

THE GEOLOGY AND PETROLOGY OF THE
ARTIÈS-SIGUER-VALIRA DEL NORTE VALLEYS, ASTON-
HOSPITALET MASSIF (FRANCE-ANDORRA)

BY

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ABSTRACT

The stratigraphic sequence of the mapped area consists of Paleozoic rocks ranging from Cambro-Ordovician to Carboniferous. Part of this sequence occurs in a regional metamorphic state as schists and migmatites. A large body of a leucocratic gneiss with granites occurs in this metamorphic series, and probably represents an orthogneiss. The petrography of the non-pelitic, mainly calcareous rocks in the schists and migmatites is described. The relations between metamorphism and folding from the Serrat area are treated.

A short description of the various structures is given in the last chapter. Pegmatites occurring abundantly in the metamorphics, are dealt with.

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1. INTRODUCTION

During the summers from 1956—1960 I mapped a sector of the western part of the Aston and Hospitalet massifs, firstly for a thesis towards a M.Sc. degree and secondly as a contribution to sheet 6 of the geologic map of the Central Pyrenees. After the completion of the work for my Master's thesis, I spent two more seasons in the field to study more closely certain rock types. The present publication is the result of this investigation.

My work was supervised by Dr. H. J. Zwart from Leiden University, with whom I made several fieldtrips, often resulting in fruitful discussions. The expenses of one fieldseason were covered by a grant from the "Molengraaff Fonds" for which I am very grateful indeed.

The Aston-Hospitalet massif is a large gneiss antiform consisting of regional metamorphic rocks, as gneisses, migmatites and schists. This large structure is dissected by the important Mérens fault. The part south of the fault is the Hospitalet, and north of it the Aston massif. The metasediments are of Paleozoic age and belong mainly to the Cambro-Ordovician. During the Hercynian orogeny these sediments were strongly folded during several phases, and at the same time metamorphosed to schists and migmatites. The leucocratic gneiss in this metamorphic sequence is probably an orthogneiss.

The morphology of the area, although not studied in any detail, offers many interesting features, which certainly deserve a better study. The landscape is for a large part modeled by glaciers. Cirques, often with lakes, hanging and U-shaped valleys, roches moutonnées, moraine walls and rocks glaciers are to be seen in every part of the area. Remnants of an older, preglacial peneplain at an elevation of 1800—2200 m occur at many places in the Aston massif, mainly on the leucocratic gneisses. In the western part of my area, consisting of schists and migmatites, this surface appears to have been eroded away.

2. STRATIGRAPHY

2.1 *Cambro-Ordovician*

The mapped area is divided in two parts by the Mérens fault and as the amount of stratigraphic displacement of this fault is unknown, but probably large, the stratigraphy of the Aston massif (the uplifted block) and the Hospitalet massif (the downthrown block) is discussed separately.

2.11 *Cambro-Ordovician of the Aston massif.* In all probability the metamorphic rocks of the Aston massif belong to the Cambro-Ordovician. The rocks underwent a regional metamorphism ranging from epi-zonal in northwest and north to meso-zonal in the centre and southeast of the area.

The leucocratic gneisses are excluded from the stratigraphic considerations, because their origin and structural position are still open to some doubt, although an ortho-origin is probable. They overlie the migmatites of sedimentary origin.

The properties of the sediments can be studied best in the western extension of the Aston- and Hospitalet massifs, where the rocks are less strongly metamorphosed. This area has been mapped by Zandvliet (1960), who made a threefold division of the Cambro-Ordovician:

1. The upper part of the Ordovician consists of quartzites and microconglomerates, alternating with slates and some limestone layers. Caradocian and Ashgillian dates from similar rocks are known from the literature (cf. Zandvliet).

2. The Pilás-Estats series is a monotonous slate-sandstone sequence, which locally consists almost entirely of slates. Zandvliet mentions a Llandeilo date from outside his area. Probably this sequence represents a large part of the Lower Ordovician.

3. The Lleret - Bayau series are the oldest rocks found in Zandvliet's area and consists of limestones with one or more intercalations of black indurated slate. Zandvliet correlates this sequence tentatively with the "Série de Canaveilles", which probably has a Cambrian age (cf. Cavet, 1958), but newer data indicate a higher stratigraphic position. According to Zwart (this issue) it concerns the Ransol formation. A subdivision similar to that of Zandvliet can be applied to the metamorphosed Cambro-Ordovician of the Aston massif.

Allaart (1954) and Destombes et Raguin (1955, 1960) already worked out the stratigraphy of the northern part of the massif. The last mentioned authors found Brachiopods in a 80 m thick sequence of "calcschistes", which crop out 900 m NE of Artiès. They consider these layers to correspond to the "grauwacke à Orthis" which is known from several places elsewhere in the Pyrenees. Below the "calcschistes" some dark grey phyllites occur, underlain by a rather thick sequence of quartzites, microconglomerates and phyllites. This whole sequence is considered to belong to the upper part of the Ordovician (Caradoc-Ashgillian).

Descending in the sequence — towards the southeast — the core of the massif — the amount of quartzites and microconglomerates decreases considerably and pelitic sediments predominate. The boundary between the upper part of the Ordovician rich in quartzites and microconglomerates and the more pelitic sequence is found south of Pradières and runs eastwards north of the Pic d'Endron and south of

the Pic de Garbie. More eastwards the upper part of the Ordovician loses its characteristics and the boundary, which is nowhere sharp, becomes completely lost.

The boundary described above does not parallel the biotite isograd or other lines of equal metamorphism.

Below the boundary some quartzites and a few calcareous horizons occur in a predominantly pelitic sediment. During the Hercynian orogeny the majority of the micaschists, sillimanite-gneisses and quartzdiorites are formed from this Al-rich sediment. More to the south, near the Mérens fault, the number of lime-silicate rocks and impure limestones (cipolins) increases. Near Etang Fourcat some lime-silicate rocks occur and in the southern part of the valleys of Gnioure and Siguer several discontinuous horizons of cipolin, limesilicate rock and gneissose amphibolite are known. The occurrence of cipolins and amphibolites near Etang Blaou reaches several tens of meters in thickness. Destombes and Raguin (1960) consider these rocks to be the equivalent of the "Série de Canveilles", which occurs in the Pyrénées Orientales in a partly volcanic facies (according to Guitard et Laffite 1956). Presumably a part of the rocks near Etang Blaou, called metavolcanics by Destombes et Raguin are rather metamorphosed marls and calcareous clays. However, this does not invalidate the stratigraphic argument as limestones also belong to the normal development of the Canveilles formation. On structural grounds the mentioned region near the faille de Mérens is assumed to be one of the deepest exposed parts of the area and it might well be that Cambrian formations crop out here.

2.12 *Cambro-Ordovician of the Hospitalet massif.* — South of the Mérens fault rocks mapped as Ordovician occur in two belts.

In the region of La Massana a belt of gray slates forms the E-W trending anticlinal core between the syncline of Tor in the north and the one of Tirvia-Llavorsi in the south. The sequence consists of gray, brown-gray or green-gray slates with a very low percentage of sandstone or sandy slate. Some sandstone-slate alternations are known along the private road to Farga, opposite Ordino.

Where the contact is exposed the Ordovician slates are bordered by the Silurian, except west of Escas where the slates are in contact with the Devonian. The contacts are sharp and steep. A small syncline of Silurian is found near Bony de las Neres, east of Ordino, but it does not reach the valley of Ordino.

The prolongation of this Ordovician belt to the west is also very poor in sandstone (Zandvliet 1960). This author includes it in his "Pilás-Estats series". This would mean that the quartzitic-conglomeratic upper part of the Ordovician is missing in this belt or that the contacts with the Silurian are tectonic ones.

A second, much broader, belt of Ordovician is situated between the Mérens fault in the north and the Tor syncline in the south. Silurian rocks which occur near Llorts in an abnormal structure probably have been pushed partly over the Ordovician towards the north. This caused a disturbing of the anticlinal character of the Ordovician belt. The Ordovician consists of slates alternating with abundant quartzites, some microconglomerates and a few limestone and black slate horizons. Thin-bedded quartzite-slate alternations are widespread, especially in the western part of the Llorts valley.

Large parts of this Ordovician anticline have been metamorphosed and deformed in several phases. Various aluminium-silicate minerals developed in the pelitic sediments.

As a whole the sedimentary sequence has the same appearance as the upper

part of the Ordovician in the Aston massif. The position below the Silurian, which, apart from the Llorts area, seems to overlie the Ordovician conformably, supports the assumption that the rocks we are dealing with, belong to the upper part of the Ordovician.

In the fault zone of the Mérens fault, at the Port de Banyells an alternation of black slate and light grey limestone layers is present (fig. 1). Zwart (1965) correlates this sequence with the Ransol formation. The black slates from the zone can be followed westwards as far as the Spanish frontier. They are folded in between the quartzites and gray slates of the upper part of the Ordovician.

It seems not very probable that in the northwestern part of Andorra any Cambrian crops out on the southern, downthrown, side of the Mérens fault.

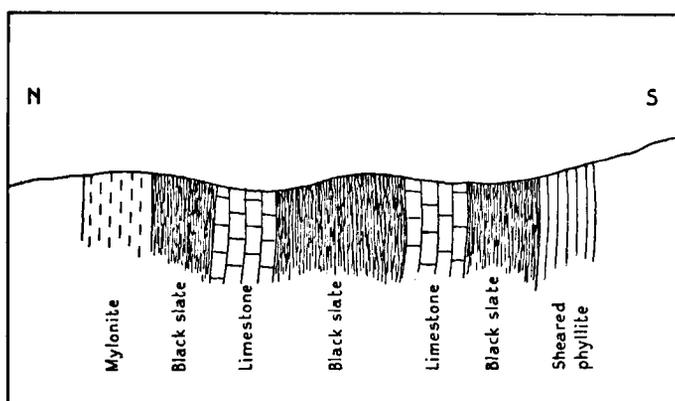


Fig. 1 Section south of Mérens fault near Port de Banyells.

2.2 Silurian

Black slates mapped as Silurian flank the Tor syncline which has a Devonian core, and the Llavorsi syncline which has a Carboniferous core and is situated more to the south. In the northern flank of the Tor syncline the Silurian underwent a tremendous tectonic thickening in the area of Llorts, where the rocks are also considerably metamorphosed.

The contact of these rocks with the Devonian is rather complicated. The boundary is steeply folded and a repetition occurs 1 km north of the Pic de Casamanya and about 1 km north of the Pic de Cap, where perhaps a very steep anticline of Ordovician is folded in between the Silurian. More to the west, on the Port Nègre, the contact of the Silurian with the Devonian is a normal stratigraphic one. The beds trend east-west with a steep dip to the north. The Silurian black slates grade via dark grey and grey limestones into the light coloured limestones mapped as Devonian. Several thick layers of the grey limestone appeared to be fossiliferous (see below).

The Silurian in the southern flank of the Tor syncline is thickened considerably southeast of Ordino. It is missing west of La Massana and probably also on the poorly exposed Col d'Ordino. In the northern flank of the Llavorsi syncline the Silurian

forms a continuous strip from Os de Civis to the Coll de Vexalis, running south of Sispony and Anyos.

Throughout a large part of the Pyrenees the Silurian is developed as a black shale with a few intercalations of dark grey limestone in the upper part. In the area under consideration the interval consists mainly of black slates and shales, with sericite and andalusite bearing black phyllites in the metamorphosed areas. A limestone intercalation was found in the valley of Llorts, but its stratigraphic position remains obscure owing to the extreme folding in that area. At several places (e.g. Port Nègre and east of Ordino) the top of the black sequence is formed by dark grey limestones.

The black colour of the slates is due to a high content of carbon and a high iron sulfide content. The aluminium content is also high and varies between 18 and 34 %; free sulphur is found occasionally.

From many places in the Pyrenees the black shales yielded abundant graptolites. Llandovery, Wenlock and Ludlow dates are known from the literature. That not a single fossil is found in the black slates of northwestern Andorra is ascribed to the intense cleavage and microfolding which disturbed all sedimentary features. Consequently it is not known whether the Silurian is complete or not, locally it is tectonically thinned or thickened but the average thickness might be even less than 100 m. Fossils have been found by J. F. Lapré and the author in the grey limestones of the Port Nègre and below that locality. Crinoid stems and Orthocerates are numerous, but besides, two crinoids with a fairly complete cup were found. Dr. A. Breimer identified the crinoids as *Sciphocrinus* sp. and thought them to be uppermost Silurian. This means that the dark limestones which occur at several places between the dark shales and the light coloured limestones belong to the Silurian system. Zandvliet (1960) wrote that also part of the light limestones mapped as Devonian might have an Upper Ludlow or Downtonian age and that the Silurian of the Central Pyrenees is merely a "rockunit", which covers the largest part of the Silurian period.

2.3 Devonian

Two east-west running belts of Devonian limestone are known in the area.

The largest belt is cut by the road from Ordino to la Cortinada and forms the core of the Tor syncline. More to the south the second strip is situated just south of Sispony and is part of the northern flank of the Llavorsi syncline. The southern strip is bordered by the Carboniferous slates in the south and by the Silurian black slates in the north. The Devonian of the Tor syncline is bordered on both sides by the Silurian as described above.

Massive limestones are the predominant rocktype of the Devonian; calcareous slates are less important while non-calcareous slates are rare. Most of the limestones are grey or lightgrey, yellowish or brownish weathering and massively bedded. Recrystallization is often observed in places where the rock is cleaved. The slates have always a distinct cleavage and have greenish or reddish weathering colours. Green marbles have been found occasionally. Nodular limestones which are considered to be typical for the upper part of the Devonian have been observed in the region above Ordino and north of the Col d'Ordino.

Apart from the Crinoids described above, very few fossil remains have been found. Some crinoid stems occur 2 km NE of la Cortinada and E of Ordino. The assumed Devonian age is based on regional lithologic correlation and the position

above the Silurian. From the Port Nègre section one would conclude that the Devonian rests conformably on the Silurian and that no hiatus is present. More to the east the contact might be a tectonic one. Whether the top of the Devonian is present, is unknown, but there seems to occur some Upper Devonian, as suggested by the presence of nodular limestone and variegated slates.

The thin Devonian of Sispony is certainly incomplete, a part of it is cut off by faulting.

2.4 Carboniferous

Besides the Quaternary fluvial and glacial deposits the youngest rocks found in the area belong to the Carboniferous. They are found in a tight syncline with a N to NW dipping axial plane and an overturned northern flank.

On the Spanish-Andorran border near Os de Civis the Carboniferous underlies the limestones of the Devonian. From there it runs eastward just N of the town of Andorra where the slates are metamorphosed to hornfels.

In the area west of the Pic d'Os the intrusive granodiorite partly takes the place of the Carboniferous. More to the west the normal continuation is found in the Llavorsi syncline. In the east the sequence is cut off by the intrusive granodiorite and by the Devonian which border the granodiorite near Escaldes. The southern boundary is formed by the granodiorite, while in the north it is bounded by the Devonian except in a part of the valley of Sispony, where the Devonian is missing and the Silurian borders the Carboniferous.

The rocks mapped as Carboniferous consist of bluish gray slates which are often somewhat sandy. A strong cleavage is always present and often some white mica. Locally quartz lenses and veins are numerous. If no muscovite is present the distinction between Carboniferous and Ordovician is difficult. However, in the area under consideration the Carboniferous has rarely more than one lineation (which is normally E-W) whereas the Ordovician does have several lineations. Also the Carboniferous is much less quartzitic.

No fossils have been found, but the Carboniferous is the normal prolongation of the Carboniferous of the Llavorsi syncline, which is mapped by Zandvliet (1960). Zandvliet has no fossil dates either and his age assignment is based on lithological and structural correlation. Whether part of the Lower Carboniferous is missing is not known, but a Lower Viséan chert horizon which is known from several places in the Pyrenees has not been found in Andorra.

3. REGIONAL METAMORPHIC CAMBRO-ORDOVICIAN: PHYLLITES, MICASCHISTS AND QUARTZDIORITES

Phyllites, micaschists, biotite gneisses, migmatites and quartzdiorites together constitute the regional metamorphic facies of the Cambro-Ordovician. The boundary with the leucocratic gneisses is generally sharp. The limesilicate rocks, cipolins and massive quartzites occurring in these metamorphics reacted in a different way upon metamorphism and deformation than the pelitic sediment and they will be reviewed in the next chapter.

The following zones, according to structure and degree of metamorphism of the rocks can be distinguished:

1. *sericite-phyllites*, often with chlorite.
2. *biotite-muscovite-schists*.
3. *andalusite-biotite-schists*, locally with staurolite rarely with cordierite and in the deeper part locally with sillimanite.
4. *feldspar bearing biotite schists*, often with a little andalusite or fibrolite.
5. a. *biotite-gneisses*.
b. *migmatites* (sillimanite-biotite gneisses).
6. *biotite-quartzdiorites*

3.1 *Areal distribution*

The phyllites occur in the northern part of the Siguer and Izourt valleys. The northern boundary is formed by the Silurian black slates in the eastern and the granodiorite in the western part of the area which has been mapped by Allaart (1954). Towards the south the phyllites grade into micaschists; the boundary indicated on the map — the biotite isograd — represents the line below which biotite is visible in the field. East of the Siguer valley this line runs E-W and is rather straight, which indicates a steep dip of the southern phyllite boundary. Towards the west the boundary turns to the south, and north of Pradières its dip apparently decreases to about 45°. Some tens to some hundreds of meters below the biotite isograd the first andalusite appears. In the Izourt valley the biotite zone is a narrow strip (see map), but in the Siguer valley it is broader because the andalusite zone is poorly developed. Below the andalusite line andalusite-micaschists occur in the whole area drained by the Izourt river and in the region around the Pic d'Albela. In the Siguer valley only locally some andalusite occurs. Fibrolite bearing micaschists are widespread in the higher metamorphic part of the andalusite zone. Their occurrence is only of local importance and mostly can be checked only microscopically.

Feldspathized micaschists are the normal link between the micaschists without feldspar and the migmatites in the Siguer valley and in the higher part of the Gnioure valley. In the Izourt valley they occur especially in the eastern part of the lake near the Pic de l'Aspre and more southward. Biotite gneisses are developed in the Gnioure valley south of the lake.

The migmatites, which often contain sillimanite are found in a long belt running eastward from the mentioned gneisses. Another migmatite area is situated NE of Et. de Gnioure. The connection between the two occurrences is formed by feldspathized micaschists west of the Et. de Gnioure. Furthermore they occur in the eastern slope of the small enclosed valley east of the Pic d'Endron.

3.2 *Mineral content and metamorphism*

The zones as distinguished above are based on the mineralogical and structural habits of the metamorphosed pelitic sediment.

3.21 *Phyllites*. — The phyllites are characterized by their distinct cleavage (or schistosity) and their silvery lustre which is caused by sericite on the cleavage plane. Lineations and microfolds are often clearly visible on those planes. The bedding is obliterated except when quartzitic layers are present.

In thin section it can be seen that the quartz grains are elongated and the material is mostly recrystallized. Locally chlorite is present in cross-cutting flakes and is apparently later than the schistosity.

Towards the micaschists the rocks become harder, the grainsize increases and already before reaching the biotite isograd as mapped in the field, microscopic small biotite is present. In the Izourt valley the biotite near the phyllite zone is post-kinematic. Synkinematic biotite appears only some tens of meters lower.

3.22 *Micaschists*. — The micaschists in the Siguer valley have a mainly synkinematic character with biotite and elongated quartz crystals forming the schistosity. Cross-cutting biotite is rare, only muscovite, where present, is frequently post-kinematic. These micaschists are homogeneous rocks, with quartz regularly distributed between the biotite foliae. Only when the sediment was quartzitic, banded schists may result. Mostly muscovite is present and sometimes chlorite, but neither in important quantities; locally garnet occurs.

Microscopically some epidote may be found; zircon, tourmaline, apatite, ore and occasionally sphene are accessory minerals. Chlorite, sericite and sagenite are common alteration products.

3.23 *Andalusite-micaschists*. — Andalusite-micaschists occur in several varieties in the Izourt valley. Just below the andalusite line the andalusite crystals are small and few in number and they hardly change the rock texture. Towards the south their number and their dimensions increase. Long crystals and crystal aggregates develop, the stalks may attain a length of 40 cm near Et. Fourcat and elsewhere. The crystals have a preferred orientation with their long axis in the schistosity plane, however, in this plane they are unoriented. Beautiful examples can be seen on the east side of the trail to Et. Izourt (fig. 2).



Fig. 2 Unoriented andalusite crystals in micaschist, Artiès valley.

Under the microscope it is evident that the andalusite started to grow after the first deformation was over. Frequently the crystals are folded and the andalusite became cracked or bent with an undulatory extinction (fig. 3). True synkinematic

andalusite with S-shaped si has not been found. Crystals extending their growth into the third folding phase have been observed. The schists have the same constituents as in zone 2, but chlorite is only rarely present. Instead staurolite is locally important, e.g. E of Et. Izourt and SE and E of Et. Fourcat. Mostly it was formed after the main deformation or during the third phase, contemporaneous with the andalusite. The staurolite is altered to muscovite to a varying extent. Cordierite is sometimes present too, but often it is completely sericitized and the range in time of the crystallization is not exactly known. A small amount of plagioclase (sodic oligoclase) is often present, but not important. Potassium feldspar is found only in schists in the neighbourhood of pegmatites, where probably supply of material took place. Excluding these occurrences near pegmatites, there is no evidence for any metasomatic introduction in the micaschists. It is believed that they originated from the pelitic Cambro-Ordovician sediments by an isochemical (except for water) metamorphism of a mesozonal grade.

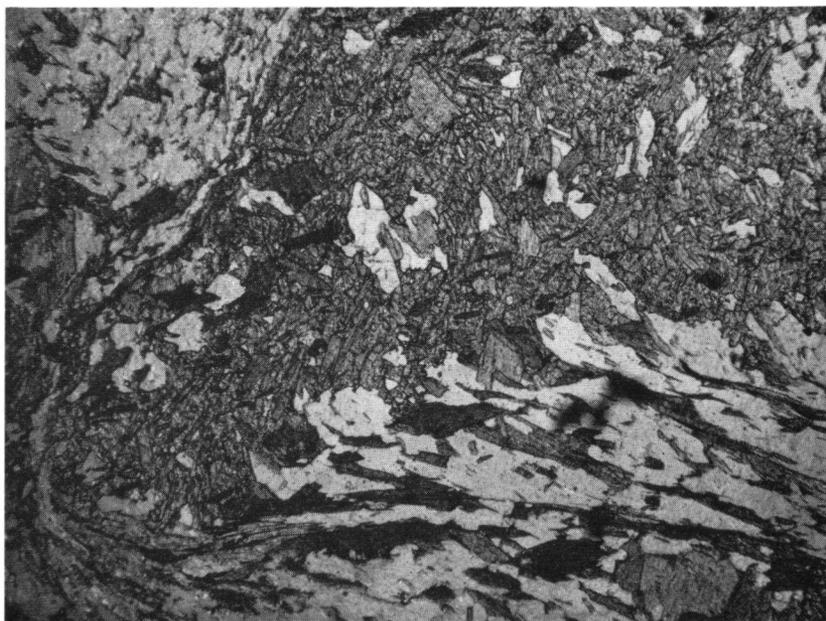


Fig. 3 Folded andalusite crystal; Artiès valley.

3.24 Feldspar bearing micaschists. — Schists which contain macroscopically visible feldspar are included in zone 4. This zone is difficult to map especially in quartz-rich schists and in areas rich in pegmatites, which may promote feldspathization of the schists. Generally the grainsize of the feldspar and the quartz is 1/4 — 2 mm.

Synkinematic feldspar (augen-shaped) is found south of Gestiès and in other places in the Siguer region. In other parts of the area the crystals are generally equidimensional and probably post-kinematic in age. Large feldspar porphyroblasts on the other hand, are relatively rare in the schists.

In the deeper part of zone 3 and in zone 4 scattered occurrences of sillimanite bearing schists are found. It is developed as fibrolite pseudomorphic after biotite. It occurs in knots always with some needles normal to the schistosity indicating the postkinematic character of the mineral (fig. 4).

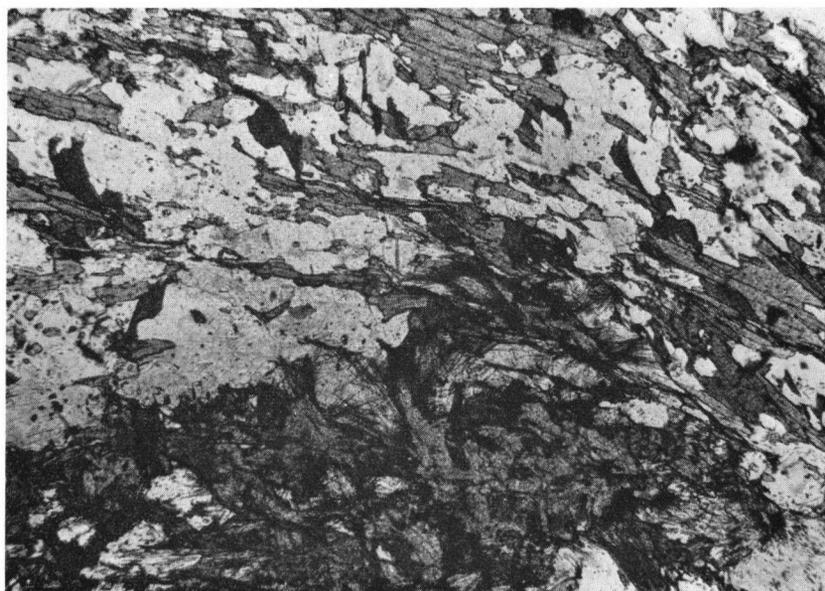


Fig. 4 Fibrolite needles in quartz; feldspar bearing micaschist.

The feldspar bearing biotite-schists grade into biotite-gneisses with an increasing amount of feldspar and with increasing grainsize of this mineral. When the feldspar and quartz attain dimensions above 2 or 3 mm the rocks are gneisses rather than schists. Two types of these gneisses can be distinguished: rather homogeneous sometimes slightly banded biotite-gneisses and gneisses with a migmatitic structure often containing sillimanite.

3.25a. *Biotite-gneisses*. — The transition from feldspathized micaschist into biotite gneiss can be seen when going from the Pic de l'Aspre to the NE or on the east slope of the Gnioure valley north of Et. de Lascours. The boundary is an arbitrary one.

The biotite-gneisses have a rather regular schistosity without folds. Augen-shaped feldspar occurs often and the biotite is always oriented so as to form the schistosity. The quartz is generally not elongated. The crystal size varies and reaches 3—4 cm in the feldspar augen.

Microscopy: Potassium feldspar (microcline) is mostly present and sometimes makes up as much as 40 % of the rock. Quartz always amounts to 20 % or more, plagioclase and biotite are also important constituents. Some muscovite is often present, sometimes sillimanite, rarely garnet or epidote. Accessories are: zircon, apatite, ore and sometimes tourmaline, while chlorite, sericite, kaolin and rutile are alteration products.

After the formation of feldspar several events took place in the rock, e.g. the recrystallization of the quartz and part of the biotite, and some alterations.

The co-occurrence of potassium feldspar and sillimanite is explained by assuming that they are not of the same age, as is suggested by the following alterations which have been observed:

- a. microcline → myrmekite (quartz + plagioclase)
- b. biotite → sillimanite (fibrolite) which preserves the structural characteristics of the biotite and inherited the zircon inclusions
- c. microcline → muscovite; along the fissures fine muscovite and sericite developed.

It is not certain if these alterations were contemporaneous but they have been observed together in the same thin section. In the same gneisses and also in other rock types it has been observed that:

- d. sillimanite → muscovite
- e. biotite → muscovite

Probably the microcline belongs to the first generation of minerals, sillimanite to the second, while the last generation always is formed by muscovite, sometimes with chlorite or biotite. Concerning the meaning of myrmekite opinions differ. It is widely accepted that the potassium feldspar is the oldest and is replaced by plagioclase and quartz. In this case plagioclase and vermicular quartz are contemporaneous. Drescher-Kaden (1948) assumed that the plagioclase is replaced and thus older than the potassium feldspar and vermicular quartz. In the biotite-gneisses it is sometimes visible that the potassium feldspar is replaced (fig. 5), the same seems to be the case in fig. 6 lower right. In the centre of this picture an idioblastic plagioclase crystal bears the vermicular quartz and seems to be attacked. With some degree of certainty it can be said that in the present rocks mostly the potassium feldspar was replaced, but in other cases this is not at all clear.

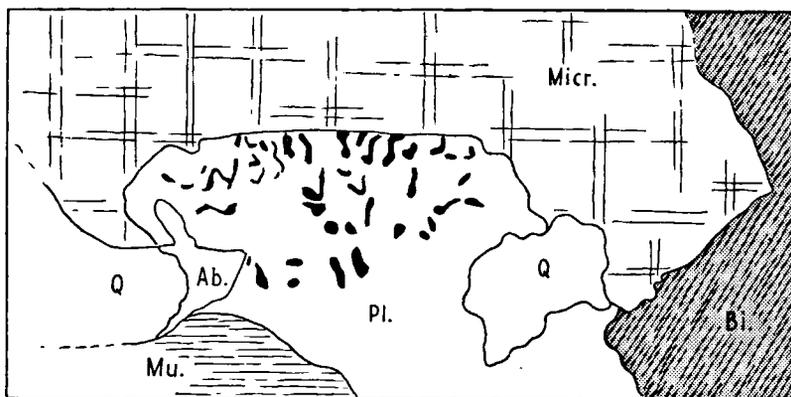


Fig. 5 Myrmekite replaces microcline; Pl = plagioclase, Q = quartz, Ab = albite, Mu = muscovite, Bi = biotite, Micr. = microcline

Roques (1955) explains how it is chemically possible that with a Si supply first myrmekite "plages" are formed, which attack the plagioclase and liberate Na and Al, and subsequently myrmekite "bourgeons" at the expense of the potassium feldspar. This second reaction liberates K and Al. It is probable that this Al contributes to the formation of fibrolite and both elements might stimulate the formation of muscovite. The more the above reactions proceed, the more the rock approaches the composition of the migmatites which are richer in plagioclase and sillimanite and contain little or no potassium feldspar.

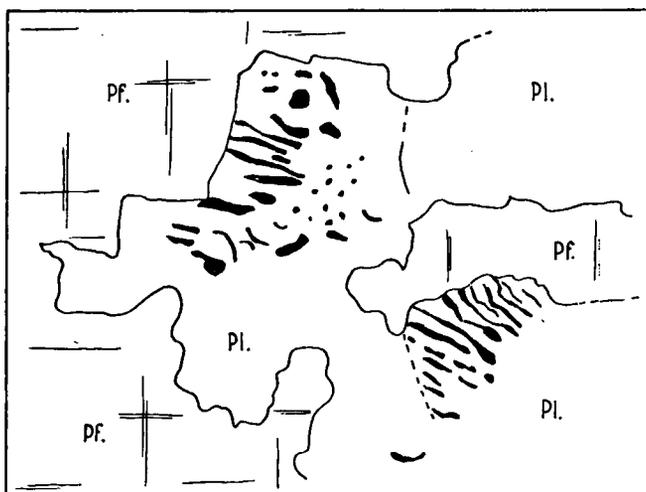


Fig. 6 Same as fig. 5. Pf. = potassium feldspar.

3.25b. *The sillimanite biotite gneisses (migmatites)*. — The boundary with the biotite gneisses (zone 5a) and the micaschists is nowhere sharp, but completely arbitrary and it was placed slightly different by Destombes et Raguin (1955). The sharp boundary with the leucocratic gneisses is discussed in chapter 6. Two components are clearly visible in the migmatites a schistose one, mainly consisting of biotite with some sillimanite, and a granitic one, composed of quartz and plagioclase with often some muscovite. These light minerals build aggregates which are lying in the schistosity, however without having a typical augen-shape and without giving the rock a lineation. The aggregates reach a width of several cm, but towards the micaschists the rock is finer grained. The schistosity is somewhat undulating.

Pegmatites occur frequently and locally muscovite concentrations have been found.

In the field it appears as if the transition from feldspar bearing micaschist to migmatite can follow two ways. First continuing feldspathization of the schist combined with metamorphic differentiation may build a migmatitic rock. Secondly small concordant pegmatite veins and streaks are locally so abundant that a migmatitic rock results (fig. 7).

Microscopy. The main constituents of the migmatites are: plagioclase, quartz and biotite. Some muscovite is always present, often also sillimanite which is concentrated in the schistose parts of the rock and mostly developed as fibrolite. Less frequent are the following minerals: potassium feldspar, albite, cordierite, epidote and garnet. The plagioclase is mostly oligoclase (up to 30 % An) except in the more calcareous rocks. The crystals are untwinned or polysynthetically twinned.

The micaceous component of the rock is strongly schistose but not linear. The quartz-feldspar aggregates have not a distinct augen shape, nor do the crystals show strong internal deformation. On the other hand the aggregates do not cut through the schistosity nor largely disturb it. Therefore it is assumed that the quartz and feldspar crystallized in late-kinematic time. Fibrolite was formed later and exhibits post kinematic relations, but mostly it is pseudomorphous after biotite. Sometimes it is clear that fibrolite was formed during the third folding phase. Locally some



Fig. 7 Concordant pegmatite veins in folded micaschist. (1x)

cross cutting biotite has been observed in the field. The muscovite is still younger and replaces sillimanite and biotite.

The facts that the migmatites are situated below the highest grade micaschists and that the transition from the schists to the migmatites is a gradual one, led to the conclusion that the migmatites are of sedimentary origin and developed from the same sediment as the micaschists. This conclusion is supported by the presence of well-rounded zircon grains in both rocks, which suggests a sedimentary origin (Verspyck 1961). Furthermore the presence of Al-silicates in micaschists (andalusite, staurolite, cordierite) and migmatites (sillimanite) may in both cases indicate an Al-rich (probably pelitic) sediment as the original material. Zwart (1959), who founded his conclusions partly on chemical analyses, presumed that the formation of the necessary quartz and feldspar is the result of metamorphic differentiation combined with the influx of some Si and Na. A small addition of these elements would be sufficient to make the transformation of a Cambro-Ordovician slate into a migmatite chemically possible.

3.26 Quartzdiorites. — The quartzdiorite occurrences are all in the migmatite area and vary in size from 1 dm to hundreds of meters of rather homogeneous rock with a granitic texture. The boundary with the migmatites, which enclose the quartzdiorites entirely, is vague and an extensive zone of nebulitic rocks is present.

The mineral composition is about the same as that of the migmatites. Only cordierite is probably more frequent but it is often difficult to recognize in the field. Fibrolite is somewhat less frequent. This mineral is often clearly pseudomorphic after biotite, but besides, real sillimanite crystals (0,3 mm or smaller) are often present; they might be older than the fibrolite. Muscovite occurs exclusively as a late alteration of other minerals. Quartz and feldspar occur in grains of a size of 0,5—5 mm. The latter mineral is mostly polysynthetically twinned; combined

Albite-Carlsbad twins occur also. If present, the potassium feldspar is somewhat coarser grained. Myrmekite is found occasionally and seems to replace the potassium feldspar. In a few cases it could be observed that the occurrence of this mineral is related to the proximity of pegmatites, which are much less frequent than in the migmatites and the nebulitic rocks. It happens frequently that in rocks which are granitoid in hand specimen locally schistose pieces or oriented biotite clusters can be seen on outcrop scale. Often also schist or quartzite patches or bands are present in the nebulite.

In the quartzdiorites the recrystallization was strong and it obliterated all traces of the main folding. The bent biotite and undulatory quartz found occasionally are ascribed to later deformations, which appeared to be often of very limited extent and locally are associated with mylonitic zones.

Rheomorphism (mobilisation), like it is known from the nearby Trois Seigneurs massif (Allaart 1958), has occurred only in one restricted area, viz. in the bedding of the Gnioure stream between 1100 and 1300 m altitude. In this area inclusions of quartzitic rock occur in nebulitic quartzdiorites. The weakly or non folded pieces have clearly rotated or moved apart by plastic flow of the rock. Probably the introduction of Si and Na necessary for the transformation of micaschists into migmatites and quartzdiorites was effectuated by aqueous solutions or gases. This H₂O richness may at the same time have promoted the mobilisation considerably (Zwart 1960).

The contact relationships and the chemical and mineralogical similarity suggest that the quartzdiorites were formed from the migmatites by post-kinematic recrystallization. The vague contacts with the surrounding rocks show that the mobilisation of the quartzdiorites was only of local importance and that the rock reached only the phase of the "autochthonous granite" in Read's granite series (Read 1957).

Only one apparently intrusive quartzdiorite dyke has been found some hundreds of meters SW of the locality of the mobilisation. Another somewhat similar dyke occurs east of the Pic Ariel; it contains plagioclase, which shows zoning and simple twinning. The source of the first dyke is presumably the mobilized quartz-dioritic material. Concerning the second dyke this seems less certain because of the great distance involved.

4. METAMORPHIC ORDOVICIAN OF THE EL SERRAT AREA

The Ordovician sediments of the Hospitalet massif have been briefly described elsewhere. The metamorphosed Ordovician rocks of the El Serrat area are bounded to the north by the Mérens fault, to the west by the sericite schists of the Arinsal valley, to the south by the Silurian and to the east by biotite-andalusite-schists. The composition of the sediments is rather similar to that of the Upper Ordovician from the Aston massif. Only microconglomerates are much rarer.

Basically the rocks around El Serrat represent an east-west trending anticline. Due to repeated intensive folding only few sedimentary horizons could be traced through the area. One is a black slate horizon running from the Pic de Cataperdis via the Pic de Varilles to the Port de Banyells; it is partly accompanied by limestones. Another one is a grey limestone bed, which can be followed from 1 km south of El Serrat eastward to the Pic de Serrera. Other limestone bodies are scattered and deformed to such extent that they cannot be correlated with certainty. A few thick quartzites can be traced over some distance, but mostly they grade into one of the numerous banded phyllite-quartzite sequences.

The Ordovician anticline is disturbed considerably by the abnormally thick Silurian of the Llorts valley. In this area the Silurian is squeezed into the Ordovician.

4.1 Folding

Like in the Aston massif the dominant trend of the folds is E-W. These folds have developed an axial plane slaty cleavage. The fold axes generally have a low dip or are subhorizontal (fig. 8) and they belong to the main deformation phase. Their character is often isoclinal in the phyllites and quartz-phyllites.

In many places this pattern is disturbed by another folding, of which the axes run NW-SE or sometimes NE-SW. This so called "cross folding" is mostly a minor folding, which deforms the s_1 planes and therefore it must be younger than the main folding. The plunge of the axes (fig. 8) depends on the local dip of the s_1 plane. Parallel to the axial plane often a crenulation cleavage is developed. This cleavage may be very coarse, the planes being 1 cm apart, in other cases it is much finer and contains oriented biotite. However, there are never more than about 6 cleavage planes per mm in contrast with the first schistosity which is a perfect mineral orientation.

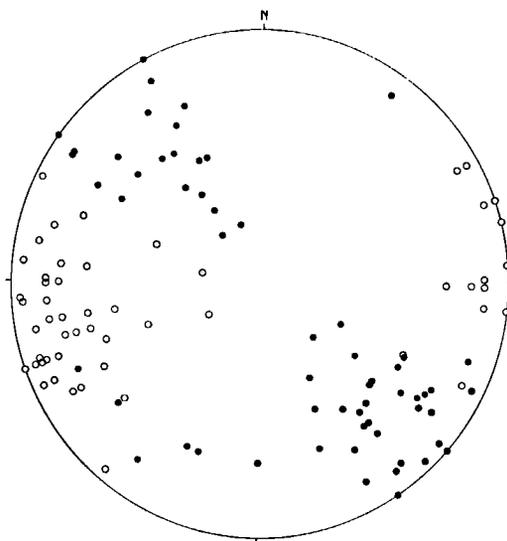


Fig. 8 Stereogram of foldaxes of first, main, phase open circles and third phase, (dots).
El Serrat area.

Locally one more set of folds is developed. It has again an E-W direction with a steep axial plane and crenulation cleavage. This folding is almost entirely confined to non- or slightly metamorphosed rocks. Only in the Comabauga creek this folding occurs in andalusite bearing rocks.

4.2 Metamorphism

The metamorphism in the El Serrat area is characterized by sericite defining the schistosity planes with cross-cutting biotite and aluminium-silicates. Andalusite is by far the most common, but staurolite and cordierite are frequent too. Staurolite often shows penetration twins (fig. 9). The rocks are phyllites with a superimposed andalusite-zone metamorphism. No feldspathic rocks have been found, apart from a few albite-bearing quartz veins.

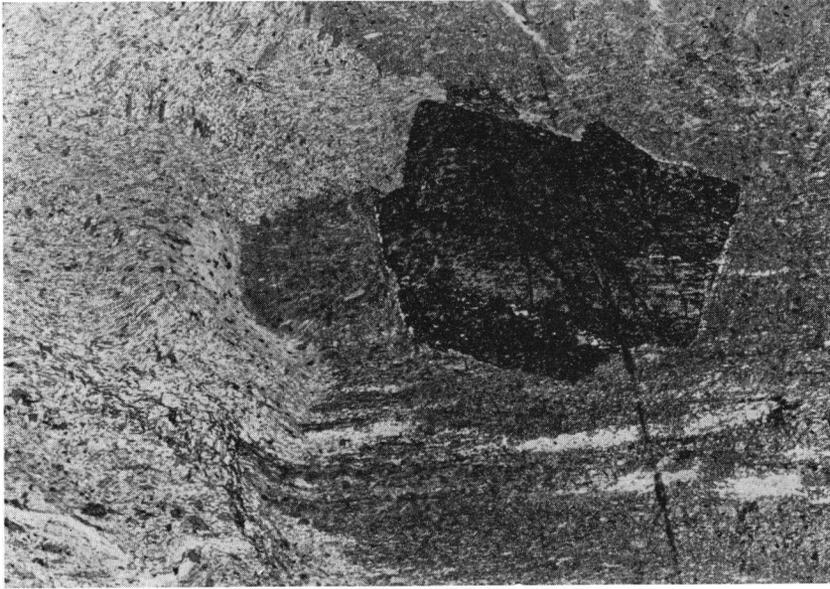


Fig. 9 Staurolite with penetration twins in cordierite crystal; both with planar *si*; matrix folded; El Serrat.

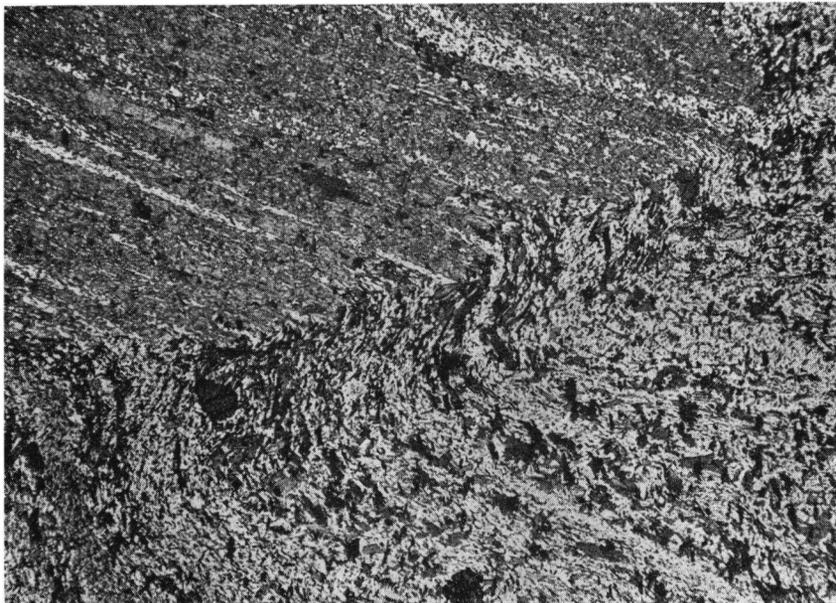


Fig. 10 Andalusite with planar *si* in folded matrix; El Serrat.

Only with the aid of the microscope the relations between folding and crystallization phases becomes more clear. Criteria as mentioned by Zwart (1960) have been used for this purpose. The main constituents of the pelitic rocks are: quartz, sericite and biotite; often also andalusite and locally chlorite. Other minerals arranged in order of descending frequency are: staurolite, cordierite, garnet, epidote.

One of the interesting specimens is pictured in fig. 10. The groundmass consists of fine grained quartz and sericite. A few quartz-rich streaks probably represent remnants of the original bedding. The schistosity (s_1) is visible as a planar si , consisting of sericite and quartz, inside the andalusite crystals. The se , however, is strongly deformed into short accordion type folds by the crossfolding. Fig. 11 shows a staurolite porphyroblast with a planar internal schistosity in a strongly folded groundmass which contains a crenulation cleavage (s_2), curving around the staurolite. In fig. 12 another thin section of the same area is depicted. In the upper part a folded andalusite crystal with wavy extinction occurs in a strongly folded matrix, in which a secondary schistosity is determined by oriented biotite crystals. The other porphyroblasts are undeformed staurolites with a planar si .



Fig. 11 Staurolite with planar si in microfolded matrix with crenulation cleavage curving around porphyroblast; El Serrat.

From these pictures it is evident that the oldest minerals are quartz and sericite, of which the latter one determines the schistosity. Andalusite, staurolite and part of the biotite developed after the end of the main folding, but before the start of the crossfolding, as is indicated by the straight si . During this time tectonic activities must have been absent.

In other cases the andalusite is rotated, but usually not broken or bent. Cordierite shows virtually the same relationship with the two folding phases. Biotite is not

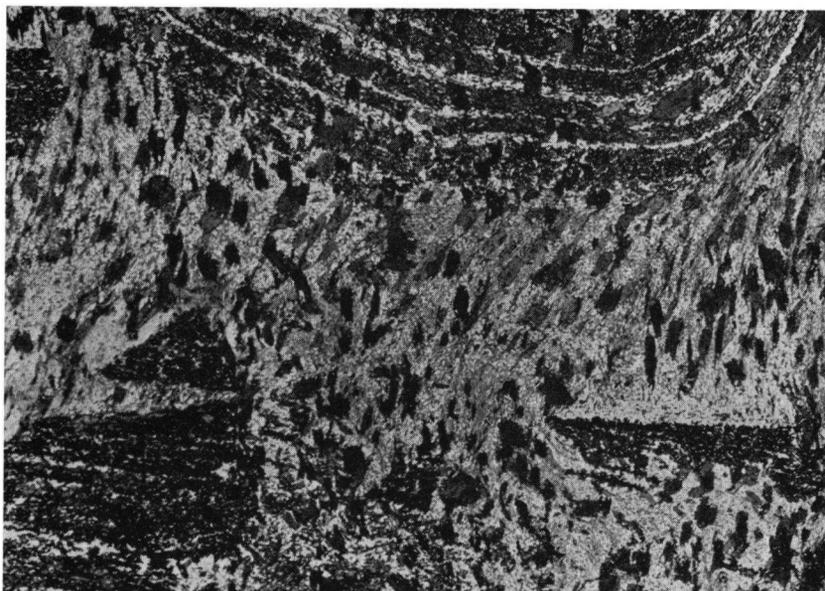


Fig. 12 Folded andalusite crystal and undeformed staurolites in folded schist with biotite indicating second schistosity; El Serrat.

found in the primary schistosity and presumably started to grow just after the main folding, like the aluminium silicates. Often biotite is found lying in the cleavage planes and rarely even crystals cutting through these planes. Sericite and muscovite occur also as alteration of the aluminium-silicates and biotite. Coarser muscovite occurs in the s_3 planes or is crosscutting. Probably the youngest mineral of the El Serrat rocks is chlorite, it occurs in s_3 planes and as alteration like muscovite, but often also as crosscutting pennine, which seems to be younger than all other minerals.

Schematically the sequence of the events described above, is as follows: during the first deformation phase muscovite-schists with a schistosity were created. After the cessation of this phase abundant porphyroblasts of andalusite, staurolite, cordierite and biotite grew in these schists, inheriting a planar s_1 . Folding with NW-SE trending axes deformed the schistose matrix, partly oriented the smaller biotite crystals and also deformed part of the andalusite and cordierite. Staurolite usually resisted the deformation. Some growth of biotite may have lasted during the NW-SE folding. This holds only for the immediate surroundings of El Serrat. To the east, for instance when going up the Sorteny valley, the pattern changes gradually. In this valley at about 2000 m altitude the first synkinematic biotite can be seen in the field. Thin sections show synkinematic sericite and fine biotite (up to 1 mm) and post-main phase coarser biotite, staurolite, cordierite and muscovite. Still higher up also part of the andalusite appeared to be oriented in the main folding schistosity. It is clear that during the first phase of deformation there was an epizonal metamorphism in the El Serrat area and an andalusite zone metamorphism in the Ransol area. Subsequently the higher grade metamorphism spread westward, reaching as far west as 1 km SW of the Pic de Varilles.

5. NON-PELITIC COMPONENTS IN THE METASEDIMENTS

5.1 *Quartzitic rocks*

This group unites pure and impure quartzites and microconglomerates. In the phyllite zone these rocks display few traces of metamorphism in the field. Thin sections show that the quartz is recrystallized and that the flattened pebbles of the microconglomerates consist of several quartz grains with parallel extinction. Some sericite is always present and in less pure quartzites sericite, chlorite and often epidote.

Towards the feldspathized micaschists and the biotite-gneisses the amount of quartzitic rocks gradually diminishes. Only the rather pure quartzites remain, but normally cannot be traced beyond a few metres. The less pure rocks probably became feldspathized and lost their characteristics. The purer quartzites, together with limesilicate rocks (see below), resisted the feldspathization and homogenisation of the rock. They survived the metamorphism up to the nebulite stage of the country rock and may be termed "resisters".

5.2 *Calcareous rocks*

This section comprises marbles, "cipolins" and limesilicate rocks. The name limesilicate rocks was used in the field for rocks distinctly more calcareous than the pelitic rocks, but not consisting mainly of carbonate.

In general the limesilicate rocks could be recognized by the occurrence of epidote and/or amphibole, by their fine grain and by their bedding even if the surrounding pelitic rocks have already completely lost the bedding. Schistosity and lineations are usually poorly developed. As with the quartzites the bedding is often disrupted and the distance over which it can be traced decreases continuously towards the migmatites.

No pure marbles or cipolins have been found in the phyllite zone proper (they do occur in the El Serrat area). In this zone limesilicate rocks occur mostly in well-separated beds. The rocks are completely recrystallized and contain chlorite, calcite, epidote and sometimes actinolite.

From the micaschists outside the pegmatite bearing area a sample of a calcisilicate rock has been studied (Table I). The rock is banded and biotite and plagioclase occur only in a few bands. The plagioclase is not abundant and occurs in small crystals (not over 0.2 mm). It appears to be labradorite. Epidote and a distinctly blue-green (Z), frequently euhedral hornblende are developed in larger blastic crystals up to 2 mm long.

Both limesilicate rocks and "cipolins" occur in the feldspar bearing micaschists (zone 4) and in the biotite-gneisses s.l. (zone 5). The marbles are always extremely folded and show characteristics of plastic flow. The layers underwent important tectonic thickening and thinning, and can be traced over a short distance only. Competent seams or veins within the marble mass were mostly torn into pieces, boudinaged and often also rolled. Some asbestos is often present in zones of intense shearing. Limesilicate rocks are widespread in zone 4 and 5. They are usually finer grained than the surrounding rocks, they rarely or never show the typical microfold lineations. Amphibole often is visible macroscopically. In zone 5 true amphibolites and amphibole gneisses occur.

The different rock types all have some variety of green hornblende (Table I) and mostly an intermediate plagioclase. The plagioclase shows different types of

twinning, often Albite-Carlsbad twins and some (001) lamellar twinning in the same thin section. Several times zoned crystals appeared to have rims 5—10 % more calcic than the cores. This may be explained by a higher second phase of metamorphism, a point supported by the presence in the surrounding rocks of abundant post-kinematic andalusite and fibrolite, which also are younger than the first of phase metamorphism.

Other amphibole bearing rocks contain plagioclase with often euhedral, basic cores (60—85 % An). This may be an indication that the rocks represent a metamorphosed volcanic rock. Only in a few cases pyroxene was found to support this view. Two of the pyroxene-bearing samples are from Et. Blaou where sediments probably belonging to the Canaveilles formation are present, which sequence is supposed to include volcanics. Several impure marbles have also been found in that region and at least a part of the limesilicate rocks probably will be of sedimentary origin. It is remarkably, however, that most of the samples with basic plagioclase cores are from the southern part of the Aston massif where the Canaveilles formation may be present.

Colourless clinopyroxene was found in an impure marble together with a colourless hornblende. This rock, which is certainly of sedimentary origin occurs southeast of Et. de Gnioure. Several specimens of limesilicate rocks have two different amphiboles (see Table I). Often one of the two is clear or uncoloured and occurs in patches in or around the green or brown hornblende and seems to replace it. Biotite pseudomorphic after hornblende is frequent. Epidote commonly replaces hornblende and basic plagioclase to strongly varying degrees.

In the quartzdiorites proper only the presence of amphibole shows that originally a somewhat more calcareous sediment was present. In the nebulitic rocks also parts of fine grained limesilicate rocks with recognizable bedding have been found. Near the northern entrance of the canyon north of Et. de Gnioure where the migmatites pass into nebulites, calcareous rocks occur in the nebulites. Bedded cipolin, limesilicate rock and amphibole gneiss are surrounded by disoriented rocks of quartzdioritic composition. The amphibole appears to be more brownish than the types mentioned above (Table I). In some samples a second amphibole occurs, like described above. No pyroxene has been found. The plagioclase has often basic cores in the labradorite-bytownite range, while the rims and unzoned crystals are of intermediate composition. Mostly the cores are euhedral and sharply separated from the rim. Carlsbad-albite and lamellar (001) and (010) twins occur. Some of the basic cores are partly altered to more acid plagioclase (not possible to measure the An-content). Epidote-albite veins in the rock and in the plagioclase have been found in one sample.

It is not clear whether here the basic plagioclase cores also indicate a volcanic source for the material, or that in this zone 6 the temperature was higher than in the other zones and that this basic plagioclase is a metamorphic one. No pyroxene occurs to support the first possibility.

In the El Serrat area of the Hospitalet massif where no feldspar bearing rocks were found cipolins and limesilicate rocks do occur. The bedding is better preserved than in the Aston massif and especially when the layers are accompanied by quartzites they can be traced quite far. From four different localities in the area an amphibole bearing sample has been studied microscopically. The amphibole is a green or bluegreen hornblende in the two higher metamorphic samples (andalusite zone) and pale green in the other ones (biotite zone). Plagioclase does not occur.

TABLE 1 *Composition of calc-silicate rocks*

ASTON MASSIF						
Rock type	Locality	No. of metam. zone (see p. 282)	Plagioclase	Amphibole	Remarks	
banded limesilicate rock	Pla Subra	2-3	small, in some bands only Labradorite	porphyroblasts often euhedral X = colourless-light Y = yellowish green Z = light bluish green	many small quartz and large epidote crystals biotite in some of the bands	
foliated limesilicate rock	Soulcem	3-4	33-38 % An, some are zoned with slightly more acid rim	X = colourless — very light yellowish green Y = green — light brownish green Z = light green — light bluish gr.	Hornblende often oriented, alters to unoriented biotite	
banded limesilicate rock	1 km SW of Etang de Siguer	4	35-40 % An, crushed, also Albite present	X = almost colourless Y = yellowish green Z = very light green (sometimes a little brownish)	Potassium feldspar in Hornblende-free bands Hornblende bands denote schistosity	
limesilicate rock	1 km NW of Pic Arial	4, near 5	34-45 % An, some zoned crystals with core 44 % An, rim 35 % An core 46 % An, rim 34 % An some Albite	few small crystals X = colourless-light yellow Y = green Z = green	much epidote. A pegmatite found in this rock has 36 % An plagioclase and no potassium feldspar	
Hornblende-biotite gneiss	*	4-5	Andesine, often zoned cores 35-38 % An rims 44-48 % An sometimes narrow Albite veins	X = very pale brown Y = green brown Z = brown green		

cores 40—48 % An
rims \pm 24 % ? An.
simple twins; Albite repla-
ces plagioclase

X = light yellowish
green
Y = light brownish
green
Z = light green

limesilicate rock	SE of Etang de Gnioure	5, near 4	Andesine 34—38 % An some zoned ones have rims of \pm 38 % An, Albite in veinlets		plagioclase builds augen, quartz in stress shadows
fine Hornblende gneiss	near Pic Arial	5, near 4	54—68 % An, some very basic rims (90 % An?)	X = pale yellow Y = dark olive gr. Z = dark green	Potassium feldspar?
limesilicate rock	S of Pic de Bourbon	5	zoned 33—46 % An rims more basic than cores few crystals have an (old?) very basic core	X = light yellow Y = light brown Z = light greenish brown X = colourless 2. Y = very pale brown Z = colourless — very pale green	Hornbl. 1 is euhedral, 2 as rims around 1, may be an alteration of 1. Pyroxene? relics
concretion in marble	2 km S of Etang de Gnioure	5	very little (bytownite?)	stalky colourless	colourless clinopyroxene
quartz bearing amphibolite	S of Etang de Gnioure	5	Andesine 40—45 % An	X = pale brown to Y = greenish brown, Z = probably due to alteration	Hornblende altered to bio- tite
nebulitic migmatite	W side of Siguer valley	5 near 6	zoned andesine cores 33—38 % An rims 40—45 % An	a little Hornblende X = colourless — light greenish yellow Y = olive green Z = green (sometimes pale, probably as a result of chloritiza- tion)	much biotite, some Pot- assium feldspar

ASTON MASSIF

Rock type	Locality	No. of metam. zone (see p. 000)	Plagioclase	Amphibole	Remarks
pegmatite in amphibolite	NE of Pic Thoumás	5	bytownite 80 % An	X = light yellow Y = green Z = blue green	colourless clino- pyroxene intergrown with Hornblende and calcite. Pegmatite part has potassium feldspar and myrmekite
amphibolite	E side of Etang Blaou	5	zoned, euhedral cores: 71—79 % An; rims and not zoned fresh crystals: 46—50 % An	X = almost colourl. Y = green-olive gr. Z = green-brown-gr.	colourless clino-pyroxene, Hornblende altered to epidote and biotite
quartzdioritic rock in migmatite	Siguer valley	5—6	a little altered plagioclase	few twins X = almost colourl. Y = olive green-brownish green Z = light green	much epidotebiotite porphyroblasts
dark nebulite	Siguer valley	6 near 5	zoned Kb—Ab twins cores 53—60 % An rims 41—53 % An	X = colourless — very pale brown 1. Y = dark olive green Z = dark green X } almost colourl. Y } Z }	Hornbl. 2 in or around 1, probably originated from 1
Hornblende-quartz-diorite; nebulitic	S of Etang de Gnioure	6	andesine, zoned 40—45 % An	X = colourless Y = light brownish-green (pale) Z = light green	Hornblende altered to biotite
Hornblende-quartzdiorite	SE of Etang de Gnioure	6	labradorite, some crystals zoned with anhedral very basic cores; rims and not zoned crystals: 52—60 % An	X = very pale brown Y % green Z = slightly brownish green parallel twins	

limesilicate rock in quartz-diorite

6

Siguer valley

zoned, euhedral cores 80-75% An; rims 40-45% non-zoned crystals of 60% An

X = colourless
1. Y = light brown
Z = very light brownish green
2. Z = more bluish green than 1

Hornblende 2 as rims around 1. Some basic plagioclase cores altered to intermediate plagioclase and epidote

6

Siguer valley (float)

zoned, some euhedral cores (86% An?), rims in labradorite range. Albite veinlets

much, some twins
X = colourless
Y = colourless-light brown
Z = colourless-light brown

Hornblende-quartzdiorite

HOSPITALET MASSIF

calcareous phyllite 1 NE of Pic Ortell
hornblende or actinolite
X = pale yellowish brown
Y = green
Z = bluish green
much epidote and calcite

limesilicate rock 2 SW of Pic de Varilles
altered
X = colourless — very pale yellowish green
Y = light green?
Z = very light bluish green
much epidote; quartz some garnet

limesilicate rock 1 +2 1/2 km NNW of Etang de Sorteny
X = light yellow
Y = green
Z = bluish green

limesilicate rock 1 +3 2 km N of Etang de Sorteny
Hornbl. (or cumingtonite?)
X = colourless
Y = colourless
Z = colourless — very pale green
much epidote, in part primary

5.3 Conclusions

From the data represented in table I it appears that no close relation exists between the metamorphic zones (mainly based on Al silicates) and the colour of the common hornblende in the limesilicate rocks. A certain trend, however, is visible.

Colourless and pale green hornblendes occur as alteration products and in the rocks of low metamorphism, blue-green hornblendes belong to approximately mesozonal rocks and green respectively brown-green and brown hornblende in rocks of the amphibolite facies proper.

Probably most of the limesilicate rocks derive from calcareous sediments, but a small portion may represent metamorphosed volcanics. Especially for amphibolites near Et. Blaou this last possibility is likely.

6. LEUCOCRATIC GNEISSES

The core of both the Aston and the Hospitalet consists of rather homogeneous gneisses, poor in mafic constituents.

The gneisses of the Hospitalet massif — situated outside the area under consideration — will not be discussed here.

In the Aston massif the leucocratic gneisses constitute the core of the dome, which in the central part has a horizontal schistosity while in the flanks, especially the northern one, this structure dips steeply. The western part of the gneiss core is situated on top of the migmatites. In most places the boundary between the gneisses and the other rocks is sharp and often even mylonitic with quartz-rich movement zones. In some places, especially north of the Pic de Bourbon, granitic rocks are situated between the gneiss and the migmatites. Just south of the Col de Laine, 3 km SE of Siguer, a nonmylonitic boundary between gneiss and micaschist is exposed. Here a micaschist with a large content of oligoclase (15—20 % An) is underlain by quartzitic rocks containing mica and feldspar. Below a pegmatite sill and a small unexposed part irregular gneisses occur with an undulating schistosity and a considerable content of potassium feldspar and plagioclase with 10 % An. Further down the gneiss becomes more regular, but over some hundreds of metres the biotite content remains higher than is normal for the leucocratic gneisses. The whole sequence dips 75° to the NW. A rather vague transition between migmatites and flasergneisses was only found on the ridge north of the Pic du Midi de Siguer.

The leucocratic gneisses are mineralogically very homogeneous, but on structural grounds several types may be distinguished.

6.1 *Medium-grained gneisses*

1. Augengneisses
2. Flasergneisses
3. Granitic gneisses (Raguin's "gneiss de Peyregrand")

In outcrop the rocks have a schistosity marked by parallel oriented biotite. Quartz, plagioclase and potassium feldspar are also recognizable in the handspecimen, and sometimes garnet or fibrolite.

Augengneisses. — In the eastern part of the Aston massif this type is abundantly represented. The gneisses are medium to coarse grained and somewhat richer in biotite than the other varieties.

Flasergneisses. — The flasergneiss is a transitional type between the augengneisses and the granitic gneisses (see map). The contacts with these gneisses are gradational. An undulating schistosity made up mainly by the biotite is often combined with the occurrence of euhedral feldspar porphyroblasts. Sometimes feldspar augen are found. The quartz and rest of the feldspar is equidimensional and has no preferred orientation. Locally small streaks or lenses of fibrolite are oriented in the schistosity plane.

Microscopy: The main components are potassium feldspar and quartz, which together make up 65 % or more of the rock, and plagioclase. Biotite and muscovite are also present. The light minerals seem to be statically recrystallized, but often the grains have a wavy extinction, especially quartz. The plagioclase has an An-content between 5 and 12 %, the albite range being the most common. Mostly albite-law lamellar twinning occurs but also untwinned plagioclase is present. Simple and complex twins are rare. Zoned crystals (normal zoning) are uncommon. Potassium feldspar is often more or less perthitic. Although the complicated quartz-feldspar relations are far from being completely understood, it is clear that quartz and probably also albite replace potassium feldspar to some extent. Myrmekite, common in other rocks in this area, occurs scarcely.

Biotite occurs in streaks but is often recrystallized. In some cases fibrolite was encountered together with the biotite, the knots are elongated in the schistosity plane, but obviously postkinematic as fine cross-cutting needles always occur. Frequently it is visible that the fibrolite is a pseudomorphic alteration of biotite. Muscovite is frequently replacing both biotite and fibrolite and is often cross-cutting. Garnet may be present, in some cases flat crystals have been observed. Secondary minerals are chlorite, sericite. Accessories are zircon, tourmaline, apatite, sphene and ore.

Granitic gneisses (Gneiss de Peyregrand). — Granitic gneisses are found in the region around the Et. de Peyregrand and reach as far west as the Et. de Gnioure. The schistosity is less undulating than in the previous type; it will be discussed below because it is more outspoken in the fine grained varieties of the gneiss. In the medium-grained gneisses foreign inclusions have been found, among which migmatite, micaschist, few amphibolite and (hornblende) quartzdiorite. Most of these inclusions are more schistose than the gneisses. The pieces never exceed a few dm in size and are very rare, a fact which stresses the homogeneity of the leucocratic gneisses.

6.2 Fine-grained gneisses and granites

In all three medium-grained gneiss types fine grained leucocratic gneissic and granitoid rocks occur. The occurrences are numerous and widespread from a few cm to over a hundred metres. Only a few large ones are shown on the map. Mostly the boundary with the normal gneisses is not sharp but the transition comes about within one metre, only where the boundary is a schistosity plane it is mostly sharp. Mineralogically the fine grained type is equal to the normal gneiss and also the chemical composition is similar. A schistosity is indistinct or absent.

In a few of the larger granite occurrences the result of rheomorphism has been observed. A beautiful example can be seen in a narrow valley south of the Et. de Milleroques. Here the fine-grained granite contains angular blocks of medium-grained gneiss with a planar schistosity, which are rotated as much as 90° in respect to each other. This feature is probably the result of plastic flow in the granitic material. At the same time this shows that the age of the mobilization and the crystallization of the fine-grained rocks is later than the main phase.

6.3 Origin of the gneisses

Originally the gneisses in the core of the Aston massif were considered to be orthogneisses (Zwart 1954) but later the possibility that the gneisses are metasediments was considered. A study of the zircon content of the gneisses (Verspyck 1961) showed that a considerable amount of well-rounded zircon grains is present in the gneisses, which militates against an "ortho" origin of the gneisses.

The supposition that the gneisses result from the same pelitic sediment as the rocks of the phyllite-quartzdiorite series, but through a more drastic metasomatism (Destombe & Raguin 1955) is not in accordance with the great homogeneity of the gneisses and with their sharp boundary with the surrounding rocks. In this case a gradual transition would be expected. The original difference of the migmatites and the gneisses is further supported by the investigation of zircons (Verspyck 1961). The zircons in the gneisses are quite different from those in the migmatites. If the leucocratic gneisses are of sedimentary origin they were a special kind of sediment, for example arkoses. Another possibility is that they were volcanic rocks, for example rhyolites. Neither of the two rocktypes is present in any appreciable quantity in its low grade or non metamorphic state in the Pyrenees, and for these reasons an ortho-origin seems the most probable. This fits best with the chemical and mineralogical homogeneity and the sharp contacts of the gneiss with the surrounding rocks.

6.4 The quartz nodule rocks in the granitic gneisses

A special problem is presented by rocks with flatquartz nodules or lenses which occur mainly in the granitic gneisses (see fig. 13).

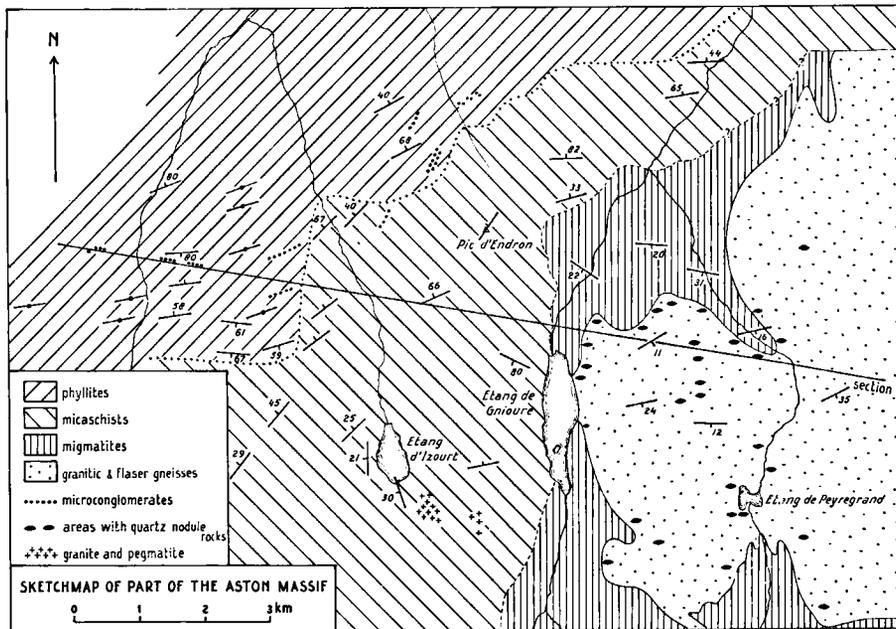


Fig. 13 Map showing distribution of quartz nodule rocks in granitic gneisses and flaser gneisses.

Where quartz lenses occur they are situated in the schistosity plane and when several lenses are close to each other they are often arranged in line and sometimes even connected. The shape of the quartz lenses may vary from almost spherical (rare) to a much flattened lens. The most common occurrence is shown in fig. 14; when many lenses occur together they may cover a large part of the outcropping schistosity plane (fig. 15). Individual lenses are often surrounded by mica or fibrolite and their appearance resembles much the fibrolite cloths of the normal gneisses, although these are less common; both occur in the most schistose part of the rock.

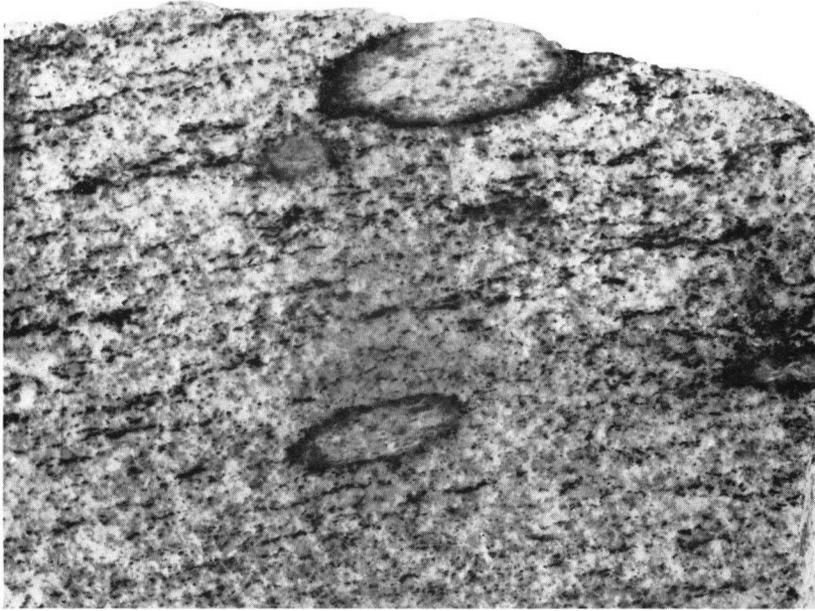


Fig. 14 Quartz-sillimanite lenses in granitic gneiss. (1×)

Several times it has been observed in the field that the quartz lenses show a fine lineation or slickensiding, some 20 measurements all showed that the direction of the lineations has a NW-SE to NNW-SSE trend, coinciding with the direction of the axes of a secondary folding known from other places in this massif.

The quartz lenses consist of numerous individual quartz crystals, which together make an elongated ellipsoidal quartz assemblage. The quartz crystals do not seem to have a preferred lattice orientation. The quartz lenses are often surrounded by fine fibrolite needles penetrating the lens between the individual quartz crystals (fig. 16). Mostly the fibrolite is in part replaced by muscovite; whether primary muscovite is also present is hard to establish. Biotite may be present. This mineral is sometimes replaced by chlorite or muscovite.

There are several objections against the idea of the quartz lenses being pebbles of granitized micro-conglomerates as suggested by Destombes and Raguin (1955).

1. In general the pebbles of the micro-conglomerates in this area are smaller than the quartz lenses described and the quartz lenses do not show any newly grown material around them.

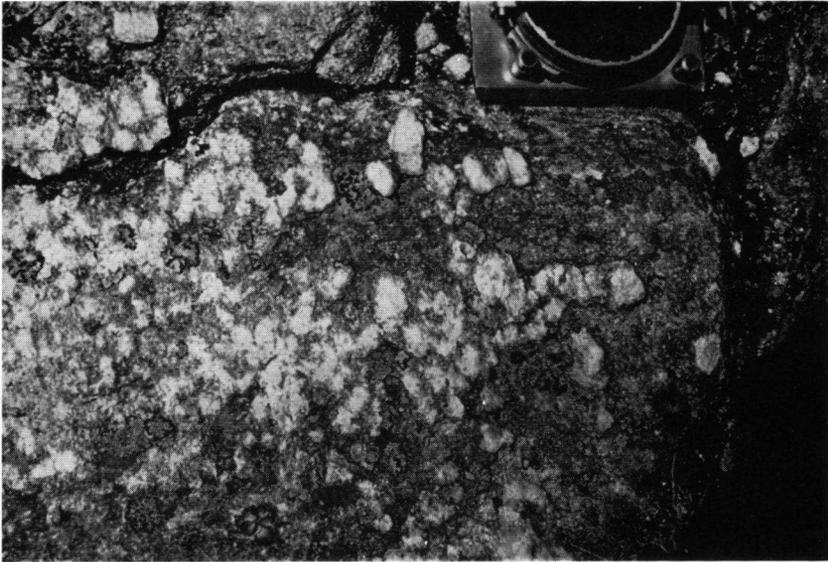


Fig. 15 Quartz-sillimanite lenses as seen on schistosity plane.

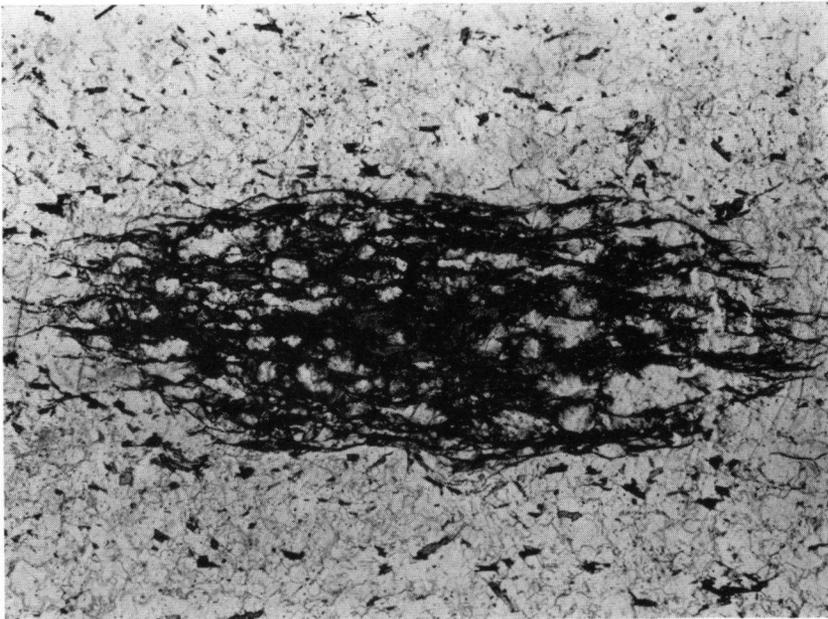


Fig. 16 Thin section of quartz-sillimanite nodule.

2. The lenses are much too few compared with the number of pebbles in an equal sized piece of conglomerate. It is not easy to understand why most of the pebbles would recrystallize and disappear and some are not attacked at all.
3. The micro-conglomerates are relatively rich in Ca, epidote-clinozoisite is common and calcite and hornblende occur also. The fine grained gneisses, the normal gneisses around them and even the pegmatites in the gneisses are very poor in Ca, so the Ca which according to Raguin was expelled from the microconglomerate layers should have been carried entirely outside the gneiss area.
4. Discordant contacts between the quartz nodule rocks and the enclosing granitic gneiss show that they are not sedimentary layers (fig. 17, 18).

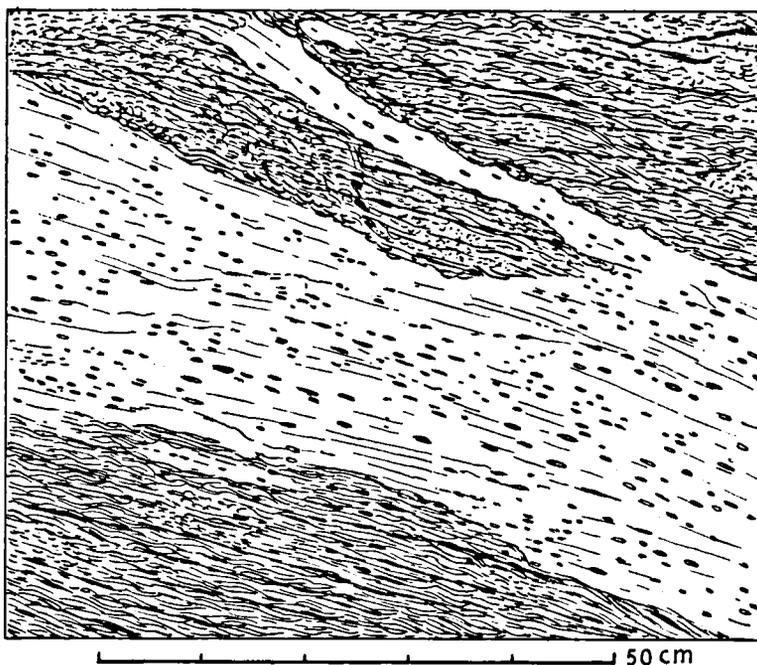


Fig. 17 Small vein of quartz nodule rock cutting discordantly through enclosing granitic gneiss.

5. Further objections concern the stratigraphic positions of both rocks discussed. The phyllites and micaschists are strongly folded mostly isoclinally, but in general the layers are parallel to the schistosity. Fig. 19 gives a profile constructed with our measurements. It is evident that the micro-conglomerates represent a much higher level than the gneisses with quartz lenses. For these reasons it is unlikely that the quartz nodule rocks are the lateral equivalent of the micro-conglomerates in the Cambro-Ordovician. In our opinion the quartz lenses with fibrolite and the fibrolite knots and streaks in the normal gneisses might be relics of planes of high differential movement, or shear zones. This hypothesis is supported by the following observations:

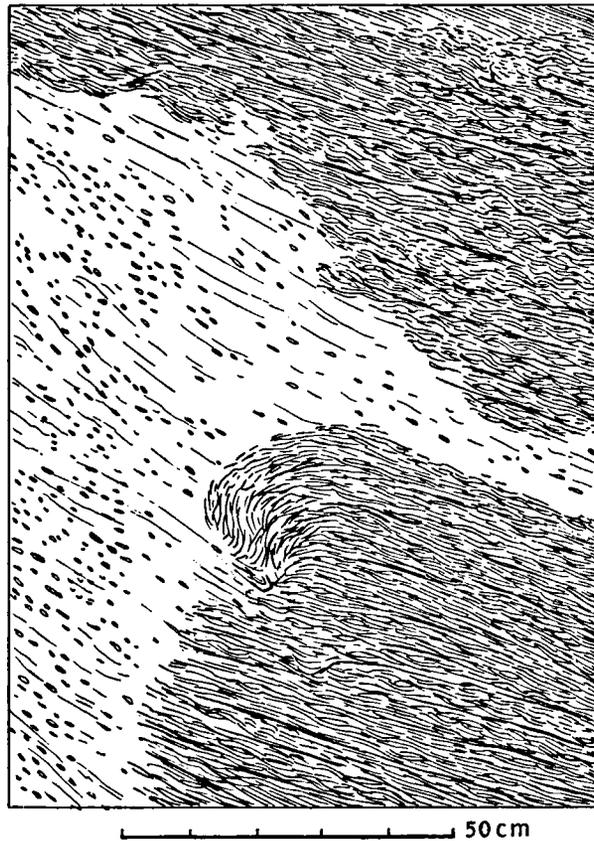


Fig. 18 Discordant contact between quartz nodule rock and granitic² gneiss; near Et. de Peyregrand.

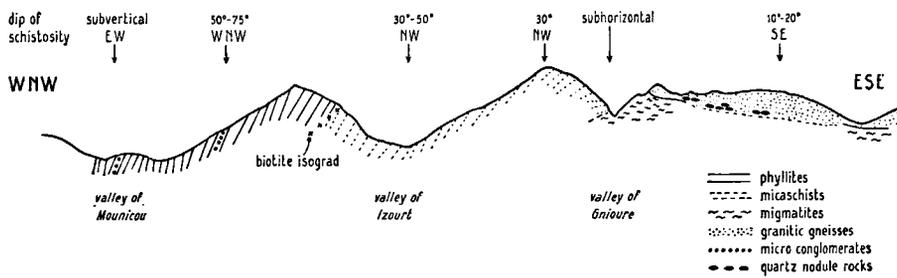


Fig. 19 Section through western part of Aston massif.

Where the contact between the fine grained type with lenses and the normal gneiss with fibrolite knots is discordant to the schistosity, the lenses are the continuation of the fibrolite knots, as can be seen on fig. 18. Apparently the fine grained type is the recrystallization product of the normal gneiss and the lenses are relics of zones of differential movement (fig. 20).

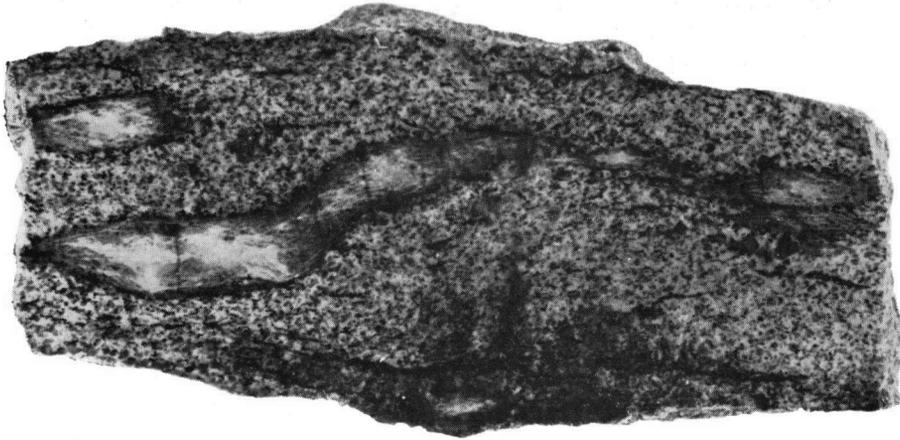


Fig. 20 Quartz-sillimanite lenses join and form a band. (1×)

As a matter of fact similar, perhaps a little more flattened lenses have been found in sheared migmatites bordering the fault of Mérens that separates the Aston- and Hospitalet massifs. Also the presence of the slickensides on the surfaces of the quartz lenses gives the impression that movement or shearing has played an important role in the formation of the lenses as they are at present. Part of the Al, which is known as one of the less mobile elements, became concentrated on the quartz lenses during the recrystallization of the rock.

Granitic gneisses with quartz-sillimanite nodules have been described by several authors. For instance the description and pictures given by Bugge from the Kongsberg Bable formation (S. Norway) are very much similar to our observations. He thinks that these rocks are of migmatic origin and that the lenses may originate from metamorphic differentiation of aluminium-rich sediments.

7. PEGMATITES

7.1 Occurrence

The pegmatites are restricted to an area with a certain degree of metamorphism, a grade slightly higher than is necessary for the growth of andalusite in the micaschists. It is not possible to indicate a certain pegmatite centre, the rocks are found throughout the whole area of the Aston massief where metamorphism was high enough.

Aplites often associated with pegmatites occur mostly as dykes, sills or irregular bodies. The thickness of the dykes varies between $\frac{1}{2}$ cm and 5 m, the range between 10 and 100 cm is the most common one. Irregular pegmatic bodies are up to several hundred metres in size.

7.2 Composition

Minerals present in all pegmatites are: quartz, plagioclase and muscovite, while potassium feldspar and biotite are mostly present too. The other minerals occurring in pegmatites are listed in order of descending frequency: tourmaline, epidote, andalusite, hornblende, garnet, sillimanite (fibrolite), calcite, apatite and cordierite. Accessories are: zircon, apatite, sphene, rutile and ore.

Sericite, muscovite, chlorite and sometimes kaolin were formed by alteration of other minerals. None of the characteristic minerals as Li-minerals, ores, etc. of magmatic pegmatites have been found.

The quartz has mostly an undulatory extinction and sometimes crushed and recrystallized zones can be seen. Plagioclase (albite excepted) is invariably twinned, mostly lamellar. Simple twins and zoned crystals are rare. The potassium feldspar is often perthitic, but the albite seems to comprise usually only a small fraction of the crystal, with the exception of the graphic granites. Potassium feldspar builds always large crystals, which show often microcline twinning. It is not found in pegmatites where biotite is one of the major constituents. Tourmaline is present in many dikes and smaller bodies and sometimes in the rim of larger bodies, rarely in irregular streaks in the core.

The degree of alteration is unequal in the different pegmatites, probably the alteration is promoted by later movements affecting the rock. In many pegmatites the plagioclase is strongly sericitized, while the potash feldspar is replaced by muscovite and quartz. The original granoblastic texture is considerably affected by such a strong alteration.

The quantities of quartz and feldspar of the rock may vary considerably. Pegmatites with up to 70 % feldspar are no exception in some areas e.g. near Et. Fourcat and just below the Soulcem in the Mounicou valley. Near the "pegmatite line" the rocks contain often very much quartz (above the line feldspar-free "pegmatites" occur). Also pegmatites with much andalusite seem to be poor in feldspar and rich in quartz. In general the aplitic rocks contain somewhat more quartz than the coarser varieties.

Pegmatite, granite and aplite often cannot be separated in the field, all transitions occur. Many of the dykes and all large bodies have a wide range of crystal sizes.

Dikes may show zoning, of which two types were recognized. A zoning in crystal size and a zoning in composition. The two types occur combined or independently.

Crystal size zoning in its most simple way is represented by dykes with a fine grained core and a coarse rim, the reversed situation is found frequently.

Two common types of compositional zoning occur:

1. quartz-tourmaline core with feldspar-muscovite rims, without notable grainsize differences.
2. coarse potassium feldspar core with finer grained material (quartz-plagioclase-muscovite) in the rim (fig. 21).

Dykes with a combination of types and with more than three different bands occur frequently especially in the area of Et. Fourcat and Et. de Siguer.

A quartz-tourmaline vein cutting through a pegmatite was observed near Et. Fourcat, proving the quartz-tourmaline vein to be the youngest.

Micrographic structure is common in pegmatites in the valley of Izourt and Mounicou, especially around Et. Fourcat and the Soulcem. The micrographic quartz is situated in potassium feldspar. Under the microscope it appears that all quartz grains situated in one potassium feldspar crystal have a parallel extinction.

The relation between the quartz and the exsolution albite lamellae can be seen in fig. 22; here the quartz was formed before the exsolution started and possibly simultaneously with the microcline.

Micrographic structure is often associated with the so called "muscovite palmé", a fan shaped arrangement of fine muscovite laths. This feature is considered to be an alteration product of potassium feldspar because it appears under the microscope that the muscovite has developed in fissuring planes or cracks of the



Fig. 21 Zoned pegmatite with quartz-muscovite core and feldspathic rim; Et. d'Izourt

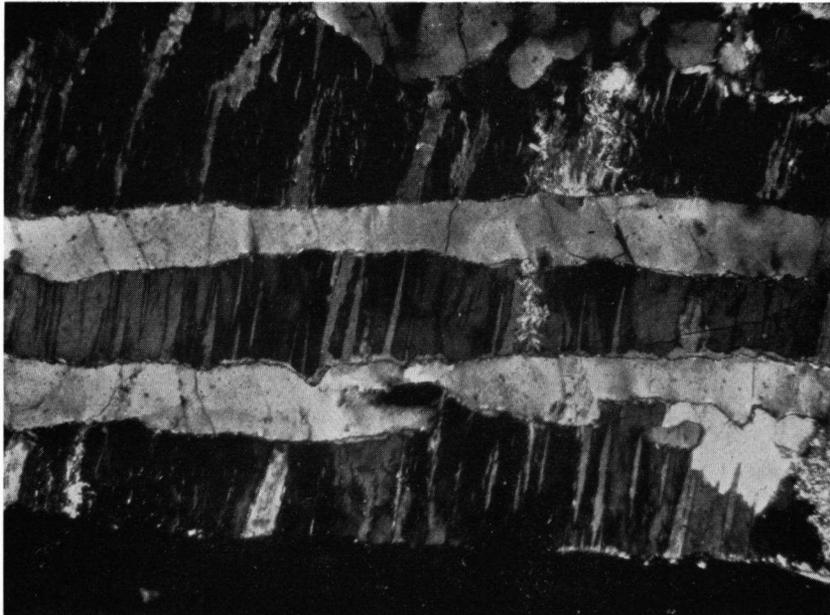


Fig. 22 Quartz lamellae in potassium feldspar

potassium feldspar, and often in fan-shaped arrangements. Sometimes the individual laths of the muscovite seem to occur in long quartz crystals building a fan shaped frame on which the muscovite developed. Why graphic granite and plumose muscovite occur so frequently together is not understood as yet.

Myrmekite is found in some pegmatites. This feature is discussed at some length in chapter 3. Also in the pegmatites it seems that mostly the potassium feldspar is replaced by plagioclase and quartz.

7.3 Relations with the country rock

7.31. *Mineralogical relations.* — In the phyllites and in the micaschists above the pegmatite line no pegmatites are found. In this region quartz, quartz-muscovite and quartz-muscovite-andalusite veins and nests occur. They have the same structural properties as the smaller pegmatites that occur in the higher metamorphic rocks.

In the phyllites south of the Mérens fault quartz and quartz-albite veins and nests are present.

Andalusite-bearing pegmatites and quartz veins occur only in andalusite bearing micaschists, not in migmatites and gneisses. Sillimanite is only found in pegmatites in sillimanite-bearing country rocks. Calcic plagioclase with more than 45 % An is found exclusively in pegmatites in the limesilicate rocks and amphibolites of the migmatite-quartzdiorite area.

These observations show that the pegmatites, according to their minerals, belong to the same metamorphic facies as the host rock. An exception of this rule was not found in the whole area.

Besides this clear conformity in facies, there is also a chemical relation between pegmatite and country rock. Albite occurs in pegmatites in the leucocratic gneisses and sodic plagioclase (up to 23 % An) is found in pegmatites of the sillimanite bearing migmatites because these rocks have respectively very low and low Ca content. Pegmatites occurring in limesilicate rocks or amphibolites in the same localities have calcic plagioclase (50—80 % An) and contain hornblende. Albite or sodic oligoclase is present in pegmatites of the granitic gneisses which are very poor in Ca.

In areas of lower grade metamorphism pegmatites in schists contain andalusite, in limesilicate rocks epidote.

These observations suggest a close chemical relation between pegmatite and host rock. Minerals which may indicate introduction of material are mentioned later on.

7.32. *Structural relations.* — The structural relation between the country rock and the larger granite-pegmatite bodies is rather uniform. The boundary with the host rock is almost invariably vague, especially in the direction of the country rock schistosity, in which direction often streaks of mica continue into the pegmatite.

On a larger scale the pegmatite masses are cutting off the schistosity and they are apparently post-tectonic in age. The undulatory extinction of quartz and the flexures in mica, phenomena often observed in thin sections, are ascribed to later movements.

Small nests, consisting merely of a few large quartz and feldspar crystals have been observed frequently. Mostly no veins or fissures, which could have served as channels for material supply are visible and the occurrences give the impression to have developed in place. From these nests to the large pegmatite bodies all transitions in size exist in the field.

Many dykes are situated in the schistosity and are not deformed internally, probably they are post-main folding features. Few concordant dykes show pinch and swell structure or boudinage, they may be syn- or late tectonic. Among the smaller veins (under 10 cm wide) pygmatic folding occurs, but it is not common. A syntectonic origin is assumed for these veins.

In the rocks with a pronounced schistosity, viz. the micaschists, migmatites and biotite gneisses, both pegmatite dykes and irregular bodies are frequent. Few dykes occur in the quartz-diorites and the leucocratic gneisses.

Structural control of pegmatites is often observed. Pegmatitic material occurs in fold hinges and axial-plane faults (fig. 7) between boudins etc. Some of the smaller dykes are situated in faults or fractures, others are not at all influenced by such phenomena.

7.4 Emplacement

Criteria as given by Niggli (1952) and Chadwick (1958) have been applied in the field, in order to decide whether the pegmatites were emplaced by intrusion or by replacement of the host rock.

1. Mostly it is impossible to fit the two walls of a dyke together and often the differences are even considerable. Only some aplitic dykes have straight walls, which fit together.
2. It has been observed frequently that parts of the country-rock structures (e.g. parts of folds, boudins) seem to have disappeared (have been replaced).
3. Often mica streaks or quartzitic bands of the country-rock continue into the pegmatite without being bent or pushed aside.
4. Crystal orientation pointing to flow of material has not been observed, on the contrary many crystals (especially in smaller dykes or dykes with zoning) have their length oriented normal to the strike of the dyke.
5. Offset criteria. When a dyke cuts an older dyke or some structure of the country-rock, the following cases have been observed and are schematically illustrated in figure 23.
 - a) No offset. In the area concerned this point is statistically of great importance, as commonly the pegmatites do not show offset (fig. 24).
 - b) A rather large amount of offset.
 - c) Offset in the "wrong" direction. Both cases b and c indicate the presence of a fault, and occur frequently.

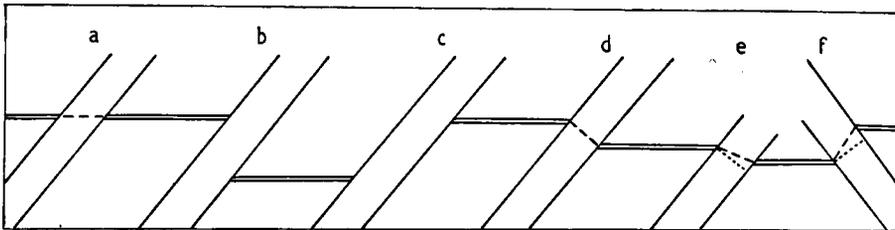


Fig. 23 Offset criteria; see text.

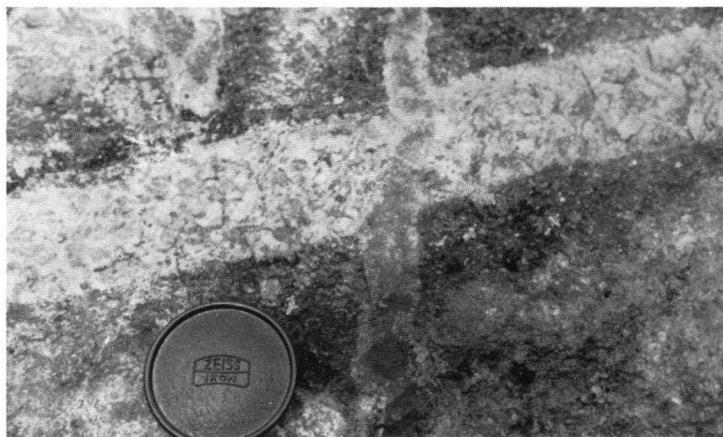


Fig. 24 Non-offsetting pegmatite

d) Offset indicating dilation normal to the walls of the dyke. This feature has been observed, but is rather rare.

e and f) A small amount of offset in either direction. This may be caused by a small fault displacement, by offset not normal to the pegmatite walls or because the observed surface cuts the direction of dilation obliquely. These features have been observed, but are considered inconclusive.

From the points mentioned above it is concluded that many pegmatitic dykes were emplaced by so-called "volume by volume replacement".

The exceptions are principally small, quartz-rich aplitic veins. Concerning the thin granitic veins, which always follow the schistosity and sometimes show pyg-matic folding, not much insight has been obtained into their emplacement. The

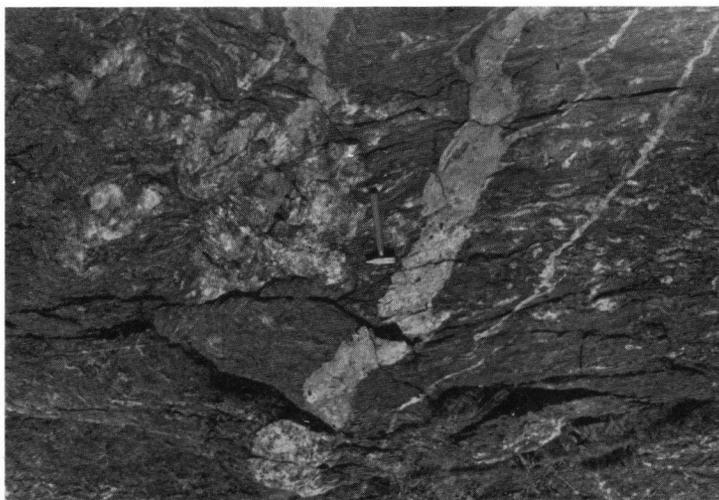


Fig. 25 Pegmatite vein and irregular pegmatite mass.

same holds for the zoned dykes, they have not been found cross-cutting, but mostly have fairly sharp and straight walls.

The irregular pegmatitic bodies (fig. 25) show all characteristics of a replacement origin. Criteria 2 and 3 can be applied here, and besides the occurrence of not disoriented pieces of country-rock in the pegmatite suggest a replacement origin.

7.5 *Origin of the material*

Owing to the coarseness and diversity of the pegmatites it is difficult to use chemical analyses to prove or disprove any introduction of material into the replacement pegmatites. As mentioned above there is a chemical relation between most pegmatites and their host rocks, which might indicate that the material incorporated in the pegmatites derives largely from recrystallization of the country-rock.

However, several minerals indicate that the formation of pegmatites was accompanied by new material.

The most striking example is tourmaline. This mineral occurs in many pegmatites and sometimes even more abundantly in the surrounding country-rock. Apparently the Boron was supplied by the pegmatites. The Boron might derive from some magma, but as marine sediments contain more Boron than a residual-magma it is more likely to consider the sediments as the source of the Boron. It is a very mobile element and may have been liberated from the sediments by hot gasses or fluids, and redistributed by the pegmatites.

Whether the mica of the other host rocks can account for the potassium present in microcline occurring in many pegmatites of the micaschist and migmatite area, is difficult to estimate. But it has been observed that micaschists and migmatites near pegmatites contain more often potassium feldspar than they do elsewhere, thus the possibility of some K-supply cannot be ruled out entirely. Even more frequently the pegmatites seem to serve as centre of plagioclase feldspatization. This can be seen clearly in the micaschists north and west of the Pic de l'Aspre.

In view of the low Na and the intermediate Si content of the pelitic sediment an enrichment of Na and Si in the pegmatites cannot be ignored.

The replacement character of the majority of the pegmatites excludes the possibility that large quantities of fluid material have been introduced. Possibly Si and Na (perhaps also K) were introduced by hot aqueous fluids or gases, of which the water content may have also promoted the formation of the late muscovite in many of the metamorphic rocks of this area. Similar presumptions can be made for the material probably introduced in the migmatites and quartzdiorites.

8. STRUCTURAL GEOLOGY

All rocks in the mapped area are strongly folded and contain at least one cleavage or schistosity plane. There is a difference to be made between the nonmetamorphic or low grade Cambro-Ordovician and Upper Paleozoic in the so called suprastructure and the mesozonal regional metamorphic Cambro-Ordovician of the infrastructure. In the first group of rocks a slaty cleavage is developed with a steep to vertical attitude. These structures occur in the Cambro-Ordovician north of the Aston massif and in the Cambro-Ordovician, Silurian, Devonian and Carboniferous south of the Hospitalet massif.

The schistosity in the micaschists and gneisses of the Aston massif has a rather flat to even horizontal dip, but these flat structures grade into the steep cleavage

structures near the biotite isograd. The metamorphics of the Hospitalet massif have an intermediate position with regard to the dip of the schistosity, but this is due to the situation of my area in the western plunge of this massif.

There can be little doubt as to the fundamental difference between the tectonics in the low grade steep structures and the higher grade flat metamorphics, although it is difficult to assess the differences. For some more information I refer to the publications of Lapré and Zwart in this issue.

The schistosity planes in the metamorphic rocks show some large but gentle folds, which probably are later than the formation of the schistosity itself. One fold, an antiform, can be followed from Etang d'Izourt to north of Et. Gnioure and then to the Pic du Midi de Siguer. Its axis plunges gently to the west. Its northern flank is rather steep and forms the northern border of the Aston massif. South of this antiform lies a synform whose axis runs through the granitic gneisses. To the south this synform is cut off by the Mérens fault, a steep upthrust with a down-thrown southern block. Beyond this important fault occurs the Hospitalet anticline which is the western continuation of the Hospitalet gneiss massif. Then follows a large outcrop of Silurian near Llorts which has partly been pushed over the Cambro-Ordovician. The mode of formation of this large mass remains, however, uncertain. The Silurian forms the northern limb of the Tor syncline with a Devonian core. Farther south occurs the Massana anticline and finally the Llavorsi syncline.

The western edge of the sheet of granitic gneisses has to be discussed in some detail, as here a few difficulties arise. This sheet clearly overlies the micaschist-derived migmatites (see chapter 3, also Raguin and Destombes, 1955). This situation remains the same all along the southern border of the Aston massif, north of the Mérens fault. Along the northern border, however, the gneisses are overlain by micaschists and consequently the whole gneiss body is a sheet with metasediments above and underneath. One should expect this gneiss unit to reappear farther west in the plunging nose of the Aston massif, but this does not happen: to the west no gneisses occur. Various explanations are possible for this situation. According to Raguin and Destombes the gneisses are metasediments and the equivalent of the Cambro-Ordovician near the Mounicou valley. To these authors this is corroborated by the presence of metaconglomerates in the granitic gneisses which are correlated with microconglomerates on the Cambro-Ordovician. This hypothesis seems very unlikely to me as the composition of the Cambro-Ordovician is strongly different from the granitic gneiss and one would have to assume very important metasomatic changes to alter the sediments in a gneiss. Moreover the gneisses are so homogeneous that it is unlikely that they are metasediments. The most probable explanation to my mind is that the gneiss is an orthogneiss, which explains its homogeneity and its granitic composition. This also explains its absence in the western part of the Aston massif as this is then simply due to the shape of the intrusive body.

Besides the large structures mentioned, many small scale folds have been observed in the area. Some of these minor folds are contemporaneous with the large structures, and with the development of cleavage and schistosity (fig. 26, 27). These folds are characterized by the relation of their axial planes to the s-planes. A large part of the minor structures postdate these main phase folds, which can be concluded from the fact that the cleavage or schistosity-plane is folded. Several sets of such folds occur, but by far the most important one is a set with NW-SE striking axial planes. Folds of this set occur for instance abundantly in the El Serrat area (fig. 8, 28) near Etang Fourcat and Etang Tristaina, in the Silurian of Llorts (fig. 29), but also elsewhere in the area. They are best developed in micaschists and phyllites. Another set, though less important, has steep E-W striking axial planes. Folds of this

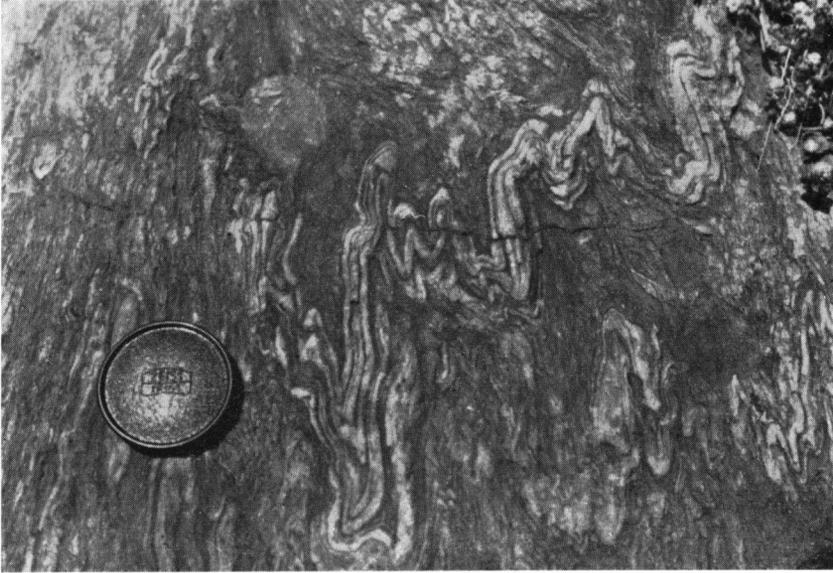


Fig. 26 Main phase folds in phyllite of Cambro-Ordovician south of Siguer.

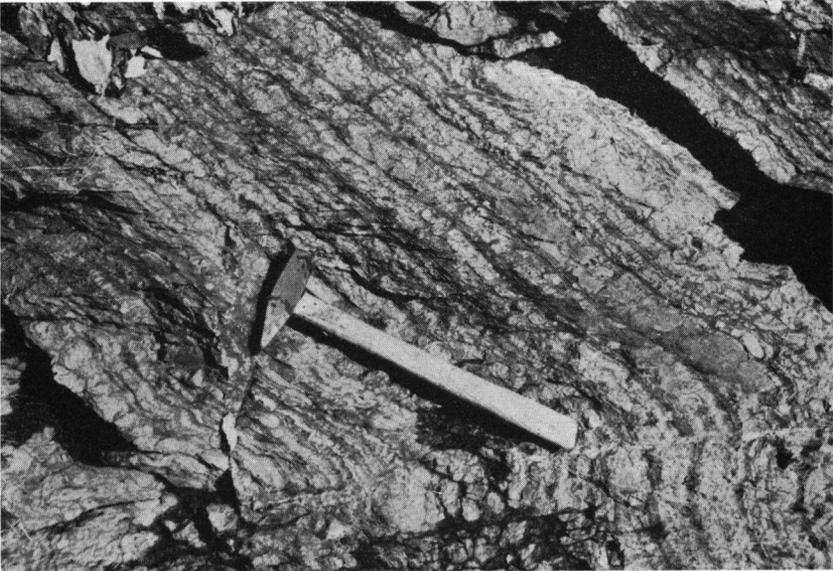


Fig. 27 Crosscutting cleavage in layered Cambro-Ordovician, near Ordino, Andorra.

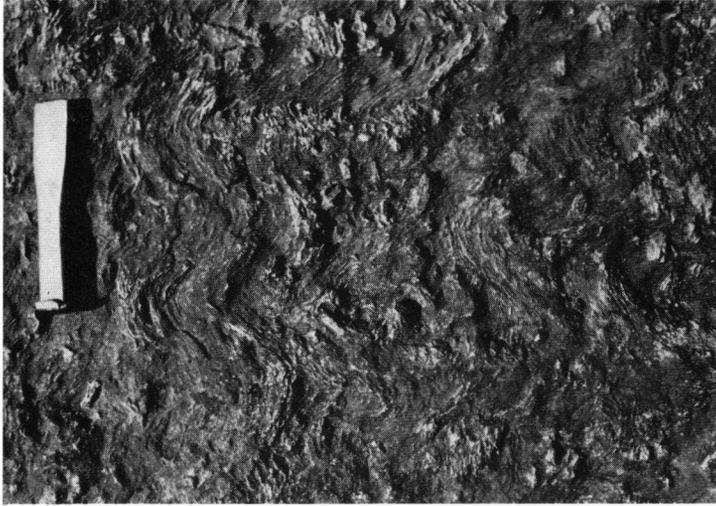


Fig. 28 Third phase folds in andalusite schist; El Serrat



Fig. 29 Silurian schist with two microfold-lineations; Llorts. (0,5 \times)

set occur scattered throughout the area, but are better developed in phyllites than in schists. In addition folds with N-S and NE-SW trending axes have been found locally. Most of these folds are accompanied by a cleavage, which microscopically proves to be a crenulation cleavage. For details I refer to Lapré.

One important fault occurs in the mapped area: the Mérens fault, which separates the Aston from the Hospitalet massif. The Aston massif is thrust up against the Hospitalet massif. The rocks north of the fault are often strongly mylonitized. In these mylonites various sets of folds occur, but they are not related to the other small scale structures in the area as schists containing these late structures are also mylonitized. South of the fault a zone with black slates and limestones occurs which can be continued to the east. They are correlated by Zwart with the Ransol formation which belongs to the Ordovician.

Small faults and macrojoints are numerous throughout the area. Their displacements are difficult to measure, but it seems unlikely that large movements are involved.

REFERENCES

- ALLAART, J. H., 1954. La couverture sédimentaire septentrionale du massif Ax-Montcalm. *Leidse Geol. Mededel.* 18, 255—271.
- 1958. The geology and petrology of the Trois Seigneurs massif, Pyrenees, France. *Leidse Geol. Mededel.* 22, 97—214.
- CAVET, P., 1958. Stratigraphie du Paléozoïque de la zone axiale Pyrénéenne à l'E de l'Ariège. *Bull. Soc. Géol. France*, (6), 8, 853—867.
- CHADWICK, R. A., 1958. Mechanisms of pegmatite emplacement. *Bull. Geol. Soc. Am.* 69, 803—836.
- DESTOMBES, J. P. & E. Raguin, 1955. Etude de la partie occidentale du massif de l'Aston. *Bull. Soc. Géol. France*, (6), 5, 101—113.
- 1960. Sept coupes à travers le massif de gneiss de l'Aston (Pyrénées de l'Ariège). *Bull. Soc. Géol. France* (7), 2, 28—37.
- DRESCHER-KADEN, F. K., 1948. Die Feldspat-Quartz Reaktionsgefüge der Granite und Gneise. Springer, Berlin, 259.
- GUITARD, G. & P. Laffitte, 1956. Sur l'importance et la nature de manifestations volcaniques dans le Paléozoïque inférieur des Pyrénées-Orientales. *C. R. Acad. Sciences*, 242, 2749—2752.
- NIGGLI, E., 1952. Stéréométrie et mise en place de filons. *Bull. Soc. Géol. France*, (6), 2, 215—223.
- RAMBERG, H., 1956. Pegmatites in West Greenland. *Bull. Geol. Soc. Am.* 67, 185—214.
- READ, H. H., 1957. The granite controversy. Murby, London, 430 p.
- ROQUES, M., 1955. Etude quantitative des myrmékites. *Sciences de la terre*, Numéro hors série. Nancy, 189—193.
- VERSPYCK, G. W., 1961. Zircons of some metamorphic and intrusive rocks from the Aston- and Hospitalet massifs (Central Pyrenees). *Geologie en Mijnbouw*, 40, 58—60.
- ZANDVLIET, J., 1960. The geology of the Upper Salat and Pallaresa valleys, Central Pyrenees, France/Spain. *Leidse Geol. Mededel.* 25, 1—127.
- ZWART, H. J., 1954. La géologie du massif du St. Barthélemy. *Leidse Geol. Mededel.* 18, 1—228.
- 1959. Metamorphic history of the Central Pyrenees, part I. *Leidse Geol. Mededel.* 22, 419—490.
- 1960. Relations between folding and metamorphism in the Pyrenees and their chronological succession. *Geologie en Mijnbouw*, 22, 163—180.
- 1965. Explanatory text to sheet 6, Aston. *Leidse Geol. Mededel.* 33.