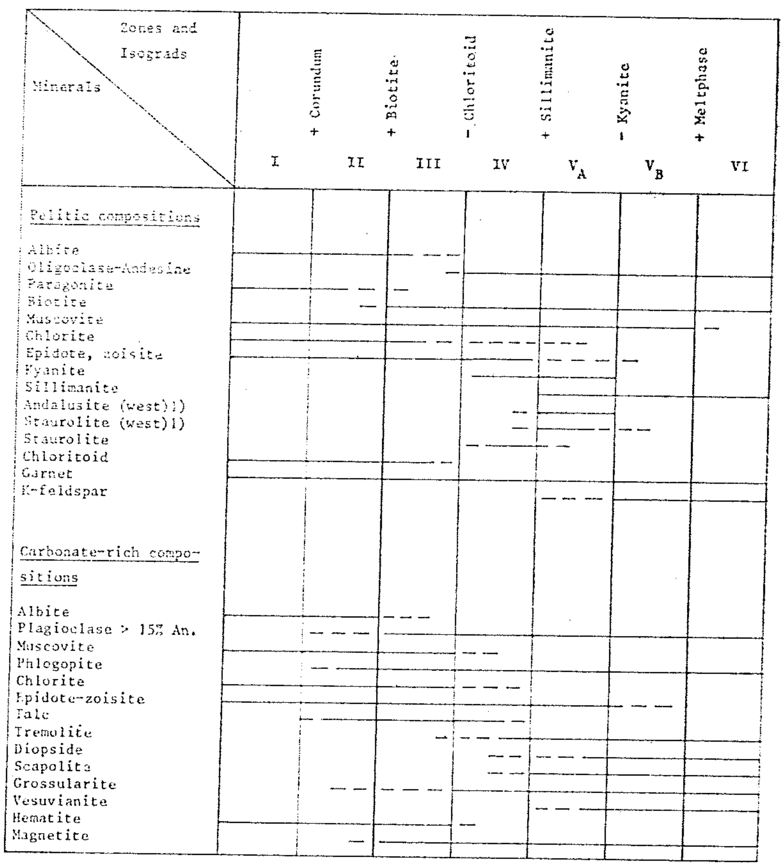


Fig. 4. Range of occurrence of distinctive minerals in rocks with pelitic- and carbonite-rich compositions in the metamorphic complex of Naxos, Greece. Dashed parts of the line indicate minor or rare occurrences. Andalusite occurs on the western side of the migmatite dome, and staurolite on that side seemingly shifts to a higher grade of metamorphism.

and zone boundaries make a division in metamorphic zones possible. For the distribution of the individual zones the reader is referred to figure 3. The metamorphic zones, together with the isograd or zone boundary that marks the beginning of the specific zone, are described in order of increasing grade of metamorphism. The range of occurrences of distinctive minerals of the pelitic- and carbonate-rich rocks is given in figure 4; of the amphibolites (and their low grade equivalents), of the metamorphosed ultramafic rocks, and of the meta-bauxites in figure 5. A selection of representative mineral assemblages observed in thin sections is arranged for each distinctive rock type in all the metamorphic zones in table 1. The location of these samples is given in figure 6. Some additional data on lithology and veining of each individual zone are given.

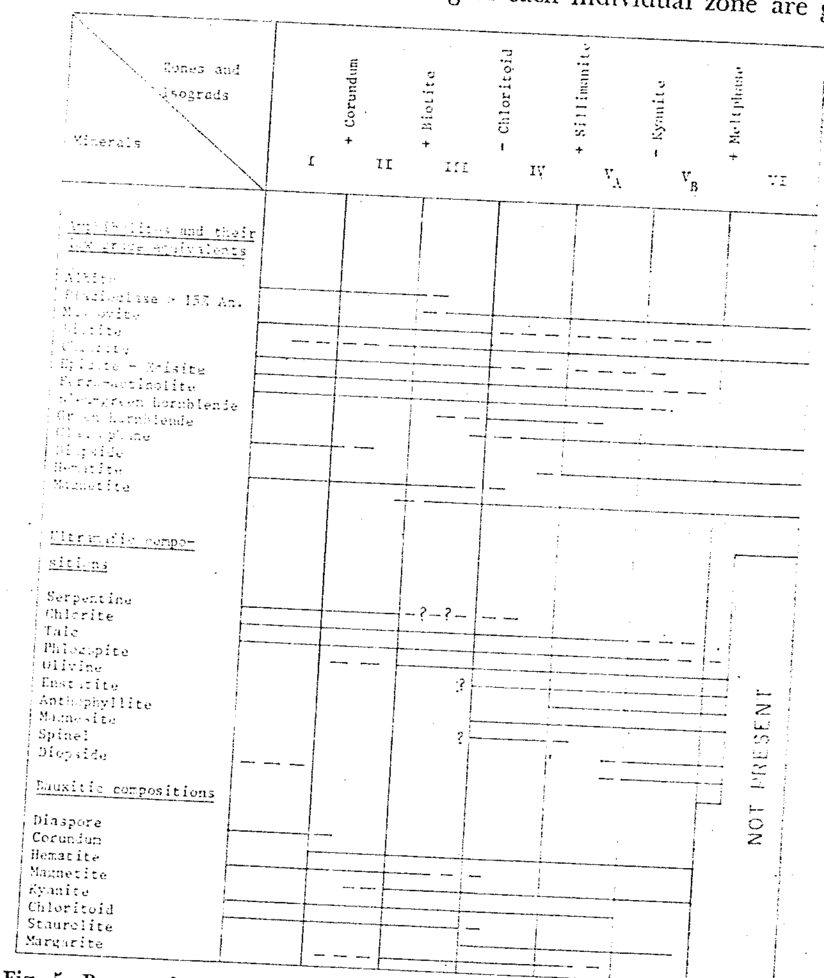


Fig. 5. Range of occurrences of distinctive minerals in amphibolites (and their low grade equivalents) and in rocks with ultramafic and bauxitic compositions in the metamorphic complex of Naxos, Greece.

The zones are named after the most characteristic or a frequently-occurring mineral or mineral assemblage.

DIASPORE ZONE (I)

The lowest grade of metamorphism is found in southeastern Naxos where the rocks consist mainly of marble with several lenses of meta-bauxite. A few pelitic schist horizons also occur, sometimes with intercalations of layers of presumably metavolcanic rock. These layers are taken to be the low grade equivalents of the amphibolites. Only one body of metamorphosed ultramafic rock is found in this zone (fig. 2).

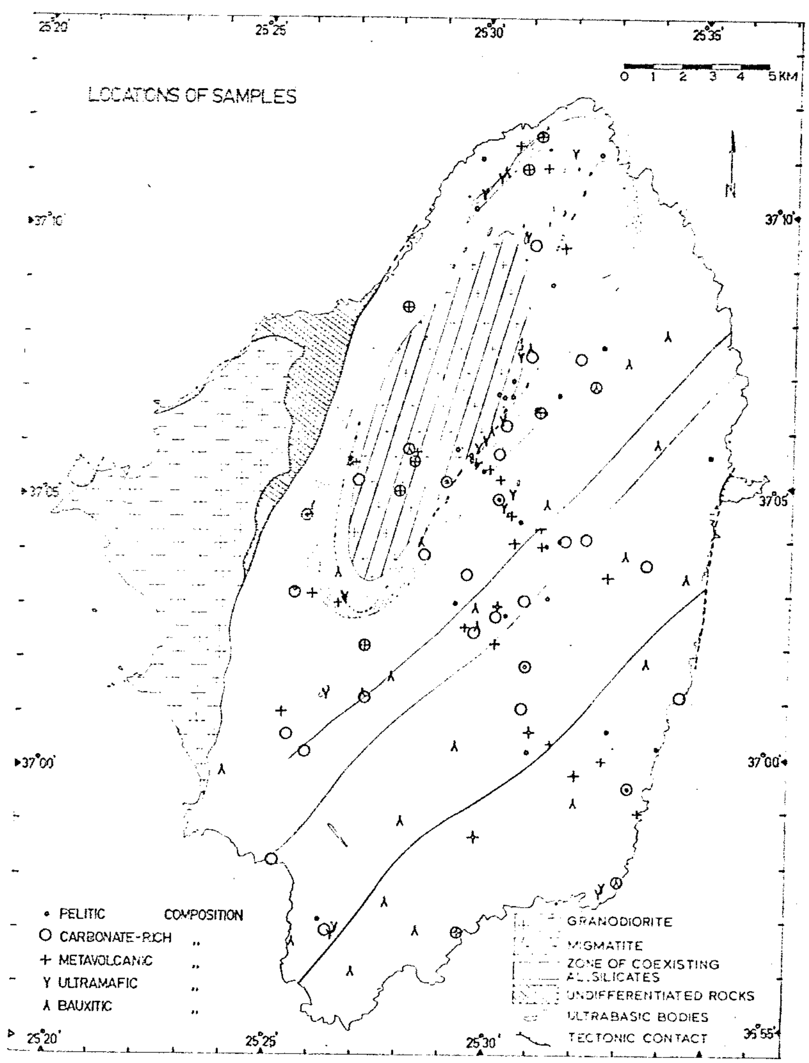


Fig. 6. Location of handspecimens mentioned in table 1.

TABLE 1

Pelitic compositions[1]																				
Zone	Sample number	Coordinates Latitude	Longitude	plagioclase	K-feldspar	paragonite	muscovite	biotite	kyanite	sillimanite	andalusite	garnet	staurolite	chloritoid	chlorite	epidote	amphibole	calcite	sphene	rutile
I	B232	37°00′10″	25°33′55″	A		o	x					x		x	o					o
	B437	36 59 30	25 33 15	A			x								x	o		o	o	
	S321B	36 56 50	25 29 25	A		o	x								x	x	x	o	o	
	S459B	36 59 45	25 32 00	A			x								x		x	o	o	
	B317	37 00 30	25 32 45				x							x	o					o
	B420C	36 58 40	25 29 40	A		o	x					x			x		x		o	
II	B510	37 00 10	25 31 00	A			x							x	x			o	o	R
	B310C	37 00 30	25 31 00	A			o								x	x	x		o	
	B224	36 57 10	25 26 10	A			x	G				x			o	x		o	o	
	Z78C	37 01 40	25 30 50	A			x					x			x	o			o	
	S183B	37 05 40	25 35 10	A			x								x				o	
	S423	37 03 00	25 31 20	A		o	x	G				x			x				o	
III	B189	37 02 40	25 30 25	A			x					x			x	x		o		
	Z54	37 04 00	25 31 20	A			x	x							x	x				
	Z56A	37 04 00	25 31 40	A			x	x						x	x					
	Z116	37 02 50	25 30 10	A			x	o							x					
IV	B239	37 02 55	25 20 20	Ol			x									x		o		R
	Z43	37 04 25	25 30 45	Ol			x	x					x		x					
	Z49A	37 04 30	25 30 30	Ol			x	x	x			x	x		x					o
	B543A	37 04 50	25 30 15	Ol			x	x				x			(o)				o	R
	S196	37 07 40	25 32 40	Ol			x	x	x			x								
	west																			
	B242	37 11 10	25 29 50	Ol			x	x	o			x	x		(o)					
	Z68C	37 03 10	25 25 40	Ol			x	x	o				o		(o)				o	
V_A	B209A	37 06 45	25 31 35	Ol	o		x	x	x	F		x	o							
	B150	37 06 25	25 31 20	Ol	o		x	x	x	x		x			(o)					
	B148	37 06 30	25 31 05	Ol	o		x	x	x	F		x								
	Z9	37 05 20	25 29 50	Ol			x	x	x	o		x								
	S74G	37 11 10	25 32 30	Ol			x	x	o	x		x								
	S119	37 08 50	25 31 25	Ol			x	x	x	F		x			(o)					o
	west																			
	B86BI	37 04 30	25 25 50	Ol	o		x	x		F	x		x							
	B86B6II	37 04 30	25 25 50	Ol			x	x		F	o	x	o			o				
	S148	37 10 10	25 29 40	Ol	o			x	x	F					(o)					
V_B	S390	37 03 05	25 26 45	Ol			x			x					(o)	x		(o)		
	Z6	37 05 30	25 29 45	Ol				x		F		x								
	Z3	37 05 35	25 29 35	Ol	x			x		x										
	Z2	37 05 40	25 29 35	Ad	x		x	x		F		o								
	S268A	37 06 45	25 30 35	Ad	x		x	x				o								
VI	S130J	37 05 10	25 29 00	Ol	x			x		F		x								
	B535	37 06 40	25 30 20	Ol	x		(o)	x		x		o								
	B542	37 07 00	25 30 30	Ol	x		(o)	x		o		x								
	B534	37 06 45	25 30 15	Ol	x		o	x				x								
	Z1	37 05 45	25 29 20	Ad	x			x		x		x								

	S74G	37 11 10	25 32 30	Ol		x	x	o	x		x					
	S119	37 08 50	25 31 25	Ol		x	x	x	F		x					
	west												(o)			o
	B86BI	37 04 30	25 25 50	Ol	o	x	x		F	x		x				
	B86B6H	37 04 30	25 25 50	Ol		x	x		F	o	x	o				
	S148	37 10 10	25 29 40	Ol	o		x	x	F					o		
													(o)			
V_B	S390	37 03 05	25 26 15	Ol		x			x				(o)	x	(o)	
	Z6	37 05 30	25 29 45	Ol			x		F		x					
	Z3	37 05 35	25 29 35	Ol	x		x		x							
	Z2	37 05 40	25 29 35	Ad	x	x	x		F		o					
	S268A	37 06 45	25 30 35	Ad	x	x	x				o					
VI	S130J	37 05 10	25 29 00	Ol	x		x		F		x					
	B535	37 06 40	25 30 20	Ol	x	(o)	x		x		o					
	B542	37 07 00	25 30 30	Ol	x	(o)	x		o		x					
	B534	37 06 45	25 30 15	Ol	x	o	x				x					
	Z1	37 05 45	25 29 20	Ad	x		x		x		x					

Carbonate-rich compositions[2]																				
Zone	Sample number	Coordinates Latitude	Coordinates Longitude	quartz	plagioclase	K-feldspar	muscovite	phlog-biotite	garnet	chlorite	epidote	hornblende	tremolite	talc	dolomite	diopside	scapolite	vesuvianite	corundum	margarite
I	B508	36°57′50″	25°33′00″	x			x			x										
	S321	36 56 50	25 29 25		A		x			x	x		x							
	B437X	36 59 30	25 33 15	x	A		x				x									
	B437W	36 59 30	25 33 15	x	A		x			x	x									
	B234	37 01 10	25 34 20		A		x			x			x							
II	B702[3]	36 56 55	25 26 20	x										x	x					
	S279	37 01 00	25 30 50	x	A		x		x	x	x									
	Z80A	37 01 40	25 30 50	x	A	o	x		o	x					o					
	Z21A	37 03 35	25 33 40	x	Ad		x			x										o
	S261A	36 58 10	25 25 10	x	A		x			x	x									
	Z58B	37 04 05	25 32 15	x	A		x	o			x									
III	B531	37 02 55	25 30 50	x	A		x	o		x										
	Z125	37 02 40	25 30 10	x	A			x	x	x	x									
	Z57	37 04 05	25 31 45	x	Ad			x		x	x									
	Z256	37 02 20	25 29 40	x	Ol		x			x										
	B493	37 00 15	25 25 50	x	A		x	o		x	x									
	B405[3]	37 01 10	25 27 10										o	x	x					

TABLE 1 (continued)

Carbonate-rich compositions[2]																				
Zone	Sample number	Coordinates Latitude	Coordinates Longitude	quartz	plagioclase	K-feldspar	muscovite	phlog-biotite	garnet	chlorite	epidote	hornblende	tremolite	talc	dolomite	diopside	scapolite	vesuvianite	corundum	margarite
IV	S408	37 00 30	25 25 30	x	x		x	x		o	x									
	B238A	37 03 30	25 29 30	x	Ad		x	x		o	x									
	B513A	37 02 10	25 27 10		Ad			x		o		x				x	x			
	B483II	37 06 50	25 32 30		La		x	x		o	o		x						x	x
	B543	37 04 50	25 30 15	x	Ad		x	x	o		x									
	B61[3]	37 03 10	25 25 40	x				x					x		x					
	B559	37 07 25	25 32 05	x	Ad					(o)		o				x	x			
V$_A$	B505A	37 03 50	25 28 30						x		x				o	x	x	x		
	B505A″	37 03 50	25 28 30						x	(o)	x					x		x		
	B505I	37 03 50	25 28 30		An						x	x				x	x			
	S116A	37 05 40	25 30 15	x	An			x				x				x	x			
	B557A	37 06 25	25 31 10		La	x	(o)	x		(o)	x	x				x				
	B203E[3] west	37 11 30	25 31 10	x						(x)			x			x				
	B86	37 04 30	25 25 50				(o)	x		(o)	x	x								
V$_B$	B495[3]	37 06 10	25 30 25	x				x		(o)			x	(o)		x				
	B486A	37 07 30	25 31 00	o	An			x					x		o	x	x	x		
	B185	37 09 30	25 31 00						x		x				o	x	x	x		
	B185X	37 09 30	25 31 00						x			x						x		
	B207 west	37 10 55	25 30 50	x	An			x		(o)						x	x			
	B545A	37 08 20	25 28 05	x	x	x		o	x		(o)					x	x			
	B545A[4]	37 08 20	25 28 05	x	x	x					x	x								
VI	S130H	37 05 10	25 29 00	x	An			x				x				x	x			
	S377	37 05 10	25 27 00						x		(o)					x	x	x		
	B551	37 05 00	25 28 00	x	An	x						x				x	x			
	B127	37 05 35	25 28 20	x	An			x				x				x	x			
	B127A	37 05 35	25 28 20	x	An				x			x				x				
	B512	37 05 45	25 28 10		An			x					x						x	(o)
	B512B[3]	37 05 45	25 28 10		An			x					x		x				x	

	[illegible]	37 05 10	25 29 00	x	An			x				x				x	x			
	S377	37 05 10	25 27 00						x		(o)					x	x	x		
	B551	37 05 00	25 28 00	x	An	x						x				x	x			
	B127	37 05 35	25 28 20	x	An			x				x				x	x			
	B127A	37 05 35	25 28 20	x	An				x			x				x				
	B512	37 05 45	25 28 10		An			x					x						x	(o)
	B512B[3]	37 05 45	25 28 10		An			x					x		x				x	

Metavolcanic compositions[4]																				
Zone	Sample number	Coordinates Latitude	Longitude	quartz	plagioclase	K-feldspar	muscovite	biotite	actinolite	hornblende	glaucophane	diopside	epidote/zoi	scapolite	calcite	garnet	chlorite	sphene	rutile	
I	B11	36°56′50″	25°29′25″	x	A		x		x		R		x		o		x		o	
	B192	36 59 00	25 33 30	x	A		x	o	x		x		x		o		x			
	B420	36 58 40	25 29 40	x	A		x	(x)			x					x	x	o		
	B201	37 00 00	25 32 35	x			x	o			x		x		x		x	x		
	S459C	36 59 45	25 32 00	x	A		x				x				x		x	o		
II	B312	37 00 20	25 31 30	x	A		x		x		R		x				x	x		
	B312Z	37 00 20	25 31 30	x					x		x		x		o		x	x		
	B310	37 00 30	25 31 00	x	A		x	o	x				x				x	o		
	Z24	37 03 20	25 32 40	x	A			o	o				x			o	x	o		
	Z231	37 02 10	25 30 10	x	A			x	o				x				x			
III	Z55	37 03 55	25 31 15	x	Ol			x	o	x			o			x	x			
	Z117	37 02 50	25 30 10	x	AOl	o		x		x			x		o	x	x			
	Z266	37 02 25	25 29 30		Ol			x		x			x			x		o		
	Z36	37 04 20	25 31 10	x	Ol					x			x			x	x			
IV	B258	37 04 00	25 30 40	x	Ad		x	(o)		x			x			o	x			
	B521	37 04 30	25 30 30	x	x		x	x		o			x		x		x			
	B513	37 02 10	25 27 10		x					x		x			x					
	S164B	37 00 55	25 25 20	x	x			o		o			x		x		o		o	
	Z15	37 05 10	25 30 20	x	Ad					x			x				(o)	x		
	B503	37 11 20	25 30 40		Ad					x		x			o	x				
	Z66	37 03 10	25 26 00	x	Ad		x	x		x			x			x	x	o		
V_A	Z11	37 05 20	25 30 00	x	Ad			x		x						x				
	B557A	37 06 25	25 31 10		Ad	x		x	x	x		x	x			o	(o)	o		
	Z64A	37 02 55	25 26 30	x	Ad			x		x			x		o	x	x	o		
	S190	37 09 25	25 31 40	x	Ad			x		x				x	o	o	(o)	o		
	B203	37 11 30	25 31 10				o			x			x		o	x	x			
	B203-4	37 11 30	25 31 10	x					x				x				o		o	
	B203-5	37 11 30	25 31 10	x	x		x	x	o				x		o	o	x			

TABLE I (continued)

Metavolcanic compositions[4]																			
Zone	Sample number	Coordinates Latitude	Coordinates Longitude	quartz	plagioclase	K-feldspar	muscovite	biotite	actinolite	hornblende	glaucophane	diopside	epidote/zoi	scapolite	calcite	garnet	chlorite	sphene	rutile
V_B	Z5	37 05 30	25 29 40		Ol			x	x	x							(o)	o	
	B569	37 10 55	25 31 20	x	Ad	x				x			x			x	(o)	o	
	B207B west	37 10 55	25 30 50	x	La			x		x		x	(o)	x	o		(o)	o	
	B544	37 05 30	25 26 55	x	x		x	x		x			(x)					o	
	B545A″	37 08 20	25 28 05	x	La	x				x		x	(o)	x	o	x		o	
VI	S395	37 05 40	25 28 20	x	x	x		(o)		x			(o)		x				
	B551A	37 05 00	25 28 00	x	An	x				x		x		x	x		(o)	o	
	B127B	37 05 35	25 28 20	x	An			o		o		x		o	o	x		o	

Ultramafic compositions																			
Zone	Sample number	Coordinates Latitude	Coordinates Longitude	olivine	enstatite	diopside	tremolite	anthophyllite	gedrite	chlorite	clinochlore	phlogopite	talc	chrysotile	antigorite	magnesite	calcite/dolomite	spinel	opaques
I	B427A	36°57′40″	25°32′45″			x				x			(o)	(o)	x				x
	B427D	36 57 40	25 32 45			x							x				x		o
	B427C	36 57 40	25 32 45				x			o			x				x		o
	B538	36 57 30	25 32 35			x				x					x				o
II	B225	36 56 50	25 26 30				x			x			x		x				o
	B704	36 57 00	25 26 35							x				(o)	x				o
	B704A	36 57 00	25 26 35				x		x			o	x	(o)	x				o
III	Not Present																		
IV	B231	37 01 20	25 26 20				x			x			x						o
	B231X	37 01 20	25 26 20					x		x					x	x			
	B231V	37 01 20	25 26 20						x	x			x		(o)				
	B231T	37 01 20	25 26 20				x	x					x						o
	B453	37 04 45	25 30 20					x					x			x			o
	B453W	37 04 45	25 30 20	x				o		o			x			x			o
	S483	37 05 00	25 30 35				x	x					x			x			o
V_A	B181	37 10 50	25 30 15	x	x		x	x			(o)	o	x		(o)				o
	B186	37 10 30	25 29 50			x	x										x		
	B203I	37 11 30	25 31 10	x	x		x			o			o		(o)				o
	B203Y	37 11 30	25 31 10				x		x	x			o						o
	B203X	37 11 30	25 31 10	x	x		(x)	x		x			x	(o)					
	S73A	37 11 10	25 31 55		x		x			(o)			(o)				(o)	o	o
V_B	B151	37 05 55	25 29 55				x	x				o	x				(o)		o
	B151X	37 05 55	25 29 55	x	x		x			x			x		(o)			o	o
	B159	37 06 20	25 30 20	x	x	x	x				x							x	o
	B159Z	37 06 20	25 30 20		x		x	x		x			x				(o)		o
	B184	37 09 40	25 30 50	x		x	x					o							o
	S390A	37 03 05	25 26 45			R	x	x			x						o		o
	S113G	37 05 50	25 29 45	x															

Zone	Sample number	Latitude	Longitude	diaspore	corundum	hematite	magnetite	An-rich plag.	calcite	margarite	chlorite	staurolite	chloritoid	kyanite	andalusite	epidote/zoisite	muscovite	biotite	spinel	pyrophyllite	tourmaline
IV	B231	37 01 20	25 26 20				x			x			x						o		
	B231X	37 01 20	25 26 20					x		x					x	x					
	B231V	37 01 20	25 26 20						x	x			x		(o)						
	B231T	37 01 20	25 26 20				x	x					x						o		
	B453	37 04 45	25 30 20					x					x			x			o		
	B453W	37 04 45	25 30 20	x				o		o			x			x			o		
	S483	[illegible]	[illegible]				x	x					x			x			o		
V_A	B181	37 10 50	25 30 15	x	x		x	x			(o)	o	x		(o)				o		
	B186	37 10 30	25 29 50			x	x										x				
	B2031	37 11 30	25 31 10	x	x		x			o			o		(o)				o		
	B203Y	37 11 30	25 31 10				x		x	x			o						o		
	B203X	37 11 30	25 31 10	x	x		(x)	x		x			x	(o)							
	S73A	37 11 10	25 31 55		x		x			(o)			(o)				(o)	o	o		
V_B	B151	37 05 55	25 29 55				x	x				o	x				(o)		o		
	B151X	37 05 55	25 29 55	x	x		x			x			x		(o)			o	o		
	B159	37 06 20	25 30 20	x	x	x	x				x								o		
	B159Z	37 06 20	25 30 20		x		x	x		x			x					x	o		
	B184	37 09 40	25 30 50	x		x	x					o					(o)		o		
	S390A	37 03 05	25 26 45			R	x	x			x						o		o		
	S113G west	37 05 50	25 29 45	x	x	x	x				x			(o)				x	o		
	B39C	37 05 35	25 26 50				x	x		(o)								o	o		
VI	S407A	37 07 30	25 30 40				x	x				(o)	(x)					x	o		
	S407C	37 07 30	25 30 40	x	x	R	x			(o)	x							x	o		

Bauxitic compositions																					
Zone	Sample number	Coordinates Latitude	Longitude	diaspore	corundum	hematite	magnetite	An-rich plag.	calcite	margarite	chlorite	staurolite	chloritoid	kyanite	andalusite	epidote/zoisite	muscovite	biotite	spinel	pyrophyllite	tourmaline
I	B508A	36°57′50″	25°33′00″			x							x				x				
	B469A	36 56 10	25 27 00	x										x						x	
	B490	36 56 50	25 28 30	x							x		x								
	B434[5]	36 59 10	25 32 00	x		x			x				x								
	B537A	36 57 25	25 27 45						x				x				x				
	B444	37 01 45	25 33 30	x	R						x		x							o	
II	S127	37 03 20	25 34 30	(x)	x	x			x	(o)	x		x								
	S255[5]	36 56 40	25 25 40		x	x	x				x		x				x				
	B368[5]	36 58 50	25 28 05	R	x	x			x				x								
	B413[5]	37 00 20	25 29 20		x	x	x						x	x							
	B270[5]	37 03 45	25 33 10		x		x		x	x							x				

TABLE 1 (continued)

Bauxitic compositions																					
Zone	Sample number	Coordinates Latitude	Coordinates Longitude	diaspore	corundum	hematite	magnetite	An-rich plag.	calcite	margarite	chlorite	staurolite	chloritoid	kyanite	andalusite	epidote/zoisite	muscovite	biotite	spinel	pyrophyllite	tourmaline
III	S93[5]	37 05 50	25 33 50		x	x	x		x	x							x	x			
	B507A	37 02 30	25 29 45		o		x			x				x		x	x	x			o
	B507D	37 02 30	25 29 45				x				x			x		x	x				
	B507E	37 02 30	25 29 45			x	x			x	x		x	x		x	x				
	B191A	37 02 50	25 29 40		x				x												
	B191AI	37 02 50	25 29 40		x					x											
	S425[5]	37 01 35	25 27 50		x		x			x			x	x				x			
IV	B492	37 01 10	25 27 10	(o)	x						x	x	x				x	x			
	B492B	37 01 10	25 27 10		x					x	x	x	x				x				
	B482	37 04 40	25 31 20	(o)	x		x							x[8]			x				o
	B501	37 07 50	25 34 05				x				x			x							
	B499	37 07 15	25 33 10		x		x		x			x		x		x	x				
	B483A	37 06 50	25 32 30		x			x	x	o	(o)						o				o
	west S466B[5]	36 59 50	25 24 00		x		x								x[9]						
V_A	B504[6]	37 04 00	25 28 30		x			x			x							x			x
	west B588[6]	37 10 50	25 30 20					x							x						x
V_B	B83[6]	37 03 30	25 26 35		x			x		(o)	x								x		
	B494A[6]	37 06 05	25 30 05		x			x								Zo		x	x		x
	B494B[6]	37 06 05	25 30 05		x		x	x									x				
	B494C[6]	37 06 05	25 30 05		x			x								Zo		x			
	S406	37 07 35	25 30 55		x		x			(o)											
	west B39[6]	37 05 35	26 26 50	(P)	x			x			x					Zo		x	x		
VI	B512E[7]	37 05 45	25 28 10		x			x	x	(o)								x	x		

S, collection of Schuiling; Z, collection of Zimmermann; B, collection of Jansen. x, major mineral; o, minor mineral; R, relict mineral; ( ), retrograde mineral; F, fibrolite; A, albite; Ol, oligoclase; Ad, andesine; La, labradorite; An, anorthite; G, green biotite, intergrown with chlorite; P, prehnite; Zo, zoisite. Notes: [1]quartz in excess; [2]calcite in excess; [3]siliceous dolomite; [4]amphibolites and their low grade equivalents; [5]selected field data; [6]desilicified pegmatite; [7]desilicified pelitic lens, locally with sapphirine; [8]fibrous kyanite along fissures; [9]andalusite is a contact-metamorphic mineral here!

The pelitic schists show the typica
quartz–chlorite–sericite, occasionall
beginning of the metamorphic r
chloritoid, and sodium-rich schists
occurs in large crystals with cube-
bles in this zone locally contain ar
or piemontite. Some small lenses
phane with epidote. Quartz, ep
muscovite are the more common n
nant iron-oxide in the low grade
stance, as a residue of organic ma
grayish marbles. Pyrite is frequent

The metavolcanic layers local
phane. The assemblage glaucopha
with chlorite and albite. Parago
occur in these schists. In the chl
crystals often show thin rims of
served in glaucophane–actinolite–c

In the metamorphosed ultram
diopside, antigorite, chlorite, an
diopside–chlorite–antigorite is cha
rocks. Magnetite (with chlorite) is
body, probably due to the impern
the preservation of a relatively low
compositions, the iron-oxide is hem

The meta-bauxites in the ma
blage diaspore–chloritoid–hematite
phyllite, kyanite, and minor kaol
mary metamorphic phase. Rarely a
along the contact of the diasporite
abundantly occurring contorted c
albite, epidote, chlorite, and calcit

CHLORITE–S

This zone starts with the +
reaction: diaspore into corundum
is defined in the field by the fir
bauxite lenses and is based on app
the lowest-grade part of zone II
spore indicate that the reaction is
that all emery deposits, up to the
contain secondary diaspore, especi
reaction may produce reverse pseu
about the same as in zone I. Me
components of conglomeratic laye
two meta-ultramafic bodies crop o

The pelitic schists show the typical greenschist facies assemblages albite–quartz–chlorite–sericite, occasionally with actinolite or epidote. From the beginning of the metamorphic range, iron-rich mica schists contain chloritoid, and sodium-rich schists contain paragonite. Pyrite sometimes occurs in large crystals with cube-faces predominant. The impure marbles in this zone locally contain ankeritic carbonate, Ni-bearing fuchsite, or piemontite. Some small lenses in the marble even contain glaucophane with epidote. Quartz, epidote, albite, actinolite, chlorite, and muscovite are the more common minerals in the marbles. The predominant iron-oxide in the low grade marbles is hematite. Graphitic substance, as a residue of organic matter, is present in these, mostly blue-grayish marbles. Pyrite is frequent along discrete horizons.

The metavolcanic layers locally consist almost completely of glaucophane. The assemblage glaucophane–epidote is normally found together with chlorite and albite. Paragonite, actinolite, and garnet may also occur in these schists. In the chlorite–epidote schists the small crossite crystals often show thin rims of actinolite. Small biotite flakes are observed in glaucophane–actinolite–chlorite schists.

In the metamorphosed ultramafic rock the minerals chrysotile, talc, diopside, antigorite, chlorite, and actinolite occur. The assemblage diopside–chlorite–antigorite is characteristic of the low grade ultramafic rocks. Magnetite (with chlorite) is present in the center of the ultramafic body, probably due to the impermeability of the rock, which permitted the preservation of a relatively low oxygen fugacity. In all the other rock compositions, the iron-oxide is hematite.

The meta-bauxites in the marble are characterized by the assemblage diaspore–chloritoid–hematite. In addition, they may contain pyrophyllite, kyanite, and minor kaolinite, which possibly represents a primary metamorphic phase. Rarely are muscovite–paragonite schists found along the contact of the diasporite with impure, schistose marbles. The abundantly occurring contorted quartz segregations frequently contain albite, epidote, chlorite, and calcite crystals.

CHLORITE–SERICITE ZONE (II)

This zone starts with the + corundum isograd, which marks the reaction: diaspore into corundum plus water. The + corundum isograd is defined in the field by the first appearance of corundum in meta-bauxite lenses and is based on approximately 25 observations (fig. 3). In the lowest-grade part of zone II pseudomorphs of corundum after diaspore indicate that the reaction is prograde. It should be noted, however, that all emery deposits, up to the highest grade of metamorphism, can contain secondary diaspore, especially along fissures, and the retrograde reaction may produce reverse pseudomorphs. The lithology in zone II is about the same as in zone I. Metavolcanic rocks are also observed as components of conglomeratic layers in a rather thick marble bed. Only two meta-ultramafic bodies crop out in this zone.

Mineral assemblages in the pelitic schists and marbles are similar to those in zone I, although on approaching the + biotite isograd paragonite disappears from the mica schists and ankeritic carbonate and Ni-bearing fuchsite from the marbles. Talc starts to form typically in the siliceous dolomitic marbles in this zone. Some phlogopite and rarely margarite are seen in the impure marbles. In the white marbles graphite locally occurs concentrated in small black spots or in thin streaks, whereas most of the marbles in this zone remain grayish colored and contain graphite dispersely distributed through the marble. Pyrrhotite appears near the + biotite isograd. Especially in the marbles and in the emery deposits, the hematite gives way to magnetite about halfway between the + corundum isograd and the − chloritoid isograd. Of course, the hematite–magnetite reaction is dependent on oxygen fugacity as well as temperature and total pressure and is, therefore, not very suitable for comparison with the mapped isograds, which are mainly dehydration reactions. Magnetite appears and hematite gradually disappears over a rather broad transition zone, which overlaps the − chloritoid isograd. In the metavolcanic components of the above mentioned conglomeratic marble crossite occurs only near the + corundum isograd. Most sodium-amphiboles show an actinolite rim, which is better developed than in zone I. The main mineral assemblage in the metavolcanic rocks is chlorite–epidote–albite–actinolite, with biotite and garnet as minor minerals.

In this zone (zone II) and in zones III and IV diopside is not found in the metamorphosed ultramafic rocks; it appears again with tremolite in the high grade part of zone V_A and with enstatite and forsterite in zone V_B. A similar disappearance of diopside in the intermediate zones is mentioned for the Swiss alps by Trommsdorff and Evans (1974). Secondary chrysotile is often observed in later fissures, a phenomenon found even in the forsterite–enstatite ultramafic rocks from the highest grades of metamorphism. The minerals talc, actinolite, chlorite, antigorite, and gedrite also occur in these low grade ultramafic rocks. The assemblages actinolite–talc, actinolite–antigorite, talc–antigorite, and chlorite–gedrite–magnetite are found. The emeries in zone II are characterized mainly by the assemblage corundum–chloritoid–hematite. Pyrophyllite and kaolinite are absent, and kyanite has been observed in only one emery deposit halfway in this zone. Margarite appears near the + biotite isograd. At the high temperature side of this zone tourmaline begins to appear in quartz veins, whereas albite and epidote disappear beyond this zone. Only in late, secondary quartz veins are albite and epidote still found.

BIOTITE–CHLORITOID ZONE (III)

Although in a few presumably metavolcanic rocks within zone I and II some local biotite occurrences have been found, the + biotite isograd can be mapped rather accurately in rocks of pelitic composition. The isograd, as drawn in figure 3, is based on approximately 30 field observations. Wherever this isograd could be studied in detail, the transition zone was not more than several tens of meters wide. In this transition zone, some beds carry biotite, whereas others do not. Very suddenly after passing through the transition zone, all the pelitic schists carry biotite, sometimes very abundantly. It seems strange that the inevitable differences in chemical composition between adjacent beds apparently had so little influence on the appearance of biotite on Naxos. There is a striking coincidence of the disappearance of the assemblage chlorite–muscovite with the + biotite isograd, suggesting that this assemblage plays a key role in the reaction by which biotite forms in pelitic schists.

The mica schists become more abundant than the marbles in the stratigraphically lower part of the metamorphic sequence (see cross section of fig. 2). This change in lithology becomes marked toward the – chloritoid isograd. Also the amphibolites gain gradually in importance, beginning a few hundred meters after passing the + biotite isograd. The transition of the greenschist facies into the amphibolite facies has commonly been situated more or less at this grade of metamorphism. In the pelitic schists, the place of albite is taken by oligoclase. Whereas the biotite–chlorite assemblage is often observed in this zone, the garnet–muscovite–biotite assemblage seems to be more frequent in the following zone. Chloritoid is only rarely observed. Albite gives way in the marbles very rapidly to plagioclase with an anorthite content of more than 15 percent toward the – chloritoid isograd. Hornblende appeared halfway through this zone, and phlogopite is more common than it is zone II. In the siliceous dolomites the assemblages talc–calcite–quartz, muscovite–dolomite–quartz, and talc–muscovite–quartz are observed.

The metavolcanic rocks in this zone are represented mainly by amphibolites. The common mineral assemblage is epidote–biotite–hornblende–oligoclase for the high grade part of this zone, the assemblages of zone II remain stable in the low grade part. There are no real meta-ultramafic bodies in this zone; only a few small phlogopite–actinolite–chlorite–talc lenses are found.

In the emery deposits within this biotite–chloritoid zone, margarite is a very common mineral. On Naxos, margarite appears in this rock type in the higher grade part of the greenschist facies and persists up into the kyanite–sillimanite transition zone of the amphibolite facies. This conflicts with the idea that margarite is confined to the greenschist facies. The experiments by Chatterjee (1974) and Storre and Nitsch (1974) show that the stability field of margarite is markedly displaced toward higher temperatures under increasing water pressure and in the absence of excess silica. The emery deposits contain characteristic mineral assemblages: corundum–chloritoid–kyanite–magnetite or hematite for the massive emery, corundum–calcite–margarite–chloritoid along the contact with the marble, and epidote–kyanite–chloritoid–margarite in schistose parts of the deposit. Most of the primary quartz veins contain minerals found also in the surrounding rocks. In secondary veins the minerals calcite, chlorite, albite, epidote, and hematite or limonite occur.

KYANITE ZONE (IV)

This zone begins with the disappearance of the chloritoid in the iron-rich pelitic schists and in the emery deposits. Although the – chloritoid isograd is based on only about 10 observations, the chloritoid seems to disappear at more or less the same grade in these two rock types with very different bulk compositions. The excess silica in the pelitic schists and the excess alumina in the emery deposits suggest we are dealing with various breakdown reactions. In the pelitic schists the breakdown reactions are not yet clear. In the emeries the appearance of staurolite coincides with the disappearance of chloritoid. Apparently the transition reaction is chloritoid + kyanite + corundum into staurolite + water. In the pelitic schists, kyanite appears a few hundred meters after passing the – chloritoid isograd, and staurolite a bit later. Chlorite is rare in zone IV—when observed it is mostly retrograde. It is remarkable that most staurolite occurrences are in the kyanite-bearing mica schists, which contain no chlorite as a primary metamorphic phase.

In the impure marbles several mineral reactions were identified at these grades of metamorphism. Of course, most mineral reactions in the impure calcitic and dolomitic marbles are not only a function of temperature and total pressure but of the H_2O/CO_2 ratio as well. Shortly after passing the – chloritoid isograd, the first appearance of the typical talc–tremolite assemblages was observed in the siliceous dolomitic marbles. The mineral talc is rare in the high grade part of this zone. Just before the + sillimanite isograd the last dolomite–quartz assemblages occur. Blue-green hornblende becomes a part of the impure carbonate assemblages at about the grade where chlorite plus actinolite disappears. On approaching the + sillimanite isograd, green hornblende, scapolite, and grossular appear. The assemblage muscovite–quartz disappears in the calcitic marbles halfway into this zone. The plagioclase becomes rich in anorthite; for example, the assemblage phlogopite–tremolite–corundum–anorthite–calcite is locally observed near the + sillimanite isograd. The calcitic marbles gradually become more and more clearly white and well crystallized. Only a few marble beds remain grayish, and most of the graphite is concentrated in gray thin streaks.

The amphibolites also show an increase of the anorthite component in the plagioclase but not as much as in the marbles; it never exceeds 70 percent An up to zone V_B. Near the + sillimanite isograd augitic diopside becomes a part of the amphibolite assemblages. The metamorphosed ultramafic rocks are characterized by the presence of magnesite. A few outcrops show the assemblage talc–magnesite and talc–anthophyllite, and other localities, closer to the + sillimanite isograd, show the parageneses talc–magnesite–anthophyllite and talc–forsterite–anthophyllite.

The emery deposits in this zone commonly contain the typical assemblage kyanite–corundum–staurolite–magnetite. Some schistose emeries consist of muscovite–staurolite–kyanite or muscovite–epidote–kyanite. Along the contacts with the marbles the margarite–anorthite–corundum

assemblage is found. Magnetite is
vite–kyanite paragenses are observ
veins through the high grade eme
of aplites and pegmatites begins a
and they become more abundan
quartz segregations in the pelitic
contain large kyanite crystals, whi
the dolomitic marbles large trem
are formed around the quartz segreg

KYANITE–SILLIMANIT

The first appearance of sillima
beginning of this zone. Although
particular interest on account of
even three, Al-silicates side by side
matite dome, the zone is exposed
this is, at least partly, an effect of t
of about 600 m is more nearly correc

In the transition zone eviden
silicates is common in the mica sch
shown by pseudomorphs of sillimar
reacted kyanite into sillimanite.
about the character of the reactio
grade) and a decrease in pressure (r
would produce the same result.
prograde or retrograde character
longs to the sillimanite- or to the ky

Typically, kyanite, sometimes
tions with sillimanite occurring in
and no reaction evidence between
This situation may indicate that a
in the mica schist, whereas the kya
taken place in the quartz segregat
small. Because the kyanite is someti
segregations are often slightly dist
kyanite was formed relatively early
reverse situation, sillimanite in the
mica schist, is rarely seen. In most
side without any evidence of react
deformed, but one should bear in
and had already undergone a part
sillimanite formed.

Near the southwest corner of
silicates occur together, sometimes
about 3 km along strike, in the
zone between the kyanite and silli
fact concentrated in seemingly late

assemblage is found. Magnetite is the predominant iron-oxide. Muscovite–kyanite paragenses are observed with tourmaline and corundum in veins through the high grade emeries. The locally frequent occurrence of aplites and pegmatites begins after passing the – chloritoid isograd, and they become more abundant toward the migmatite dome. The quartz segregations in the pelitic schists of the kyanite zone commonly contain large kyanite crystals, which are often deformed. Typically in the dolomitic marbles large tremolite crystals with clear white calcite are formed around the quartz segregations.

KYANITE–SILLIMANITE TRANSITION ZONE (V_A)

The first appearance of sillimanite, in most cases fibrolite, marks the beginning of this zone. Although the zone is limited in width, it is of particular interest on account of the fact that it contains two, locally even three, Al-silicates side by side. Near the northeast side of the migmatite dome, the zone is exposed over a width of more than 1 km, but this is, at least partly, an effect of topography. An average true thickness of about 600 m is more nearly correct.

In the transition zone evidence of reaction between the two Al-silicates is common in the mica schists and quartz segregations, and it is shown by pseudomorphs of sillimanite after kyanite and by incompletely reacted kyanite into sillimanite. These features do not say anything about the character of the reaction. An increase in temperature (prograde) and a decrease in pressure (retrograde), or both changes combined, would produce the same result. Strictly speaking, it depends on the prograde or retrograde character of the reaction whether the zone belongs to the sillimanite- or to the kyanite zone.

Typically, kyanite, sometimes bent, occurs in small quartz segregations with sillimanite occurring in the directly surrounding mica schists, and no reaction evidence between the two Al-silicates can be observed. This situation may indicate that a sillimanite producing reaction occurs in the mica schist, whereas the kyanite–sillimanite transition has not yet taken place in the quartz segregation, because the overstep is still too small. Because the kyanite is sometimes bent and deformed and the quartz segregations are often slightly distorted, it can be concluded that the kyanite was formed relatively early with respect to the sillimanite. The reverse situation, sillimanite in the quartz segregation and kyanite in the mica schist, is rarely seen. In most cases the two Al-silicates occur side by side without any evidence of reaction. The kyanite is more frequently deformed, but one should bear in mind that the kyanite formed earlier and had already undergone a part of the metamorphic history before the sillimanite formed.

Near the southwest corner of the migmatite dome, the three Al-silicates occur together, sometimes even in a single hand specimen, for about 3 km along strike, in the direct prolongation of the transition zone between the kyanite and sillimanite. Some of the andalusite is in fact concentrated in seemingly late quartz segregations, but in the same segregations kyanite also occurs, commonly without evidence of reaction. The andalusite in these segregations in very well crystallized and not deformed, whereas the kyanite crystals may be deformed. Only in a few thin sections does andalusite form a rim around the kyanite. The partial reaction of kyanite into sillimanite, commonly observed in other parts of the transition zone, is rare in this andalusite-bearing part of the zone. In a few cases the fibrolite is found as pseudomorphs after kyanite. No textural relations between fibrolite and andalusite have yet been observed. The Al-silicate-bearing schists in this part of the zone are intercalations in amphibolites and in Al-poor biotite-schists, and, therefore, no continuous cross sections through the zone can be studied. Nevertheless the andalusite in the low grade of the zone is always found with or near kyanite, whereas in the high grade part always near sillimanite. In both circumstances the andalusite is the less deformed and more euhedral Al-silicate. Most observations indicate that we are dealing with a "retrograde direction" of reaction, and probably we may apply this idea for the whole transition zone. As proposed by Thompson (1970) andalusite may form when the pressure drops to the point where kyanite and sillimanite are coexisting phases, and this seems to be the case in the metamorphic complex on Naxos. In the southeastern part of the zone the metamorphic conditions were near or just above the triple point of the Al-silicates, whereas in the rest of the zone the conditions were so far away from the triple point (parallel to the kyanite–sillimanite reaction curve) that during a pressure drop no andalusite could form.

In order to prevent misunderstanding, the formation of the andalusite and the sillimanite in the southwestern part of the transition zone is in no way related to later contact metamorphism caused by the granodiorite (fig. 3). Between this zone of coexisting Al-silicates and the granodiorite, the metamorphic grade decreases considerably, moreover the contact metamorphism is restricted to a few tens, with a maximum of a hundred, of meters along the border of the granodiorite. If the andalusite formed at some later time during the contact metamorphism, then it would be a strange coincidence that the andalusite so faithfully follows this transition zone of the Al-silicates.

In the pelitic schists staurolite, not uncommonly associated with the three Al-silicates, disappears on the southwestern side of the migmatite dome near the − kyanite isograd, whereas it has disappeared on the eastern side more or less with the + sillimanite isograd. The apparent shift of the staurolite stability to higher grade metamorphism on the southwest side of the dome is due to the fact that the transition zone of the Al-silicates is displaced toward a lower grade of metamorphism. Cordierite is not found anywhere in the transition zone, despite careful search. Almandine-rich garnet commonly occurs in this zone. The main mineral assemblage is quartz–plagioclase–muscovite–biotite–kyanite–sillimanite–garnet or staurolite. The paragenesis kyanite–almandine–biotite suggests that the staurolite–muscovite–quartz already reacts in the kyanite zone.

The marbles, in general, contain mineral assemblages similar to those found in the high grade part of zone IV. In addition they show vesuvianite and xanthophyllite. The mineral vesuvianite is found in this zone with grossular–diopside–calcite–epidote and quartz, and the xanthophyllite with grossular–vesuvianite and calcite. The frequent occurrence of aplite veins at these locations may indicate that silica and water were added to the marble during the formation of vesuvianite and xanthophyllite. Plagioclase with scapolite is almost pure anorthite; muscovite disappears from the impure marbles. The siliceous dolomites start to contain the assemblage diopside–calcite with or without quartz. Forsterite has not been observed, probably due to the high aluminum content of most dolomite–calcite marbles in the high grade parts of the metamorphic complex on Naxos.

The amphibolites contain more and more diopside, K-feldspar, and grossular. The hornblende is green colored. Scapolite appears. Although enstatite–magnesite might still be a stable association in the ultramafic rocks of this zone, magnesite has not yet been observed. One of the ultramafic rocks shows the enstatite–talc assemblage in a matrix of olivine–tremolite–anthophyllite. Enstatite–talc represents the high pressure equivalent of anthophyllite, while that mineral is found in the same outcrop. In outcrops near the — kyanite isograd the assemblages diopside–tremolite and enstatite–forsterite are found. Antigorite is secondary. In gabbroic parts along these ultramafic bodies the assemblage garnet–chlorite–zoisite is found. The emery deposits in this zone are rare. They show the high grade mineral assemblage of zone IV. Indeed, they still contain staurolite in contrast to the pelitic schists. This fact seems to suggest that staurolite plus corundum has a larger stability field than staurolite plus quartz. Pegmatites in this transition zone become abundant, and their mineralogy is dependent on the country rock. Pegmatites in the pelitic schists contain quartz, oligoclase, K-feldspar, muscovite, biotite, garnet, tourmaline, and beryl. Those cross cutting the marbles contain quartz, andesine, K-feldspar, biotite, garnet, diopside, and tourmaline.

SILLIMANITE ZONE (V_B)

The disappearance of kyanite from the pelitic schists and gneisses forms the low grade boundary of this zone. The zone is extremely variable in thickness, depending on the location of the + meltphase isograd. As might be expected, sillimanite-rich layers are very common in the mica schists. The sillimanite normally remains fibrolitic in this zone. Locally, rocks of pelitic composition become gneissic and sometimes develop into augen-gneisses, in which K-feldspar is the predominant mineral. The assemblage K-feldspar plus sillimanite is found locally. However, the muscovite–quartz assemblage in other places remains stable into the outer parts of the migmatite dome.

In the marbles epidote becomes rare toward the higher grade of this zone, and blue-green hornblende disappears from the amphibolite streaks in the marbles. The assemblage biotite–diopside–hornblende–anorthite–calcite is frequently observed. The marbles in this zone are very well crystallized, and they occasionally contain gray, graphite-rich streaks. Pyrrhotite seems to be more frequent than pyrite. The amphibolites are similar to those of the previous zone. Epidote and chlorite become rare toward the migmatite dome.

The ultramafic rocks start to contain the assemblages enstatite–forsterite–green spinel, clinochlore–brown spinel, and enstatite–diopside. Cummingtonite, anthophyllite, and actinolite are observed as intergrown assemblages. The highest grade emery deposits are found in the sillimanite zone, and their mineralogy is the same as in the previous zone. Staurolite remains stable in these outcrops. Margarite as a primary phase has disappeared before the – kyanite isograd. Along the border with the marble anorthite–calcite–corundum assemblage is observed. The pegmatites cross cutting the ultramafic rocks are strongly desilicified and they consist of the minerals phlogopite, anorthite, corundum, chlorite, zoisite, tourmaline, and beryl.

MIGMATITE ZONE (VI)

Evidence for anatexis marks the beginning of this zone, which is represented by the migmatite dome in figures 2 and 3. The classical phenomena of anatexis that can be observed here include irregularly contorted and disrupted schistosity planes and separation of the gneissic rock into dark, biotite-rich patches (still showing schistosity) together with light-colored isotropic granitic parts. Anatexis is also expressed by flow folding and by the development of pegmatitic pods, which may or may not show sharp boundaries with their surroundings. All these phenomena are considered as evidence for the appearance of a new phase in the pelitic rocks, namely a granitic melt, and the beginning of these phenomena has been marked as the highest metamorphic isograd on the map. As regards the mineral assemblages of the migmatitic gneisses, it should be noted that some parts are the crystallization products of granitic melts, while others are dark restites, and still others are formed by late stage pegmatitic pods. The following major assemblages have been found: oligoclase–quartz–K-feldspar–muscovite or biotite, occasionally with tourmaline or garnet for the pegmatoid rocks, oligoclase–quartz–K-feldspar–biotite for the granitic parts, and oligoclase–biotite–sillimanite–quartz–garnet with commonly K-feldspar for the pelitic restites.

Siliceous carbonate assemblages in the migmatite are characterized by occurrences of diopside, hornblende, anorthite, grossular, phlogopite, and quartz. In addition, biotite, scapolite, and sphene have been found. Sulphide assemblages such as pyrrhotine–pyrite are present in these high grade marbles. All these minerals commonly occur in dark streaks parallel to the boundaries of the marble layers. These streaks, therefore, probably represent the original sedimentary layering, possibly enhanced

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CONDITIONS O

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by metamorphic differentiation. The marbles within the migmatite contain virtually no primary epidote. The calcite marbles are extremely well crystallized and probably metasomatically cleaned. Graphite is not found in these high grade marbles, they are all clear white. Probably the graphite is more or less oxidized gradually throughout the metamorphic sequence and completely burnt away in this migmatite dome. The observations of the reduction of more and more ferric iron toward the higher grades is compatible with this gradual disappearance of graphite. In one desilicified schist lens in dolomitic marbles near the deepest-exposed part of the gneiss dome, some sapphirine has been found. The association consists of sapphirine–spinel–corundum–anorthite–calcite and phlogopite. Apparently the desilification is a consequence of the abundant formation of phlogopite and tremolite at the contact between the mica schist lens and the surrounding dolomitic marble (although most of the tremolite here may be of later formation).

Amphibolites consist of the same minerals as described for the siliceous streaks in the marbles. The mineralogy of the few metamorphosed ultramafic rocks preserved in this zone is similar to the mineralogy of the rocks occurring in the preceding zone. Chlorite and talc are secondary minerals here. Within the migmatite dome no emery deposits have been found. Beryl, topaz, monazite, and xenotime commonly are accessory minerals in some of the larger pegmatitic dikes.

CONDITIONS OF METAMORPHISM

In this section, an attempt will be made to estimate the pressure-temperature conditions of the regional metamorphism on Naxos. The isograds, as mapped in the field (fig. 3), can be used to determine pressure-temperature conditions on the basis of relevant experimental data. Some additional petrologic information is also used for the estimates. The pressure-temperature relations for the metamorphism in some areas of Naxos are converted in figure 7 into inferred vertical geothermal gradients. The temperature estimates are based on the assumption that the total pressure, being locally more or less equal to the water pressure, was around 5 to 7 kb, as explained below.

The diaspore dehydration curve is based on data from Kennedy (1959), Fyfe and Hollander (1964), and Haas (1972). The temperature for the + corundum isograd is estimated at about 420° to 440°C for the relatively high pressure involved (from the kyanite occurrences it can be inferred that the total pressure was at least about 3 kb). The + biotite isograd is taken to be about 500°C based on the experimental metamorphism of pelitic clays (Winkler, 1957; and unpublished work at our laboratory). The disappearance of chloritoid in the metamorphic series takes place in the temperature range of 540° to 580°C. The − chloritoid curve, as given in figure 7, is based on several reaction curves established by Richardson (1968) and Hoschek (1969). The Al-silicate diagram as given in figure 7 is intermediate between the diagrams of Althaus (1967) and Richardson, Gilbert, and Bell (1969) and is consistent with the

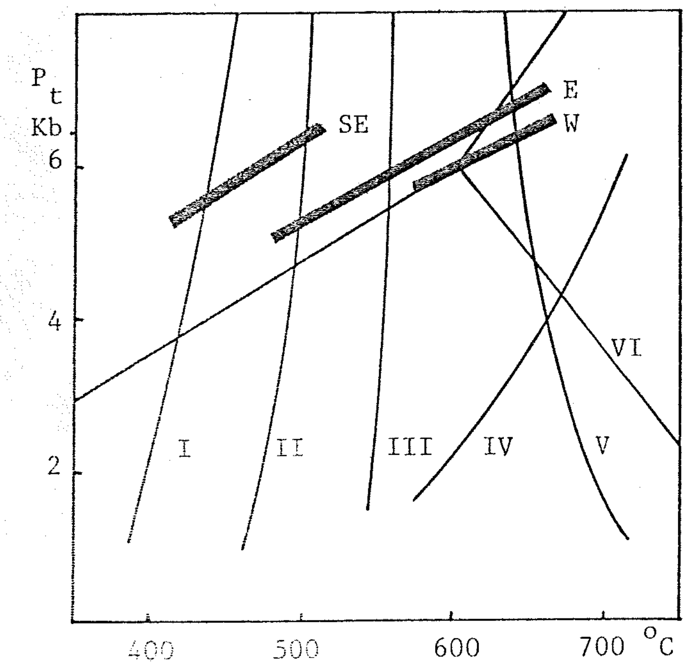

Fig. 7. Equilibrium curves of relevant mineral reactions under condition P water equals P total, and estimated P–T relations in different parts of Naxos, Greece, SE, E, and W indicate inferred vertical geothermal gradients during metamorphism for respectively SE-, E-, and W- part of Naxos. I, dehydration curve of diaspore (Haas, 1972). II, formation of biotite in pelitic rocks (Winkler, 1957; and our data). III, compilation of breakdown reactions of chloritoid (Richardson, 1968; and Hoschek, 1969); IV, breakdown reaction of muscovite plus quartz (Evans, 1965); V, minimum melting curve of granite (Luth, Jahns, and Tuttle, 1964); VI, average phase diagram of the Al-silicates (Althaus, 1967; and Richardson, Gilbert, and Bell, 1969). For further explanation, see section "Conditions of Metamorphism."

experimental data of Newton (1966). This diagram fits better with the other P-T estimates for the metamorphic conditions on Naxos, but, especially due to the disappearance of chloritoid and to the appearance of staurolite, it differs from earlier estimates made by one of us (Schuiling, 1957, 1962b) as well as from the experimental work of Holdaway (1971). The temperature of the transition zone of kyanite into sillimanite can be estimated at about 620° to 660°C. Under the condition water pressure equals total pressure, the beginning of melting of granite would be around 630° to 640°C for the high pressures involved (Luth, Jahns, and Tuttle, 1964). Our present estimate for the beginning of melting at Naxos is between 660° and 690°C, because CO_2 was almost certainly a major component of the fluid phase (Kreulen, personal commun., 1975).

All the mineral occurrences and mineral assemblages mentioned in the petrological description of metamorphic zones I to VI are consistent with the general trend of temperature estimates (Jansen and Schuiling,

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in preparation). The temperature estimates from experimental data are also in good agreement with oxygen isotope work. Noticeable discrepancies in metamorphic temperatures, as determined by fractionation of the oxygen isotopes for a few mineral pairs, occur at higher grades of metamorphism. These minerals, especially the micas, have apparently undergone isotopic reequilibration during cooling, or possibly a later alpine thermal event (Rye, Schuiling, Rye, and Jansen, 1976).

Our pressure estimates are based on the following observations. The assemblage of glaucophane with epidote and calcite suggests moderately high pressures for the low grade rocks in Naxos (Ernst, 1968, and personal commun.). The distribution of the Al-silicates provides a good estimate of these relatively high pressure conditions, especially since most of the Barrovian-type of metamorphism on Naxos has taken place under conditions on the high-pressure side of the triple point. On the west side of the dome the conditions must have been confined to a bit lower P-T conditions near the triple point. The kyanite–epidote assemblages, observed in the emery deposits near the — chloritoid isograd on the east side of the dome may suggest, according to Storre and Nitsch (1974), that we are dealing with a pressure of more than 5 kb. The fact that the disappearance of the primary muscovite plus quartz assemblage coincides roughly with the beginning of melting in pelitic rocks around the gneiss dome, although in some parts of the anatectic rocks muscovite plus quartz remains stable, provides a local water pressure estimate of more than 3.5 kb. This estimate becomes on the order of 4 kb, taking into account the composition of plagioclase (about 10 percent An). The occurrence of kyanite close to the beginning of melting in the southeast corner of the migmatite dome points to a minimum total pressure of about 6 kb. The assemblage talc–magnesite–anthophyllite in zone IV and the Mg-rich sapphirine + calcite occurrence in the deepest part of the migmatite dome may be used as indications for local high CO_2 pressures under the temperatures and total pressures involved (Johannes, 1969; Seifert, 1974). In other places assemblages like anorthite–zoisite–corundum in zone V_B and vesuvianite–occurrences in zone V_A, sometimes associated with diopside–epidote–grossular, indicate that here the mole fraction of water in the fluid must have been very high (Storre and Nitsch, 1972; and Johannes, personal commun., 1975).

The conclusion of this section is that the metamorphism on Naxos took place within a gradually increasing temperature range from about 400°C in southeastern Naxos to about 700°C in the deepest part of the migmatite dome, while the total pressure must have been on the order of 5 to 7 kb for the whole complex. In a number of cases it can be shown that the composition of the fluid phase was locally buffered and varied from place to place, although in most cases a pervasive fluid phase in which both water and CO_2 played a major role, was presumably present.

GEOTHERMAL GRADIENTS DURING METAMORPHISM

Naxos, and as we know, the Cyclades in general, are unusual in showing a transition range of metamorphic facies from glaucophane schist into the migmatites of the amphibolite facies. The two facies are commonly considered to belong to different metamorphic facies series (Miyashiro, 1961). None of the Cyclades shows the complete, continuous range of metamorphism, although Naxos displays almost the complete range. Fieldwork in the Cyclades by Marinos (1942), Davies (1966), Dixon (ms), and by the department of geochemistry of the Vening Meinesz Laboratory has shown that on some islands a true high pressure glaucophane schist facies has been developed. On Ios, glaucophane, columnar calcite taken to be pseudomorphic after aragonite, and pseudomorphs after lawsonite have been found. In a meta-bauxite lens the mineral assemblage diaspore–kyanite–chloritoid occurs. On Siphnos, glaucophane and jadeite-rich pyroxene are found, and on Syros, lawsonite, glaucophane, and the jadeite–quartz assemblage occur. On both islands columnar calcite is typical of the high-pressure assemblage, and omphacite with almandine for eclogitic rocks. On other islands normal greenschist facies or amphibolite facies, with or without migmatites, are found. For instance, on Kythnos the assemblage quartz–albite–chlorite–sericite–actinolite characterizes the greenschist facies whereas no glaucophane is found. The occurrence of migmatite and of the minerals staurolite and sillimanite on Paros indicates a metamorphism in the amphibolite facies.

At present there are few petrologic indications of more than one cycle of regional metamorphism on the Cyclades, except for a weak, probably very late Alpine overprint. There is some evidence for changing conditions on Ios and southeast Naxos, as can be seen from the existence of actinolitic rims around glaucophane crystals. Preliminary Rb-Sr muscovite ages also seem to indicate a rather long Alpine metamorphic history, but data are as yet insufficient to sketch the different stages of metamorphism in time. The close spatial relation of otherwise rather different metamorphic facies can be explained in our opinion by accepting that the geothermal gradients present during the regional metamorphism of the Cyclades varied over rather short distances. A model to explain such variation has already been set up by Richardson (1970), and the Cyclades may be a very good example for this model. In figure 8 we have attempted to sketch the distribution of isotherms during the final stage of metamorphism on Naxos, after the updoming of the migmatite, as well as their distribution in the high pressure metamorphism as found on Ios.

The diapiric rise of the migmatites and the updoming of the metamorphic rocks around it have the effect of enhancing the contrast between metamorphic rocks, both as regards the temperatures of metamorphism and as regards their metamorphic facies. Over horizontal distances of 10 to 15 km on Naxos we find regionally metamorphic rocks

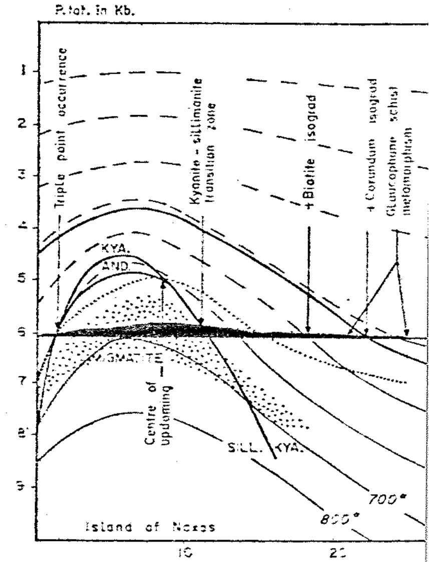

Fig. 8. Distribution of isotherms during t
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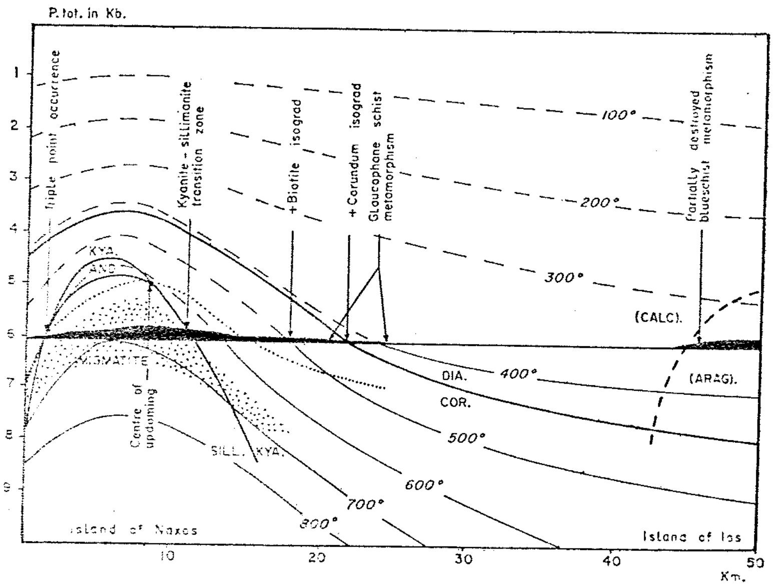

Fig. 8. Distribution of isotherms during the metamorphism in the central Cyclades after the updoming of the migmatites on Naxos. The heavy stippled line indicates the diapiric rise. Note that the center of the uplift does not coincide exactly with the focal point of the thermal rise, resulting in an asymmetric pattern of the metamorphic zones. Heating of the coolest part of the section causes the partial destruction of the glaucophane schist metamorphism (pseudomorphs of aragonite and lawsonite, and overgrowth of actinolite on glaucophane). The heavy dashed line indicates the stability field of aragonite before the heating of the coolest part. Erosion, finally, of the overlying metamorphic rocks exposes the section that can be studied at present on the islands of Naxos and Ios (black).

which crystallized under geothermal gradients ranging from 30°C/km for the western side of the migmatite dome to 22°C/km in southeast Naxos (fig. 7). Over distances of only a few tens of kilometers from Naxos to the islands of Syros, Siphnos, and a part of Ios, the geothermal gradients may drop to 15°C/km, or even less. Unfortunately, the fragmentary exposure of the geological elements of the Cycladic Massif and our resulting incomplete knowledge do not allow a conclusion on whether the Cycladic Massif is an example of an ancient paired metamorphic belt.

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REFERENCES

Althaus, Egon, 1967, The triple point andalusite–sillimanite–kyanite: Contr. Mineralogy Petrology, v. 16, p. 29-44.

Anastopoulos, J., 1963, Geological study of Antiparos Island group: Geol. Geophys. Research, Athens, v. 7, no. 5, p. 285-375 (English summary).

Boettcher, A. L., and Wyllie, P. J., 1968, The calcite–aragonite transition measured in the system $CaO–CO_2–H_2O$: Jour. Geology, v. 76, p. 314-330.

Chatterjee, N. D., 1974, Synthesis and upper thermal stability limit of 2 m-margarite, $CaAl_2(Al_2Si_2O_{10}/(OH)_2)$: Schweizer. min. pet. Mitt., v. 54, p. 753-767.

Davies, E. N., 1966, Der geologischer Bau der Insel Siphnos: Geol. Geophys. Research, Athens, v. 10, no. 3, p. 161-200 (German summary).

Ernst, W. G., 1968, Amphiboles: Monograph series of theoretical and experimental studies, v. 1: New York, Springer-Verlag, 125 p.

Evans, B. W., 1965, Application of a reaction-rate method to the breakdown equilibria of muscovite and muscovite plus quartz: Am. Jour. Sci., v. 263, p. 647-667.

Fyfe, W. S., and Hollander, M. A., 1964, Equilibrium dehydration of diaspore at low temperatures: Am. Jour. Sci., v. 262, p. 709-712.

Haas, Herbert, 1972, Diapore–corundum equilibrium determined by epitaxis of diaspore on corundum: Am. Mineralogist, v. 57, p. 1375-1385.

Holdaway, M. J., 1971, Stability of andalusite and the aluminum silicate phase diagram: Am. Jour. Sci., v. 271, p. 97-131.

Hoschek, G., 1969, The stability of staurolite and chloritoid and their significance in metamorphism of pelitic rocks: Contr. Mineralogy Petrology, v. 22, p. 208-232.

Jansen, J. B. H., 1973, Geological map of Naxos: Athens, Inst. Geol. and Mineral Researches.

Johannes, W., 1969, An experimental investigation of the system $MgO–SiO_2–H_2O–CO_2$: Am. Jour. Sci., v. 267, p. 1083-1104.

Kennedy, G. C., 1959, Phase relations in the system $Al_2O_3–H_2O$ at high pressures and temperatures: Am. Jour. Sci., v. 257, p. 563-573.

Luth, W. C., Jahns, R. H., and Tuttle, O. F., 1964, The granite system at pressures of 4 to 10 kilobar: Jour. Geophys. Research, v. 69, p. 759-773.

Marinos, G., 1942, Contribution à la pétrologie du système cristallophyllien du sud-est de la Grèce. L'île d'Ios: Ann. Géol. Pays Helléniques, Athens, v. 1, 42 p. (French summary).

Marks, P., and Schuiling, R. D., 1965, Sur la présence du Permien Supérieur non-métamorphique à Naxos: Prakt. Akad. Athens, v. 40, p. 96-99.

Miyashiro, A., 1961, Evolution of metamorphic belts: Jour. Petrology, v. 2, p. 277-311.

Newton, R. C., 1966, Kyanite–andalusite equilibrium from 700° to 800°C: Science, v. 153, no. 3732, p. 170-172.

Newton, R. C., and Kennedy, G. C., 1963, Some equilibrium reactions in the join $CaAl_2Si_2O_8–H_2O$: Jour. Geophys. Research, v. 68, p. 2967-2983.

Papastamatiou, J., 1963, Sur la présence de roches sédimentaires d'âge prétriassique à Mykonos (archipel des Cyclades, Grèce): Acad. Sci. Paris Comptes rendus, v. 256, p. 5167-5169.

Papavasiliou, S. A., 1909, Ueber die vermeintlichen Urgneise der Kykladen: Deutsche geol. Gesell. Zeitschr., v. 61, p. 34-201.

Priem, H. N. A., Boelrijk, N. A. I. M., Hebeda, E. H., Verdurmen, E. A. Th., and Verschure, R. H., 1969, Rb–Sr investigations on the island of Naxos (Greece), *in* Annual report ZWO Laboratory of Isotope Geology: Amsterdam, p. 76.

Richardson, S. W., 1968, Staurolite stability in part of the system Fe–Al–Si–O–H: Jour. Petrology, v. 9, p. 467-488.

———— 1970, The relation between a petrogenetic grid, facies series and the geothermal gradient in metamorphism: Fortschr. Mineralogie, v. 47, Heft 1, p. 65-76.

Richardson, S. W., Gilbert, M. C., and Bell, P. M., 1969, Experimental determination of kyanite–andalusite and andalusite–sillimanite equilibria, the aluminum silicate triple point: Am. Jour. Sci., v. 267, p. 259-272.

Rye, R. O., Schuiling, R. D., Rye, D., and Jansen, J. B. H., 1976, Carbon, hydrogen and oxygen isotope studies of the regional metamorphic complex at Naxos, Greece: Geochim. et Cosmochim. Acta, v. 40, p. 1031-1049.

Schreyer, W., and Seifert, F., 1969, High pressure phases in the system $MgO–Al_2O_3–SiO_2–H_2O$: Am. Jour. Sci., v. 267-A (Schairer v.), p. 407-443.

Schuiling, R. D., 1957, A geo-e
anite, andalusite): Koninkl.
3, p. 220-226.
———— 1962a, On petrology,
(SW-Turkey): Bull. M.T.A.,
———— 1962b, Die petrogene
Neues Jahrb. Mineralogie, v.
———— 1966, Continental d
p. 1027-1028.
———— 1969, A geothermal
Mijnb. Genoot., Proc., v. 26,
Seifert, F., 1974, Stability of sap
$MgO–Al_2O_3–SiO_2–H_2O$: Jour.
Storre, B., and Nitsch, K. H., 19
1 Calcit + 1 H_2O: Beitr. Min
———— 1974, Zur Stabilität
Mineralogie Petrographie, v.
Thompson, J. B., Jr., 1970, Ge
Cosmochim. Acta, v. 34, p. 52
Trikkalinos, T. K., 1947, Beiträg
lands. II. Ueber den tektoni
ques, 1, p. 7-37.
Trommsdorff, V., and Evans, B.
Schweizer. min. pet. Mitt., v.
Winkler, H. G. F., 1957, Expe
Metamorphose karbonatfreie
———— 1973, Petrogenesis of
237 p.

Schuiling, R. D., 1957, A geo-experimental phase-diagram of Al_2SiO_5 (sillimanite, kyanite, andalusite): Koninkl. Nederlandse Akad. Wetensch., Proc., ser. B, v. 60, no. 3, p. 220-226.

——— 1962a, On petrology, age and structure of the Menderes migmatite complex (SW-Turkey): Bull. M.T.A., v. 58, p. 71-84.

——— 1962b, Die petrogenetische Bedeutung der drei Modifikationen von Al_2SiO_5: Neues Jahrb. Mineralogie, v. 9, p. 200-214.

——— 1966, Continental drift and oceanic heat-flow: Nature, v. 210, no. 5040, p. 1027-1028.

——— 1969, A geothermal model of oceanization: Koninkl. Nederlandse Geol. Mijnb. Genoot., Proc., v. 26, p. 143-148.

Seifert, F., 1974, Stability of sapphirine: a study of the aluminous part of the system $MgO–Al_2O_3–SiO_2–H_2O$: Jour. Geology, v. 82, p. 173-204.

Storre, B., and Nitsch, K. H., 1972, Die Reaktion 2 Zoisit + 1 CO_2 = 3 Anorthit + 1 Calcit + 1 H_2O: Beitr. Mineralogie Petrographie, v. 35, p. 1-10.

——— 1974, Zur Stabilität von Margarit im System $CaO–Al_2O_3–SiO_2–H_2O$: Beitr. Mineralogie Petrographie, v. 43, p. 1-24.

Thompson, J. B., Jr., 1970, Geochemical reactions and open systems: Geochim. et Cosmochim. Acta, v. 34, p. 529-551.

Trikkalinos, T. K., 1947, Beiträge zur Erforschung des tektonischen Baues Griechenlands. II. Ueber den tektonischen Bau der Insel Naxos: Ann. Géol. Pays Helléniques, 1, p. 7-37.

Trommsdorff, V., and Evans, B. W., 1974, Alpine metamorphism of peridotitic rocks: Schweizer. min. pet. Mitt., v. 54, p. 333-352.

Winkler, H. G. F., 1957, Experimentelle Gesteinsmetamorphose I; Hydrothermale Metamorphose karbonatfreier Tone: Geochim. et Cosmochim. Acta, v. 13, p. 42-69.

——— 1973, Petrogenesis of metamorphic rocks, 3d ed.: New York, Springer-Verlag, 237 p.