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# HIGH-GRADE METAMORPHIC ROCKS OF THE MELLID AREA, GALICIA, NW SPAIN

BY

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## ABSTRACT

This study concerns the petrology of the Mellid area, the SE portion of the outer zone of the Ordenes Complex which is one of the upthrusted Precambrian complexes in the axial zone of the Hercynian orogen in Galicia, NW Spain. An eugeosynclinal rock sequence is found containing units with different metamorphic evolutions. All units were affected by Precambrian tectonization and retrogressive metamorphism. This orogeny may be subdivided into three metamorphic and four deformation phases which caused definite changes in the mineralogical composition and the texture of the rock.

The sequence of metamorphic phases, established in a granulite facies unit is as follows: the first phase of Precambrian metamorphism is characterized by the (hornblende-)granulite facies, more precisely the (hornblende-)clinopyroxene-garnet-sodic plagioclase subfacies of the kyanite-bearing granulite facies. At that time, $P_{H_2O}$ must have been very low locally. The second and third phases were marked by the hornblende-clinopyroxene-garnet-sodic plagioclase subfacies and the amphibolite facies, respectively. The other units bear witness to lower grade metamorphic activities.

The granulite facies unit comprises metamorphosed basic lavas, metapelitic rocks, garnet-bearing metagabbros and garnet-bearing peridotites. The metapelitic rocks (kyanite-garnet-orthoclase-sodic plagioclase-biotite) and the metamorphosed basic lavas (clinopyroxene-garnet-sodic plagioclase-amphibole) are described in detail. The latter rocks contain Ca-rich inclusions, displaying scapolite-bearing mineral assemblages. The inclusions can be ascribed to deuteric alteration or incipient metamorphism in the basic lavas prior to the granulite facies metamorphism.

The other units contain metasedimentary rocks and granitic and granodioritic orthogneisses.

Metamorphic conditions during the Hercynian orogeny did not go further than the lower amphibolite facies. Therefore, retrogradation of the Precambrian units continued but a clear conversion of the rock texture cannot be discerned. The most important Hercynian event in the Mellid area was the emplacement of an ophiolitic rock suite.

## RESUMEN

Este estudio trata de la petrología de la región de Mellid, la parte sureste de la periferia del Complejo de Ordenes, uno de los complejos Precámbricos de la zona axial del orogenio Hercínico en Galicia, España. Se han encontrado una secuencia eugeosinclinal, en la cual se pueden distinguir tres unidades con una evolucíon metamórfica diferente. Durante el Precámbrico todas las unidades sufrieron una tectonización en cuatro fases y un metamorfismo retrógrado en tres fases. La retrogradación trajó consigo unas reconstituciones de la composición mineralógica y de la micro-estructura de las rocas.

La secuencia de las fases metamórficas en la unidad de la facies granulítica es la siguiente: la primera fase del metamorfismo Precámbrico era al alcance de la facies granulítica, más exactamente la subfacies de (hornblenda-)clinopiroxena-granate-plagioclasa sódica de la facies granulítica del tipo distena. En aquella fase las presiones de vapor de agua han sido localmente muy bajas. La segunda y tercera fase eran caracterizadas por la subfacies de hornblenda-clinopiroxena-granate-plagioclasa sódica y la facies anfibolítica respectivamente. Las otras unidades sufrieron un metamorfismo menos elevado.

La unidad de la facies granulítica contiene lavas básicas metamórficas, rocas metapelíticas, gabbros y peridotitas metamórficos con granate. Las lavas básicas metamórficas contienen inclusiones muy abundantes en Ca, mostrando parageneses con scapolita. Estas inclusiones se han formado por una alteración deuterica o un metamorfismo incipiente en las lavas, antes del metamorfismo de la facies granulítica.

Las otras unidades contienen metasedimentos y ortogneises.

Las condiciones metamórficas durante la orogénesis Hercínica no exedieron la parte baja de la facies anfibolítica. El fenómeno Hercínico más importante en la región de Mellid es el emplazamiento de una serie de rocas ofiolíticas.

## CONTENTS

# CHAPTER I

# INTRODUCTION

## I.A. GEOGRAPHY

The investigated area surrounds the township of Mellid in the southeastern part of the province of La Coruña, Galicia, NW Spain. The southern part of the area extends beyond the border river, Rio Ulla, into the province of Pontevedra. Mellid is located at the crossroads Lugo-Santiago de Compostela (C. 547) and Betanzos-Lalín (C. 540).

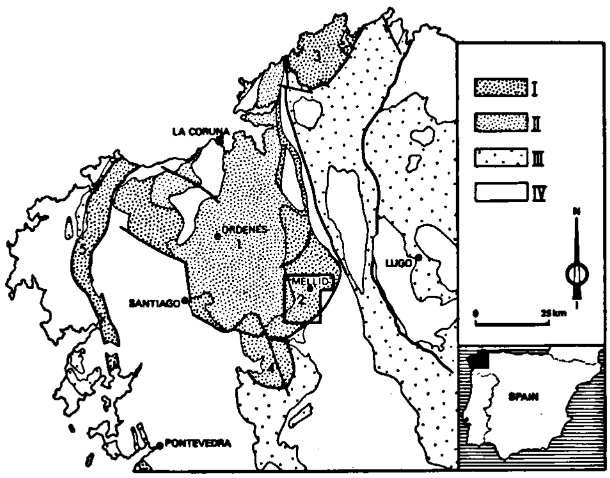


Fig. 1. Simplified geological map of northwestern Galicia, from the 'Carte Géologique du Nord Ouest de la Péninsule Iberique' (1967). Legend:

I: Precambrian polyorogenic complexes:
1: Ordenes Complex; 2: area investigated (part of 1); 3: Cabo Ortegal (Vogel, 1967); 4: Lalín unit (Hilgen, 1971); 5: blastomylonitic 'Graben' (den Tex & Floor, 1967)
II: metasedimentary central portion of the Ordenes Complex
III: Upper Precambrian and Lower Paleozoic (meta)sedimentary rocks
IV: granites, granodiorites and migmatites

The topographical data on which the enclosed geological map is based were taken from the maps edited by the Cartografia Militar de España (scale 1:25,000): sheet 96 (quadrants I, II, III, IV) and sheet 122 (quadrants I, IV). The new road from Mellid southwest to Las Cruces is roughly indicated on the geological map on the basis of field observations.

## I.B. GEOLOGICAL SETTING

The Mellid area forms part of the outer zone of the Ordenes Complex, formerly called the Ordenes 'Basin' because of its generally inward-dipping loop structure (Fig. 1). Parts of this mainly mafic and ultramafic rock-bearing outer zone were described by Koning (1966), Warnaars (1967) and van Zuuren (1969). Parga-Pondal (1956) described the rocks surrounding the metasedimentary core of the Ordenes Complex as 'rocas del lopolito', because of their quasi-circular configuration on the map.

Parts of the outer zone of the Ordenes Complex belong to the upthrusted Precambrian, polyorogenic complexes in the axial zone of the Hercynian orogen in NW Spain, also called the Hesperian Massif. Comparable Precambrian complexes in the Hesperian Massif are found at Cabo Ortegal (Vogel, 1967; Engels, 1972; den Tex, Engels & Vogel, 1972) and at Morais-Lágoa and Bragança-Vinhais, NE Portugal (Anthonioz, 1969,1970). Evidence for a Precambrian age of the Cabo Ortegal Complex of 900 ± 30 m.y. was presented by Vogel and Abdel-Monem (1971). The complexes show rather similar lithologies and were all affected by a Precambrian high-pressure metamorphism, which preceded the low-pressure Hercynian metamorphism.

The most recent study of the Hesperian Massif was presented by Parga & Vegas (1971) and Bard, Capdevila & Matte (1971). The geology of western Galicia has recently been reviewed by den Tex & Floor (1971).

## I.C. THE (HORNBLENDE-)CLINOPYROXENE-GARNET-SODIC PLAGIOCLASE SUBFACIES OF THE KYANITE-BEARING GRANULITE FACIES

The most high-grade metamorphic conditions to be recognized in the Mellid area are those which prevailed during the first phase of Precambrian metamorphism. The conditions were those of the granulite facies. The following mineral associations led to this assumption:

1) clinopyroxene-garnet-oligoclase/andesine-hornblende in metabasic rocks.

2) kyanite-garnet-orthoclase-oligoclase/andesine-biotite in metapelitic rocks.

These assemblages have also been reported in other Precambrian complexes in the Hesperian Massif, viz. Cabo Ortegal (Vogel, 1967; den Tex et al., 1972) and the complexes of Morais-Lagoa and Bragança-Vinhais (Anthonioz, 1969).

If considered separately, both the basic and the pelitic parageneses have been reported in high-grade areas, but a joint occurrence has seldom been recorded in other granulite facies complexes. As far as we know, comparable parageneses have only been described for the granulitic complex of Stary Gieraltow, East Sudetes, Poland, by Kozlowski (1961). Therefore, these contemporaneously occurring stable parageneses require special attention. To designate the subfacies defined by the associated assemblages, the author proposes the term (hornblende-)clinopyroxene-garnet-sodic plagioclase subfacies. The reason why current names are not suitable for this purpose will be discussed below.

Engels (1972) and den Tex et al. (1972) recognized the subfacies under discussion as part of a high-pressure intermediate- temperature facies series (den Tex, 1971). They reported a prograde sequence from the staurolite-almandine-muscovite and the kyanite-almandine-muscovite subfacies of the almandine-amphibolite facies, via the subfacies in question, to the eclogite facies.

Actually, the most striking feature is the absence of sillimanite and orthopyroxene in the pelitic and basic rocks, respectively. This must be due to a low geothermal gradient. Green & Ringwoord (1967) demonstrated that the orthopyroxene + anorthite $\leftrightharpoons$ clinopyroxene + garnet + quartz transformation proceeds to the right with decreasing temperature or increasing pressure in rocks of basaltic composition. Turner (1968) reviewed such well-described granulite facies areas as the Adirondack massif (USA), Ceylon and Broken Hill (Australia); Winkler (1967) also discussed the Saxon granulite range and similar complexes in the Moldanubicum. It appears from their considerations that kyanite, although much less frequent than sillimanite, may be present in granulite facies rocks but not in association with basic rocks displaying a stable coexistence of clinopyroxene-garnet-sodic plagioclase.

In their comprehensive surveys, Miyashiro (1961) and Hietanen (1967a) showed that the high-temperature extremes of most recorded facies series penetrate the sillimanite stability field. The series discussed here obviously extends well into the kyanite stability field, which characterizes its low geothermal gradient. Therefore, this facies series could be designated as the high-pressure or kyanite type of regional metamorphism. It can be fitted into Miyashiro's (1961) scheme as a high-pressure intermediate series between the kyanite-sillimanite and the jadeite-glaucophane type. On the basis of experimental data, den Tex et al. (1972) deduced the rather low geothermal gradient of about 16–18 °C/km for this series.

The low-pressure extreme of the facies series is not very well represented in the Galician complexes, but it may be comparable to series described by Hashimoto (1972). He reported two facies series of the high-pressure intermediate group (Miyashiro, 1961) ending in the epidote-amphibolite zone.

An often-quoted subdivision of the granulite facies, proposed by de Waard (1965), is the (hornblende-)clinopyroxene-almandite and the (hornblende-)orthopyroxene-plagioclase subfacies in order to distinguish between high-pressure and low-pressure granulite facies metamorphism. In this scheme however sillimanite should always be the $Al_2SiO_5$ polymorph. The kyanite-orthoclase association in pelitic rocks is not critical: for example it also occurs in the Saxonian type (Hietanen, 1967a), which is characterized by orthopyroxene and plagioclase in mafic rocks and even in high-pressure amphibolite facies rocks. Therefore, the name should define the mafic mineral assemblage and the frequent and stable occurrence of kyanite in the pelitic rocks. Consequently, the author proposes the (hornblende-) clinopyroxene-garnet-sodic plagioclase subfacies of the kyanite-bearing granulite facies. Brackets around hornblende indicates the incidental absence of hydrated minerals.

## I.D. GEOLOGY OF THE MELLID AREA

The surroundings of Mellid have been investigated by graduate students of Leiden University since 1965. This survey resulted in four unpublished theses by Saltet (1968), van Scherpenzeel (1969), Hubregtse (1970) and Dantuma (1972). Using data from these reports to support additional laboratory and field work, the present author reported on the gabbroic rocks (Hubregtse, 1973) and element distribution in the granulite facies rocks (Hubregtse, in press) of the Precambrian part of the Mellid area. The present study deals with the regional geology of the area. Permission to use hand specimens, thin sections, maps and other data by the above-mentioned authors is gratefully acknowledged.

### I.D.1. *Explanation of terms*

The following subdivision is used for grain sizes:

| | |
|---|---|
| >3 cm | very coarse-grained |
| 3 cm–5 mm | coarse-grained |
| 5 mm–1 mm | medium-grained |
| 1 mm–1/3 mm | fine-grained |
| 1/3 mm–1/100 mm | microcrystalline or aphanitic |

When the colour of a pleochroic mineral is mentioned, it is always that of $\gamma$.

Foliation and schistosity are used to describe a planar mineral orientation in non-fissile and fissile rocks, respectively. Layering denotes a compositional heterogeneity of the rock. Banding and lamination indicate layering on a macro and micro scale, respectively.

Where rocks of the granulite facies are named, the proposals of Behr et al. (1971) and Mehnert (1972) are followed.

Phases of metamorphism and deformation are designated as M and F, respectively. The suffixes represent the relative ages of these events.

Local rivers, villages, etc. are indicated on the enclosed detailed geological map (Enclosure 1).

All the described samples are stored at the Rijksmuseum voor Geologie en Mineralogie in Leiden. The samples are numbered from RGM 129080 to RGM 129156. Their locations are indicated on the sample map (Enclosure IV). The sample numbers are used in the text without the prefix 'RGM'.

### I.D.2. *Evolution of the investigated area*

In the following, the detailed geological map (Enclosure 1) and the map units in Fig. 2, which is a simplified representation of the geology of the investigated area, are frequently used for reference. The main subject of this study is the Precambrian rock complex, comprising roughly the area northwest of the Rio Ulla and the Rio Furelos: map units 1, 2 and 3 in Fig. 2.

The geological history of the area started with the development of an eugeosyncline in Precambrian times. Semi-pelitic rocks, greywackes, etc. (parts of 1b and 3) were deposited and subsequently intruded by a calcalkaline granite-granodiorite series (2 and part of 3). Basic lavas (parts of 1), belonging to an ophiolitic suite, were also deposited prior to the Precambrian orogeny. In addition ultramafic rocks derived from the mantle (largely 1a) were emplaced during this orogenic event (Maaskant, 1970); minor gabbro intrusions were emplaced at different levels (in 1b, 2 and 3; Hubregtse, in press). The conditions ranged from granulite to greenschist facies. The present appearance of the Precambrian rocks is highly variable owing to superimposed retrograde metamorphic and tectonic activities.

#### I.D.2.1. *Precambrian tectonic and metamorphic phases.*

– The sequence of Precambrian retrograde metamorphic phases has been deduced from the relative ages of mineral assemblages. Dating of mineral assemblages may be possible on the basis of their:

relationship to folding phases, e. g. oriented mineral growth,

occurrence as cataclastic mineral grains or aggregates in subsequently recrystallized rocks,

replacement by other mineral associations.

Three metamorphic and four tectonic Precambrian phases have been detected in this way. Their relative ages

| deformation phases | metamorphic phases | highest grade of metamorphism |
|---|---|---|
| $F_1$ | $M_1$ | (hornblende-) granulite facies |
| $F_2$ | | |
| $F_3$ | $M_2$ | hornblende granulite facies |
| $F_4$ | $M_3$ | amphibolite facies |

Table I. Synopsis of the relationship between metamorphic and deformation phases and highest grade of accompanying metamorphism.

are shown in Table I. These events gave rise to mylonitization, isoclinal folding and recrystallization of the rocks.

Textural relics are the only record of the activities of $M_1$, $F_1$ and $F_2$. Megascopic structures due to these phases have not been found. $F_1$ and $M_1$ resulted in medium- to coarse-grained granofelsic textures without a clearly preferred mineral orientation. $F_2$ textures have

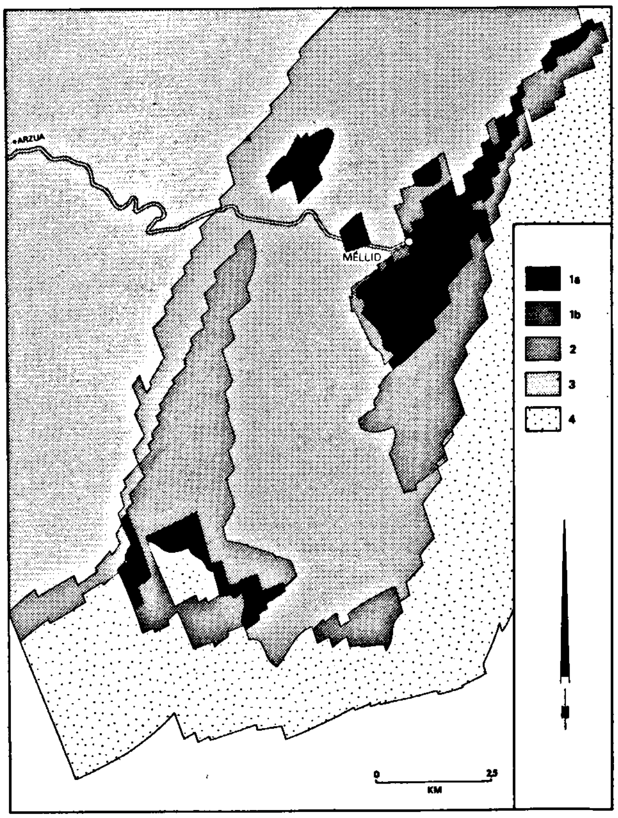


Fig. 2. Simplified geological map of the Mellid area, see I.D.2.1 and I.D.2.2.

1: high-grade unit; a: predominantly garnet-bearing peridotites; b: metabasic and metapelitic rocks

2: intermediate-grade unit: orthogneisses

3: low-grade unit: metasediments of the Ordenes Complex and orthogneisses

4: Hercynian ophiolitic rocks

only been found in the rocks of map unit 1. Samples of this textural type display a weak foliation and a manifest partial recrystallization of the (hornblende-)granulite facies mineral assemblages.

The formation of NW–SE trending $F_3$ structures coincided with $M_2$. $F_3$ structures are preserved as lineations and isoclinal folds with axial planes which dip NW in map unit 3, and as isoclinal folds with subhorizontal axial planes in map unit 1.

$M_3$ was initiated by the development of NNE–SSW trending $F_4$ structures, recognized as subhorizontal lineations in map unit 2 and as isoclinal folds with subhorizontal axial planes in map unit 1.

Both $F_3$ and $F_4$ led to the present blastomylonitic aspects of the high-grade rocks: the textural types will be discussed in chapter II.

I.D.2.2. *Units of different metamorphic grade.* – In the Mellid area all facies boundaries are marked by a tectonic discontinuity because of faulting during both the Precambrian and the Hercynian orogenies. Thus, parts of the Precambrian complex were brought into juxtaposition, resulting in a telescoped facies series of retrogressive metamorphism. Fig. 2 shows a division of the Precambrian rock complex into units according to the grade of metamorphism:

1) high-grade metamorphic unit, comprising the catazonal rocks: metabasites, paragneisses, metagabbros and metaperidotites.

2) intermediate-grade metamorphic unit, consisting of granodioritic orthogneisses with augen textures and metagabbros.

3) low-grade metamorphic unit, containing the metasedimentary rocks of the Ordenes Complex, granitic orthogneisses with cataclastic textures and minor metagabbros.

This subdivision holds in fact only for the metamorphic conditions during $M_2$, as will be discussed below.

Syntectonic movement during the Precambrian led to different grades of metamorphism in the separate units during the same event. A sequence of facies during the Precambrian is outlined in Table II.

In the high-grade unit, the metamorphic conditions during $M_1$ were those of the (hornblende-)clino-pyroxene-garnet-sodic plagioclase subfacies (see chapter II). Although difficult to detect, amphibolite facies conditions apparently prevailed at that time in the lower grade units. Gabbro intrusions were emplaced prior to or during $M_2$. Adjustment to the regional metamorphic conditions prevailing in each unit is reflected in the mineral assemblages of the gabbroic rocks (Hubregtse, in press). The following characteristics are worth mentioning:

The gabbros of the high-grade unit display high-pressure granulite facies assemblages, i. e. incompatible calcic plagioclase and igneous pyroxenes yielded a jadeitic diopside-pyralmandine-oligoclase-rutile association, later including amphibole. Orthopyroxene has never been found.

| | high-grade unit | intermediate-grade unit | low-grade unit |
|---|---|---|---|
| $M_1$ | (hornblende -) granulite facies | amphibolite facies | |
| $M_2$ | hornblende granulite facies | amphibolite facies | greenschist facies |
| $M_3$ | amphibolite facies | | greenschist facies |

Table II. Sequence of facies in units of different metamorphic evolution.

In the intermediate-grade unit, the igneous mineral assemblages of the gabbro were converted to amphibolite facies associations. Orthopyroxene was partially replaced by amphibole-biotite-quartz symplectites, which were also recognized by Katz (1968) and Griffin & Heier (1969) as products of retrograde amphibolite metamorphism. A second generation of garnet was formed in rocks of both the high-grade and the intermediate-grade units during $M_2$.

The low-grade unit is devoid of second generation garnet. Moreover the gabbros display greenschist facies mineral assemblages.

Consequently, it is warranted to state that all three units suffered different grades of retrograde metamorphism during $M_2$. The original difference in depth between the high-grade and the intermediate-grade units must have been cancelled by thrust-faulting prior to $M_3$, because similar amphibolite facies mineral assemblages formed in both units during that phase. The thrust plane that separates the units was later folded during the Hercynian and accentuates the megastructures due to that orogeny. No significant mineralogical changes due to $M_3$ are found in the low-grade unit.

I.D.2.3. *Hercynian orogeny; fault systems.* – The effect of the Hercynian orogeny on the Precambrian complex was considerably less. It caused large-scale folding, steep faulting and the formation of secondary mineral assemblages, but there was not an obvious reconstruction of the microscopic rock fabric. The grade of Hercynian metamorphism is not quite clear, but it certainly did not go further than the lower amphibolite facies.

The most important Hercynian event in the Mellid area was the emplacement of an ophiolitic rock suite. A Silurian age has been attached to a comparable suite at Cabo Ortegal by Ho Len Fat, Koning & van der Meer Mohr (in press). The ophiolitic rocks, metagabbros, serpentinites and chlorite peridotites (map unit 4 in Fig. 2) form an arcuate zone outside the older complex. The ophiolites probably facilitated the upthrust of the Precambrian complex along NE–SW trending fault planes which dip NW.

The southern limit of the ophiolitic rock complex is also a NE–SW fault, locally marked by talc and graphite schists and quartz veins. Only its ENE–WSW trending southern extremity is shown on the map. A N to NW dipping fault plane separates the ophiolitic rocks from a zone of greenschist facies rocks (Hilgen, 1971).

Elsewhere in W-Galicia two main Hercynian deformation phases are known to have caused the formation of N–S trending fold axes with subhorizontal and subvertical axial planes, respectively (den Tex & Floor, 1971). This relationship is rarely found in the Mellid area. Most Hercynian fold axes trend NNE–SSW to ENE–WSW, and are subhorizontal with steep NW dipping or subvertical axial planes. These folds developed in the ophiolitic rocks and in the schists and phyllites of the low-grade unit. The large- scale NE–SW trending vaulted structures of the intermediate and high-grade units may possibly be due to the same deformation phase. The structures are largely parallel to the marginal faults of the Precambrian complex as well as the ophiolitic rock suite. Therefore we assume an intimate relationship between these structures and the upthrust of the Precambrian rock complex. Furthermore the Precambrian units must have been brought up to the same level along the westernmost NE–SW fault during this event.

The two-mica granite in the southeastern part of the mapped area is a product of Hercynian magmatism. It is part of the large and widespread granite series in the Hercynian metamorphic belt.

One of the latest Hercynian events is the development of numerous steep NNW–SSE faults. They caused a deflection to the north of the trend of the lithological boundaries as well as older structural trends. In addition porphyries were locally emplaced.

Other parts of Galicia, in particular the eastern region, are more appropriate for a detailed study of the effects of the Hercynian orogeny, in which mainly acid plutonic and sedimentary rocks were involved. For more detailed information the reader is referred to papers by Floor (1966), Matte (1968), Capdevila (1969) and Arps (1970).

# CHAPTER II

# HIGH-GRADE UNIT

## II.A. METABASIC ROCKS

### II.A.1. *Petrography*

This group comprises all basic rocks of the high-grade unit, with the exception of the peridotites and the metagabbros which have been extensively described by Maaskant (1970) and Hubregtse (1973, in press), respectively. These greenish black basic rocks occur in two main N–S trending zones: one in the central part of the area, the other E of Mellid in the region of the Rio Furelos. In addition some isolated outcrops are found on both banks of the Rio Ulla. Their composition is olivine-tholeiitic. This will be discussed in II.A.3. Retrogradation caused a diversified appearance both mineralogically and texturally. This is why the general term clinopyroxene-garnet-amphibole gneiss is used in a separate paper (Hubregtse, 1973) on the mineral chemistry of this rock. Macroscopically, the metabasic rocks are striped due to the development of zones with considerable amounts of epidote minerals during retrogressive metamorphism, and also as a result of initial differences in the chemical composition (see II.B).

Comparable metabasic rocks have been studied by Vogel (1967) and Engels (1972) from Cabo Ortegal and by van Zuuren (1969) from an area E of Santiago de Compostela, both in W Galicia. Vogel and van Zuuren used the term pyrigarnite. Anthonioz (1969) reported similar rocks from the complexes of Morais and Braganca in NE Portugal.

The metabasites showing mineral assemblages representative of granulite facies conditions will be considered first.

II.A.1.1. *Granulite facies metabasic rocks.* – These rocks are most abundant in the two N–S trending zones, particularly near the Rio Catasol. Three types are distinguished: the granofelsic type, the granoblastic type and blastomylonitic type I.

The *granofelsic type* displays the texture and mineral assemblage of the first metamorphic phase, $M_1$. Generally this type did not survive the subsequent deformations and retrogradations. Only polymineralic aggregates, up to 1.5 cm across and embedded as relics in the recrystallized counterpart, give an impression of what it once was: a medium- to coarse-grained rock without a clearly preferred orientation of the minerals. According to Vogel (1967) and Engels (1972), this textural type is due to tectonic phase $F_1$. Macroscopical structures of this type are rather well-preserved in the complex of Cabo Ortegal. According to the 'proposal for a general definition of granulite' (Mehnert, 1972), these rocks should be called granofelsic pyrigarnites. Specimen 129083 is a good illustration of the microscopical appearance of the granofelsic type; see Fig. 3.

Light green diopside ($2V_\gamma \approx 61°$, $[001] \wedge \gamma \approx 43°$) and pink hypidiomorphic garnet are the main constituents. Plagioclase, about 32% An in this case but generally 26–36% An, and greenish brown amphibole ($[001] \wedge \gamma \approx 17°$) are less abundant. Accessory minerals are quartz, rutile, sphene, apatite and opaque minerals. Pistacite is a secondary mineral.

Rocks of this type are more abundant in the metabasites near Sobrado, N of the Mellid area (H. Koning, pers. comm.).

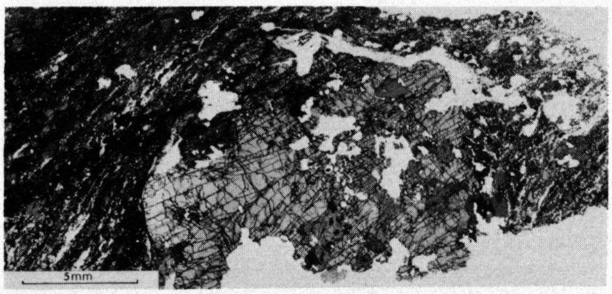


Fig. 3. Granofelsic inclusion in a blastomylonitic hornblende pyrigarnite. Thin section RGM 129083.

The *granoblastic type* is locally preserved near the Rio Catasol, both N and S of the Mellid-Arzua road. Mineralogically it does not differ from the granofelsic type. The texture however has changed due to a second tectonic phase, which caused a foliation in the granofelsic pyrigarnites. The prevailing metamorphic conditions were still those of the (hornblende-) clinopyroxene-garnet-sodic plagioclase subfacies, $M_1$. Recrystallization of the main constituents into elongated mosaics warrant this conclusion (Fig. 4). Garnet did not recrystallize, but there is some evidence of renewed growth. The overall grain size is reduced in comparison with that of the granofelsic pyrigarnites. For this reason, neither granofels nor, because of the large quantity of ferromagnesian minerals, granulite is an appropriate name for this rock. Therefore the term granoblastic pyrigarnite will be used. Representative samples are 129088, 129084 and 129086.

As stated previously, conditions were unfavourable for extensive amphibole formation. Specimen 129088 is a good illustration in support of this assumption. The sample consists almost completely of poikiloblastic garnet, embedded in recrystallized clinopyroxene (Fig. 4); minor amounts of plagioclase, quartz, rutile, titanite and opaques were also found. The garnets contain inclusions of biotite and amphibole with the same greenish brown colour as in the granofelsic type. The newly formed 'dry' mosaics of recrystallized clinopyroxene point to local dehydration of amphibole, which was an important constituent as indicated by the inclusions in garnet.

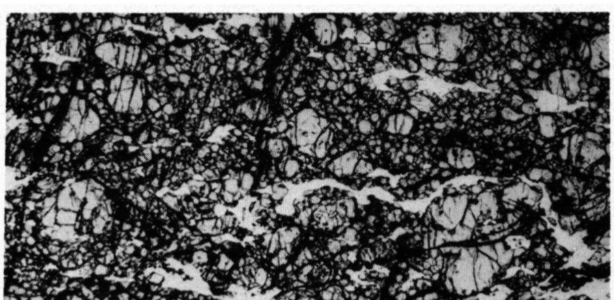


Fig. 4. Garnets, larger grains, in mosaics of recrystallized clinopyroxene. Thin section RGM 129088.

The other samples of granoblastic pyrigarnites, 129084 and 129086, also have relatively low modal contents of amphibole, about 30 vol% for both. In specimen 129084 the clinopyroxene is colourless and the amphibole shows a deep orange tinge. The colours of the minerals in specimens 129088 and 129086 do not deviate from those reported for the granofelsic type.

Engels (1972) recognized a very similar sequence of tectonic and metamorphic events in the catazonal complex of Cabo Ortegal. However in that region the second tectonic phase, $F_2$, caused cataclasis only; recrystallization of the mafic minerals was not evident. In this respect the order of events is obviously different in the Mellid area, as described above. It is the author's opinion that prolongation of the (hornblende-)clinopyroxene-garnet-sodic plagioclase subfacies conditions, $M_1$, was related to the emplacement of the garnet-bearing peridotite, which is in direct contact with the metabasic rocks in the northeastern part of the Mellid area. The following considerations support this hypothesis.

Samples of granoblastic pyrigarnites were found in the immediate surroundings of the peridotite body. The recrystallization of the ferromagnesian minerals and the dehydration process near the contact and in the general neighbourhood of the ultramafic body suggest that the peridotite itself exerted some influence on the metamorphic conditions of the surrounding rocks. If the peridotites were emplaced during $F_2$, the reigning phase of regional metamorphism $M_1$ in the (hornblende-) clinopyroxene-garnet-sodic plagioclase subfacies could have been prolonged and intensified locally to reach the clinopyroxene-garnet-sodic plagioclase subfacies by the excess heat supplied by the ultramafic body. Maaskant (1970) postulated temperatures of about 800°–900 °C during the crustal emplacement and catazonal retrogradation of the peridotites. Although minimum temperatures of about 600 °C were estimated for the regional rocks (Hubregtse, 1973), values of 800 °C seem too high for the regional metamorphism. Consequently, it appears likely that excess heat from the peridotites inhibited the formation of amphibole. However judging from the coefficients of distribution for the major elements among clinopyroxene, garnet and amphibole (Hubregtse, 1973), in both the granoblastic specimens and the other samples of metabasic rocks the ion exchanges appear to have reached a state of equilibrium under equal physical conditions. Probably the temperature was lowered again during the subsequent (hornblende-)clinopyroxene-garnet-sodic plagioclase subfacies, $M_2$, before the ion exchanges were arrested.

The *blastomylonitic type I* displays a preferred orientation of appropriate minerals such as brown to greenish brown amphiboles due to a third deformation phase, $F_3$. Granulite textures according to the definitions of Behr et al. (1971) and Mehnert (1972) are rarely seen. Specimen 129081 (Fig. 5) is representative of blastomy-

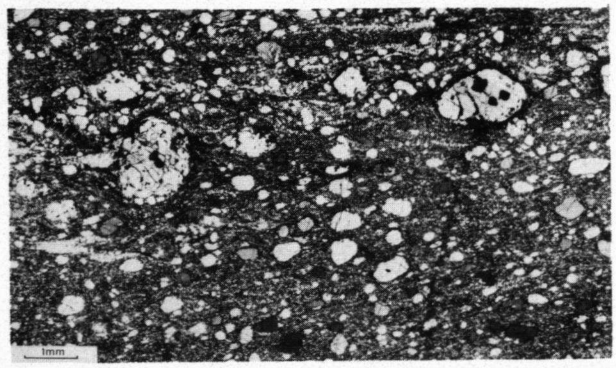

Fig. 5. Blastomylonitic hornblende pyrigarnite, containing garnet, larger grains, and clinopyroxene porphyroclasts and hornblende porphyroblasts, dark grey. Thin section RGM 129081.

lonitic type I. Its microscopical appearance may be described as follows.

Remnants of the older granofelsic type are embedded in a recrystallized matrix consisting of microcrystalline plagioclase, brownish green amphibole and sometimes small interstitial growths of scapolite in addition to the common accessory minerals. This type is best characterized by the abundant, poikiloblastic, fine- to medium-grained, brownish green to greenish brown amphiboles, which form a stable association with more or less equally sized cataclastic grains of clinopyroxene and garnet. As a result of the slight increase in $PH_2O$, the amount of amphibole increased at the expense of clinopyroxene. Normally however clinopyroxene and garnet suffered mechanical disruption without neo- or recrystallization. This is clearly visible in a mylonitized, isoclinally folded band consisting almost entirely of clinopyroxene, less garnet and minor amounts of plagioclase (Fig. 6). This feature stresses the irregular distribution of the free aqueous phase during the deformation and metamorphism of this particular rock. Here, on a scale of 2 cm, two metamorphic facies lie adjacent: an amphibole-poor assemblage belonging to the dry granulite facies and an amphibole-rich assemblage representing the water-deficient granulite facies.

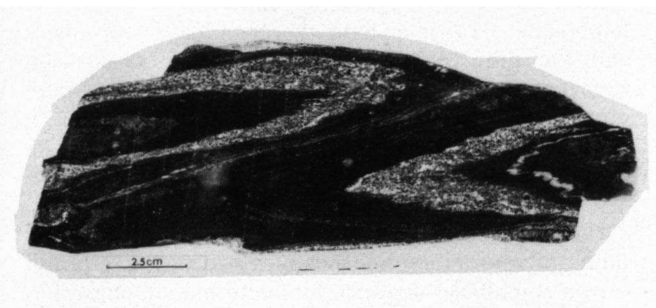

Fig. 6. Folded hornblende pyrigarnite. Amphibole-poor assemblages lie adjacent to amphibole-rich assemblages. Specimen RGM 129081.

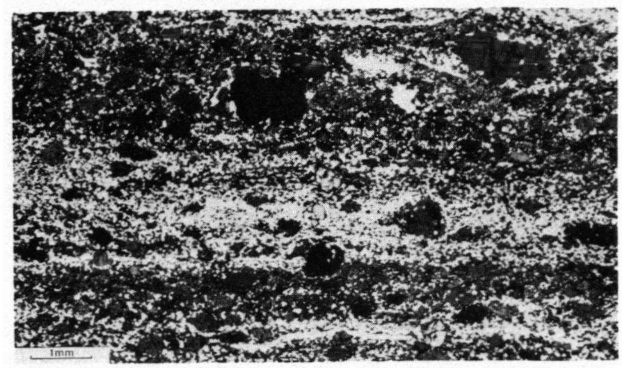

Fig. 7. Porphyroblasts of bluish green amphibole in a garnet amphibolite. Thin section RGM 129092.

II.A.1.2. *Amphibolite facies metabasic rocks.* – Essentially these rocks are granulite facies metabasites which did not survive retrogradation during $M_3$. The beginning of this phase was accompanied by tectonic phase $F_4$, which primarily caused cataclasis resulting in the blastomylonitic aspect of the rocks. The attending isoclinal folding produced preferred metablastic growth of amphibole. The paragenesis garnet-clinopyroxene-sodic plagioclase-amphibole then became unstable.

All metabasic rocks which were affected by $M_3$ are grouped together as *blastomylonitic type II.* Rocks of this type are most abundant near the Rio Ulla and in the zone W of Varelas. Representative samples are 129089, 129092, 129098 and 129100. The microscopical appearance of blastomylonitic type II (Fig. 7) shows:

A completely recrystallized microcrystalline to fine-grained matrix of amphibole, plagioclase (oligoclase-andesine) and minor amounts of quartz; this matrix envelops fine- to medium-grained metablasts of brownish to bluish green amphibole and cataclastic grains of garnet and clinopyroxene. A new generation of garnet developed locally. The titanium-bearing phase is titanite. Other accessories are apatite, zircon and opaque minerals.

This mineral association clearly points to amphibolite facies conditions. Alteration to assemblages of secondary minerals such as epidote, chlorite and blue-green to colourless amphibole was widespread. The conditions subsequent to $M_3$, during the formation of these minerals, were obviously those of the greenschist facies.

In general the whole rock composition determines whether garnet or clinopyroxene can be in stable coexistence with amphibole and plagioclase during amphibolite facies metamorphism. This is illustrated in Fig. 8a, an ACF diagram in which 20 whole rock analyses have been plotted. The plots for two garnet-amphibole-clinopyroxene triads from specimens 129084 and 129087 are also shown. The mineral plots and tie lines for other samples would fall within the area enclosed by the tie lines given in the diagram. The hornblende pyrigarnites, in which stable garnet-clinopyroxene coexistence can still be observed, cover an area

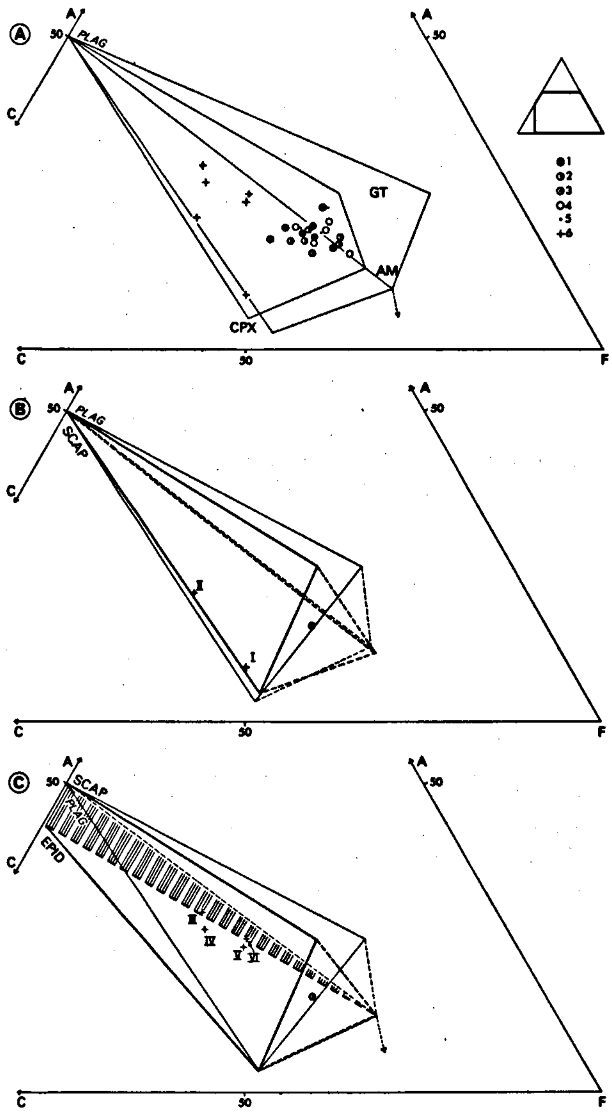


Fig. 8., Projection points of the chemical compositions of metabasic rocks and Ca-rich inclusions in ACF diagrams.
1: hornblende pyrigarnite, am ≈ gt ≈ cpx vol%; 2: hornblende pyrigarnite, am > gt ≈ cpx vol%; 3: clinopyroxene amphibolite; 4: garnet amphibolite; 5: amphibolites; 6: Ca-rich inclusions.
a: all compositions; projection for the amphibolite facies; the arrow indicates a change in the amphibole composition during retrogressive metamorphism.

b: composition of hornblende pyrigarnite (RGM 129081), inclusion I (RGM 129102) and garnet-clinopyroxene layer II (RGM 129104).

c: composition of hornblende pyrigarnite (RGM 129083), inclusion III (RGM 129101) and scapolite-bearing gneiss IV (RGM 129103); the plots of epidote-amphibolites, with amphibolite facies mineralogies, V (RGM 129108) and VI (RGM 129107) are also shown; the arrow indicates a change in the amphibole composition during retrogressive metamorphism.
For b and c: projections for the granulite facies are drawn in solid lines and in dashed lines for the hornblende- granulite facies conditions, thin lines for metabasic rocks and thick lines for Ca-rich inclusions.

on both sides of the plagioclase-amphibole join. This tie line gradually becomes significant during $M_2$ and $M_3$ as a result of the increase in $PH_2O$. The hornblende pyrigarnites are now retrograded to (garnet-bearing) clinopyroxene amphibolites and (clinopyroxene-bearing) garnet amphibolites.

The clinopyroxene amphibolites are on the cpx-side of the amphibole-plagioclase join as was expected. On the other hand, the garnet amphibolites lie on both sides of the plagioclase-amphibole join. Thus it seems that garnet is more persistent than clinopyroxene or, in other words, that the am-cpx-plag field is reduced during retrograde metamorphism. This could be due to the changing composition of amphibole. Unfortunately, there are no analyses of amphiboles from the amphibolite facies metabasites, but it seems warranted to state that the Al and alkali content of amphibole decreased simultaneously with the grade of metamorphism (Hietanen, 1972). Therefore the amphibole composition will change in the direction of the arrow, to a more aktinolitic composition, resulting in an enlargement of the am-gt-plag field.

The $\gamma$-colours of the amphiboles point to varying Al contents. Starting with the orange-brown, brown and greenish brown amphiboles of the granofelsic and the granoblastic type and blastomylonitic type I, their colour changes to light greenish brown and a green tinge in the amphibolite facies metabasites. A comparable shift in colour with decreasing grade of metamorphism was reported by Shidô & Miyashiro (1959). Shidô (1958) however states that only the alkali content and not the tetrahedral Al content of amphibole varies with the degree of metamorphism. In any event, the amphiboles in the metabasites from the Mellid area display a relationship between colour and the Al as well as the alkali content, as shown in Fig. 9. It should be noted that the amphiboles with the lowest Al and alkali contents are from specimens 129080 and 129087. There is microscopical evidence that these samples, in particular 129087, were affected more intensively by retrograde metamorphism than other specimens. Con-

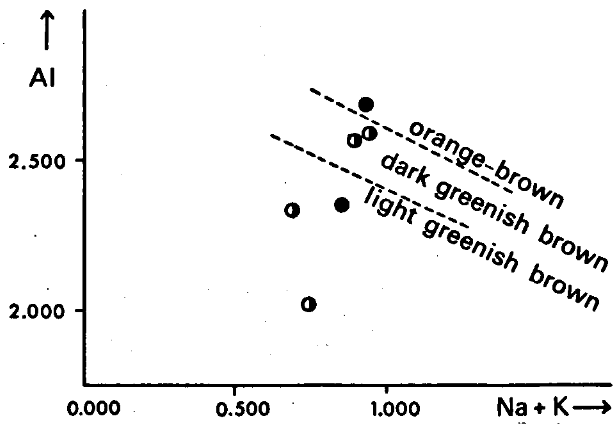


Fig. 9. $\gamma$-colours of amphiboles from metabasic rocks; Al and Na+K are the amounts of aluminum and alkali in the structural formula based on 23 oxygens; symbols as in Fig. 8.

sequently there is some confirmation of simultaneous change in metamorphic conditions and the Al and alkali contents of amphibole, accompanied by a shift in colour.

II.A.2. *Composition of the coexisting ferromagnesian minerals and the distribution of major elements over these minerals*

The chemical analyses of clinopyroxene, garnet and amphibole from 6 samples *), in which the stable coexistence of this triad can still be established, are given by Hubregtse (1973). The compositional characteristics are summarized as follows:

The clinopyroxenes are diopsides according to Poldervaart & Hess (1951). Their jadeite mol% varies between 8.0 and 15.5 and the $Mg/(Mg + Fe^T + Mn)$ ratio ranges from 0.643 to 0.750 ($Fe^T = Fe^{3+} + Fe^{2+}$). The garnets are rich in almandine. The variation in composition expressed in end-member molecules is as follows: $Gr_{16.5\text{-}33.5}Al_{44.0\text{-}60.0}Sp_{2.0\text{-}5.5}Py_{8.5\text{-}31.0}$. The amphiboles are classified as hornblendes, pargasitic hornblendes and pargasites according to Leake (1968). The $Mg/(Mg + Fe^T + Mn)$ ratio varies between 0.499 and 0.663. The amounts of Ti and Al per 23 oxygens in the structural formula range from 0.169 to 0.258 and 2.016 to 2.686, respectively.

For the distribution of $Mg\text{-}Fe^T\text{-}Mn$ and Ca-Na-K over clinopyroxene, garnet and amphibole, the following means and standard deviations of the distribution coefficients ($K_D$) were calculated (Hubregtse, 1973):

$K_{D\,mg}^{gt\text{-}cpx} = 0.190 \qquad s = 0.046$

$K_{D\,mg}^{am\text{-}cpx} = 0.622 \qquad s = 0.108$

$K_{D\,mg}^{gt\text{-}am} = 0.304 \qquad s = 0.047$

$K_{D\,ca}^{am\text{-}cpx} = 0.340 \qquad s = 0.050$

The uniform element distribution led to the conclusion that a chemical equilibrium between clinopyroxene, garnet and amphibole was closely approximated within the metabasic rocks. In this respect no differences were noted between samples of the granofelsic, granoblastic and the blastomylonitic types. The same applies for the distribution of Ca in garnet with respect to clinopyroxene and amphibole. In these calculations the structurally non-equivalent lattice sites of the monoclinic minerals were taken into consideration.

Comparison with $K_D$ data from the literature yielded evidence for (hornblende-)granulite facies conditions when the ion exchanges considered were arrested.

II.A.3. *Chemical composition of the metabasic rocks*

The chemical analyses of 20 samples are listed in Enclosure IIa. The compositions were plotted in an ACF triangle (see II.A.2). According to Kuno (1960), these

| | 1 | 2 | 3 |
|---|---|---|---|
| $SiO_2$ | 48.28 | 46.89 | 43.20–49.52 |
| $TiO_2$ | 1.73 | 1.83 | 1.20– 4.77 |
| $Al_2O_3$ | 15.43 | 14.29 | 12.63–15.87 |
| $Fe_2O_3$ | 2.64 | 3.63 | 2.54– 7.54 |
| FeO | 8.60 | 10.05 | 8.57–13.72 |
| MnO | 0.17 | 0.25 | 0.20– 0.30 |
| MgO | 8.42 | 7.20 | 5.92– 8.90 |
| CaO | 10.22 | 11.56 | 9.75–14.25 |
| $Na_2O$ | 2.27 | 2.61 | 1.50– 3.58 |
| $K_2O$ | 0.64 | 0.21 | 0.05– 0.34 |
| $P_2O_5$ | 0.23 | 0.17 | 0.01– 0.36 |
| $CO_2$ | n.d. | 0.13 | 0.09– 0.27 |
| $H_2O^{-+}$ | 0.88 | 1.09 | 0.54– 1.85 |
| Sum | 99.51 | 99.91 | |

Table III. Olivine tholeiites, chemical analyses.
1: Arithmetic mean of 182 olivine tholeiitic basalts (Manson, 1967)
2 and 3: Average and range of 20 metabasic rocks from the Mellid area
n.d.: not determined

rocks belong to the tholeiitic parentage. If compared with the average olivine tholeiitic composition according to Manson (1967), the wt% of $SiO_2$, $Al_2O_3$ and MgO are somewhat lower. On the other hand, the values for $Fe_2O_3$, FeO and CaO are higher in the metabasic rocks of the Mellid area (Table III). Some samples (129083, 129098 and 129090) display alkali olivine basaltic affinities. The normative composition of 10 samples is given in Enclosure IIb. Only three of them contain normative nepheline. A differentiation index was calculated to characterize the chemical variation within the analyzed samples. Based on the cation percentages and a total of 100.00 %, (1/3Si + K)–(Ca + Mg)-indices were derived (Enclosure IIa). This index was modified according to Nockolds & Allen (1953); oxygen was disregarded.

It may be significant that the lowest index values were calculated for the samples from the northern and northwestern exposures of the Mellid area. The samples of the southern and eastern exposures yielded the highest index values. The structure of the Mellid area is defined by mainly NW dipping foliation and fault planes. This may indicate that the metabasites with lower index values are situated on top of their counterparts with higher index values.

## II.B. Ca–RICH INCLUSIONS OF THE METABASIC ROCKS

II.B.1. *Petrography*

Inclusions and streaks were encountered in all exposures of the metabasic rocks; they consist of Ca-minerals such as garnet, clinopyroxene, amphibole, scapolite, epidote and titanite. Mineralogically they are easily distinguished from the host rock. As far as the textures are concerned however, there is a differentiation between types parallel

*) 129080, 129081, 129083, 129084, 129086 and 129087.

to that within the metabasic host rock. The origin of the Ca-rich inclusions will be discussed in II.D. Because of the presence of Ca-rich granofelsic inclusions in the metabasites deformed by tectonic phase $F_3$, the chemical heterogeneity must have already been present during $M_1$, the (hornblende-)clinopyroxene-garnet-sodic plagioclase subfacies.

The augen-like inclusions have sharp, tectonized boundaries with the host rock. The streaks are parallel to the foliation of the host rock and cause a greenish banding in the black metabasites. The following paragraphs deal with some textural types and their typical mineralogies:

II.B.1.1. *Scapolite-salite-granofels.* – The granofelsic type is well-preserved in specimen 129102, which is an inclusion about 3.5 cm in diameter embedded in a metabasite of blastomylonitic type I (sample 129081). The microscopic appearance of the inclusion (Fig. 10), a scapolite-salite granofels, may be described as follows:

Blue-green salite ($2V\gamma \approx 58°$, $\gamma \wedge [100] \approx 43°$) is obviously the main constituent. It forms fine-grained mosaics without preferred orientation. Rounded grains of titanite and interstitial growths of scapolite are also primary minerals. The grain boundaries of clinopyroxene are marked by microcrystalline opaque grains. Accessories are opaque minerals, apatite and dark green amphibole, which is intergrown with clinopyroxene. Other accessories are dark reddish brown, isotropic allanite and light green pleochroic pistacite. Allanite occurs both as inclusions in clinopyroxene and as interstitial grains. Pistacite seems to replace the interstitial allanite and is assumed to be younger than the paragenesis described so far. A striking secondary phase is an unidentified, aphanitic, deep bluish green, prismatic mineral in which Cu is the major cation (microprobe analysis). Microcrystalline aggregates of non-pleochroic turbid epidote, untwinned plagioclase and opaque minerals replace the clinopyroxene-scapolite assemblage.

II.B.1.2. *Scapolite-bearing gneisses.* – Sample 129103 displays a more varied mineralogy and should be called a garnet- clinopyroxene-scapolite-epidote gneiss (Fig. 11). Its relation to the host rock could not be established in the field. The texture however shows a vague foliation revealed in the weak planar orientation of titanite and clinopyroxene. Scapolite and epidote display some lattice orientation. It is difficult to decide which textural type should be ascribed to this rock. Under the microscope, the most abundant minerals are deep pink garnet, bluish green clinopyroxene and scapolite. Other major phases are non-pleochroic epidote, titanite and plagioclase (about 23 % An). Dark green amphibole occurs in minor amounts. All minerals form a fine-grained mosaic. There are no signs of disequilibrium between these phases. Accessory minerals are deep reddish, isotropic allanite, apatite and opaque minerals. The same alteration occurs in this rock as described for the salite-scapolite fels.


Fig. 10. Granofelsic scapolite-salite inclusion, upper half, in a blastomylonitic hornblende pyrigarnite, lower half. Thin section RGM 129102.

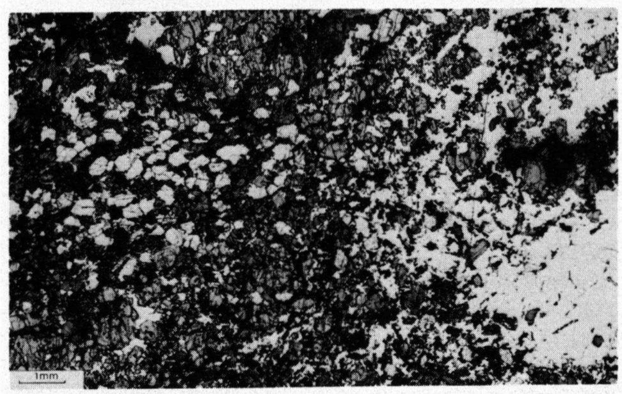


Fig. 11. Granoblastic scapolite-bearing gneiss. White grains on the left are epidotes, white mosaics on the right contain scapolite. Garnets, light grey; clinopyroxene, dark grey. Thin section RGM 129103.

A streaky band, 129101, in specimen 129083 reveals a mineral composition similar to that in sample 129103. The boundary with the host rock is blurred. The Ca-rich streak displays a vague foliation parallel to that of the enveloping metabasite. The host rock and this streaky Ca-rich band belong to blastomylonitic type I. The mineralogical appearance is as follows:

Medium- to coarse-grained porphyroblasts of deep pink garnet enclose fine-grained bluish green clinopyroxene, non-pleochroic epidote and titanite. The grains of the latter minerals are elongated parallel to the foliation. Dark greenish brown amphiboles show the same lattice orientation as those in the host rock. Highly turbid spots containing secondary minerals, probably epidote and perhaps also prehnite, are common. Zeolitic material – the optical properties vaguely suggest thompsonite – has blurred the scapolite and plagioclase grain boundaries to a large extent. This alteration into zeolite has never been established elsewhere in the granulites of the Mellid area and is assumed to be a local process that does not fit in the Precambrian regional retrogressive metamorphism.

II.B.1.3. *Scapolite-bearing mineral assemblages.* – The chemical analyses of the Ca-rich inclusions and streaks discussed are plotted in Fig. 8 and listed in Enclosure II.C. The compositions of the minerals are presented in Enclosure III. These tables include some data from the literature. Two stable assemblages are recognized in the scapolite-bearing inclusions and streaks.

1) salite-scapolite (68.5 % Me)-titanite(-magnesian hastingsite) in 129102, shown in Fig. 8b.

2) diopside-scapolite (66.4 % Me)-garnet (36.5 % Gr)-titanite-epidote (17.0 % Ps)(-ferroan pargasite) in 129101 and 129103, shown in Fig. 8c.

The mineral associations of the enclosing metabasites, also plotted in Figs. 8b and c, are clearly representative of the (hornblende-)granulite facies. From the contemporaneous development of similar textures in inclusions and host rock, it can be concluded that similar facies conditions should be reflected in their assemblages. The absence of garnet in assemblage 1) may be due to the small size of the inclusion. But specimen 129104, a more voluminous layer consisting of mylonitic material rather similar to that of assemblage 1), is scarce in garnet. This layer consists of porphyroclasts of bluish green clinopyroxene in a microcrystalline matrix; only a few garnet porphyroclasts are visible. Both compositions are shown in an ACF-triangle (Fig. 8a en b) and fall very close to the scapolite-clinopyroxene join, which explains the absence or scarcity of garnet in rocks of this composition. The plots of the garnet-bearing samples 129101 and 129103, on the other hand, lie further from the clinopyroxene-scapolite join. Because of the progressive increase in the amphibole and epidote contents with retrograde metamorphism, the following 'dry' assemblage is taken as representative for the Ca-rich rocks during $M_1$: clinopyroxene- garnet-scapolite-titanite(-allanite? ), perhaps also minor amounts of plagioclase, amphibole and epidote. In Fig. 8b the mineral assemblage recognized in 129102 and 129104 is plotted with the 'dry' paragenesis of host rock 129081. When amphibole appears, the dashed joins gradually become more significant. Fig. 8c represents the paragenesis of the epidote-bearing inclusions 129101 and 129103 together with the assemblage of host rock 129083. The dashed lines represent the conditions when amphibole is abundant, which is clearly observed in inclusion 129101.

II.B.1.4. *Hornblende-grunerite inclusions.* – Textures as mentioned for the granofelsic type persist during retrogradation. Two samples, 129105 and 129106, illustrate this phenomenon. The sharp boundary between the inclusion and the host rock is still manifest. A web of black opaque grains marks the former clinopyroxene mosaic (Fig. 12), which is completely replaced by an aggregate of two coexisting amphiboles, viz. a grunerite and a hornblende. In the core of the inclusions, a colourless prismatic amphibole with $\gamma \wedge [001] \approx 17^\circ$, $\Delta \approx 0.035$ and $2V\,\gamma \approx 90^\circ$ is by far the most abundant. The optical properties point to a member of the cummingtonite-grunerite series with about 35 mol % of the magnesian end-member. The grunerites (nomenclature after Kisch, 1969) probably grew co-axially after clinopyroxene and formed pseudomorphic aggregates, dotted with microcrystalline black opaques. Towards the rim of the inclusion, a light green amphibole (optically negative, $\gamma \wedge [001] \approx 16^\circ$ and $\Delta \approx 0.022$) becomes increasingly abundant. Its optical properties indicate normal hornblende with about 50 mol% of the magnesian end-member. The colour of the hornblendes also changes gradually to brownish green near the edge. Finally the outer rim consists of amphiboles which are optically similar to those of the metabasic host rock. Another striking feature is rutile instead of titanite, as the Ti-bearing phase among the brownish green amphiboles of the outer rim.

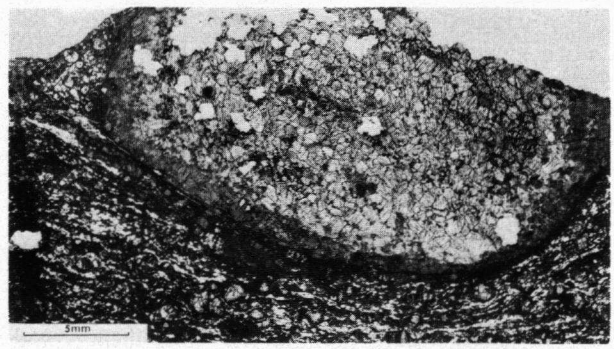

Fig. 12. Hornblende-grunerite inclusion in a blastomylonitic garnet amphibolite; hornblende mainly found near margin. Thin section RGM 129106. Compare Fig. 10.

These retrograde phenomena are explained as follows. Assuming an original clinopyroxene with a composition approximating that of the salite from inclusion 129102, it is clear that the Al content was insufficient for isochemical replacement by amphiboles with a composition comparable to those reported from the same Ca-rich inclusion (Enclosure III). Therefore Ca- and Al-poor amphibole, viz. grunerite, was formed which required a considerable amount of excess Fe + Mg ions from the environment, in this case the metabasite. An approximately equal quantity of surplus Ca ions migrated into the host rock. This model of substantial Ca $\leftrightharpoons$ Fe + Mg exchange is to be preferred over an Al exchange. The latter element does not migrate as easily as Ca and particularly Fe + Mg under the prevailing metamorphic conditions (Hubregtse, 1973). As mentioned above, a normal hornblende is abundant in the rim of the inclusion close to the metabasite. The Al deficiency should have been replenished up to and including this zone.

Recently Smithson, Fikkan & Houston (1971) reported amphibolitized margins of diopside granofelses enclosed by metasedimentary rocks, metamorphosed under uppermost amphibolite facies conditions. Although the origin of the rock is different, the amphibolitization process is rather similar: mutual Ca = Fe + Mg

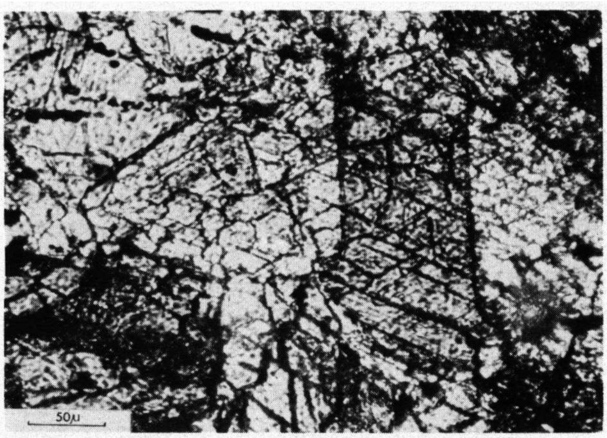

Fig. 13. Grunerite, light grey, and hornblende, dark grey, intergrowth parallel to (010). Thin section RGM 129106.

yielded an amphibolite rim with a basaltic composition.

The coexistence of grunerite and hornblende in the altered inclusions is illustrated in Fig. 13.
The photomicrograph shows distinct, rather broad intergrown lamellae of both amphiboles parallel to (010). The rough Fe/Mg ratios, based on the optical properties, make it possible to calculate the coefficient of Mg distribution between both amphiboles. Kisch & Warnaars (1969) reported several $K_{D\,mg}^{am\text{-}cum}$ values for different mineral assemblages. The value for the (plagioclase-free) inclusion of the Mellid sample is 1.9, which falls within the range 1.45–2.03 as calculated for comparable assemblages by the authors quoted.

Although gradual dispersion of the Ca-rich streaks and inclusions is to be expected as a result of their repeated retrogradation, they are still conspicuous in metabasites of blastomylonitic type II with an amphibolite facies mineralogy. A new assemblage of Ca minerals was formed during $M_3$. Representative samples were collected from the southern exposures in the region of the Rio Ulla (specimens 129107 and 129108). Their mineralogy is very simple: fine- to medium-grained non-pleochroic pistacite or clinozoisite and blue-green amphibole are the chief constituents. Minor amounts of garnet, plagioclase (about 34 % An) and titanite are also found. A foliation is well-developed.

### II.B.2. *Mineralogy of the Ca-rich inclusions*

In the following, the compositions of the minerals will be compared with those from similar assemblages reported in the literature: the clinopyroxene-garnet-plagioclase-scapolite granulites from Ruby Hill (R 165) and from Delegate (R 130), Australia (Lovering & White, 1964) and the garnet-hornblende-pyroxene-scapolite gneiss from Mampong, Ghana (von Knorring & Kennedy, 1958). We shall refer to these rocks as granulites R 165 and R 130 and the Mampong gneiss, respectively. The geological environment of granulites R 165 and R 130 clearly differs from that of the samples from the Mellid area, because of their occurrence with eclogitic and ultrabasic rocks as inclusions in breccia pipes. The Mampong gneiss, on the contrary, is found together with clinopyroxene-bearing garnetiferous hornblende gneisses. The following considerations deal with scapolite, the most conspicuous mineral in these rocks.

II.B.2.1. *Scapolite; composition and equilibrium with plagioclase.* – The scapolites of samples 129101 and 129102 are mizzonites with 68.5 % and 66.4 % Me, respectively (calculated according to Shaw, 1960); they are quite similar to those of the Mampong gneiss and granulite R 165. The scapolites of the latter rocks, however, have higher S and lower Cl contents when compared with the Mellid samples, in which the amount of $H_2O$ and $CO_2$ have not been measured (Enclosure IIIa). The deficiency of the anion group is about 0.880 and 0.734 in 129102 and 129101, respectively. Judging from the $H_2O$ content of the Mampong scapolite, about 0.500–0.650 C should be added to complete the anion group. These assumptions are rather speculative because in general the totals for the anion groups listed by Shaw (1960) are considerably less than 1.000. In any event, the presumed C contents are acceptable when compared with the Mampong specimen and a scapolite, Q 85, with 66.2 % Me reported by Haughton (1971). Lovering & White (1964) stated that the amount of S varies with the metamorphic grade. According to their view, lower metamorphic conditions have to be accepted for the Mellid scapolites than for those of the Mampong gneiss and granulite R 165 in particular. Both scapolites from the Mellid area have highly divergent Cl/S ratios. In sample 129102, the amount of S is about 4–5 times less than in sample 129101. For Cl, the situation is just the reverse. Whether this heterogeneity is original or caused by a subsequent redistribution is not clear. In this respect the different bulk compositions of the two samples should be remembered (Enclosure IIc and Fig. 8). In any case, amounts of S ranging from 10 to 30 % of the anion group should be representative for hornblende granulite subfacies conditions (Lovering & White, 1964) and the scapolite from 129101 meets this requirement quite well.

In specimen 129103 the plagioclase coexisting with the scapolite contains 23 % An. The association of Ca-rich scapolite with Ca-poor plagioclase is frequently found in rocks of various metamorphic grades. Shaw's (1960) data however do not reveal a uniform relationship. On the other hand, Hietanen (1967) and Ekström (1972) reported a regular Ca distribution between plagioclase and scapolite in amphibolite facies rocks. In addition to the higher Ca/Na ratios of scapolite, both authors observed a clear positive relationship between the varying Ca/Na ratios of the two minerals involved. Hietanen (1967) related the increasing Ca contents of plagioclase and scapolite to the increase in the metamorphic grade from the greenschist facies up to the upper amphibolite facies. On the basis of several scapolite analyses, Haughton (1971) stated that the tem-

| | An% | Me% | K' | S | $K_{D\,mg}^{gt\text{-}cpx}$ |
|---|---|---|---|---|---|
| Mellid | 23 | 66.4 | 0.15 | 0.248 | 0.17 |
| Mampong gneiss | 34.1 | 65.8 | 0.27 | 0.486 | 0.30 |
| granulite R 165 | 65 | 72 | 0.76 | 0.72 | 0.46 |

Table IV. Relationship between the compositions of scapolite and plagioclase, K' and $K^{gt\text{-}cpx}$.

$K' = \left[\frac{Ma\%}{Me\%}\right]^{scap} \cdot \left[\frac{An\%}{Ab\%}\right]^{plag}$ ;

$K_{D\,mg}^{gt\text{-}cpx} = X_{mg}^{gt}/X_{mg}^{cpx}$, $X_{mg} = Mg/Fe^T + Mn$.

Because of insufficient data from granulite R 165, this value is calculated from the analyses of another clinopyroxene- garnet-plagioclase-scapolite granulite, R 130 (Lovering & White, 1964). Haughton (1971) used the reciprocal value of K'.

perature dependence of the Ca distribution becomes umpredictable under granulite facies conditions. This conclusion, however, was based on analyses of plagioclase-scapolite pairs from various rocks of variable chemical and mineralogical bulk composition. For restricted bulk compositions, a correlation between the Ca distribution in the plagioclase-scapolite pairs and the grade of metamorphism could have become apparent. The compositions of Mellid sample 129103, the Mampong gneiss and granulite R 165 are of the basaltic parentage and display a similar mineralogy. The Me% and An% of the coexisting scapolite and plagioclase are listed in Table IV together with the amounts of S and the coefficients of Ca distribution (K') as used by Haughton (1971). Although von Knorring & Kennedy (1958) reported oligoclase in the Mampong gneiss, Lovering & White (1964) measured three plagioclase grains with an average of 34.1 % An in the same rock. The coefficients of distribution for Mg between the coexisting garnet and clinopyroxene were calculated. Banno (1970) found that $K_{D\,mg}^{gt\text{-}cpx}$ increases with temperature. According to Lovering & White (1964), the same holds true for the S content of scapolite. Here the simultaneous increase of K' is manifest. So it seems warranted to state that K' increases with the temperature in rocks of comparable chemical and mineralogical composition.

II.B.2.2. *Titanite and epidote.* – Titanite is the only stable titaniferous phase in the Ca-rich rocks, while both rutile and titanite occur in the enveloping metabasite. The same holds for the pyrigarnites of the catazonal complex at Cabo Ortegal (Vogel, 1967). In contrast, rutile is reported from the scapolite-bearing Mampong gneiss and granulite R 165. The view that rutile rather than titanite is representative for higher grades of metamorphism is generally accepted and is supported by the combined data taken from von Knorring & Kennedy (1958), Lovering & White (1964) and this paper.

Epidote is abundant in samples 129101 and 129103. There is textural evidence that it did not form part of the dry granofelsic assemblage. Together with amphibole, epidote became a stable phase during the initial phases of retrogradation ($M_2$ and $M_3$). A chemical analysis of the epidote of specimen 129101 is given in Enclosure IIIa. Expressed in end-member molecules of clinozoisite and pistacite, the composition is $Cz_{83}Ps_{17}$. This value falls in the immiscibility gap reported by Strens (1964, 1965), which extends from $Cz_{76}Ps_{24}$ to $Cz_{87}Ps_{13}$. The top of the solvus is supposed to be located exactly at $Cz_{83}Ps_{17}$, when $T \approx 550$ °C and $P \approx 5$ kbar. Hietanen (1972) reported coexisting epidote and clinozoisite at temperatures up to 600 °C. Although the pressures quoted by these authors are considerably lower than those assumed for the Mellid area, temperatures exceeding 550°–600 °C during the first phases of retrogradation agree quite well with other data. Unfortunately, the effect of pressure on the immiscibility gap is not known.

II.B.2.3. *Ferromagnesian minerals.* – The garnet compositions and the clinopyroxene and amphibole compositions are listed in Enclosures IIIb and c, respectively. Because the element ratios (Mg/Fe and others) of minerals depend on the host rock composition, the distribution of the elements among coexisting minerals is a better indication of physical conditions than a chemical peculiarity of a single mineral. The use of $K_{D\,mg}^{gt\text{-}cpx}$ was mentioned previously in II.B.2.1. The clinopyroxenes of the Mellid samples and the Mampong gneiss have a low Tschermak component substitution when compared with the clinopyroxene of granulite R 130. The view that a high Tschermak component substitution should be an indication of high P and T (White, 1969) agrees with the $K_{D\,mg}^{gt\text{-}cpx}$ values calculated. With the exception of their Fe/Mg ratios, the clinopyroxenes and the amphiboles of the Mellid samples are quite comparable in composition to those of the Mampong gneiss.

II.B.3. *Chemical composition of the Ca-rich inclusions*

The chemical analyses of 6 inclusions are given in Enclosure IIc. Two compositions (129101 and 129102) were calculated from the modal contents and the chemical analyses of the minerals. The compositions of these minerals are given in Enclosure III. In general the most striking differences with respect to the average composition of the metabasic host rock are the lower values for $SiO_2$, MgO and alkalis. Usually higher values are found for $TiO_2$, $Fe_2O_3$, FeO and, in particular, CaO. The Niggli mg values are conspicuously lower: 0.25–0.45 for the inclusions versus 0.45–0.52 for the metabasites. These chemical characteristics lead to rare normative compositions (Enclosure IIc). Only one sample has some normative hypersthene, the others contain nepheline. Titanite was calculated instead of ilmenite. Considerable amounts of wollastonite, in one specimen even calcium-orthosilicate, are present in the norm. Substantial amounts of normative anorthite and diopside further emphasize the Ca-richness of the inclusions.

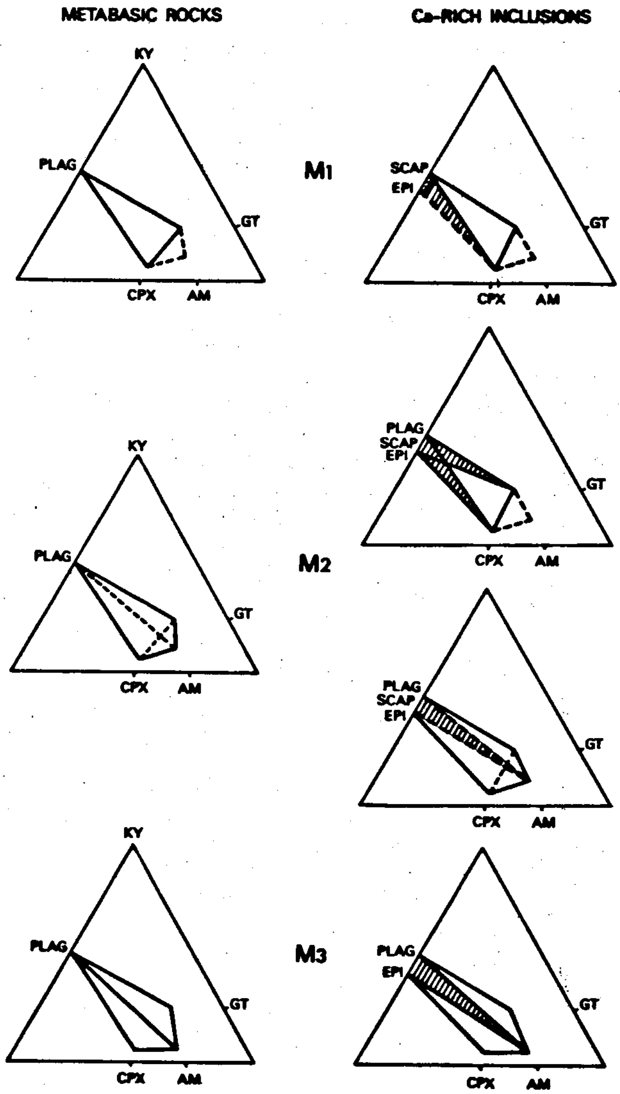


Fig. 14. ACF diagrams, a sequence of subfacies characteristic for M1, M2 and M3 as recognized in the metabasic rocks and the Ca-rich inclusions.

## II.C. COMPILATION OF THE MINERAL PARAGENESES

The mineralogy of the metabasic host rock indicates a retrogradation from the (hornblende-)clinopyroxene-garnet-sodic plagioclase subfacies ($M_1$) of the granulite facies via the hornblende-clinopyroxene-garnet-sodic plagioclase subfacies ($M_2$) to the amphibolite facies ($M_3$). A simultaneous change in the mineralogy of the Ca-rich inclusions could be established. If compared with the gneiss from Mampong which is very similar in composition and environment to the samples described in this paper, the conclusion may be drawn that equilibrium among the chief constituents was approached under somewhat lower grade metamorphic conditions within the granulite facies rocks of the Mellid area. Fig. 14 is a graphic summary of the parageneses as recognized in the metabasic rocks and their inclusions. From $M_1$ onward, hydrated minerals became increasingly important with continued retrogradation. More varied mineralogical assemblages developed in the Ca-rich inclusions and streaks. Scapolite disappeared gradually, while plagioclase became more important. In the early stages ($M_1$ and $M_2$) epidote was the predominant hydrous phase.

Kyanite is added in the left column in order to stress its stable existence in the high-grade metamorphic pelitic rocks (I.C. and II.E.).

## II.D. PROPOSED ORIGIN OF THE Ca-RICH INCLUSIONS

The foregoing sections dealt with the chemical and mineralogical heterogeneity between the host rock and its inclusions. Since the various characteristics persisted during several metamorphic and tectonic stages, it may be postulated that the inclusions were more abundant in earlier times, viz. before the (hornblende-)granulite facies metamorphism. Furthermore the chemical heterogeneity might have been even greater. The cause of this phenomenon is the question in point. If the chemical heterogeneity was established prior to the imprint of the highest degree of metamorphism, the origin may be sought in primary processes such as differentiation and consolidation of the magma, or secondary processes such as deuteric alteration and incipient metamorphism of the metabasic rocks.

A magmatic process seems unlikely for the following reasons: 1) The small amounts of Ca-rich material are regularly distributed over the metabasic rock. 2) The group of Ca-rich inclusions has a very distinct chemical composition and shows a considerable variety in itself. 3) The compositions are markedly undersaturated in silica and therefore unlikely to be magmatic products.

In our opinion post-magmatic processes such as the redistribution of chemical components during the formation of new mineral assemblages as caused by alteration or incipient metamorphism are more plausible. This idea fits in with the uniform distribution of Ca-rich material. Also the chemical variety of the group as a whole could be due to local factors.

Starting with the possibility of superficial alteration by weathering, one may distinguish between subaerial and submarine conditions. A geochemical investigation of rocks weathered under humid subtropical climatic conditions by Levi & Melfi (1972) demonstrated depletion and enrichment of elements in weathered basic rocks similar to that noted in the Ca-rich inclusions with respect to the host rock. The Al and in particular the Ca contents of the Mellid inclusions, however, are not in agreement with their data. A report on the palagonitization of submarine basalts by Moore (1966) demonstrated a gain of K, Ti and Fe and loss of Ca and Na during the replacement of glassy parts. Thus, the decrease in Ca does not fit in with the data presented in this paper.

Studies on chemical variation in lavas by Vallance (1965, 1969) demonstrated that the diversity is due to local, volcanic or postvolcanic hydrothermal alteration grading into diagenesis and burial metamorphism. Vallance suggested redistribution of basalt components due to hydrolitic reactions, especially in the glassy selvages of pillowy lavas. If pressures and temperatures are sufficiently high, a redistribution of chemical components should take place during devitrification of glass in the presence of excess water. Depending on $PCO_2$ and the pH of the water, certain elements will pass into solution from the selvages. In this way the chemical and mineralogical heterogeneity between selvages and the cores of the pillows was explained by Vallance (1965, 1969). He illustrated this difference with several examples from ancient pillow lavas. All selvages were depleted of $SiO_2$ and $Na_2O$. Furthermore there are Ca-rich and Ca-poor selvages which apparently are governed

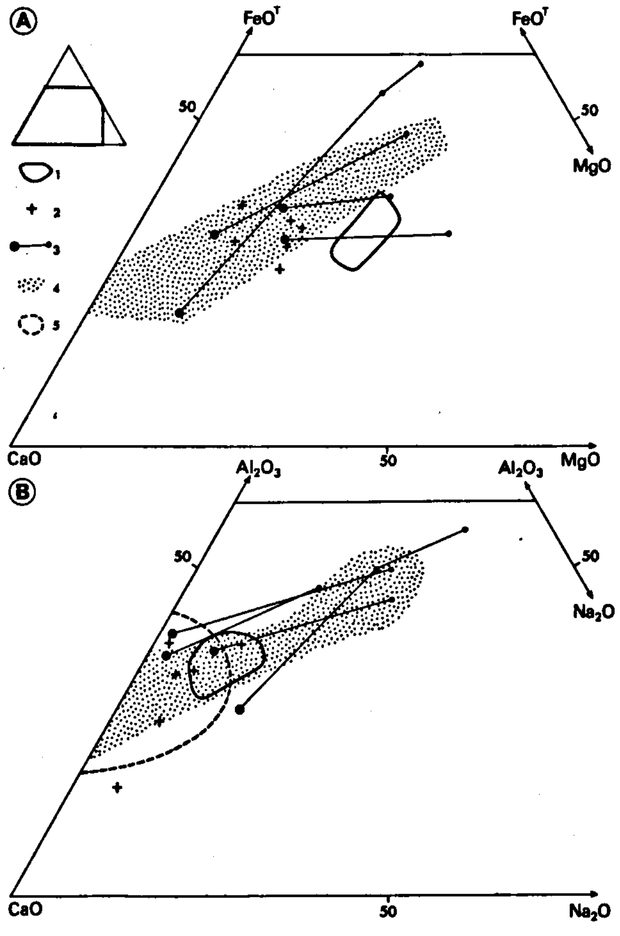


Fig. 15a. $FeO^T$-CaO-MgO diagram, $FeO^T$= FeO + $Fe_2O_3$.
15b. $Al_2O_3$-CaO-$Na_2O$ diagram; mol proportions.
1: field of the metabasic rocks from the Mellid area
2: plot of a Ca-rich inclusion from the Mellid area
3: core (small dot) and selvage (large dot) of pillow; data from Vallance (1965, 1969)
4: field of 59 plots of altered basic lavas (Smith, 1968)
5: field of pumpellyite and epidote domains, data from Jolly & Smith (1972)

by the general relationship Ca-rich core – Ca-poor selvage and Ca-poor core – Ca-rich selvage. Two types of Ca-rich selvage, which deserve our special attention, were recognized. The chemical and mineralogical characteristics are:

1) Si- and Na-poor, Ca-, $Fe^{3+}$- and Al-rich; pumpellyite, epidote and prehnite; if Ti-rich titanite is also present.
2) Si- and Na-poor, Ca-rich, substantial Fe and Mg, reduced Al; tremolitic-aktinolitic amphibole.

A review of the Ca-rich inclusions from the Mellid area (Enclosure IIc) shows that five samples meet the chemical requirements of type 1. One sample (129102) matches type 2.

To visualize this conclusion, some data taken from Vallance (1960, 1969) are plotted together with the compositions of the metabasites and their Ca-rich inclusions from the Mellid area. The Ca enrichment and Na depletion in the selvages of ancient pillows is demonstrated in $FeO^T$-MgO-CaO and $Al_2O_3$-$Na_2O$-CaO triangles, respectively. The compositional ranges cover the chemical variety within the Mellid samples very well (Fig. 15).

Smith (1968) reported on the alteration of basic lavas from New South Wales, Australia. The chemical diversity found for 59 analyses is characterized at best by an antipathetic change of Ca and ($Fe^T$, Mg, Na, K), leading to the formation of Ca-enriched domains and parts with spilitic compositions. Another detailed study on this subject was provided by Jolly & Smith (1972). They reported metamorphic differentiation of olivine tholeiitic lavas from Michigan, U.S.A. Monomineralic domains of pumpellyite and epidote were recognized. The bulk compositions of these domains, enriched in Ca en Al, closely resemble those of the Ca-rich inclusions of the Mellid area. Both Smith (1968) and Jolly & Smith (1972) presume that the departure from the original chemical and mineralogical composition was the result of adaptation to the conditions of the prehnite-pumpellyite facies of burial metamorphism. For comparison, the compositional fields of the altered basic lavas (Smith, 1968) and the pumpellyite and epidote domains (Jolly & Smith, 1972) are given in Fig. 15.

On the basis of these considerations, an early redistribution of Ca-rich phases within the metabasic rocks of the Mellid area may explain the presence and nature of the inclusions. This process could have been due to deuteric activities or low-grade metamorphism during the burial of these rocks. Whether or not the devitrification of glassy selvages facilitated the alteration processes, and what the mineralogical compositions of the inclusions were originally remain unanswered, because one can only determine the present bulk compositions.

As argued in the previous paragraph, the Ca-rich inclusions already existed before the main phase of (hornblende-)granulite facies metamorphism ($M_1$). Therefore, we conclude with Vallance (1969) that the chemical heterogeneity is due to adjustment to deuteric conditions and/or incipient metamorphism rather than high-grade metasomatic activities.

## II.E. PARAGNEISSES

These gneisses are found in association with the metabasic rocks. They are located mainly in the N–S trending zone of granulite facies rocks, E of the road Mellid-Las Cruces, north of the Rio Ulla. The gneisses are also found near the Rio Ulla. On the southern bank of the Rio Catasol is another small outcrop.

### II.E.1. *Petrography*

The rocks display grey, brown, purple-brown and black colours. A foliation is always present. Although often massive, a distinct lamination could sometimes be observed macroscopically. The gneisses are fine- to medium-grained and have planar and planolinear textures.

Some chemical analyses of the paragneisses (Enclosure IId) point to semi-pelitic and greywacke compositions.

II.E.1.1. *Granulite facies paragneisses.* – Because of their association with granulite facies basic and ultrabasic rocks, the paragneisses must have developed under quite similar physical conditions. Their original paragenesis, formed during $M_1$ – the (hornblende-)granulite facies – can only be surmised from mainly monomineralic porphyroclasts. These remnants do not provide more information about the original rock texture than that the grain size was considerably coarser. One may conclude that the paragneisses were coarse- to medium-grained rocks containing kyanite, garnet, biotite, plagioclase, orthoclase, quartz and rutile. Strictly speaking, the granulite facies paragenesis of these rocks is not as typical as that of the 'dry' granofelsic metabasites. The presence of orthoclase and kyanite porphyroclasts could indicate granulite facies conditions (Hietanen, 1967).

In the case of the paragneisses, the sequence of textures and the distinct change in mineralogical composition caused by retrogradation cannot be as easily detected as in the metabasic rocks. A certain synchronous development can however be seen in the textures of the two rock types.

The Precambrian tectonic phases $F_3$ and $F_4$ and their contemporary phases of metamorphism caused different preferred orientations of appropriate minerals such as biotite, muscovite and quartz. On this basis a number of textural types will be described. Although the pelitic counterpart of the granofelsic type of metabasic rock did not survive retrogradation, certain characteristics are still recognizable. Cataclastic grains of kyanite (up to 2 mm) contain quartz, biotite, euhedral garnet and also rutile. Larger garnets display biotite, plagioclase, quartz and rutile inclusions. Potash feldspar porphyroclasts show flame and hair microperthite; twinning is seldom observed. Plagioclase (20–35 % An) shows pericline and albite twins. Porphyroclasts of the two feldspars (Ø up to 1 cm) contain patches of quartz.

The blastomylonitic type I (related to $F_3$, see II.A) is well-developed locally. Representative samples are 129117, 129124 and 129156. The cataclastic minerals mentioned in the previous paragraph are embedded in a recrystallized microcrystalline groundmass consisting of elongated aggregates of feldspar or biotite and feldspar alternating with ribbons of discoid quartz. These laminae form clear garlands around augen-like feldspar porphyroclasts in sample 129156 (Fig. 16). Parallel to the lamination, a second generation of microcrystalline kyanite can be discerned. The younger kyanites are particularly well- developed in sample 129124. In the same hand-specimen the older generation kyanites are greatly bent and kinked (Fig. 17). A second generation of garnet is also evident locally. Rutile, zircon, allanite, apatite, opaques and, if present, muscovite are accessory minerals. Some samples of this type come from the few rocks of the Mellid area to which the name 'granulite' can justifiably be applied, according to the definitions of Behr et al. (1971) and Mehnert (1972).

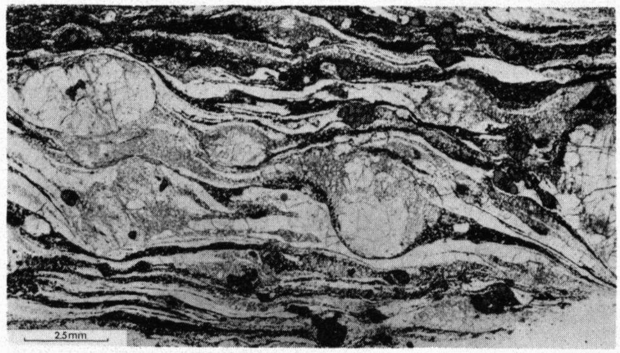


Fig. 16. Blastomylonitic kyanite-garnet-biotite gneiss. Thin section RGM 129156.

II.E.1.2. *Amphibolite facies paragneisses.* – Retrograde metamorphism during $M_3$ led to amphibolite facies mineral parageneses in the metabasic rocks. Such a mineralogical change is less clear in the paragneisses. Feldspars, micas and quartz recrystallized again and show preferred orientation due to tectonic phase $F_4$. The planolinear, sometimes linear, textures range from gneissic to blastomylonitic. Because of their intimate


Fig. 17. Kinked kyanite in a blastomylonitic kyanite-garnet-biotite gneiss. Thin section RGM 129124.

location with respect to comparable metabasites, paragneisses of this textural and mineralogical type will also be designated as blastomylonitic type II (see II.A). Paragneisses of this sort are exceedingly abundant. Muscovite becomes more important. It grew at the expense of first generation kyanite, which is reduced in grain size and shows microcrystalline to fine-grained augen. Second generation microcrystalline kyanite is rarely found in feldspar porphyroclasts; generally these kyanites have disappeared.

Tourmaline is normally absent but one sample, 129123, contains a dark green variety. It occurs not only as recrystallized small grains (ca. 0.5 mm) in the groundmass, but also in a porphyroclastic oval inclusion which consists of a tourmaline porphyroblast (Ø about 4.0 mm) surrounded by a kyanite-quartz-plagioclase mosaic.

Biotite-free varieties are rarely found. The few samples display the properties of blastomylonitic type II, i. e. muscovite in abundance, showing a preferred orientation due to $F_4$. Sample 129129 is representative. Fine-grained kyanites are included in porphyroclastic feldspar grains. Quartz, plagioclase (about 15 % An) and potash feldspar form recrystallized aggregates. The new generation of potash feldspar differs in appearance from clastic microperthite. The older potash feldspar is mainly untwinned; sometimes it shows a weak polysynthetic twinning. The recrystallized potash feldspar shows a distinct tartan twinning, yielding evidence of microcline. Increasingly distinct microline twinnig during retrogradation is rather common (Katz, 1968 and others). Well-twinned microcline is typical of amphibolite facies rocks (Heier, 1961).

The mineralogical composition of the paragneisses is somewhat variable. Thus, the amount of potash feldspar may be low in some samples and kyanite is not always present. Staurolite was never found, although the physical conditions were appropriate during the amphibolite facies retrograde metamorphism. A thorough study of the chemical analyses of the paragneisses revealed that the compositions were not suitable for staurolite formation.

II.E.2. *Chemical composition of the paragneisses and their ferromagnesian minerals*

Garnets and biotites of four specimens (129121, 129122, 129123 and 129124) were chemically analyzed. The analyses are listed in Enclosure III.

Expressed in end-member molecules, the chemical variation of the garnets can be written as $Gr_{5.8\text{-}11.9}Al_{64.7\text{-}76.5}\ Sp_{2.6\text{-}3.8}Py_{12.9\text{-}20.1}$. Almandine-rich garnets are common in metapelitic rocks. Comparable compositions have been reported by Matějovská (1970) from kyanite-bearing metapelites in the Moldanubicum, as well as others.

The analyzed biotites are all armoured inclusions in porphyroclastic garnets. This guarantees that the chemical compositions are representative for the generation of biotite formed during the (hornblende-)granulite facies metamorphism. The Mg/Mg + $Fe^T$ ratios for the biotites range from 0.464 to 0.634. They are true biotites according to Foster (1960). The $TiO_2$ wt% varies between 1.80–3.40. The amount of tetrahedral aluminum in the structural formula on the basis of 22 oxygen ranges from 2.476 to 2.646.

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| $SiO_2$ | 62.26 | 58.98–68.50 | 64.2 | 58.10 |
| $TiO_2$ | 0.99 | 0.76– 1.07 | 0.5 | 0.65 |
| $Al_2O_3$ | 18.05 | 15.95–20.72 | 14.1 | 15.40 |
| $Fe_2O_3$ | 1.55 | 0.41– 3.12 | 1.0 | 4.02 |
| FeO | 5.87 | 3.98– 7.38 | 4.2 | 2.45 |
| MnO | 0.15 | 0.10– 0.23 | 0.1 | n.d. |
| MgO | 2.51 | 1.80– 3.19 | 2.9 | 2.44 |
| CaO | 1.57 | 1.23– 2.00 | 3.5 | 3.11 |
| $Na_2O$ | 2.10 | 1.59– 2.86 | 3.4 | 1.30 |
| $K_2O$ | 2.44 | 1.94– 2.89 | 2.0 | 3.24 |
| $P_2O_5$ | 0.13 | 0.09– 0.18 | 0.1 | 0.17 |
| $CO_2$ | 0.12 | 0.03– 0.22 | 1.6 | 2.63 |
| $H_2O^{-+}$ | 1.99 | 0.79– 2.87 | 2.2 | 5.00 |
| Sum | 99.73 | | 100.0 | 100.00 |

Table V. Paragneisses, chemical compositions.
1 and 2: average and range of 7 kyanite-garnet-biotite gneisses from the Mellid area
3: greywacke, average (Pettijohn, 1957)
4: shale, average (Clarke, 1924; from Pettijohn, 1957)

The chemical analyses of 11 rock samples are listed in Enclosure IId. The average and the range of the compositions of 7 kyanite-bearing gneisses are presented in Table V. For comparison purposes the average for greywacke and shale compositions are also given. The kyanite-bearing gneisses from the Mellid area are rather Al-rich. Although somewhat richer in $SiO_2$, the compositions of the kyanite-free samples are fairly close to the average composition of the sediments.

II.E.3. *Gabbro intrusions in the paragneisses*

Minor gabbroic bodies are widespread throughout the paragneisses. The composition of the gabbros approximates that of high- alumina basalts with olivine-tholeiitic affinities. These gabbros have high-pressure mineral assemblages: garnet-clinopyroxene (up to 24 wt% jadeite)-plagioclase (25–30% An)-amphibole-rutile. They are presumed to have been intruded during (hornblende-)granulite facies conditions and then to have cooled down under the prevailing high pressure. Some samples still show ophitic textures. The petrogenesis of the gabbroic rocks is discussed in a separate paper (Hubregtse, in press). The distribution of major elements over the major constituents is also reported elsewhere (Hubregtse, 1973).

In weathered cross-sections, the gabbros are visible as globular bodies. As far as can be seen, the diameter of the bodies does not exceed 10 m. On the whole, the type of intrusion resembles that defined by Berthelsen (1970) as globulith: an intrusive body or a group of closely associated bodies of globular shape. Because of their resistance to weathering, some gabbroic rocks remain as boulders on the present erosion surface.

If the dry gabbros were emplaced at liquidus temperatures of about 1200 °C, the heat surplus and the heat liberated during crystallization could have raised the regional temperature locally, the latter being at least 600 °C (Hubregtse, 1973). If this were the case, a rise in temperature of about 200 °C would be sufficient to cause incipient melting in the paragneisses (Brown & Fyfe, 1970), a feature which has been noticed near the gabbroic rocks. The paragneisses have been reconstituted into layered migmatite-like gneisses. Quartzofeldspathic streaks enclose dark restites, consisting of garnet, biotite, rutile and amphibole. Kyanite did not survive the thermal increase. Other results of the gabbroic activity are widespread assimilation phenomena. There is a complete range from metagabbros to assimilated paragneisses to unaffected kyanite-garnet-biotite gneisses. The chemical variety within the assimilated rocks is best characterized by an antipathetic change in the Niggli k and mg values.

## CHAPTER III

## INTERMEDIATE-GRADE AND LOW-GRADE UNITS

Both units will be considered in one chapter because they consist largely of orthogneisses which probably originated from one single plutonic body. Therefore the earlier part of the history will be reviewed for both units together.

### III.A. ORTHOGNEISSES

The composition of the orthogneisses *) ranges from granitic, mainly in the low-grade unit, to granodioritic, chiefly in the intermediate-grade unit. Chemical analyses of 8 samples are listed in Enclosure IIe. The compositional diversity is clearly illustrated in Fig. 18. The larger amounts of FeO, $Fe_2O_3$ and MgO for the granodiorites is revealed in the preponderance of biotite over muscovite. Muscovite is much more abundant in the granitic orthogneisses.

A preliminary investigation of the morphology of zircons from the orthogneisses yields evidence for their igneous origin (C. E. S. Arps, pers. comm).

After intrusion into the country rock (see III.B), the orthogneisses were affected by the first phase of Precambrian metamorphism ($M_1$) resulting in a medium- to coarse-grained rock with a granofelsic, granite-like texture. Although a compositional zoning of the intrusive body gave rise to the present chemical difference between the orthogneisses of both units, equal mineral assemblages were formed. They can easily be reconstructed from cataclastic components in the subsequently deformed orthogneisses: plagioclase (mainly oligoclase), potash feldspar (with tartan twinning), garnet, reddish brown biotite, muscovite and quartz; accessories are apatite, zircon and opaque minerals. This assemblage points to amphibolite facies conditions during $M_1$ and is also encountered in the adjoining schists of the Ordenes Complex. Potash feldspar however occurs much less frequently in the latter rocks (see III.B). Pre-$F_3$ structures have not been observed in the orthogneisses.

*) These rocks are not to be confused with Upper Ordovician calcalkaline granites from other parts of Galicia (den Tex & Floor, 1967; 1972). Those series yielded Rb/Sr whole rock isochrons indicating ages between 460 and 430 m.y. (Priem et al., 1970). The Rb/Sr ratios measured in the orthogneisses in question are not suitable for accurate age determinations.

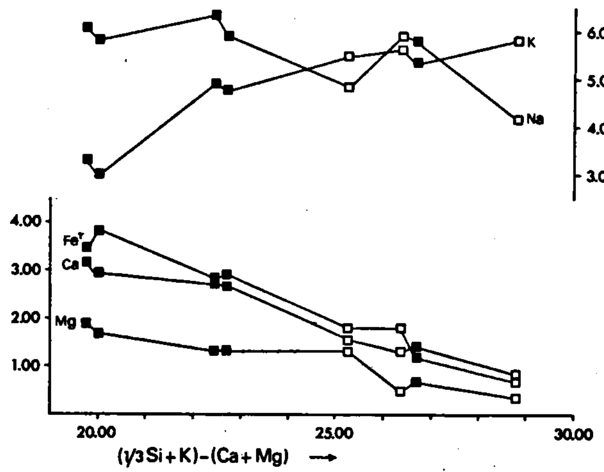


Fig. 18. Chemical variation within the orthogneisses of K, Na, Ca, Mg and $Fe^T$ (atomic proportions) with (1/3Si+K)–(Ca+Mg), index modified according to Nockolds & Allen (1953); solid squares, orthogneisses from the intermediate-grade unit; open squares, orthogneisses from the low-grade unit.

The contrasting types of deformation and metamorphism due to $F_3$ and $M_2$, respectively, mark a divergence in the evolution of the orthogneisses of the two units.

In the granodioritic orthogneisses of the intermediate-grade unit, $F_3$ and $M_2$ gave rise to coarsely foliated and highly recrystallized orthogneisses. A second generation of garnet developed at the expense of older biotite. These observations indicate amphibolite facies metamorphism.

The granitic orthogneisses of the low-grade unit show NW–SE trending lineair structures due to severe mylonitization during $F_3$. The incomplete subsequent recrystallization points to a lower grade than amphibolite facies metamorphic conditions.

At the time of $F_3$ and $M_2$, a considerable difference in tectonic level must have led to the separate textural and metamorphic evolution of the two units.

## III.B. LOW-GRADE UNIT

The low-grade unit is situated in the western part of the Mellid area (Fig. 2). It comprises the western half of the orthogneiss body and the metasedimentary rocks of the inner part of the Ordenes Complex, hereinafter called the Ordenes schists. A NE–SW fault separates this unit from the intermediate-grade unit in the east and south-east.

The Ordenes schists *) are microcrystalline to fine-grained schistose rocks. Their mineraloy points to a sedimentary origin. The main constituents are muscovite-biotite-quartz-plagioclase, and less frequently some potash feldspar. Garnet occurs in the eastern part of the Ordenes schists. Accessory minerals are tourmaline, zircon and opaque minerals. Depending on the relative amounts of constituent minerals, these rocks may be designated as schists, sometimes paragneisses and, particularly in the western part of the unit, phyllites.

The intrusive relationship between the Ordenes schists and the granitic orthogneisses is practically obliterated by the NNW–SSE fault system. Nevertheless, intrusion phenomena were found about 2 km W of Boente. There a hollow road shows a weathered section in which the regional schistosity cuts across an irregular intrusive contact. Deformed apophyses have been found in this exposure as well as further north, halfway between Boente and Golán. Xenoliths of schists (specimen 129146) also occur sporadically in the orthogneisses near the contact.

The paragenesis garnet-biotite-muscovite-plagioclase-quartz represents the highest grade of metamorphism in the Ordenes schists. From other parts of the Ordenes Complex, E of Santiago de Compostela (van Zuuren, 1969) and S of Touro (H. Koning, pers. comm.), kyanite- and staurolite-bearing paragneisses have been reported. In this part of the Ordenes Complex, however, the latter minerals have not been found. Moreover the grade of metamorphism decreases in a westerly direction. The northwesternmost part of the mapped area contains phyllites with chlorite, sericite and quartz as main constituents. Whether the boundary between the garnet- and biotite-bearing schists and the phyllites is a fault or a normal metamophic isograd could not be established in the field because of scanty exposures.

The garnet-biotite association is postdated by an isoclinal folding phase on NW–SE axes. Its NNE–SSW trending axial plane dips in a northwestern direction. In analogy with the fold directions of the high-grade unit, these structures are ascribed to $F_3$. A good example is specimen 129154.
It shows a N–S trending pre-$F_3$ schistosity plane in which garnet and biotite porphyroblasts have grown. Isoclinal, NW rending microfolds accompany the formation of a new schistosity plane also trending NNE–SSW. The contemporary phase of metamorphism ($M_2$) gave rise to retrogradation of the older assemblage through the formation of chlorite and sericite.

The granitic orthogneisses *) are devoid of isoclinal microfolds. Nevertheless, the NW trending $F_3$ microstructures are well-developed. They represent the intersection of a set of shear planes, which are recognizable microscopically as mylonitic garlands consisting of microcrystalline quartz, biotite, muscovite, feldspar and garnet. These mylonitic ribbons enclose porphyroclasts of sodic plagioclase and potash feldspar. Only muscovite displays recrystallization phenomena. Aggregates of highly undulose quartz subgrains show an originally uniform optic orientation.

In any event, the lack of recrystallization of garnet and biotite during $M_2$, and all subsequent metamorphic phases, indicates the low-grade metamorphism which characterizes this unit. The same may be concluded from the mineral assemblages of the metagabbro, emplaced in this unit prior to $M_2$ (see I.D.2 and Hubregtse, in press).

The NNE–SSW trending $F_4$ structures do not occur in the low-grade unit. This may be due to obliteration by overprinting of a NE–SW trending vertical crenulation cleavage during the Hercynian. This cleavage is well-developed in the Ordenes schists where it locally forms the axial plane cleavage of large-scale chevron folds with subhorizontal fold axes.

## III.C. INTERMEDIATE-GRADE UNIT

The simplified geological map in Fig. 2 shows the close spatial association of the intermediate-grade and the high-grade units. The former overlies the latter. Both units were brought into juxtaposition prior to $M_3$. The units together are limited on both sides by NE–SW trending faults. The western fault is exposed along the road Mellid–Las Cruces, E of Busel. Going in westerly direction, one may observe a progressive gradual reduction in the grain size of the orthogneisses in the shear zone. The eastern fault is exposed E of Barazón. Both faults have a variable (20° – 70°), but mainly steep (about 55°), dip in a northwestern direction.

Only two rock types are found in the intermediate-grade unit. The granodioritic orthogneisses are by far the most abundant. Gabbro intrusions were emplaced prior to $M_3$ (see I.D.2 and Hubregtse in press). Although they cover a vast area, three main sites were established, viz. the regions W of Mellid, S of the Rio Catasol near the road Mellid-Lalín, and between this road and the Rio Furelos in the neighbourhood of Barazón.

The coarsely foliated orthogneisses**) with mineral assemblages representative of the amphibolite facies conditions of $M_2$ were mentioned previously in III.A. In essence this textural type survived tectonic phase $F_4$ only locally. $F_4$ caused the formation of the characteris-

*) 10 representative samples are 129145 up to and including 129154.

*) Representative samples are 129130, 129131, 129132, 129133, 129136 and 129144.

**) Representative samples are 129137, 129138 and 129142.

tic augen texture. *) The resulting lineair structures trend NNE–SSW, which is also seen in the high-grade unit (see II.A.1.2). The prevailing metamorphic conditions in both units were those of the amphibolite facies $M_3$. In contrast to the low-grade unit, the $F_4$ deformation was highly penetrative in the intermediate-grade and high-grade units.

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## SAMENVATTING

Het gebied rondom Mellid in Galicië (N.W. Spanje) behoort tot het zuidoostelijke gedeelte van de buitenste zone van het Ordenes Complex, één van de omhoog geschoven polyorogene Precambrische complexen in de axiale zone van het spaanse Hercynische orogeen. De petrologie van het gebied is het onderwerp van deze studie. De mineraal-chemische aspecten van bepaalde gesteenten zijn elders beschreven (Hubregtse, 1973).

De ontwikkelingsgeschiedenis van het gebied rondom Mellid kan als volgt worden samengevat:

Na afzetting van pelieten, grauwackes e. d. en uitvloeiing van basische lavas gedurende een Precambrische eugeosynclinale fase vond een orogenese plaats, tijdens welke achtereenvolgens granitische-granodioritische gesteenten en gabbros intrudeerden. De plaatsname van peridotieten, gegenereerd uit de mantel (Maaskant, 1970), geschiedde eveneens tijdens deze orogenese. De gesteenten waren onderhevig aan drie fasen van metamorfose en vier deformatie fasen, welke veranderingen in de mineralogische samenstelling en de textuur tot gevolg hadden. Men kan in het complex van Precambrische gesteenten drie eenheden onderscheiden, die een verschillende ontwikkeling hebben doorgemaakt.

De opeenvolging van fasen van metamorfose zoals die vastgesteld is in een granuliet facies eenheid, is als volgt: de eerste Precambrische fase werd gekenmerkt door de (hoornblende-)klinopyroxeen-granaat-Na-rijke plagioklaas subfacies van de distheenhoudende granuliet facies. De tweede en de derde fase werden gekarakteriseerd door achtereenvolgens de hoornblende-klinopyroxeen-granaat-Na-rijke plagioklaas subfacies en de amfiboliet facies. De beide andere eenheden ondergingen een opeenvolging van fasen van lager gradige metamorfose.

De granuliet facies eenheid omvat metamorfe basische lavas, metapelitische gesteenten, granaathoudende metagabbros en metaperidotieten. De metapelitische gesteenten (paragenese: distheen-granaat-orthoklaas-Na-rijke plagioklaas-biotiet) en de metamorfe basische lavas (paragenese: klinopyroxeen-granaat-Na-rijke plagioklaas-amfibool) worden uitvoerig beschreven. De laatstgenoemde gesteenten bevatten Ca-rijke insluitsels, welke vaak skapoliethoudende mineraalgezelschappen bevatten. Het ontstaan van de insluitsels wordt toegeschreven aan deuterische omzetting of zeer laaggradige metamorfose in de lavas voorafgaande aan de granuliet facies metamorfose.

De beide andere eenheden bevatten metasedimenten en granitische en granodioritische gneizen.

De condities van de Hercynische metamorfose worden niet hoger geschat dan die van de lagere amfiboliet facies. Alhoewel de retrogradatie van het Precambrische complex gecontinueerd werd, vond er geen duidelijke verandering van de gesteentetextuur plaats. Tijdens het Hercynicum geschiedde de plaatsname van een ophioliet serie en voltrok zich de opschuiving van het Precambrische complex.

Elektronenmikrosonde-analysen werden gemaakt van de belangrijkste mineralen van de granuliet facies gesteenten, teneinde de verdeling van hoofdelementen over de coexisterende mineralen te bestuderen. Hiertoe werden uit de metamorfe basische lavas en de granaathoudende metagabbros, granaten, amfibolen en klinopyroxenen geanalyseerd. De verdelingscoëfficienten van Fe, Mg en Mn alsmede van Ca, Na en K over de mineralen geven, na vergelijking met literatuurgegevens, aanleiding te veronderstellen dat de temperatuur en druk condities van de Precambrische granuliet facies metamorfose minimaal 600 °C en 10 kbar geweest moeten zijn. De verschillende genetische ontwikkelingen van de granaathoudende metagabbros en de metamorfe basische lavas zijn af te leiden uit de zonering van de mineralen. De verdeling van Ca, Fe en Mg over amfibool-granaat en klinopyroxeen-granaat, de verschillende roosterposities van de monokliene mineralen daarbij in acht nemende, vertoont hetzelfde beeld.

De chemische analysen van een veertigtal gesteentemonsters en de elektronenmikrosonde-analysen van granaten en biotieten uit de metapelitische gesteenten alsmede skapolieten en een epidoot uit de Ca-rijke insluitsels, worden ter aanvulling gepresenteerd.

*) Representative samples are 129134, 129135, 129139, 129140 and 129143.

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