LEIDSE GEOLOGISCHE MEDEDELINGEN, Deel 50, Aflevering 1, pp. 1-74, 1-11-1979

THE GEOLOGY OF THE CENTRAL PYRENEES

BY

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CHAPTER 1

INTRODUCTION

The present memoir is the final report of a geological mapping project which has been executed by staff and students of the Department of Structural Geology of the Geological Institute of Leiden University.

The project was initiated by professor Dr. L. U. de Sitter, to whom this memoir is dedicated. In the first instance this project was intended as a field training program for undergraduate and M. Sc. students, but at the same time many important contributions to the geological knowledge of this mountain chain were obtained by these students, by Ph. D. students and by staff members.

The first visit to the Pyrenees was made by de Sitter with a small group of undergraduates in 1948 when fieldwork was done in the Arize massif. The present author started mapping in the next year in the Saint Barthélemy massif, first as a M. Sc., later as a Ph. D. student. From then on an increasing number of students has been busy mapping during the summer months of each year, up to 1968.

After the first two years in the North-Pyrenean massifs, the so-called axial zone was tackled in the valley of the Salat in 1950, the Lez in 1951, and near St. Béat in the Garonne valley in 1952. The next year the work was extended to the Spanish side in the Valle de Arán. From then on, the mapping was mainly concentrated in the Spanish Pyrenees.

Were the initial plans for working in the unmetamorphosed Palaeozoic only, on the initiative of the present author, medium and high grade rocks were also investigated, first in the Saint Barthélemy massif, and later in the Trois Seigneurs, Aston and Hospitalet massifs, and the Bosost area.

The first maps, published in 1954 were special maps, for example of the Arize massif (Keizer, 1954), St. Barthélemy massif (Zwart, 1954), Ariège valley (Allaart, 1954), and a preliminary map of the northern part of the Central Pyrenees (de Sitter, 1954c). All these maps were on different scales, but in about 1958 it was decided to publish a series of maps on a scale 1:50,000 of the Central Pyrenees, extending from the meridian of Ax-Les-Thermes in the east to the meridian of Bagnères de Luchon in the west (Fig. 1). Altogether 10 sheets have been printed now, of which the last one, sheet 9 is for the first time published in this memoir. All other sheets have been published before, either as a map accompanying a Ph. D. thesis or as a special publication in the Leidse Geologische Mededelingen, with the exception of sheet 7 which, due to special circumstances, has been distributed on a very limited scale only. In addition two maps on 1:25,000 and one on 1:20,000 have been prepared, two of which accompanied Ph. D. theses (Zandvliet, 1960; Mey, 1968), whereas one (Roberti) has not been published before. A map of the entire Pyrenees on 1:200,000 has been compiled and is also included in this volume. Finally a set of profiles on 1:50,000 was prepared and printed.

Much of the collected material is incorporated in unpublished reports, but eight Ph. D. theses have appeared besides a large number of publications in various journals. Moreover six Ph. D. theses on sedimentological, palaeontological and petrographical subjects are also to be considered as a result of this project. Altogether more than 100 students have participated in the field work producing more than 60 M. Sc. theses. The total number of man-years spent in the Pyrenees is at present hard to estimate.

Mapping was terminated in 1968 so that the entire fieldproject took 20 years. Together with publishing this final memoir more than 30 years have gone. For most of this time the project was directed by, and has greatly profited from the enthusiasm and wide experience of professor de Sitter, without whom this project never would have started. Between 1948 and 1965 he spent each year several months in the Pyrenees, mainly checking the work of his students. From the large number of maps in this volume it can be concluded that making reliable and detailed geological maps was his great passion.

It goes without saying that in such a large time span as was required to produce these maps, the knowledge of geology in general, and that of the Pyrenees in particular, has greatly increased. This accounts for the fact that the last maps (sheets 7, 8, 9 and 10) are much more detailed than the first ones (sheets 1, 2 and 3). Also the understanding of in particular small scale structures, polyphase folding, origin of cleavages was much larger during the second half of the project, and we have no doubt that a number of important features have been overlooked on the first three sheets. Certainly, future work in these regions will result in a better understanding of the stratigraphy and the structure. The same holds true, of course for the other sheets, but we feel that our work is more up to date there. Furthermore it is needless to say that we were unable to solve a number of problems about the structure of the Pyrenees, and we

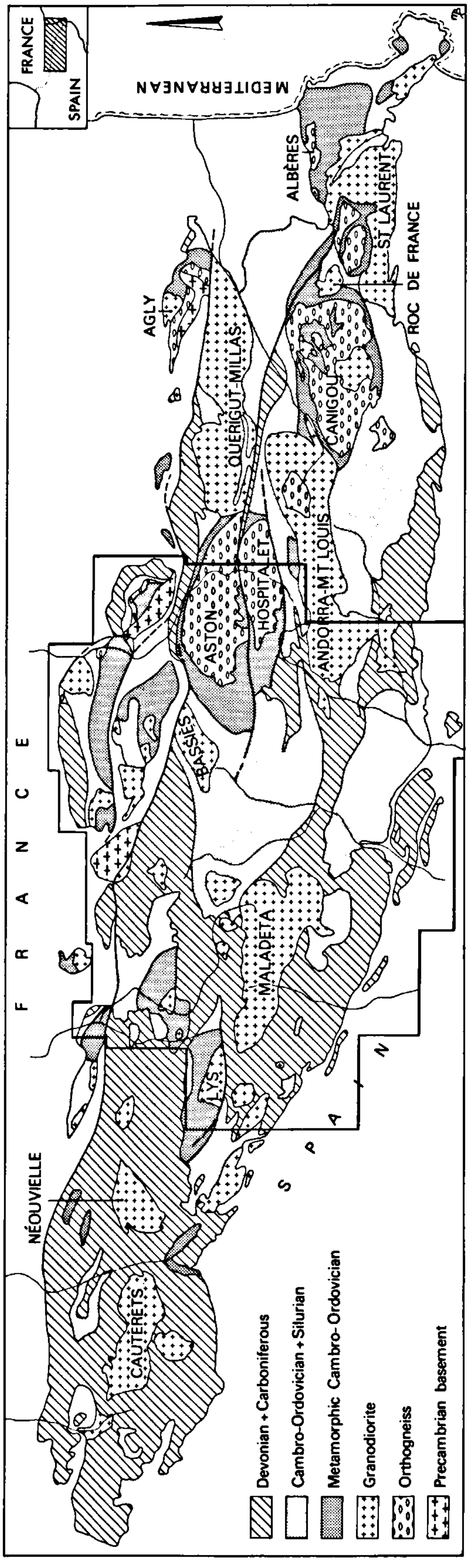


Fig. 1. Map of the Variscan Pyrenees, with outline of region treated in this publication. Major part of this map, except isolated massifs in north and south, forms axial zone.

do hope that future work will benefit from the work we have done and the maps that have been made.

The work by the Leiden group is mainly concerned with those rocks folded, metamorphosed and intruded during the Variscan orogeny, and little attention has been paid to Mesozoic and Tertiary rocks, especially on the northern side, which at that time were investigated by M. Casteras. Somewhat more work was done on post-Variscan stratigraphy and structure on the south-side, especially since much less was known there.

The study of the geology of the Pyrenees dates back already to the eighteenth century, and up to the beginning of this century several publications by various authors appeared, culminating in six large volumes by J. Carez 'Géologie des Pyrénées françaises' (1906). In the following years two names have to be mentioned, viz. L. Bertrand who tried to prove that the structure of the Pyrenees is characterized by large nappes, and M. Casteras who proved the contrary. After Worldwar II the study was intensified, mainly by three groups, a French-Spanish group with P. Cavet, G. Guitard, A. Autran and J. M. Fontboté in the eastern Pyrenees, our group from the University of Leiden in the Central Pyrenees, and a third group with M. Clin, R. Mirouse and J. Muller amongst others in the western Pyrenees. But of course many more geologists have been working in this mountain chain.

At first sight the Pyrenees look like an intercontinental mountain chain, bordered on its northern side by the Aquitanian basin and on the southern side by the Ebro basin. However, the situation is more complicated, due to the fact that the Pyrenees were built in at least two orogenic periods, the Variscan and the Alpine orogeny. At present the main body of the Pyrenees, the central zone, usually called axial zone, consists almost entirely of rocks folded, metamorphosed and intruded during the Variscan orogeny in Carboniferous times, whereas both to the north and to the south, narrow strips of Mesozoic rocks have been folded during at least two phases of the Alpine orogeny (Fig. 1). In recent years it has become evident that the axial zone has also been influenced by the Alpine orogeny. Rocks similar to those in the central zone are to be found north of the Pyrenees in the Mouthoumet, the Montagne Noire, and in boreholes in the Aquitanian basin, and to the south along the Mediterranean and elsewhere in Spain. These rocks are, however, mostly covered with epi-continental sediments of Mesozoic and Tertiary age. Consequently the Variscan Pyrenees as they stand out at present are nothing but a large uplifted block with two subsiding regions to the north and south. This uplift took place in the Tertiary. For the Alpine Pyrenees the situation is different in sofar that folding is restricted to those regions in and close to the axial zone. However, one could hardly consider the Pyrenees as belonging to the Alpine chain proper, although it has a number of characteristics of an orogenic belt.

As has been explained already, the backbone of the Pyrenees consists of Palaeozoic rocks and was mainly formed during the Variscan orogeny. The commonly used name 'axial zone' for this backbone is misleading, as it has nothing to do with the axis of the Variscan orogen. To the north this central zone is cut off by a large fault, the North-Pyrenean fault, to the north of which Palaeozoic rocks are exposed in the so-called North-Pyrenean massifs, surrounded by folded Mesozoic rocks. Although strike-slip movements on this fault may have occurred, it certainly has a large vertical component of movement, locally amounting to five or more kilometres. For this reason some of the deepest exposed rocks occur in the North-Pyrenean massifs.

The structure on the southside is even more complex. Although the axial zone there is unconformably covered by Stephanian, Permian or Triassic, faultblocks of Palaeozoic and younger rocks occur immediately to the south in the so-called Nogueras zone (enclosures: Plate 1).

CHAPTER 2

STRATIGRAPHY

INTRODUCTION

The existence in the Pyrenees of a reasonably well dated stratigraphic sequence comprising Upper Ordovician, Silurian, Devonian and Lower Carboniferous is known already for a long time. Work of the Leiden group has especially tried to unravel the litho-stratigraphy of the upper part of the Ordovician (Hartevelt, 1970), the Devonian and the Carboniferous (Mey, 1968; Roberti, internal report; Hartevelt, 1970). More recently the biostratigraphy of the Devonian has been studied with the aid of conodonts by Boersma (1973), whereas more recent work is being done by other groups.

The dated Upper Ordovician is underlain by a thick sequence of slates, phyllites, quartz-phyllites, quartzites and some marbles which has yielded no fossils. Based on comparisons with the Montagne Noire, Cavet (1957) has suggested the Cambrian to be present in this sequence. On the maps in this memoir we have used the term Cambro-Ordovician for this unit, supposing that Cambrian is present. The absence of fossils is in part due to the increase in grade of metamorphism. In many cases the Cambro-Ordovician phyllites grade into micaschists and then into migmatites, or they are underlain with a sharp contact by felsic augen- or orthogneisses like in the Canigou and Aston massifs. As metamorphic isograds cut across the stratigraphy, there is no doubt that Cambro-Ordovician rocks are present as migmatites in amphibolite facies, for example in the Trois Seigneurs and Aston massifs. In other areas the phyllites are not deeply enough exposed to see a substratum. This is the case in the Orri dome near Seo de Urgel, in the Garonne dome in the Valle de Arán, and other exposures of Cambro-Ordovician in the Central and western Pyrenees.

PRECAMBRIAN

Naturally the question arises as to the nature of the basement on which the Cambro-Ordovician was deposited. This question was first considered by the present author who in his thesis on the St. Barthélemy massif (1954) proposed that a series of high grade paragneisses forming the lowest unit in that area, constitute the basement on which the Cambro-Ordovician sediments were laid down. This conclusion was based on degree and type of metamorphism differing from the Variscan metamorphics, and a different set of structures. However, the supposed unconformity later proved to be a fault, and moreover it was difficult to explain why Precambrian rocks should have retained their original mineral parageneses and structures, whereas the overlying Lower Palaeozoic was strongly deformed and metamorphosed under amphibolite facies conditions during the Variscan orogeny.

The presence of a Precambrian basement was strongly disputed by French geologists (Guitard, 1955), and also the present author in a later publication (Zwart, 1959a) doubted his original interpretation. However, a few years later the presence of a Precambrian basement was accepted by French geologists (Guitard, 1970; Autran & Guitard, 1969) who not only considered the high grade gneisses of the Agly massif, which are similar to and occur in the same position as the paragneisses of the St. Barthélemy massif, but also the orthogneisses of the Canigou, Roc de France and Albères massifs as a Precambrian basement. Autran et al. (1966) even published an article in which they describe an unconformity between the Cambro-Ordovician and such a basement. However, this unconformity is not very convincing and can be interpreted in a different way. Anyway the preservation of an original unconformity after the strong Variscan folding and metamorphism seems not very probable.

The orthogneisses of the Canigou massif are interpreted by Guitard (1970) as the Precambrian cores of large recumbent folds, comparable to those of the Alpine Pennine nappes. However, rigorous stratigraphic proof, like in the Alps, is lacking in the Pyrenees, and according to the author Guitard's opinion is just one possible interpretation.

Recent radiometric work on these gneisses in the Canigou massif has given ages of about 535–570 m.y. and fall in the Lower Cambrian or near the base of the Cambrian (Vitrac & Allègre, 1971). Ages of similar orthogneisses in the Aston-Hospitalet massif (Jäger & Zwart, 1968) are 475 m.y. old, that is Ordovician, and conform with a large number of dates from similar gneisses throughout the Variscan belt of Europe. Therefore another interpretation is favoured by the author, namely intrusion of Cambrian or Ordovician granites in Cambro-Ordovician sediments.

A different position is taken by the high grade paragneisses occurring only in the North-Pyrenean massifs as Agly, St. Barthélemy, Trois Seigneurs, Arize and Castillon. Here we are dealing with rocks which seem difficult to place in the Variscan orogen, and the original interpretation by the writer, namely a Precambrian basement is considered to be correct. Radiometric age dating (Vitrac & Allègre, 1971) seems to confirm this.

The Precambrian of the Arize and Saint Barthélemy massifs has been called the old paragneisses (Zwart, 1954) and the basal gneiss series by de Sitter & Zwart (1959) and Zwart (1959a), to which I refer for more details. In the St. Barthélemy and Arize massifs the following succession from top to bottom is found:

1) granitic biotite-muscovite-gneiss
2) linear and folded garnet-augengneiss
3) schistose garnet-bearing granitic gneiss
4) leucocratic granite and gneissose granite.

The upper member is a fairly homogeneous, usually rather melanocratic granitic gneiss, which near its top grades into sillimanite-bearing migmatites. Amphibolite inclusions are common.

The mineralogical composition is quartz, oligoclase, potassium feldspar, biotite, muscovite, sillimanite, cordierite and garnet. The last-mentioned mineral is absent in the overlying migmatites. The lower boundary is drawn at the disappearance of muscovite, which is not a constituent of the linear garnet-augengneiss. In this gneiss the augen are formed by feldspar, cordierite or garnet around which quartz and biotite are arranged in a flow-like pattern. The microstructure is blastomylonitic. Minerals are quartz, sodic plagioclase, potassium feldspar, biotite, garnet, and occasionally cordierite, sillimanite, kyanite and orthopyroxene. The lineation is due to the elongate shape of the augen-forming minerals. Sometimes the schistosity is folded and foldaxes are parallel to the N–S directed lineations. Towards the base the feldspars of the augengneisses lose their linear and eyeshaped character, and a schistose gneiss results with the same mineralogy as the augengneisses. Downwards these schistose gneisses become more granitic, but they are still quite heterogeneous. The mineralogical composition is the same as in the overlying gneisses, although orthopyroxene is more common.

In this unit several outcrops of calcsilicates and marbles occur in which the following minerals have been found: calcite, diopside, hypersthene, grossularite, spinel, forsterite, hornblende, scapolite, plagioclase and phlogopite. Amphibolites are common throughout the basal gneiss series, and consist of calcic plagioclase, darkgreen hornblende, biotite and sometimes clinopyroxene. In the eastern part of the Trois Seigneurs massif, garnetiferous gneisses have been described by Allaart (1959). They occur within a large unit of migmatites and sillimanite-gneisses. Possibly the garnet-gneisses are comparable to the basal gneisses of the St. Barthélemy massif and may belong to the same Precambrian unit.

The largest occurrence of high grade gneisses is undoubtedly in the Castillon massif which consists solely of gneisses and migmatites. They have been described by Zwart (1959a) and more recently by Roux (1977). The rocks are mineralogically and structurally similar to those in the St. Barthélemy massif, but there is a larger variation of rock types in the Castillon massif. According to Roux there are four main rock types, viz. leptinites, cordierite-bearing granitized leptinites, kinzigitic gneisses and migmatitic gneisses. The most common minerals are quartz, plagioclase, potassium feldspar, garnet, cordierite, biotite, kyanite, sillimanite and andalusite. Locally charnockitic rocks occur with hypersthene, clinopyroxene, garnet, amphibole, plagioclase (up to 65% An), quartz, biotite and garnet.

Like in the other massifs we are dealing here with Precambrian rocks containing old, granulite facies assemblages. Roux (1977) describes a complex metamorphic history with a first phase of intermediate pressure granulite facies metamorphism, producing hypersthene-kyanite-gneisses with mafic granulites. Strong deformation produced a schistosity and N–S directed lineations. Then anatexis took place, followed by a period of recrystallization resulting in the formation of garnet.

A second phase of synkinematic metamorphism resulted in further flattening and stretching of the garnets. This metamorphism took place under amphibolite facies conditions, and is again accompanied by migmatization. This metamorphic phase involved lower pressures as shown by the development of sillimanite, cordierite and biotite. This second phase is supposedly of Variscan age.

The Agly massif has been studied by Fonteilles (1970), Guitard and Raguin (1968). The Bélesta gneiss is similar to the garnet-gneisses of the St. Barthélemy massif and contains the following minerals: quartz, plagioclase, potassium feldspar, biotite, garnet, sillimanite, cordierite, and occasionally kyanite. Amphibolite and calcsilicate layers occur in these gneisses. The principal minerals are besides hornblende and plagioclase, cummingtonite, hypersthene and clinopyroxene. Another type of gneiss is called the Caramany gneiss and consists of an alternation of fine- and coarse-grained gneisses. Layers of amphibolite and calcsilicates occur frequently. Marble layers contain clinopyroxene, garnet, scapolite, forsterite, chondrodite, spinel, phlogopite and pargasite. A small body of charnockite, near Ansignan, contains besides quartz and feldspar, hypersthene, garnet and biotite.

According to Fonteilles (1970) the gneisses are of Precambrian age, but have acquired a new Variscan mineral assemblage, with the preservation of only few relics of kyanite of the original Precambrian mineral association.

Like in the St. Barthélemy massif we are dealing here with granulite facies rocks, which are overprinted by Variscan amphibolite facies metamorphism. As the Lower Palaeozoic metasediments are metamorphosed only in the low pressure amphibolite facies, the present author believes that the granulite facies metamorphism is of Precambrian age.

CAMBRO-ORDOVICIAN

As has been indicated in the previous section, a large part of the Cambro-Ordovician is unfossiliferous, and the oldest palaeontologically dated rocks are of Caradocian age.

For the eastern Pyrenees the Cambro-Ordovician has been described by Cavet (1957), for the Central Pyrenees a comprehensive review is given by Hartevelt (1970), whereas in the western Pyrenees only very little Cambro-Ordovician is exposed.

Cavet divided the Cambro-Ordovician into two formations, a lower one, the Canaveilles series, and an upper one, the Jujols series. The Canaveilles series whose main occurrence is around the metamorphic Canigou massif consists of phyllites, quartz-phyllites and graphite-phyllites with local intercalations of limestone usually not exceeding 10 metres in thickness. These rocks are in sharp contact with the augengneisses of the Canigou massif.

According to Cavet the thickness of the Canaveilles series is at least 2000 m. However, the rocks are strongly folded and cleaved, and any estimate of their original thickness is necessarily dubious. Like in the Jujols series there are no marker beds in the Canaveilles series, the limestone intercalations are always lenticular and cannot be correlated from one locality to the other. This makes the unravelling of the structure an impossible task.

The Jujols series consists of slates and phyllites with thin (10–50 cm) quartzite layers and lacks limestone intercalations. The same observations as to folding, original thickness and lack of marker beds as in the Canaveilles series can be made.

The Jujols series is overlain by a sequence of sandy slates, often calcareous and locally with thin limestone layers. The calcareous component of the rock is often weathered out at the surface, giving rise to a rock with small holes ('schistes troués') or 'grauwacke à Orthis' of Depéret (in Cavet, 1957).

Fossils have been found frequently in this formation and have yielded an Ashgillian age. According to Cavet, volcanic rocks, called porphyrite, and possibly of dacitic nature, occur east of the Canigou massif as layers in the Upper Ordovician rocks.

In the Central Pyrenees the Cambro-Ordovician shows many similarities to the development in the eastern Pyrenees. Locally a mappable stratigraphy in the Upper Ordovician occurs. It has been worked out by Hartevelt (1970) on sheet 10, and that author has also compiled the data outside his own map area. In addition observations on the stratigraphy of the Cambro-Ordovician have been made by Zwart (1954, 1965), de Sitter & Zwart (1962), Zandvliet (1960), Kleinsmiede (1960) and Mey (1968).

Zandvliet (1960) divided the Cambro-Ordovician in the Lleret-Bayau series and in the stratigraphic higher Pilas-Estats series. The Lleret-Bayau series consists of an important limestone layer associated with black slates, quartzite layers and calcareous slates. This unit occurs in an anticlinal structure and belongs to the deepest exposed rocks on sheet 5. Zandvliet tentatively correlated this unit with the Canaveilles Formation of Cavet (1957). However, later work by Zwart (1965) has shown that the continuation of this unit on sheet 6 lies not very deep underneath the Silurian and that correlation with the Canaveilles Formation is unlikely. This unit was named Ransol Formation by Zwart (Fig. 2). According to him it is improbable that the Canaveilles Formation occurs in the Central Pyrenees. Also the occurrence near the Puymorens mine claimed by Cavet to belong to the Canaveilles Formation, is probably of Devonian age. All this is in good agreement with the westward axial plunge of the Pyrenees, which is responsible for the occurrence of the deepest exposed rocks in the eastern Pyrenees.

The Pilas-Estats series of Zandvliet consists of slates, quartzites, quartzitic slates and microconglomerates, and is comparable to the Jujols series of Cavet. Hartevelt renamed this monotonous unit the Seo Formation, which includes the Ransol Formation. We follow Hartevelt's nomenclature here.

An important limestone unit occurs in the upper part of the Seo Formation on sheets 4 and 5. In the French literature this limestone is called the 'calcaire métallifère' because of the frequent lead-zinc deposits in this limestone. Its maximum development is reached south of

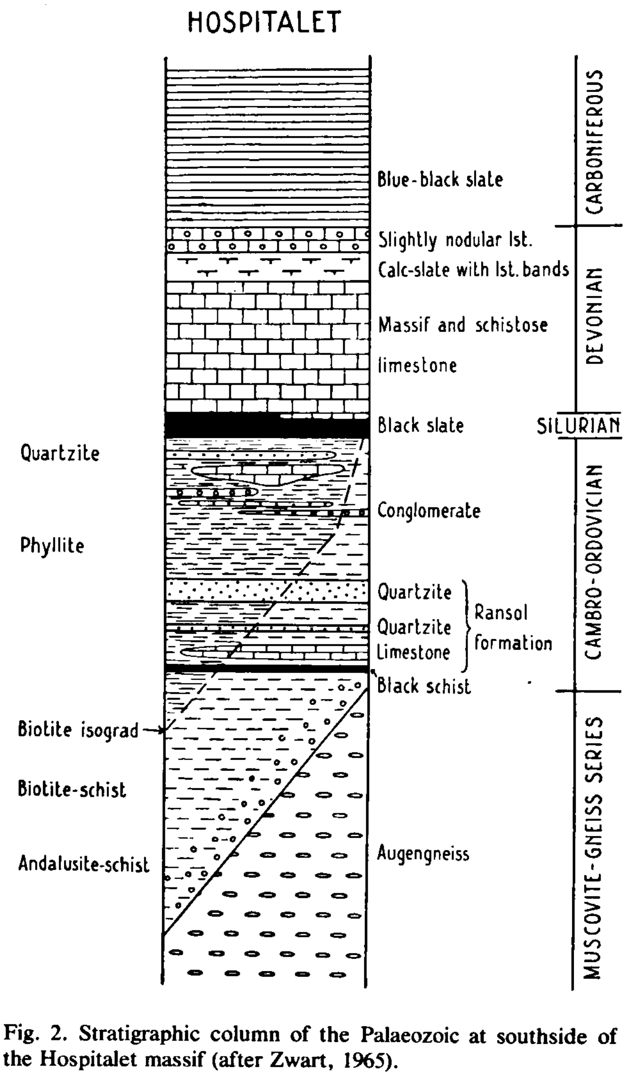


Fig. 2. Stratigraphic column of the Palaeozoic at southside of the Hospitalet massif (after Zwart, 1965).

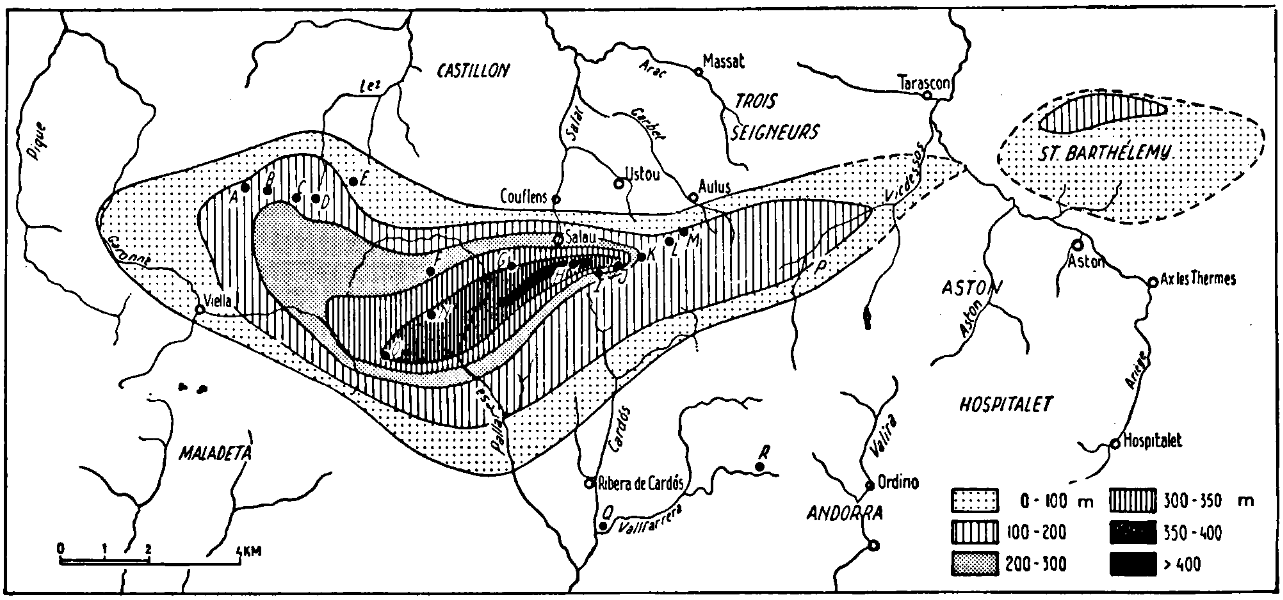


Fig. 3. Isopach map of limestone in Upper Ordovician (metalliferous limestone) (after Zandvliet, 1960).

Salau* (Fig. 3) where it is up to 400 m thick. The limestones around the Marimaña granite shown on sheets 4 and 5 as being of Ordovician age, have been reinterpreted by Hartevelt (1971) as Devonian. We have followed this reinterpretation on the 1:200,000 map. In the southeastern Pyrenees also limestones occur in the upper part of the Cambro-Ordovician (Trouw, 1969).

The bulk of the Seo Formation everywhere in the Pyrenees consists of dark-coloured or grey phyllites, quartz-phyllites, and light-coloured quartzite layers from a few millimetres to a few metres thickness.

The upper boundary of the Seo Formation is defined by the presence of a conglomerate horizon, occurring in many areas in the Pyrenees. This horizon is called the Rabassa conglomerate by Hartevelt. It consists of pebbles of quartz, quartzite, slate, and occasionally black schists or gneisses in a matrix of slate or sandy slate. The size of the pebbles varies from a maximum of 50 cm to the size of large sandgrains. It is absent in the northern part of the axial zone (sheet 3). Its thickness and pebble size increase towards the south, reaching a maximum on sheet 10, namely more than 100 m (Fig. 4).

The Rabassa conglomerate is overlain by the Cavá Formation which is only developed on sheet 10. This formation consists from bottom to top of greywackes, red and green slates, siltstones and a purple quartzite. It is possible that the greywackes are tuff-derived. On the southern part of sheet 10 the Cavá Formation may reach a thickness of more than 850 m, but it wedges rapidly out to the north, where it grades laterally into the time-equivalent black slates of the Ansobell Formation which is more widely developed in the Pyrenees (Fig. 5). Fossils (brachiopods, tentaculites, cystoids and Bryozoa) have been found and indicate an Upper Caradocian age.

The next higher unit is the Estana Formation and consists of marls and limestones. It occurs in the upper part of the Cambro-Ordovician above the Rabassa conglomerate. On the southern part of sheet 10 its stratigraphic position is defined as occurring between the Cavá and Ansobell Formations. Its thickness is limited to a few tens of metres and decreases to the north.

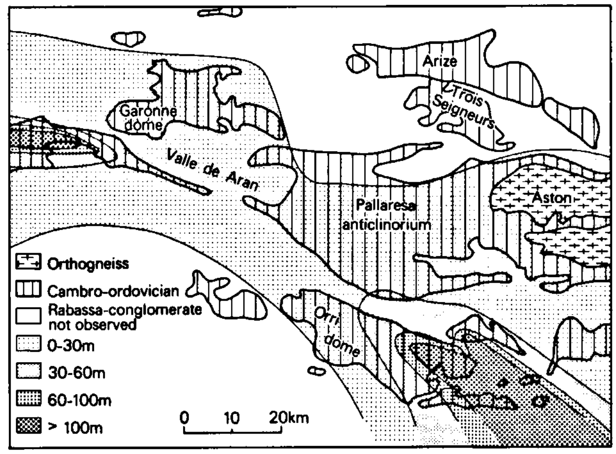


Fig. 4. Isopach map of Ordovician Rabassa conglomerate (designed by U. Dornsiepen).

* Recently Derré & Krylatov (1976) ascribed this limestone to the Lower Devonian, based on lithological comparison with the Devonian in the Valle de Aran area and on some conodonts from a drill core near Carboire. The Devonian age is not confirmed by surface samples. It may be possible that the limestone which has direct contact with confirmed Devonian, is also of Devonian age. But certain structural and cartographic difficulties arise if the whole mass of limestones in this area is attributed to the Devonian.

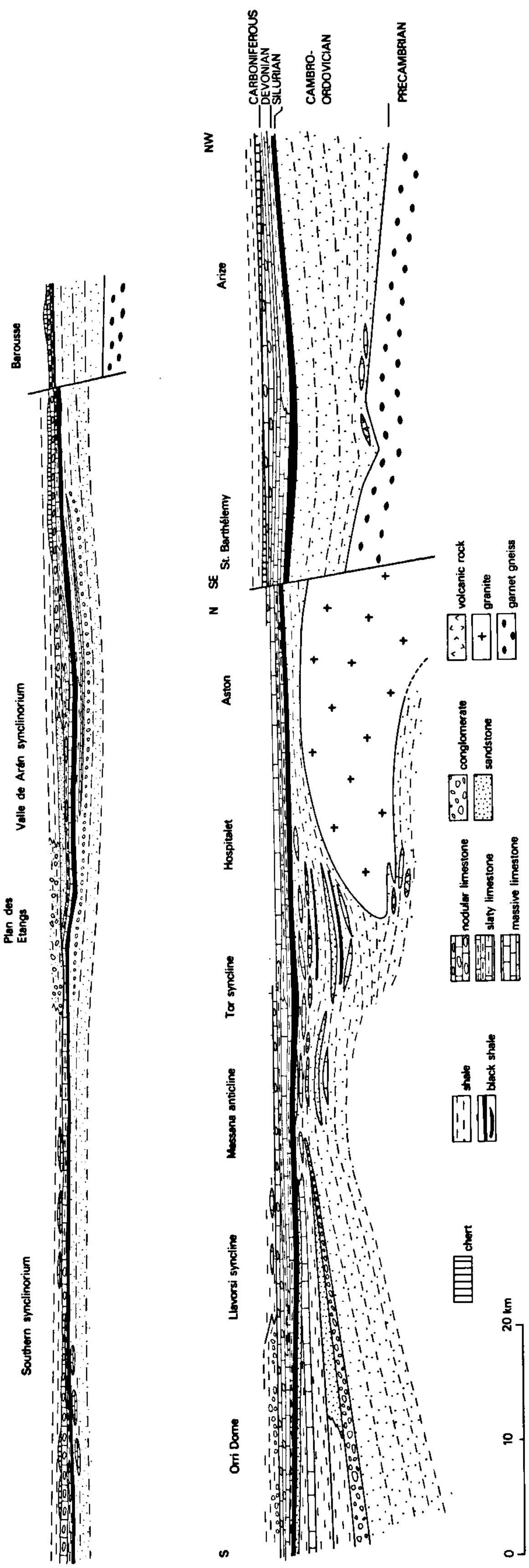


Fig. 5. Palinspastic profiles through the Pyrenees. For location see Plate 1, vertical scale 2–3 times exaggerated (designed by U. Dornsiepen).

A much thicker development occurs in the Fresser valley, outside our map area in the eastern Spanish Pyrenees (Trouw, 1969). Various fossils (brachiopods, bryozoans, conodonts) have been found and indicate an uppermost Caradocian to Lower Ashgillian age (Schmidt, 1931; Boissevain, 1934; Solé Sabaris & Llopis Llado, 1946; Hartevelt, 1970).

The Ansobell Formation, also defined on sheet 10, occurs usually directly on top of the Rabassa conglomerate, and only in the south it is underlain by the Estana and Cavá Formations. This formation consists of rather uniform black slates, locally with some sandy layers. Its thickness varies from 20–320 m. Fossils are quite rare and hence the exact age of this formation is uncertain, but probably it is Ashgillian. At most places the Ansobell slates grade into the black carbonaceous slates of the Silurian. However, on sheets 6, 9 and 10 a thin (8–18 m) quartzite unit, the Bar quartzite, is intercalated between the Ansobell Formation and the Silurian. Some fossils are found in the quartzite but no definite age can be given. According to Hartevelt (1970) an Upper Ashgillian age seems likely.

The Cavá, Estana and Ansobell Formations can be correlated with the 'schistes troués' and 'grauwacke à Orthis' of Cavet (1957) and other French authors.

Summarizing it can be stated that the Seo Formation, consisting mainly of phyllites, quartzites and quartz-phyllites was deposited during pre-Caradocian Ordovician times and possibly during part of the Cambrian. Marker beds other than of restricted occurrence are absent. Conditions of deposition were probably shallow marine. Conditions apparently changed towards the end of the Ordovician, as indicated by the widespread occurrence of the Rabassa conglomerate, which according to Hartevelt (1970) is at least partly a mudflow deposit, originating in the south. Although the conglomerate indicates less stable conditions with higher relief and emergence, the succession is always concordant. Consequently there is no relation with a Caledonian folding phase, as sometimes is reported in the literature.

Conditions of sedimentation remained variable till Silurian times, with marine, fluviatile and volcanic influences. The total thickness of the Cambro-Ordovician is unknown in the axial zone as the basement is not exposed. A minimum thickness of about 2000 m seems indicated.

SILURIAN

The Silurian epoch in the Pyrenees is characterized by a very uniform development of black, carbonaceous shales, which in the field are easy to recognize. It thus divides the Palaeozoic into a lower Cambro-Ordovician, and an upper Devono-Carboniferous part. Although the Silurian is usually very thin, it has due to its softness and incompetent behaviour in deformation a profound influence on the topography and on the tectonic style of the folds, where it may cause decollement and diapiric structures.

When fresh, the Silurian consists of black, staining and often strongly contorted slates. On weathering the rocks become rusty brown, yellow and white due to the oxidation of pyrite and the occurrence of sulphur or alum. Small streams are often iron stained. Near the top a black, fossiliferous, sometimes nodular, limestone has been found frequently.

Chemical analyses of the black slates show a high aluminium content with Al_2O_3 up to 35%, a low SiO_2 content, varying between 40 and 60%, and a carbon content varying from 0.3 to 6%.

As a result of strong folding, flattening and contortion, the original depositional thickness of the Silurian remains uncertain. At some places its present thickness is reduced to almost zero, and in other places large accumulations of Silurian slates occur. A maximum thickness of 200–250 m has been assigned to this system by de Sitter (1954) and Hartevelt (1970).

Fossils have been found at many localities. In the black slates the fauna consists almost exclusively of various species of *Monograptus*. In the limestones *Orthoceras* spec. and *Cardiola interrupta* are recorded from different places. In Andorra a *Scyphocrinus* spec. was discovered, indicating an uppermost Silurian age. The graptolites have given Llandoverian, Ludlovian and Wenlockian ages, and seen over a large area the whole of the Silurian seems to be present. However, due to strong tectonization nowhere a complete and uninterrupted sequence of the Silurian has been recorded. For more palaeontological details I refer to Zandvliet (1960), Kleinsmiede (1960), Zwart (1965) and Hartevelt (1970).

The facies of the Silurian indicates very quiet and slow sedimentation in a marine euxinic environment.

The Silurian from the eastern and western Pyrenees is essentially the same, as described here. Reference can be made to the publications of Cavet (1957) and Mirouse (1966).

DEVONIAN

In contrast to the Ordovician and Silurian, the Devonian system is very variable throughout the Pyrenees. Thickness and facies changes, both along and across the strike of the Pyrenees, are common, and although a great deal of stratigraphical and palaeontological work has been done during the last two decades, we are still far from a complete synthesis. Major contributions have been made by Mirouse (1966) and Mey (1967a, b, 1968). The latter author proposed four facies areas for the Devonian and Carboniferous, a southern, a central, a northern and a western facies area. Some of these have been subdivided into subfacies areas. Another important paper is from Boersma (1973) who gave Mey's lithological classification a sound palaeontological footing. It is unfortunate that not more work, especially on conodonts, has been done in the remainder of the area mapped by the Leiden group. Although a lithological division of the Devonian has been made, the scarcity of microfossils in the Central and northern Pyrenees makes any assignment to the biostratigraphic time scale necessarily dubious. Also more detailed mapping is responsible for a better knowledge of the Devonian of the southern part with regard to the central and northern areas. However, an attempt to a correlation will be made here.

An E–W directed zonation into facies areas, as has been proposed by Mey (1968) is applied here with some modifications (Fig. 6). The southern facies area, extending from the eastern Pyrenees (Fresser valley) to about 10 km west of Benasque in the Esera valley, and subdivided in two of Mey's four subfacies, remains unchanged after this author.

The stratigraphy of the central facies area has been modified considerably as a result of new work by Krylatov & Stoppel (1969, 1971) who discovered that the Las Bordas sandstones, described by Kleinsmiede (1960) occur further west in an area south of Col de Peyresourde and SE of Laruns in the western Pyrenees, the so-called Agudes and Sia Formations. In addition Krylatov and Stoppel have on the basis of conodonts been able to date these rocks as Frasnian and possibly Lower Famennian, but Corsin et al. (1973) give palaeontological evidence that part of the Sia Formation is Carboniferous. This formation then occurs along strike of the Pyrenees over a distance of at least 100 km, and like the southern facies area clearly testifies to the parallelism of the sedimentary facies to the large scale structure.

On the 1:200,000 map the formations of Sia and Agudes are shown as Carboniferous, as the discoveries of Krylatov and Stoppel were partly made after the drafting of this map.

The northern facies area comprises the Devonian and Carboniferous of the axial zone north of the central facies area, and a fourth North-Pyrenean facies area only occurring in the North-Pyrenean massifs can be identified. Further westward, but south of the central facies area, the Devonian is different from the southern facies. Consequently for this region Mey's division in a western and southern facies is retained.

North-Pyrenean facies area (Fig. 6)

Devonian and Carboniferous rocks occur almost exclusively in the Arize and St. Barthélemy massifs. In the Arize massif the Devonian consists from bottom to top of 80–100 m grey calcslates and limestones of probably Lower Devonian age; 30–120 m blue limestone and 100–110 m grey dolomite, belonging to the Middle Devonian, and 60–70 m calcslate and 100–120 m nodular limestone of Upper Devonian age according to Keizer (1954). The lowest part of the Devonian in the St. Barthélemy massif is unfossiliferous and consists of 200–300 m slates and calcslates. These rocks are covered by a blue limestone correlated with the dated Middle Devonian of the Arize massif. The contact with the nodular limestones of probably Frasnian or Famennian age (Mangin, 1967) is always faulted so that a continuous section of the Devonian does not exist. The nodular limestones have a thickness of about 180 m. Some volcanic rocks, spilites and keratophyres have been reported from the eastern part of the St. Barthélemy massif (Zwart, 1954; Krylatov, 1964).

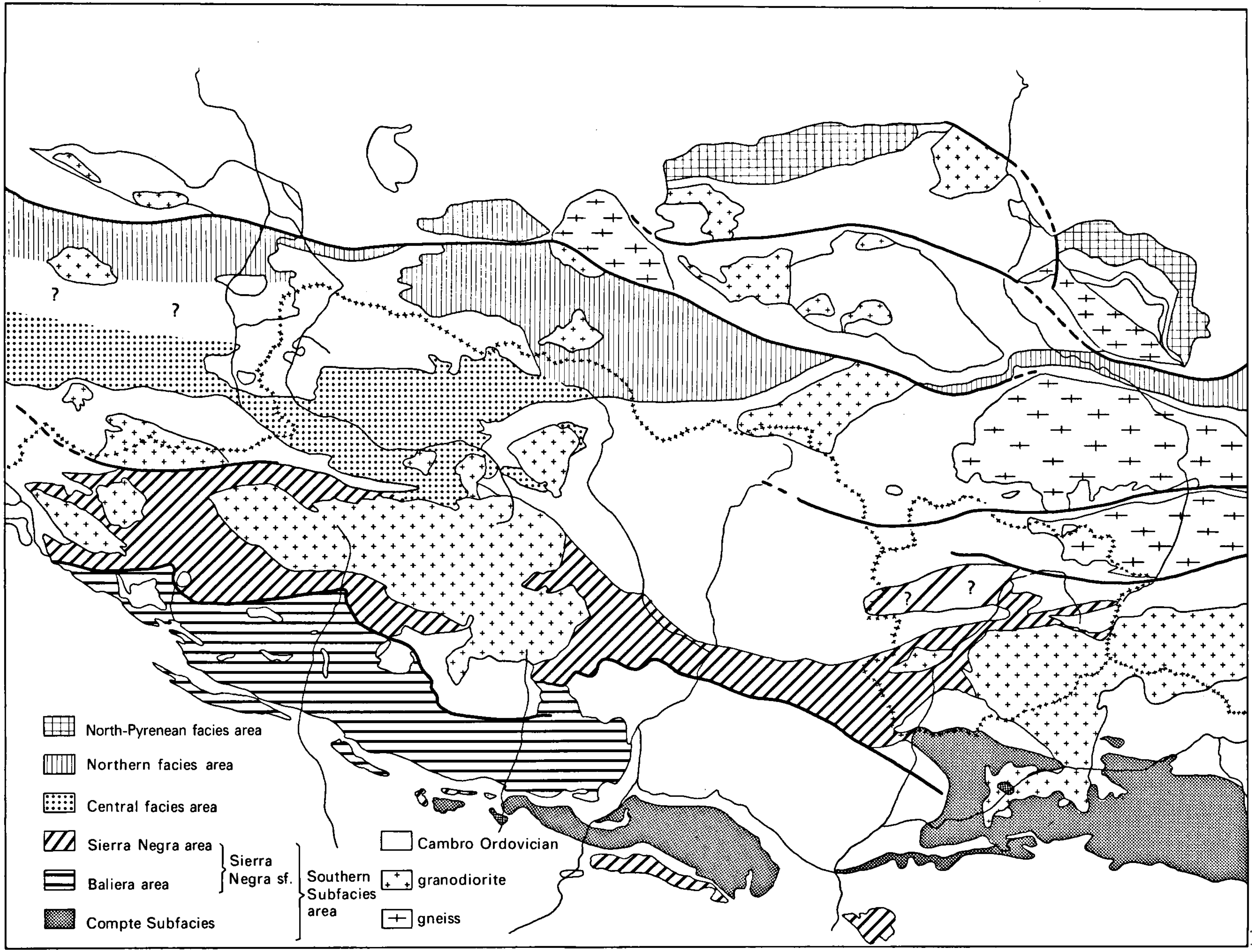


Fig. 6. Map of Devonian facies areas in the Central Pyrenees.

According to Mangin (1967) the nodular limestone is covered by 30 m breccious limestone of Upper Famennian age. The total thickness of the Devonian in the North-Pyrenean facies area is quite small and does not seem to exceed 500 m, although Zwart (1954) arrived at slightly higher figures.

Northern facies area (Fig. 6)
The following units belong to this area: Villefranche syncline in the eastern Pyrenees, the cover of the Aston massif, the area around the Riberot granite, and the area south of the Barousse massif.

The Devonian of the Villefranche syncline, outside our map area, has been described by Cavet (1957). According to his data the Devonian resembles that of the North-Pyrenean facies area. The lower part of the Devonian, often with tectonic intercalations of Silurian consists of calcslates and thinly bedded grey limestones. The Middle Devonian, dated by trilobites as Eifelian, consists of rather monotonous grey limestones and dolomites up to 400 m thick. The Upper Devonian (Frasnian–Lower Famennian) occurs again as red or grey nodular limestones (griotte) and variegated calcslates. Its thickness does not exceed 100 m.

Both, thickness and facies of the Devonian are rather similar to those in the St. Barthélemy and Arize massifs.

The Devonian of the remainder of the northern facies area has thus far yielded only very few fossils, and consequently the division is a lithostratigraphic one.

On the northern side of the Aston massif the Devonian consists mainly of calcslates and black slates with some sandy layers, often with intercalations of limestones. Grey or white limestones of a thickness of several hundreds of meters occur in the calcslates and can possibly be correlated with the Middle Devonian of the North-Pyrenean zone.

Nodular limestones occur only at one place near Lordat. Due to strong, almost isoclinal folding and flattening with the developments of cleavage, the thickness of the Devonian is difficult to estimate. Possibly it was not much more than 500 m.

The Devonian of the Salat area has been mapped and described by Zandvliet (1960). Above the Silurian an alternation of 30–160 m thick blue or brown-grey limestones and dark grey and blue slates has been mapped by him. Five or locally six limestones are present. The uppermost of these limestones is followed by the well-known Upper Devonian nodular limestone. The total thickness has been estimated at 1400 m. According to Zandvliet the fourth and fifth limestone may be correlated with the Middle Devonian limestone of the Arize massif. However, lack of fossils makes this correlation not very reliable.

North of the Silurian anticline of Couflens and around the granite of Riberot, the Devonian is somewhat different and less complex than in the Salat region. The Devonian starts with dark slates, grey calcslates with some limestone intercalations, banded sandwich limestones and siltstones. Then follows an important limestone horizon, which may be equivalent to the Middle Devonian of the Arize massif and to one or more of Zandvliet's limestones. The next unit is a dark grey, blue or black slate, similar to the slates between the limestones of the Salat area. The Devonian terminates with the nodular limestones of Upper Devonian age. They are well developed in this region and have been exploited in some quarries, like near Estours. The development of the Devonian in this region is again rather similar to that of the Arize and St. Barthélemy massifs.

Further west, in the area of the Pique River the Devonian starts with black and grey slates and calcslates, followed by blue-grey slates with an Eifelian trilobite fauna. The next unit is a limestone with a thickness of 30–50 m possibly also of Middle Devonian age. The Upper Devonian is represented by variegated shales and limestones, and a nodular limestone with a Famennian fauna (Destombes, 1953).

Central facies area (Fig. 6)
The development of the Devonian in the central facies area is strongly different from the northern area. As type locality the Valle de Aran section described by Kleinsmiede (1960) can be used. The Devonian here starts with a limestone with some chert and sandstone intercalations at the base, and some slate intercalations near the top. Its thickness is about 100 m and is followed by a sequence of blue and black slates with thin limestone layers. The next higher unit is quite important as it is typical for the central facies. Contrary to most of the Devonian in the Pyrenees it is a sandy sequence, starting with sandstones and slates, a quartzite, some non-graded sandstones, and then a fairly thick (200 m) unit consisting of graded sandstones, called the Las Bordas sandstones. At the top a thin green slate unit has been found. Absence of fossils has thus far prevented palaeontological dating.

However, further west Krylatov & Stoppel (1969, 1971) have found the same unit near Col de Peyresourde and between Laruns and Gavarnie. With the aid of conodonts they have been able to date this sequence as Frasnian and Lower Famennian. In that area a thin layer of Upper Devonian nodular limestone has been found. Further east of the Valle de Arán the sandstone sequence has not been found. As has been stated by Krylatov and Stoppel the occurrence of this sandy unit has some important palaeogeographical implications. One is that the central part of the Pyrenees, north of the central anticline, has not emerged during the Upper Devonian. Another implication is that the occurrence of a basin, closely parallel to the strike of the Pyrenees, seems to be confirmed (Fig. 12).

The total thickness of the Devonian in the Valle de Arán varies from 500 metres to a maximum of 1000 m.

South of the Valle de Arán synclinorium in which the Las Bordas sandstones are found, the central anticline with an Ordovician core separates this synclinorium from the Plan des Étangs area, again with Devonian and Carboniferous. The Devonian here consists solely of a limestone, probably equivalent to the basal limestone of the Valle de Arán region. It occurs as a white marble, due to contact metamorphism of the Maladeta granodiorite. It is covered with a disconformity by dark-coloured greywackes, ususally rather coarsegrained and micaceous, and containing pebble beds. Probably a large part of the Devonian is missing, or present as a condensed sequence, and an emergence during the later part of the Devonian in this area seems to be indicated. It is clear, however, that there is no angular unconformity between the Carboniferous and the Devonian, and there is no intra-Devonian folding phase.

Southern facies area (Fig. 6)

The southern facies area has been described by Mey (1968), Hartevelt (1970) and Boersma (1973), and is relatively better known than the more northerly area. This is due to a well recognizable lithostratigraphy, better mapping and more palaeontological work, mainly on conodonts (see also legend sheet 9).

Mey (1968) has made a division in four subfacies areas, which later has been amended by Boersma (1973). I shall follow the suggestions of the last-mentioned author, and only retain a division in two subfacies: the Sierra Negra and the Compte subfacies (Fig. 7).

In the Sierra Negra subfacies, which includes Mey's Baliera and Renanué subfacies, the following formations from bottom to top have been recognized: the Rueda Formation, locally subdivided in the Aneto and Gelada members, the Basibé Formation, the Fonchanina Formation and the Mañanet griotte Formation.

The difference between the Sierra Negra area and the Baliera area lies mainly in the thickness of the Devonian. In the first named area the Devonian is 200–270 m thick, in the last named area up to about 750 m. In the Baliera area a quartzite unit, the San Silvestre member occurs in the Basibé Formation which lacks in the Sierra Negra area (Figs. 8, 9). There is also a difference in structure. The Baliera area is characterized by large, NE–SW trending early folds, overprinted by E–W to ESE–WNW striking cleavage folds. The early set of folds is absent, or at least much more inconspicuous in the Sierra Negra area. The Sierra Negra subfacies occurs south of the Maladeta granodiorite, in the Llavorsi syncline and in the Nogueras zone (Las Paules block, Gotarta-Malpas block, Las Iglesias-Sta. Coloma block, Castells block and the Montsech de Tost, Fig. 9, enclosures: Plate 1).

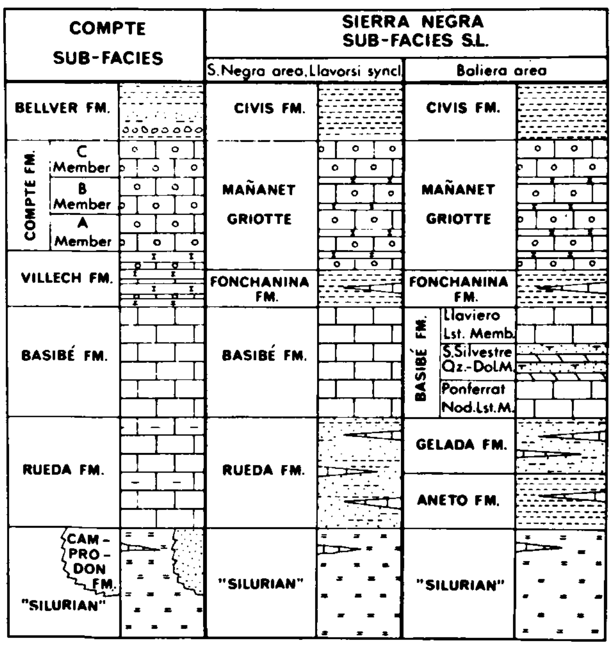


Fig. 7. Lithostratigraphy of Compte subfacies and Sierra Negra subfacies s.l. (after Boersma, 1973).

The Aneto member of the Rueda Formation (which does not occur everywhere) consists of dark slates with argillaceous limestones and thin calcslates.

The Gelada member contains calcareous quartzwackes, siltstones, argillaceous limestones and slates (Fig. 10). The thickness of the formation is about 200 m. According to Boersma (1973) the age of the formation is Gedinnian based on conodonts (see sheet 9).

The Basibé Formation is a characteristic and well mappable unit, as it consists of a conspicuous, well-exposed limestone. In the western part of the area on sheet 8 it contains a quartzite-dolomite member described by Habermehl (1970), the San Silvestre member (Figs. 8, 11). The thickness of this formation varies from 20 to 175 metres with the quartzite member up to 60 metres. The age of the Basibé Formation is Siegenian-Lower Emsian (Boersma, 1973).

The Fonchanina Formation consists of well-cleaved dark-coloured slates with thin limestone layers, and is up to 70 m thick. Conodonts indicate an Upper Emsian-Eifelian age.

The Mañanet Griotte Formation consists of multicoloured, often red or green, nodular limestones with intercalations of slaty limestone. It has been recognized not only in the southern facies area, but also, as apparent from the previous description, in the central and northern facies areas (Fig. 13). The thickness of this formation varies from about 40 to 300 m. According to Boersma (1973) the Mañanet Formation of the Sierra Negra subfacies area was deposited in Givetian, Frasnian and Lower Famennian time.

The Compte subfacies area (Mey, 1968) occurs south of the Llavorsi syncline in the region from the Andorra granodiorite to the valley of the Fresser in the eastern Pyrenees, and in the Erdo and Feixa blocks of the Nogueras zone (Fig. 6, Plate 1).

The lower part of the Devonian in the Compte subfacies area is rather similar to that of the Sierra Negra area.

The Rueda Formation, in which no longer the two members can be recognized and consisting of dark argillaceous limestones with thin calcslate layers, and the Basibé Formation with massive dark limestones and thin calcslate layers, occur also in the Sierra Negra subfacies. The next higher unit is the Vilech Formation, and consists of reddish slates, calcslates, limestones with locally greenish slates, nodular limestones and yellow limestone layers. Its thickness lies between 80 and 125 m. The age of the Vilech Formation as based on conodonts (Boersma, 1973) is Upper Emsian and Eifelian. The Vilech

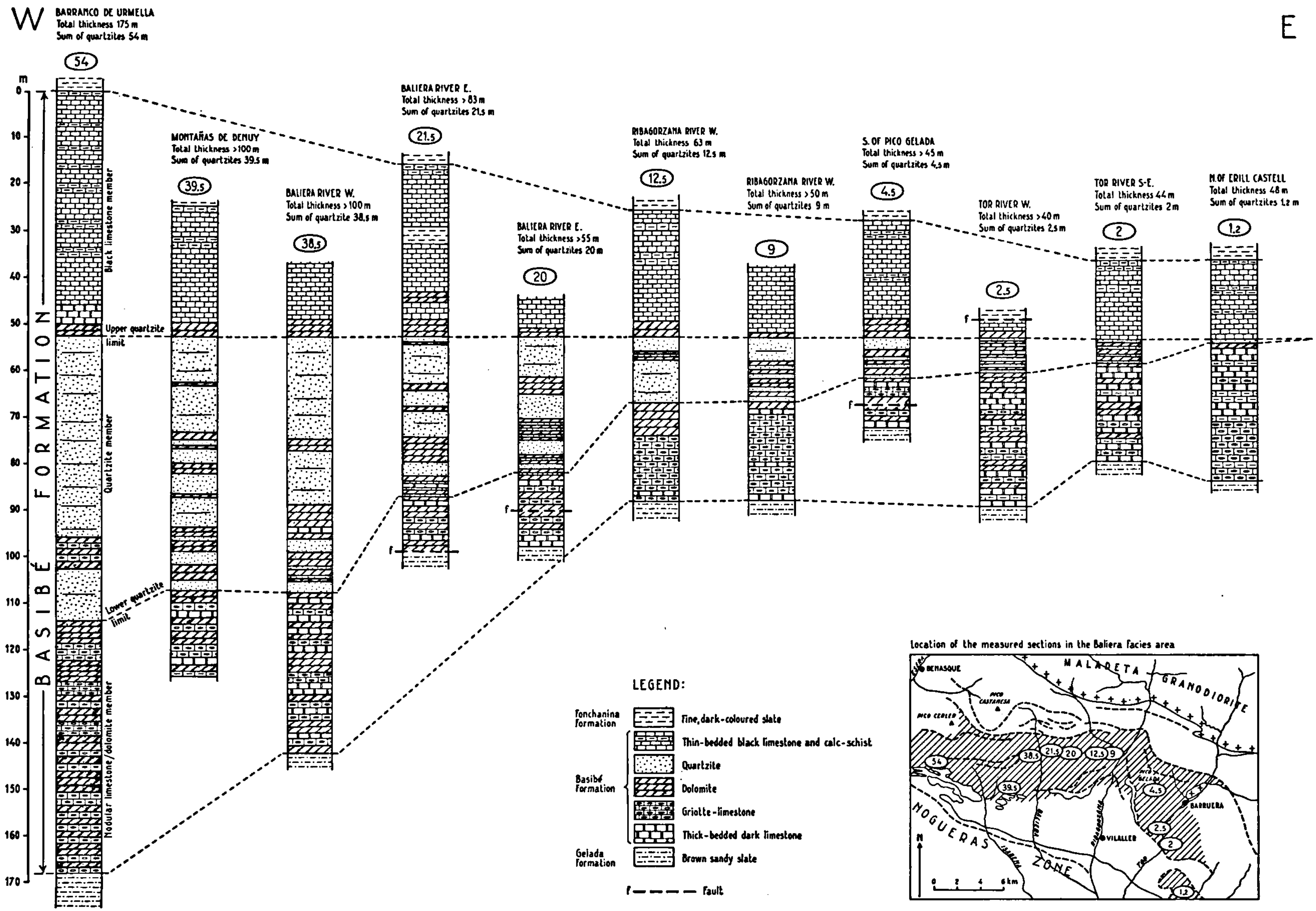


Fig. 8. Lithostratigraphic sections of Devonian Basibé Formation (after Mey, 1967).

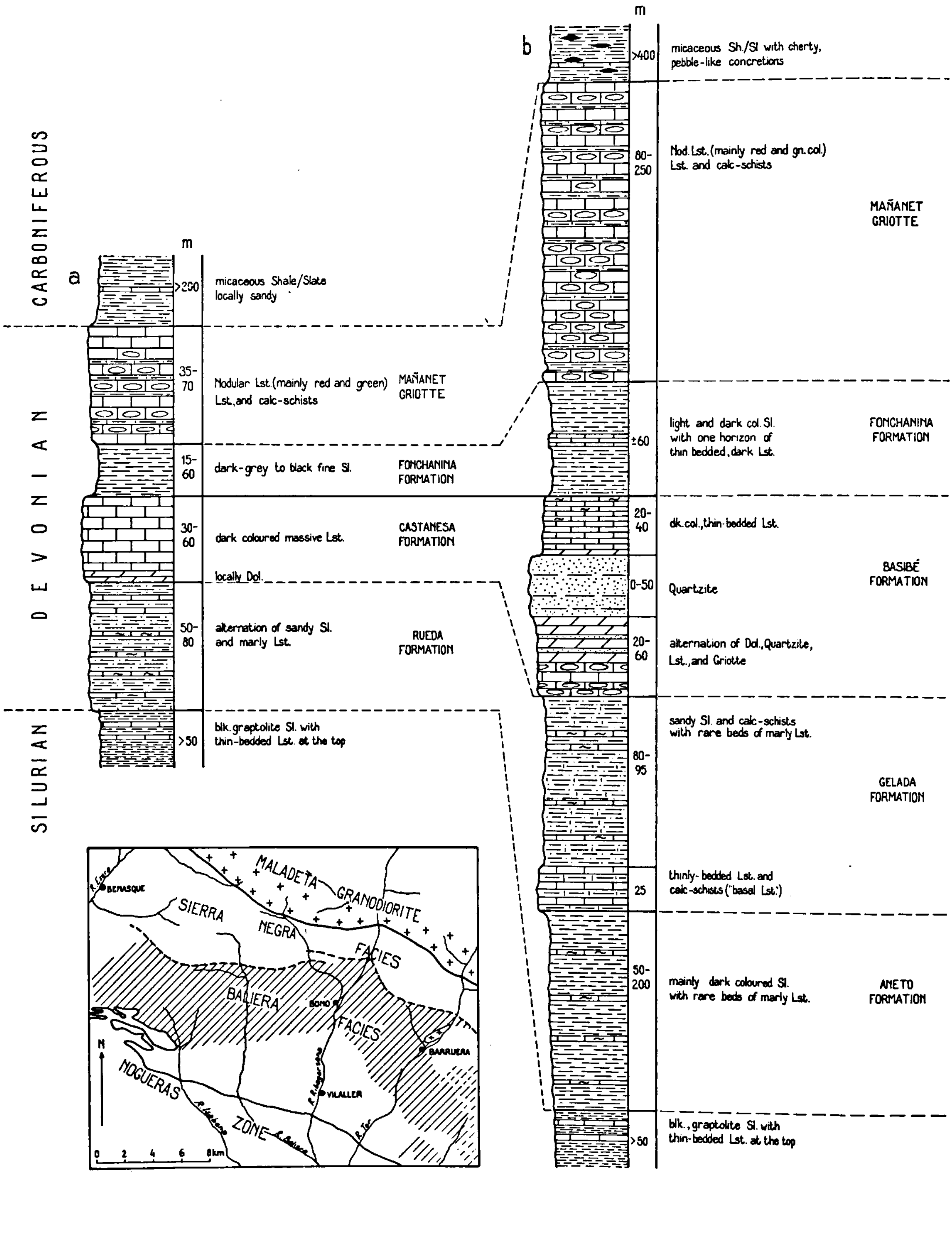


Fig. 9. Lithostratigraphy in Sierra Negra area (left column) and Baliera area (right column) of Sierra Negra subfacies (after Mey, 1967).

Formation is followed by the Compte Formation which consists of three members. The oldest member is formed by light grey, sometimes nodular limestones, in the middle member occurs a dark red nodular limestone, and the top member is again a light grey limestone. Thickness of the formation lies between 80–200 m. According to Boersma (1973) the lower member has a Givetian age, the middle member a Frasnian age, and the upper member a Famennian age. Consequently it can be correlated with the Mañanet Formation.

PRE-VARISCAN CARBONIFEROUS

The Carboniferous in the Pyrenees will be treated in two sections, viz. those deposits taking part in the major Variscan orogeny, and those deposits occurring unconformably on the folded Palaeozoic. The Dinantian, Namurian and Lower Westphalian participated in the major orogenic movements and are referred to as pre-Variscan Carboniferous, whereas the Upper Westphalian and Stephanian are post-Variscan, although not wholly unaffected by some tectonic movements. Like the Devonian, the pre-Variscan Carboniferous occurs in certain facies areas which follow those of the Devonian. Three of such areas can be recognized, a northern facies area including the North-Pyrenean massifs and the northern part of the axial zone, a central facies area south of the central anticline, and a southern facies area, which can be split in two subfacies, the Sierra Negra and the Compte subfacies.

In the northern facies area, sediments of Carboniferous age are mainly restricted to the Dinantian and Lower Namurian. The sequence begins sometimes with nodular limestones, much like in the Upper Devonian, but more often the lower part of the Carboniferous is characterized by a black chert layer of a few metres to some tens of metres thickness. Interlayering of chert and nodular limestone is also reported. This chert horizon carries occasionally Radiolaria, and more often phosphate nodules and manganiferous layers. According to new palaeontological data (Bouquet & Stoppel, 1975) the often described hiatus in the Tournaisian seems to be non existent as one is dealing with a condensed sequence. In this case the lowermost Tournaisian consists of nodular limestones whereas the cherts where deposited during the remainder of the Tournaisian. The cherts are followed by a rather monotonous sequence of shales and siltstones which has been dated as Visean at several localities (Dalloni, 1930; Keizer, 1954). The Ardengost limestone just west of the Garonne valley and outside our map area is of Upper Visean to Lower Namurian age. To the east of the map area, conglomerates, carrying pebbles of gneisses, have been found. They seem to be comparable to those of the Bellver Formation.

In the central facies area the Carboniferous of Plan des Étangs in the upper Esera and Artigua de Lin valleys north of the Maladeta granodiorite has been studied by several geologists (Dalloni, 1910; Schmidt, 1931; Kleinsmiede, 1960, and Waterlot, 1965, 1969). In this area the Devonian basal limestone is directly overlain by a sequence of thick-bedded micaceous, coarse-grained greywackes, grading into fine- to coarse-grained conglomerates. Fine-grained greywackes and slates are in the minority. Most of these sediments are dark coloured and contain abundant but badly preserved plant fragments. Determinations by Schmidt (1943) and Waterlot (1969) indicate a Namurian to Lower Westphalian age. The contact with the underlying Devonian has been described as an unconformity by Kleinsmiede (1960), but he probably meant it to be a disconformity, as there are no signs of folding of the Devonian before the deposition of the Carboniferous. There is, however, no palaeontological evidence for a large hiatus between the Devonian and Carboniferous. The Carboniferous rocks of the Plan des Étangs area occur in a syncline (Profile 6) which in the east is cut off by granite bodies associated with the Maladeta batholith. East of this large gra-

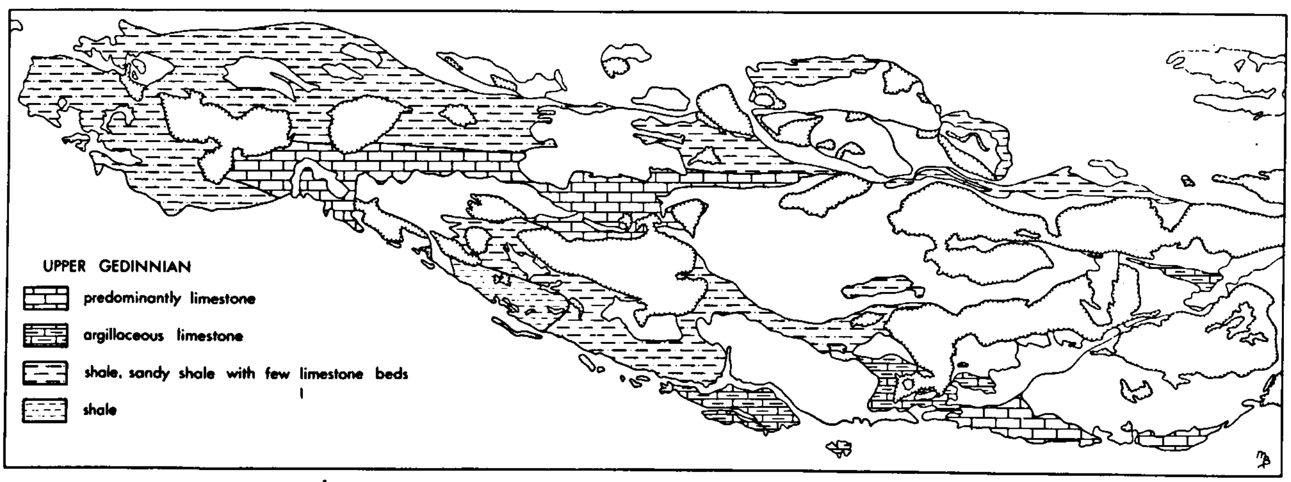


Fig. 10. Qualitative litho-facies map of Gelada Formation, Gedinnian (after Boersma, 1973).

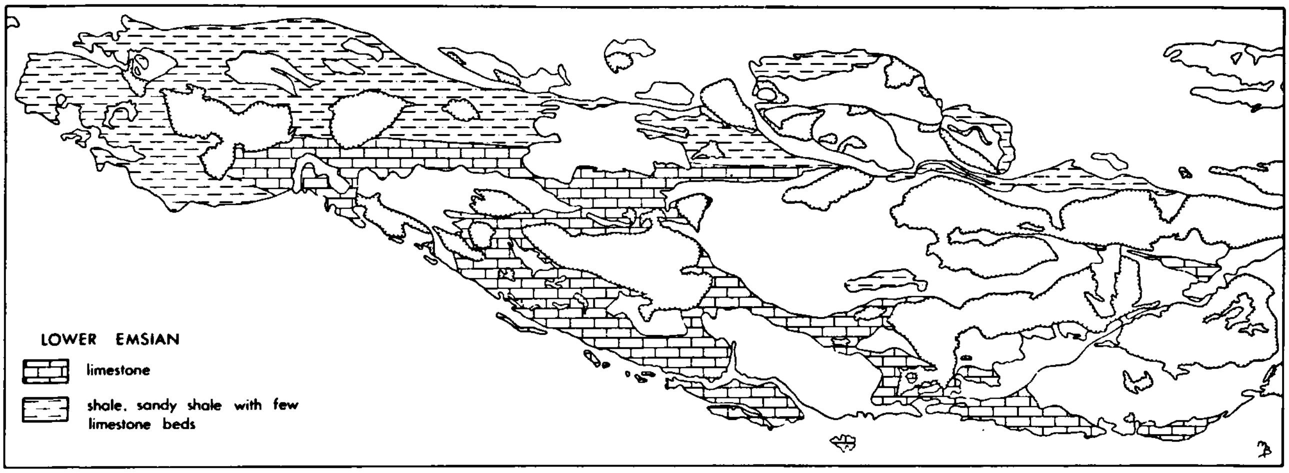


Fig. 11. Qualitative litho-facies map of Basibé Formation, Lower Emsian (after Boersma, 1973).

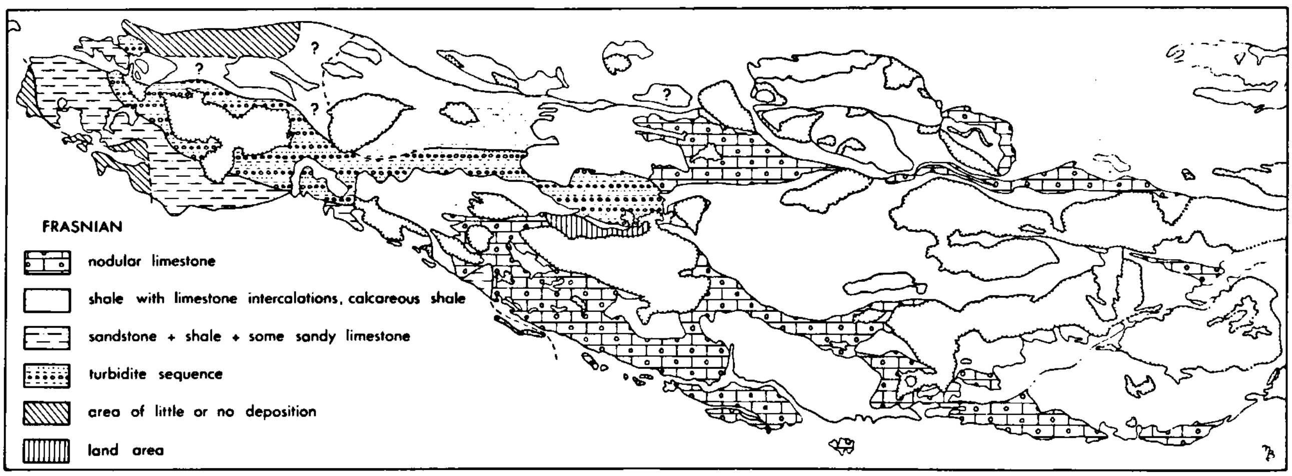


Fig. 12. Qualitative litho-facies map of the Frasnian (a.o. Las Bordas sandstone and Sia Formation, after Boersma, 1973).

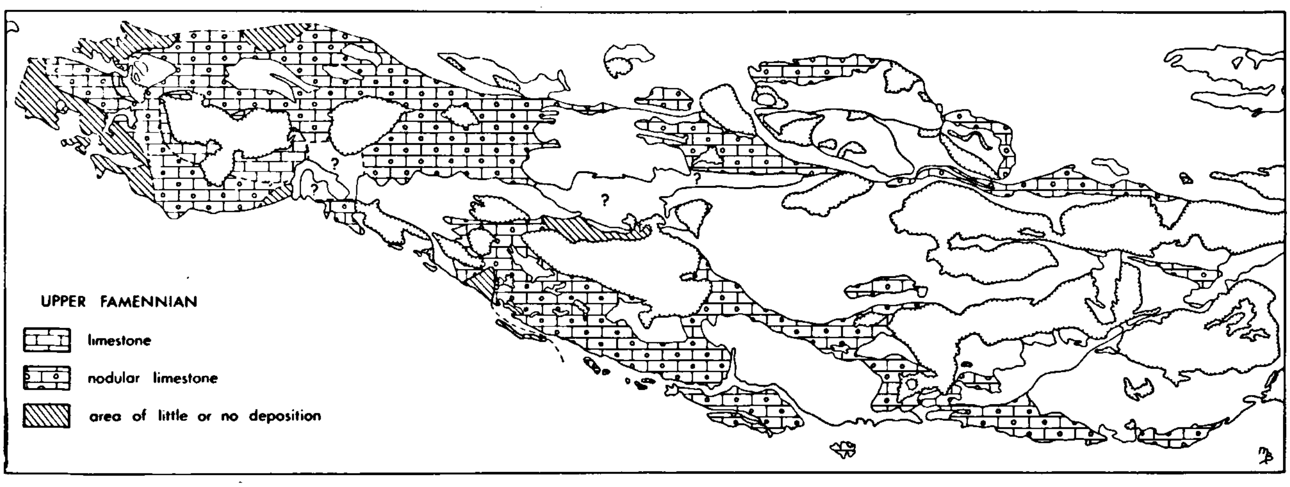


Fig. 13. Qualitative litho-facies map of Upper Famennian (Mañanet Formation, after Boersma, 1973).

nite appears another syncline with Carboniferous in the core: the Llavorsi (also Tirvia-Espot) syncline. Despite the apparent continuity, there is little reason to assume a connection between the two synclines, as the lithologic differences are rather large.

The Carboniferous of the Sierra Negra subfacies area is referred to as the Civis Formation. Like in the central facies area the chert horizon at the base is absent, and the Devonian nodular limestones are directly overlain by a sequence of dark grey to black, micaceous, often chloritoid-bearing slates in the western part of the Llavorsi syncline. Towards the east the sequence becomes more sandy, but still contains clastic micas and new chloritoid. In Andorra impure limestones occur in the slates, in which some fossils, tentaculids, gastropods and crinoids, have been found by Hartevelt (1970). Traditionally the clastic rocks of the Civis Formation have been considered as belonging to the Carboniferous. However, due to the scarcity of fossils, the age of the Civis Formation is uncertain. Groos-Uffenorde et al. (1972) and Buchroithner (1976) have shown that the Devonian directly adjacent to the slates is of Middle-Lower Devonian age, and they propose that the Civis Formation should also be Devonian. In the Civis Formation outside the Llavorsi syncline, near Aguiró in the Flamisell valley, limestones near the base have yielded Devonian fossils (tentaculids, conodonts, ostracods, Groos-Uffenorde et al., 1972). Together with the tentaculids found by Hartevelt (1970) there is therefore some evidence that at least the base of the Civis Formation is Devonian. However, the strong folding and shearing especially at the contact of limestones and slates, prevent in many cases to sample undisturbed profiles, and the proof that all clastic rocks of the Civis Formation are Devonian has as yet to be given. It is assumed here that the major part of that formation is Carboniferous. The Civis Formation occurs furthermore along the southern border of the axial zone, in the Flamisell valley, in the Ribagorzana valley near Villaller, and in the Esera valley near Benasque. The rocks are rather similar to those in the Llavorsi syncline (see Mey, 1968). According to Delattre & Waterlot (1967), cherts and nodular limestones are found at the base of the Civis Formation west of the Segre valley, indicating its probable assignment to the Carboniferous.

In the southeastern part of the map area, west of the Pallaresa valley, in the Segre valley, and from there further east, the Carboniferous in the Compte subfacies area is again different, and is called the Bellver Formation (Hartevelt, 1970, sheets 9, 10). The development is quite similar to the northern facies area. The Carboniferous begins with a chert horizon, followed often by a thin limestone layer and then a sequence of black and brown shales with sandstone layers and lenses of conglomerates. The pebbles in the conglomerate consist of quartz, quartzite, chert, gneiss, porphyrite, granite and limestone. Fossils indicate a Tournaisian, Visean and possibly Namurian age (Boissevain, 1934; Waterlot, 1969; Boersma, 1973). According to Buchroithner & Milan (1977) these conglomerates have a southerly origin.

POST-VARISCAN CARBONIFEROUS (sheets 7–10, Table 1)*

The post-Variscan Carboniferous deposits in the area of study are strictly non-marine and largely fluvial. Volcanic deposits (Erill Castell volcanics) occur on top of strata dated as Westphalian D (Aguiró Formation) and below strata having a Stephanian age (Malpas Formation). Only a few levels have yielded fossils, mainly in the form of plant imprints (Dalloni, 1910, 1930).

Aguiró Formation (Westphalian D) (Fig. 14)

The formation has been named after the village of Aguiró, several hundred metres west of which occurs a well-exposed and characteristic section.

The Aguiró Formation is composed of mainly coarse conglomerates with locally, at the base, a horizon of breccia and some coal stringers. It rests unconformably on folded and cleaved Lower Carboniferous and older rocks. The upper boundary is drawn at the base of the first pyroclastics (Erill Castell volcanics).

Erill Castell Volcanic Formation (Stephanian) (Fig. 15)

The formation is named after the small village of Erill Castell, in the surroundings of which this formation is complete and well exposed.

The Erill Castell Volcanic Formation is composed of mainly light-coloured tuffs containing bombs of up to 1 m, which in the Erill Castell-Peranera area are overlain by a dark-green, massive basaltic-andesite sheet.

Malpas Formation (Stephanian) (Fig. 15)

The formation has been named after the village of Malpas. In that area the formation as well as the intercalated coal seams are thickest. The Malpas Formation comprises a series of dark-grey and brownish fluvial sediments with intercalations of coal seams of up to 2 m thickness, and some carbonate beds.

PERMIAN

Peranera Formation

The formation has been named after the Peranera River. In the Peranera valley the formation is thin but typical, and here both the lower boundary and the unconformable relation with the overlying Bunter are clearly exposed.

The Peranera Formation consists of a monotonous greyish-red alternation of calcareous mudstones, siltstones, sandstones, and tuffs, transported tuff material, breccia beds, and nodular limestone/dolomite beds.

The upper boundary is drawn at the sharp contact with the first coarse quartz-sandstone or pebble-bed belonging to the Bunter. In many areas this contact is a conspicuous unconformity, its angle varying from 10° to about 30° locally.

* This section is mainly adopted from a publication by Mey et al. (1968).

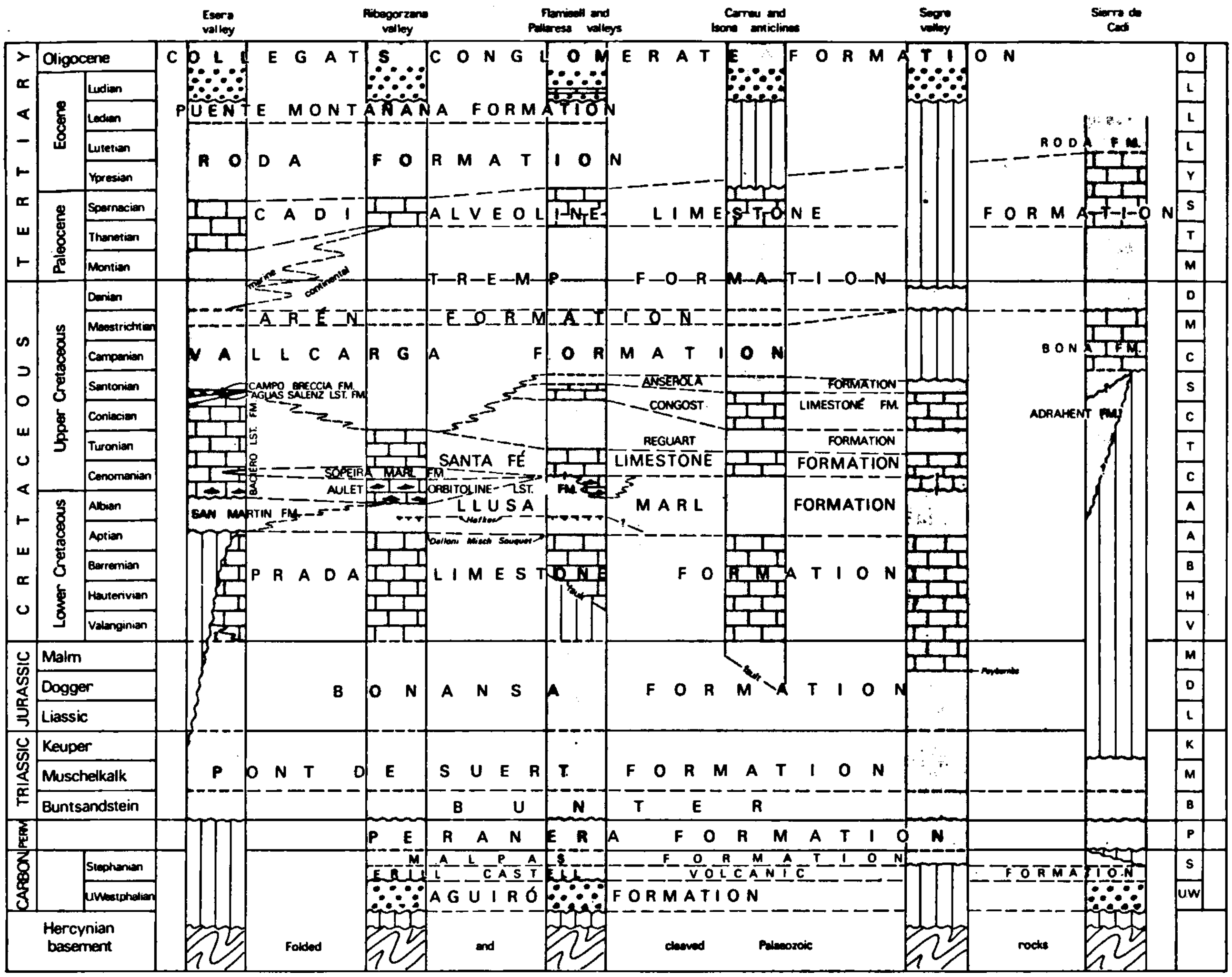


Table 1. Post-Variscan stratigraphy in the southern part of the Central Pyrenees (after Mey et al., 1968).

MESOZOIC

The Mesozoic deposits start with fluvial red beds of the Bunter, but the overlying deposits are all marine, mostly limestones. The uppermost Mesozoic was a period of regression; near-shore and fully continental deposits are of widespread occurrence (Arén Sandstone Formation, Tremp Formation).

Bunter (Lower Triassic)
For this Lower Triassic rock unit with its very uniform development over large parts of the European continent the introduction of a local geographic name would be undesirable. The name Bunter (Buntsandstein) is used as a formation as done by many authors.

The Bunter consists of a predominantly greyish-red sequence of conglomerates, coarse-grained quartzose sandstones, silt- and mudstones, the medium-grained rocks commonly being micaceous.

The upper boundary is drawn at the contact of reddish, greenish or dark-grey shales of the uppermost Bunter with dolomitic marl, gypsum or limestone/dolomite; this contact is, however, often faulted.

Pont de Suert Formation (Middle and Upper Triassic)
The formation is named after the village of Pont de Suert. The Pont de Suert Formation consists of dark-greyish to black micritic limestone, fine-grained dolomite, grey, red and green dolomite marls, cavernous dolomite, and vividly-coloured gypsum and rock-salt. Bodies of crystalline basic rock (ophite) generally occur mainly in the gypsiferous part of the formation. Almost invariably the lower as well as the upper boundaries of this formation are tectonically disturbed. When not faulted or truncated by an unconformity the upper boundary is the contact of gypsum, dolomitic marl or cavernous dolomite with a competent carbonate unit in which thinly stratified, non fossiliferous marly and dolomitic limestones and dolomites alternate (Bonansa Formation).

Bonansa Formation (Lias-Dogger)

The formation is named after the village of Bonansa, 2½ km east of which a very well-exposed continuous outcrop occurs along the road.

The Bonansa Formation consists in its lower part of a competent, fine-grained carbonate unit in which alternate thinly-stratified marly and dolomitic limestones and dolomites, the latter locally being cavernous. The central part contains dark shales, marls and marly, very fossiliferous limestones, and is followed by a competent, coarse-grained sequence of mainly massive dolomites with a characteristic dark-grey to black weathering.

The transition into the overlying Prada Formation is rather gradual, but the boundary is drawn at the top of a sequence in which dolomites dominate over limestones.

Prada Limestone Formation (Malm-Upper Aptian: 'Urgonian')

The formation is named after the Sierra de Prada, a limestone mountain ridge west of the Segre River. A very thick and characteristic section is exposed along the road from Orgaña to Seo de Urgel.

The Prada Limestone Formation is a massive, dark-grey to black, fossiliferous and predominantly micritic rock sequence overlying the Bonansa Formation. In its lower part dolomitic and breccious intercalations abound, and in its upper part occur some intercalations of marly limestone and marl.

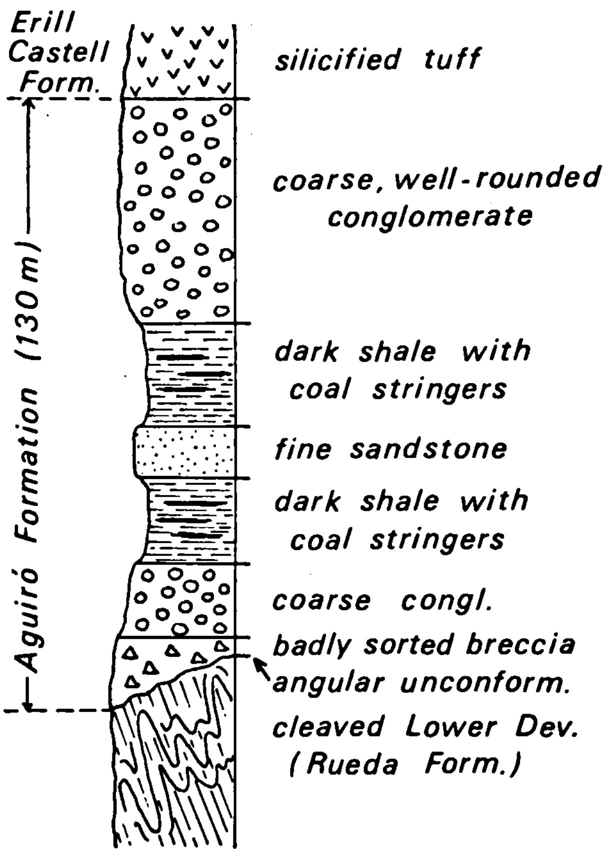


Fig. 14. Stratigraphic section of Aguiró Formation, west of village of Aguiró (after Mey, 1968).

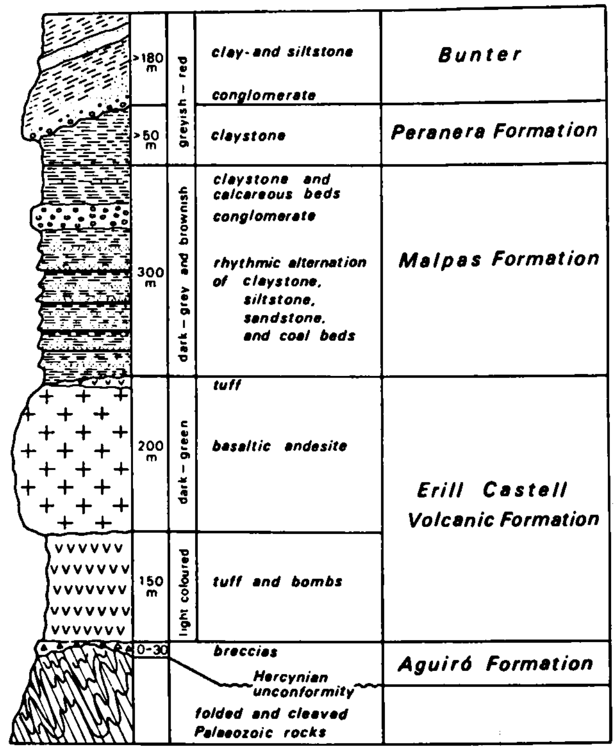


Fig. 15. Stratigraphic section of post-Variscan Carboniferous and Permian (after Mey, 1968).

Llusa Marl Formation (Upper Aptian-Lower Cenomanian)

The formation is named after the hamlet Llusa in the Flamisell valley, which is surrounded by a well-exposed and typical sequence of these marls.

The Llusa marls consist of a greyish, monotonous alternation of fossiliferous shaly marls and silty marls (often nodular) with only a few intercalations of marly limestone.

San Martin Formation (Albian)

The formation is named after the village of San Martin in the Nogueras Zone. In the San Martin Formation alternate light-coloured quartz-sandstones, quartz conglomerates, shaly oyster-rich marls, coal seams, and a few thin-bedded bituminous black limestones. Locally only sandstones and conglomerates are developed.

Generally the lower boundary is an unconformity, the formation resting with a sharp contact on rocks of the Prada, Bonansa or Pont de Suert Formations. Frequently the upper boundary is also an unconformity, the base of the overlying Baciero Formation being a fragmental, light-grey limestone or a yellowish to reddish sandy orbitoline limestone.

The San Martin Formation is the lateral equivalent of the Llusa marls.

Aulet-Orbitoline-Limestone Formation (U. Albian-L. Cenomanian)
The formation is named after the Sierra de Aulet, an E–W striking mountain ridge formed by a very thick (almost 900 m) sequence of these limestones, in which the Ribagorzana River carved a narrow gorge. In the upstream part of the gorge the Escales gravity-dam has been constructed.

The Aulet-Orbitoline-Limestone Formation mainly consists of a yellowish to reddish-brown coloured, coarse-grained bioclastic limestone with an abundance of orbitolines. The limestones, apart from being breccious and dolomitic, are generally sand-bearing (except in the Pallaresa valley), and alternate with dark-coloured marls and marly limestones.

Sopeira Marl Formation (Upper Cenomanian)
The formation is named after the village of Sopeira in the Ribagorzana valley which is built in a depression several hundred metres wide between the Aulet limestone in the north and the Santa Fé limestone in the south. The occurrence of this formation is restricted to the Ribagorzana and Isabena valleys.

The Sopeira Marl Formation consists of a light-coloured sequence in which alternate regularly and thinly-bedded sandy marls and nodular argillaceous limestones, spotted with glauconite and pyrite. An abundance of echinoids and ammonites is characteristic.

Santa Fé Limestone Formation (Upper Cenomanian-Turonian)
The formation is named after the Peña de Santa Fé, an impressive mountain which is crowned by a shallow syncline of Upper Cretaceous limestones. The Santa Fé limestone is the highest individual limestone unit, about 20–40 m thick, which lies just below the steep upper scarp formed by the thicker Congost limestone.

The Santa Fé Limestone Formation is a competent limestone unit composed of a light-grey to beige micritic carbonate, characterized by an abundance of prealveolines and miliolids. The upper part of the formation, in which fissurines and globigerines abound, is generally marly. In the Ribagorzana valley the middle and upper part of the formation is strongly slumped and brecciated; in the Flamisell valley one slump horizon has been observed in the upper part.

Reguart Formation (Turonian-Coniacian)
The formation is named after the hamlet of Reguart in the Flamisell valley, southwest of which an easily accessible and typical sequence of mainly marls and nodular limestones is exposed.

The Reguart Formation consists of a regular, grey sequence in which alternate shale, shaly marl and nodular argillaceous limestone.

Congost Limestone Formation (Coniacian-Santonian)
The name of this formation is derived from the narrow gorge in these rocks about 5 km north of Pobla de Segur, through which the Flamisell River has made its way. This formation wedges out west of the Flamisell River.

The Congost Limestone Formation is a complex light-grey limestone unit, consisting of micritic limestones, coral and algal limestones (small bioherms), reef talus and argillaceous limestones. Locally the limestones are glauconitic and/or sandy. A bed with *Hippuritus* is frequently developed at the top.

Anserola Formation (Santonian)
The formation is named after a small tributary of the Flamisell River, running in the strike of these rocks just southwest of the above-mentioned Congost gorge.

The Anserola Formation consists mainly of an alternation of shaly marls and nodular argillaceous limestones in which echinoids abound. A high content of glauconite is characteristic for this rock unit. In the upper part of the formation, slump structures are frequently found.

In the area west of the Isabena River the entire lower part of the Upper Cretaceous (Cenomanian-Santonian) is developed as a limestone, in which only locally slightly marly intercalations can be mapped separately. In that area (sheet 7) the entire carbonate complex above the San Martin Formation and below the Vallcarga Formation is called the Baciero limestone.

Baciero Limestone Formation (Cenomanian-Santonian)
The formation is named after the Sierra de Baciero east of the Esera River, where these limestones form a steep mountain ridge. This formation is restricted to the area west of the Isabena River.

The Baciero Limestone Formation is defined as a complex limestone unit, whose lower part is a grey and reddish, orbitoline-bearing, locally sandy, coarse bioclastic limestone, occasionally followed by a horizon of a very fine-grained argillaceous limestone with a nodular habit.

The upper part is formed by a dark-grey coarse bioclast, which may be sandy or even conglomeratic, and which contains a mappable horizon with black chert concretions.

Vallcarga Formation (Santonian-Maastrichtian)
The formation is named after the small tributary of the Pallaresa River just north of Pobla de Segur (sheet 9). In this streamlet an almost complete section of the lower part of the Vallcarga Formation is exposed.

The Vallcarga Formation consists in its lower part of characteristically yellow-brownish weathered turbidites (containing quartz and calcareous fragments of strongly varying coarseness). The middle part of the formation contains mudflows and olistostrome levels; in the upper part homogeneous bluish-grey marls predominate. Glauconite and wood fragments are characteristic components of the entire formation.

Adrahent Formation (Albian?-Santonian?)
The formation is named after the village of Adrahent south of Seo de Urgel, in the vicinity of which a typical development is found. The Adrahent Formation consists of white quartz conglomerates, sandstones, and occasionally thin shale intercalations. Plant remains, so far indeterminable, may occur throughout the entire formation.

Bona Formation (Campanian-Maastrichtian)
The formation is named after the streamlet Bona southeast of Seo de Urgel, the upper course of which forms a narrow gorge in this limestone unit.

The Bona Formation consists of bioclastic limestones, which are highly fossiliferous (mainly rudists). The deposits locally are marly or sandy; the latter especially at the base.

Arén Sandstone Formation (Maastrichtian)
The name of this formation has been derived from the village of Arén in the Ribagorzana valley. The name 'Arén sandstone' for this rock unit has been widely used before in the literature. The Arén Sandstone Formation consists in its lower part of coarse-grained calcarenites and marls, and in its upper part of fine- to coarse-grained, well-sorted quartz-bearing calcarenites, usually showing large-scale cross-bedding.

Tremp Formation (Upper Cretaceous-Lower Paleocene)
The formation is named after the small town of Tremp in the Pallaresa valley, in the surroundings of which many good exposures occur.

The Tremp Formation contains in its lower part black shales, coal beds and non-marine limestones (e.g. Isona), and in its upper part reddish-brown conglomerates, sandstones and mudstones, non-marine limestones and gypsum beds.

The lower part, as mentioned above, is not developed in the Ribagorzana valley.

TERTIARY

In the Tertiary a major transgression is recorded by the marine Alveoline limestones and the Roda Formation, on top of which a depositional regression followed (Puente Montañana Formation). After the Pyrenean phase of the Tertiary orogeny pre-Ludian extensive sheets of conglomeratic piedmont deposits were formed (Collegats conglomerates).

Cadi Alveoline-Limestone Formation (Thanetian-Lutetian)
The formation is named after the Sierra de Cadi, the steep mountain ridge just southeast of Seo de Urgel, dominating the Upper Segre valley; the Cadi Formation forms the crest of this mountain ridge.

The Cadi Alveoline-Limestone Formation consists mainly of limestone sometimes argillaceous with an abundance of alveolines and nummulites.

The lower part may be developed as a limestone and quartz conglomerate with Foraminifera in the matrix. The top of the formation can be sandy and even sandstones may occur.

Roda Formation (Sparnacian-Lutetian)
The formation is named after the village of Roda de Isabena where various good sections are found.

The Roda Formation consists largely of medium-grey to dark-grey calcareous claystones between which thin calcarenite beds are intercalated. Both litho-facies may contain numerous marine microfossils.

Puente Montañana Formation (Ledian)
The formation is named after the village of Puente Montañana in the surroundings of which many exposures are found.

The Puente Montañana Formation consists of brownish-grey to grey claystones and mudstones often containing calcareous concretions, poorly sorted sandstones and conglomerates in discontinuous beds. The fauna is largely non-marine; some marine levels occur.

Collegats Conglomerate Formation (Ludian-Oligocene)
The formation is named after the narrow gorge in the Pallaresa valley north of Pobla de Segur (sheet 9).

The Collegats Conglomerate Formation consists of a thick sequence of largely conglomeratic piedmont deposits. Colours vary from light-grey to reddish-brown. Locally marls and coal beds are intercalated near the base. Facies changes in a southerly direction are shown by a lateral transition into fine-grained sandy and evapontic deposits.

The unconformity at the base of the formation develops a sharp relief near the axial zone of the Pyrenees. The upper boundary, in the area under consideration, is the present topography.

CHAPTER 3

STRUCTURAL GEOLOGY*

INTRODUCTION

Except for the Precambrian rocks in the North-Pyrenean massifs, all rocks involved in the Variscan folding are of Cambro-Ordovician to Carboniferous age. The whole sequence of Ordovician, Silurian, Devonian and Carboniferous is conformable, and despite the presence of conglomerates in the Upper Ordovician, there is no Caledonian folding phase. The statements made by Ravier et al. (1975) about Caledonian folding, based on the flatlying metamorphic Ordovician are not related to a Caledonian orogeny, but to disharmonic Variscan folding.

The age of the main folding can be established in the map area. The youngest rocks involved in major folds with development of a cleavage are of Westphalian A age, whereas near Aguiró, Westphalian D is already unconformable (Fig. 14). Elsewhere, the Stephanian volcanics and sediments are not involved in major folding and rest unconformably on older rocks. The main folding in the Pyrenees took therefore place in the Westphalian.

Weak late Variscan folding is witnessed by the unconformity between Triassic and Permian rocks. However, folds in the Stephanian and Permian generally are devoid of well-developed cleavages, and the folding was not very intense. Alpine folding has however influenced these folds.

Although the whole area has been mapped on a scale of 1:25,000 to 1:50,000, detailed structural analysis has not been done everywhere, as only several years after the beginning of the project the usefulness of small scale structures was recognized. Therefore a more detailed structural analysis is available only from the western part of the Aston-Hospitalet massif and its continuation in the Pallaresa antiform (Lapré, 1965; Oele, 1966; Zwart, 1965), the Bosost area (Boschma, 1963; Zwart, 1963a, b), the Lys-Caillaouas massif (Wennekers, Zwart, unpublished), sheet 8 (Boschma, 1963; Mey, 1967a, 1968) and sheet 10 (Hartevelt, 1970). Observations have shown that in other areas similar structures can be found and that a correlation between the well-studied areas is possible. However, some caution in these correlations is certainly necessary.

MAJOR STRUCTURES AND STRUCTURAL UNITS

The large structural units are represented on Plate 1 (enclosures). These are metamorphic massifs as Aston-Hospitalet, Garonne dome and Lys-Caillaouas, a number of syn- and anticlinoria in the Palaeozoic sequence, the North-Pyrenean massifs, the Nogueras zone, and a number of granite to granodiorite massifs.

The large scale structures can best be seen on the Profiles 1–6. From these profiles it is evident that two structurally different domains occur in the Pyrenees, viz. regions with high grade rocks like the Aston-Hospitalet massif, the Garonne dome with Bosost area, and some of the North-Pyrenean massifs and regions with low grade rocks, where metamorphism does not reach beyond the upper greenschist facies. In the first domain, structures are mainly characterized by recumbent folds and flatlying schistosities, in the second by steep folds and cleavages, although in the southern part of the axial zone, cleavages have quite a low dip as well. This is, however, attributed to a late effect of tilting.

In the metamorphic regions, usually within the biotite isograd, the layering of the rocks and the schistosity are gently undulating as shown on Profiles 1, 2, 5 and 6. It could be shown that the undulation is due to one or two late folding phases and that the original position of the schistosity must have been close to horizontal. Large scale structures have not been found in these metamorphic regions.

In the Hospitalet-Aston massif the attitude of the schistosity forms a major antiform. The Garonne region has a large flat dome structure. In earlier publications I have called the regions with flatlying schistosity, the infrastructure. However, its position is not directly related to depth of burial, but more to higher temperature, as from east to west in the Pyrenees the infrastructure is gradually lying higher in the stratigraphic sequence. The relationship between the infrastructure and the overlying steep suprastructure will be discussed in a later section.

In the low grade, suprastructural Palaeozoic rocks a number of roughly E–W trending major structures can be distinguished. They are anti- and synclinoria involving Cambro-Ordovician, Silurian, Devonian and Carboniferous sediments.

The Silurian has played a particular role due to its special nature, causing it to be a very incompetent layer between the Cambro-Ordovician and the Devonian. In general it acted as a decollement surface, and the structures above and below the Silurian are strongly disharmonic. This is especially clear in the Garonne dome where it forms the boundary between the infra- and suprastructure (Fig. 16), but also elsewhere a strong disharmony in the structures is present. This behaviour is also very outspoken in the Llavorsi syncline (sheet 10) where elongate diapiric structures of Silurian slates (with graptolites) occur within the Carboniferous (Hartevelt, 1970).

In the axial zone two large anticlinorial structures in Cambro-Ordovician rocks can be distinguished: the Pallaresa anticlinorium and the Orri-Payasso dome. The Pallaresa anticlinorium grades towards the east in an infrastructure with the high grade rocks of the Aston-Hospitalet massif. To the west the anticlinorium becomes

* For this chapter the maps and especially the coloured profile sheet should be consulted. Whenever Profile 1, 2 etc. is mentioned it refers to one of the profiles of this sheet.

very narrow and is known as the Central anticline, which further west includes the Lys-Caillaouas granite body.

The Orri dome with low-dipping cleavages described by Hartevelt (1970) plunges westwards underneath the Devonian, but reappears again in the small Payaso dome. Due to the lack of marker beds in the Cambro-Ordovician, the detailed structure of these anticlinoria cannot be worked out, and on the profile sheet the structure is depicted in a schematic manner. It is clear that many small scale structures occur in the Cambro-Ordovician, but they cannot be continued beyond the scale of one outcrop. The fact that large scale structures do occur in these Cambro-Ordovician regions, is shown by the Siluro-Devonian Tor syncline in the southern part of the Pallaresa anticlinorium.

These two large structures and the metamorphic massifs are separated by synclinoria with Silurian, Devonian and Carboniferous rocks. Due to the fact that there is a well mappable lithology in these systems, also large scale structures in these synclinoria can be deciphered. Along the northern side of the axial zone and north of the Pallaresa and Garonne unit lies the northern synclinorium, of which Profiles 4 and 5 give a good view of the structure.

Similar structures as in the northern synclinorium can be found in the Valle de Arán synclinorium, between the Garonne dome and the Pallaresa anticlinorium (Profiles 3, 4). The northern and Valle de Arán synclinoria are, at least in part, overlying the Garonne dome.

Between the Pallaresa anticlinorium and the Orri dome appears one of the largest structures of the map area, the Llavorsi syncline, which can be followed along strike from Andorra to the Maladeta granodiorite. It is a tight to almost isoclinal structure, which must reach to a depth of several kilometres. Its axial plane dips 30-50° to the north. Many small folds accompany this large structure. To the east it branches into several synclines which abut against the Andorra-Mont Louis granodiorite. In westerly direction the Llavorsi syncline is cut off by the Maladeta granodiorite, and it does not reappear west of this batholith.

South of the Orri dome and the Maladeta granodiorite occurs the southern synclinorium. In its eastern part it has horizontal to gently north dipping folds (Profiles 3, 3a), but to the west the structure becomes complicated as can be seen from the map pattern, sheets 8, 9, and the Profiles 4, 5 and 6. At least part of the stratigraphic sequence there is inverted. These structures will be discussed later.

From the profiles it is evident, that the axial planes of the folds, and the accompanying cleavage in the suprastructure in the northern part of the axial zone are steep to vertical, and that towards the south the attitude of these planes is gradually less inclined to the north, so that finally recumbent structures are found. Looking from this point of view a major structure is present in the axial zone, namely half a fan which is independent of the described syn- and anticlinoria (Fig. 30).

Crosscutting through all these structures occur a number of large granodiorite bodies, which show, however, the same general trend as the major E–W structures.

FOLD GENERATIONS (see also Table 6)

Although on the small scale geological map (1:200,000) the major structure seems to be quite simple, many detailed investigations have shown that several fold generations occur in most Palaeozoic rocks. It could be

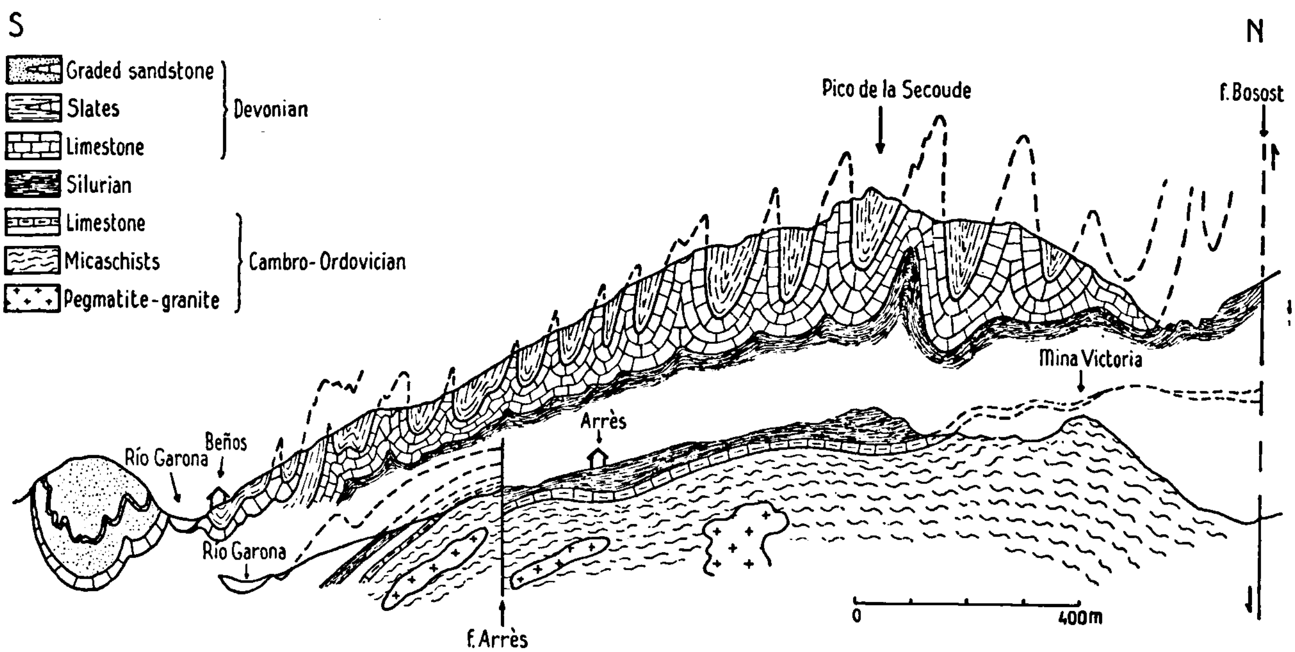


Fig. 16. N–S section east of Bosost (sheet 4) showing steep structures in Devonian and flatlying structures in Cambro-Ordovician (after Kleinsmiede, 1960).

shown that folding took place already before the development of the E–W cleavage folds, whereas several other fold generations postdate it.

In our previous work, and also in this paper it is assumed that the major structures as outlined above, all belong to one folding phase, usually called the main phase as it is for a large part responsible for the map pattern. We have also referred to it as F_1, and to the corresponding cleavage or schistosity as S_1. After we had started to use these terms, it appeared that the main phase was preceded by at least one generation of folds, which has a large distribution in the Pyrenees. In order not to confuse terms used in this and earlier publications, we retain F_1 and S_1 for the major folds described already, and use F_0 and S_0 for these earlier folds.

In the following discussion the F_1 structures are used as a kind of datum plane, but contemporaneity throughout the Pyrenees of this folding phase can, of course, not be proved. However, the large scale structures and the occurrence of a cleavage in the axial planes of minor and major folds, make the main phase to a convenient frame of reference (Table 6).

Pre-main phase (F_0)

Folds predating the E–W cleavage folds occur at many places in the Pyrenees, especially near the southern and northern border of the central zone. They were first discovered by Boschma (1963) in a dome-shaped structure in the Ribagorzana River (sheet 8). They have also been described from the region near Estours (de Sitter & Zwart, 1962, sheet 2) and they have been extensively treated by Mey (1967a: sheet Ribagorzana, and 1968: sheet 8). Furthermore they have been discussed by Hartevelt (1970). On sheets 7, 8, 9 and Flamisell-Mañanet it is clearly visible that the map pattern is different from regions further north like in the Valle de Arán (sheet 4) and Salat (sheet 5), where E–W trending structures form the dominant pattern. On these first mentioned sheets there is instead an intricate pattern, without such dominant E–W trend.

On all four maps N–S to NE–SW trending structures can be recognized, as well as a typical 'Schlingenbau' and some completely closed structures. The map pattern already suggests that it is due to the interference of at least two fold generations, and Boschma (1963) and Mey (1967a, 1968) have brought forward ample evidence that two fold generations are indeed responsible for this pattern. These are an early generation with N–S to NE–SW trending folds with steep axial planes, and a later generation with an E–W to NW–SE direction and N to NE dipping axial planes. The second set is accompanied by a well-developed cleavage, and is correlated with the main (F_1) phase elsewhere in the Pyrenees. In the field it is clear that the N–S to NE–SW striking limbs of the first generation folds are cut by an E–W cleavage and small scale folds (Fig. 17). The cleavage cuts undisturbed through the limbs and the nose of the early folds (called henceforth F_0). The cleavage–bedding intersections show a great circle distribution in stereograms (Fig. 18).

Fig. 19 represents two stereograms taken from the thesis of Boschma (1963). The locality is a small dome

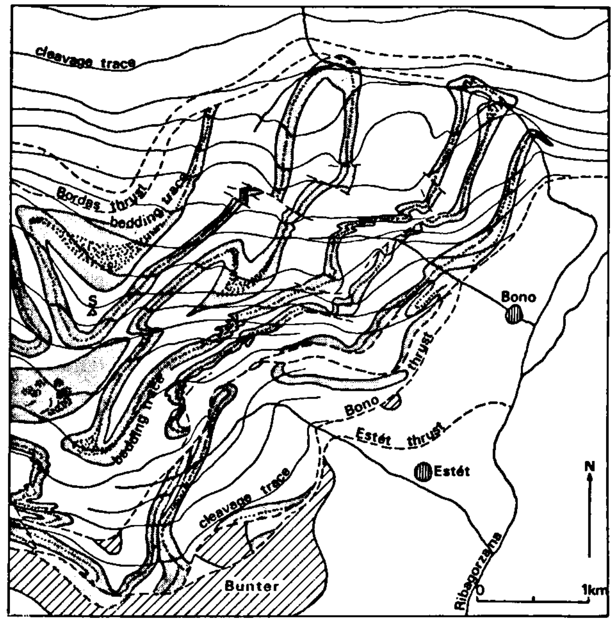


Fig. 17. Outcrop pattern of Basibé Formation (sheet 8, Ribagorzana-Baliera) showing major F_0 and small F_1 folds (after Mey, 1968).

structure cut by the main road in the Ribagorzana valley near the village of Viñal. Measurements of the bedding plane clearly show the dome shape. The S_1 cleavage has one maximum with an average dip of 34° to the north. The δ-lineations (intersections cleavage–bedding) spread within this plane. Obviously the dome is the result of the interference of an early N–S fold with an E–W cleavage fold.

On the map (Fig. 20) made by Mey (1968) the relationship between the F_0 folds, indicated by the contours of the Basibé Formation or the hingelines of anti- and synclines, and the S_1 cleavage is shown. A detail of this map is reproduced as Fig. 17 on which major F_0 folds and smaller F_1 cleavage folds are visible. The four stereograms (Fig. 18) confirm this. There is a large spread of the bedding; S_1 is dipping to the NNE and has one maximum, the intersections between cleavage and bedding are spread in the S_1-plane, but with a maximum in NNE direction, indicating the trend of the early folds.

It should be noticed here, that on sheet Flamisell-Mañanet the N–S trending folds in the Devonian on the western part of the map are steep pre-main phase folds, whereas the folds of Devonian-Carboniferous in the Flamisell valley in the eastern part are recumbent F_1 folds, without any visible influence of the early fold set, and it seems that the F_0 folds disappear in this area (see also Plate 1).

In the central part of the axial zone the map pattern is distinctly E–W. Even here, however, there is evidence for the presence of F_0 folds. In the Llavorsi syncline (sheet 10) elongate structures closing on both sides are found. They are probably very flat dome or sheath structures (Fig. 21). This is confirmed by the fact that cleavage–bedding lineations are not E–W, but are lying

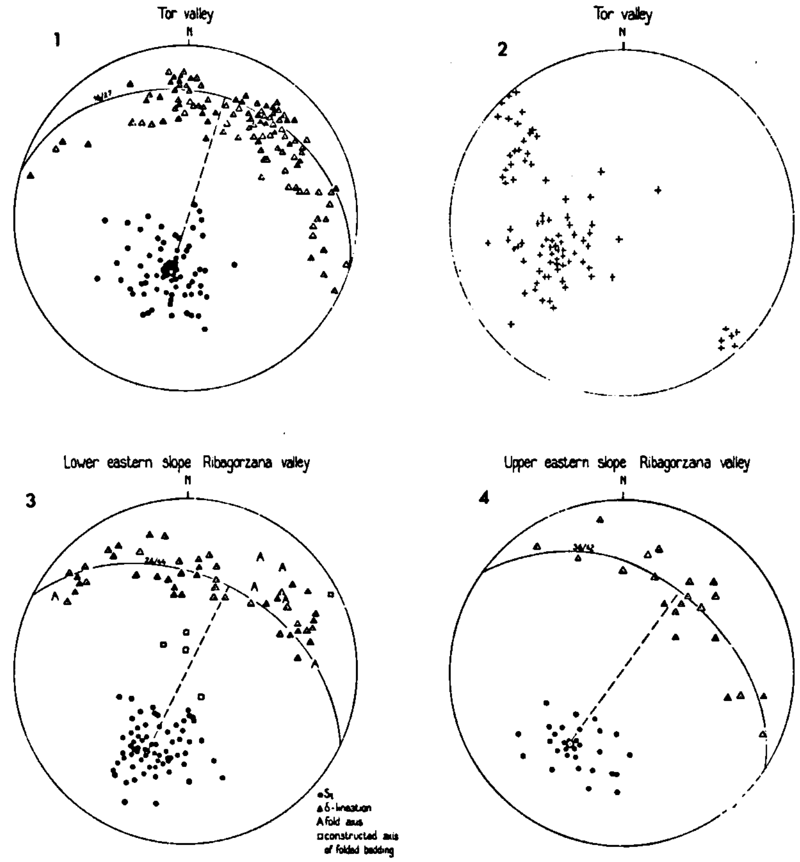


Fig. 18. Stereograms of Tor and Ribagorzana valleys with distribution of poles to bedding (crosses, diagram 2), poles to S_1 cleavage, and small foldaxes and cleavage-bedding intersections (δ-lineations); (after Boschma, 1963 and Mey, 1967).

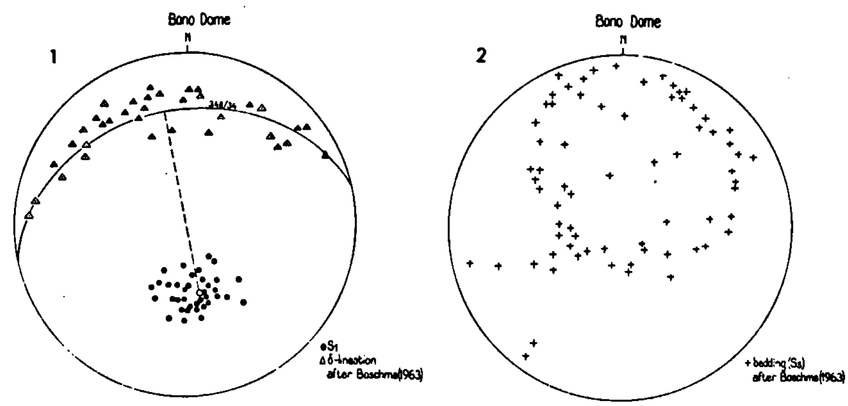


Fig. 19. Stereograms of poles to bedding, S_1 cleavage and δ-lineations of dome structure south of Bono (after Boschma, 1963).

within the cleavage plane and plunge to the north (Hartevelt, 1970, p. 203). The same applies to the Devonian of the Valle de Arán, where no interference pattern can be discerned on the map. Nevertheless the δ-lineations spread in a great circle, with many steeply plunging lineations in the vertical cleavage plane.

In most cases the F_0 folds are not accompanied by a cleavage, reason why they previously have been called pre-cleavage folds. However, in some cases, an axial plane cleavage in pelitic rocks has been observed, like near Durro (sheet 8).

All these structures, described so far, occur in Devonian and Carboniferous rocks, so that the dating of this early fold phase gives no problem: it must have taken place after the deposition of the Lower Carboniferous and before the main phase.

In the Cambro-Ordovician rocks this early folding phase has also been described. From the map pattern no

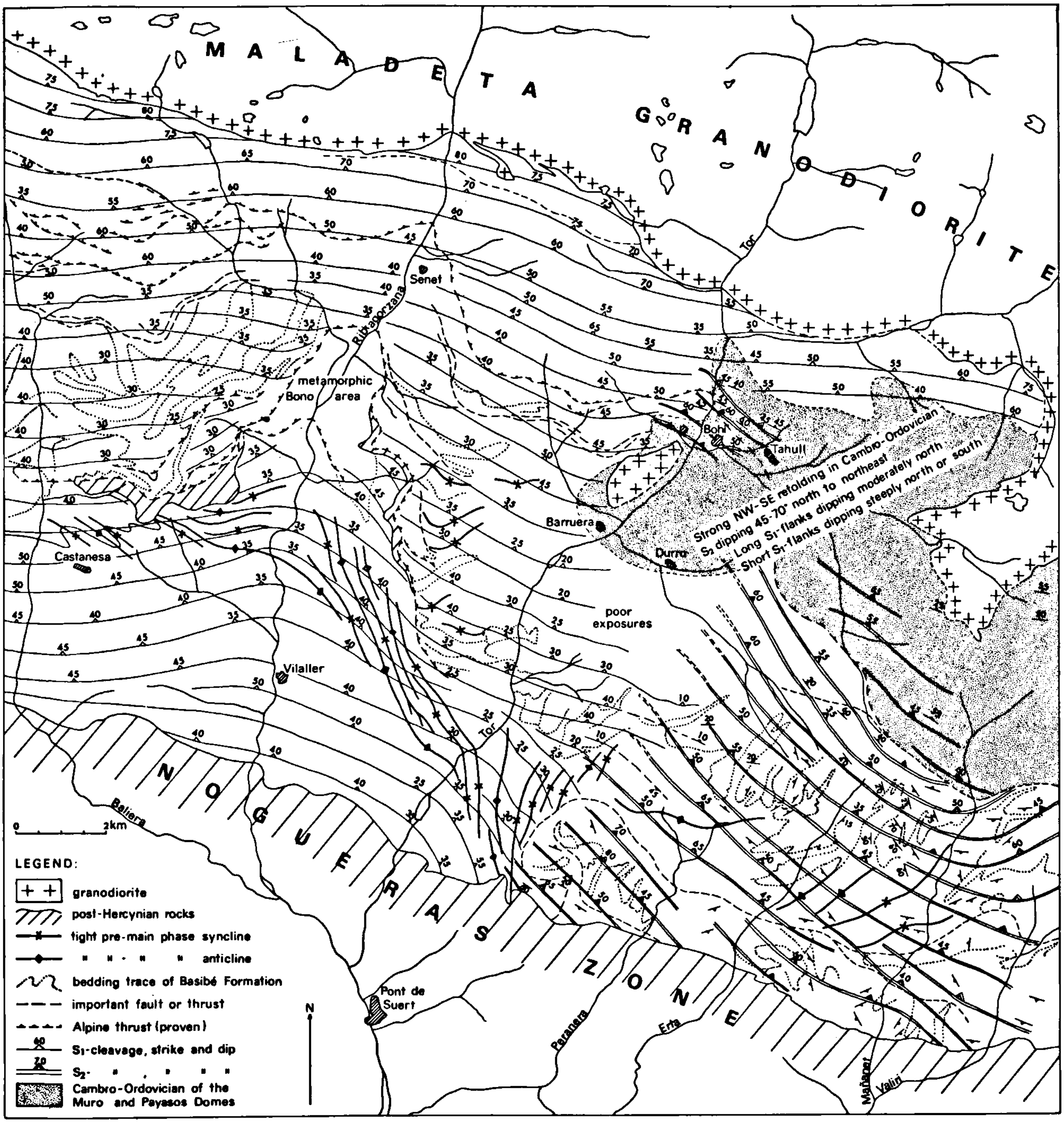


Fig. 20. Map of part of sheet 8 indicating strike lines of S_1 cleavage and S_2 cleavage, trace of Basibé Formation and F_0 anti- and synclines (after Mey, 1968).

conclusions can be drawn, due to the lack of marker horizons. At several localities, however, like in the Orri dome and the Massana anticline, it has been noticed that there is a strong variation of the plunge of minor fold axes and cleavage–bedding intersections within the main phase cleavage plane (Hartevelt, 1970). This can again be interpreted as being due to an earlier oblique set of folds on which the F_1 folds are superposed (Fig. 22). In the Cambro-Ordovician of the Pallaresa anticlinorium, however, Zandvliet (1960) and Oele (1966) have not found such variations of F_1 lineations, and for that region there are no arguments to assume the presence of this early folding phase.

In the metamorphic Cambro-Ordovician west of the Lys-Caillaouas massif, microstructures indicate the presence of two cleavage planes of which the first one probably can be correlated with S_1 further east. This early cleavage has been described by Trouiller (1976) who

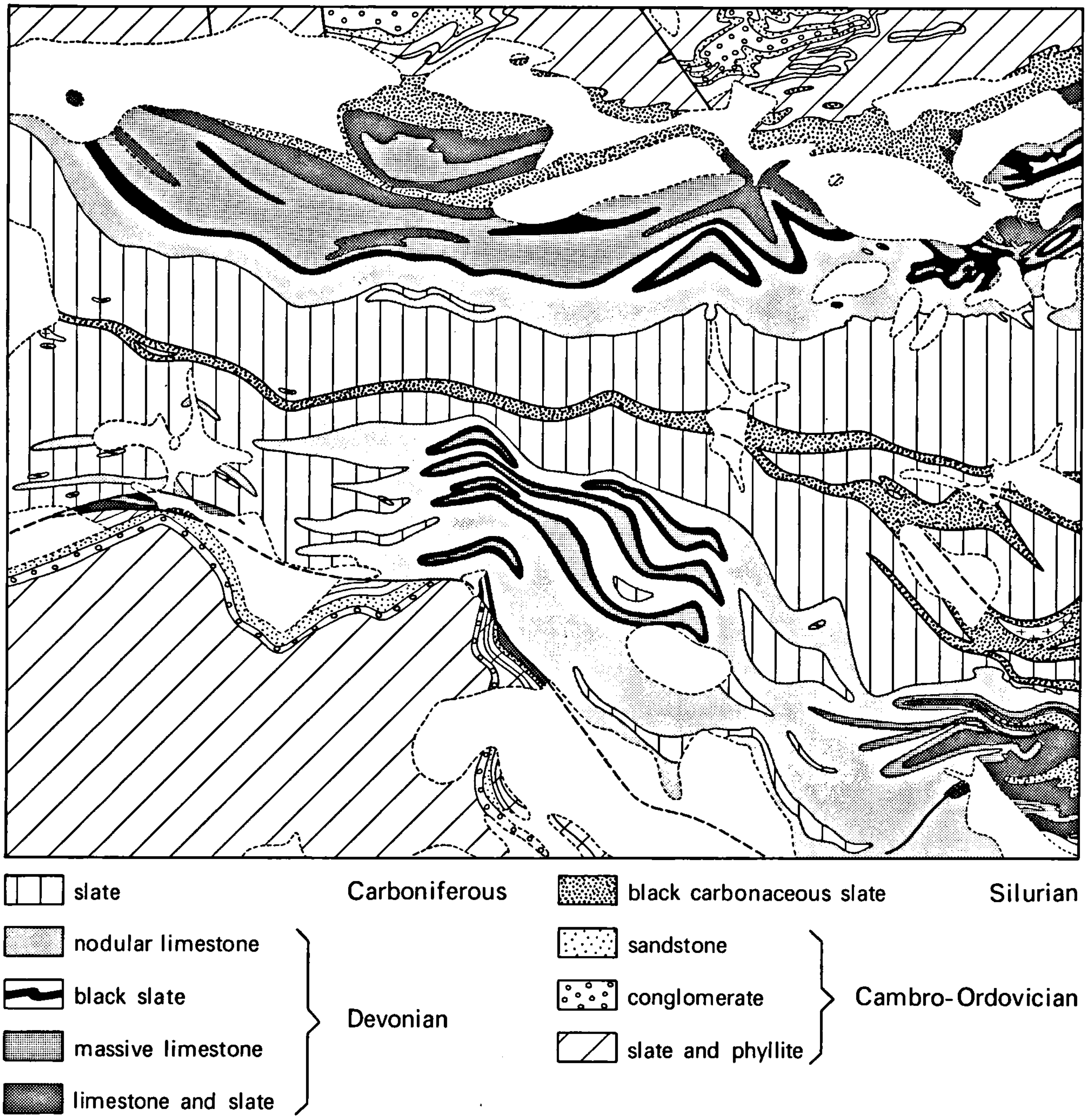


Fig. 21. Geological map of part of the Llavorsi syncline (sheet 10) showing closed structures in the Devonian (sheath folds).

found E–W trending folds corresponding to this cleavage. In the Gavarnie area a similar early fold set accompanied by a slaty cleavage or schistosity in regional metamorphic rocks belonging to the Cambro-Ordovician has been described by Majesté-Menjoulas (1979). According to his interpretation the folds connected with this early cleavage have a N–S direction and were originally recumbent. In that region the second folding phase has the properties of F_1 folds in this publication (they are called F_1 and F_2 by Majesté-Menjoulas). Up to now it is not possible to correlate these early synschistose folds in regional metamorphic Cambro-Ordovician rocks with the ones described here in the Devono-Carboniferous. In the section on microstructure these F_0 and F_1 schistosities shall be further discussed. From work outside our map area, an early folding phase in Devonian rocks has been described by Muller & Roger (1977) for the region to the west of our map area. These two authors have demonstrated the presence of major early, pre-cleavage folds, also with the aid of interference patterns and overprinting relationships. To the east Guitard (1962) has mentioned the occurrence of pre-cleavage

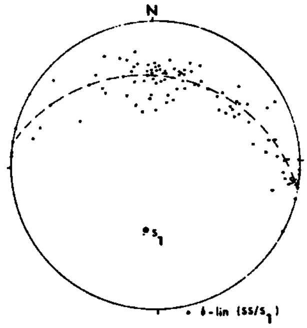


Fig. 22. Distribution of δ-lineations (cleavage–bedding intersections) in Cambro-Ordovician of Orri dome (sheet 10, after Hartevelt, 1970).

folds in the micaschists of the Canigou massif. Summarizing it can be stated that traces of an early, pre-main cleavage fold generation have been found in most of the axial zone.

Main phase (F_1)

As has been said already, the main phase is responsible for most major folds in the Palaeozoic of the Pyrenees, and hence also largely for the map pattern. Furthermore a distinction has to be made for the low grade suprastructure and the high grade infrastructure with respectively steep- and low-dipping folds and cleavages.

Suprastructure. – In the section on major structures and structural units the map pattern and the six profiles through the Central Pyrenees have already been briefly discussed. Further details can be gathered from cross sections accompanying some of the 1:50,000 sheets, and from the map descriptions published earlier. Two sets of sections from the Valle de Arán synclinorium (sheet 4, Kleinsmiede, 1960) and from the northern synclinorium (sheet Salat, 1:20,000, Zandvliet, 1960) are reproduced here as Figs. 23 and 24.

From these sections it appears that in the Devonian-Carboniferous sequence the folds are usually very tight to almost isoclinal. In well-bedded Devonian rocks chevron folds are common. The same applies to the Llavorsi syncline (sheets 9, 10, Hartevelt, 1970). Minor folds show the same style. As has been remarked before, the attitude of these folds varies. They are vertical in the northern part of the axial zone and in some of the North-Pyrenean massifs, inclined to the north in the more southerly part of the axial zone, for example in the Llavorsi syncline, and locally are recumbent as in the Flamisell valley (sheet 9).

The structure of the Flamisell-Mañanet sheet on a scale 1:25,000, is quite complicated. As this sheet has not been published before, a special set of profiles through that map has been drawn by K. Roberti, who compiled the map and sections (Plates 2, 3, 4 (enclosures)). From these sections it is evident that in the western part of the map, major F_0 and F_1 folds dominate the picture, whereas in the eastern part, in the Flamisell valley, only recumbent F_1 folds occur. The early folds apparently disappear within a rather short distance. From the western part also E–W profiles have been drawn to show the shape of the F_0 folds (Plate 4). Further complications arise from the fact that in the western part the sequence is inverted, and all antiforms are synclines and vice versa. In addition in this area a late folding phase with a new crenulation cleavage disturbs the F_1 cleavage, and is perhaps responsible for a change in the average dip of S_1 towards the south (see also Profiles 4 and 5). In areas further towards the west and east this late folding is less intense or absent, and there S_1 has a shallow dip to the north. From the relationships between NNE–SSW trending folds with F_1 folds and S_1 cleavage, the superposition of two sets is obvious. The inversion is to our opinion due to a major F_1 fold, of which the hinge, however, has not been located. The sequence of events would then be: first formation of fairly large, tight F_0 folds with an approximate N–S trend and with only locally the development of a cleavage. Then cleavage folding, producing in the western part of the Flamisell-Mañanet sheet a major fold of which the lower, inverted limb is preserved. In the eastern part of this area a number of large recumbent folds were formed. In the west a third foldset affected the earlier folds and caused the S_1 plane to rotate to a southerly dip.

In the Cambro-Ordovician there is also a large variation in size of the folds, and in these rocks there is an axial plane cleavage as well. It is, however, much more difficult to trace out major folds, as marker beds do not exist in the monotonous quartz-phyllites.

In the Pallaresa antiform, small-scale folds are tight to isoclinal. In the south, for example in the Orri dome, folds in the Cambro-Ordovician slates are less oppressed, usually no longer isoclinal but close to tight. In the profiles (see profile sheet) this is only schematically portrayed.

As has been mentioned before, the F_1 folds are all accompanied by an axial plane cleavage in pelitic, in psammitic and in calcareous rocks. The nature of the cleavage depends largely on the rock composition. In pelitic rock, up to Carboniferous age, north of the southern limb of the Llavorsi syncline, S_1 is a slaty cleavage partly determined by a parallel arrangement of phyllosilicates (Fig. 33). However, in some rare cases it can be seen that the cleavage is a very fine crenulation cleavage, which probably developed on a fine-grained sedimentary fabric (Fig. 46). In many cases, however, the S_1 cleavage has been folded by one or two later crenulation cleavages.

In pelitic rocks of Cambro-Ordovician age south of the Llavorsi syncline, the foliation is usually a crenulation cleavage, according to the author's opinion developed on a sedimentary fabric (see also section on microstructures). In quartzrich rocks a fracture cleavage or a differentiatied layering is often present, whereas in calcareous rocks the cleavage is usually accompanied by a differentiated layering.

The change of the slaty cleavage in the north to the crenulation cleavage in the south is rather abrupt, but

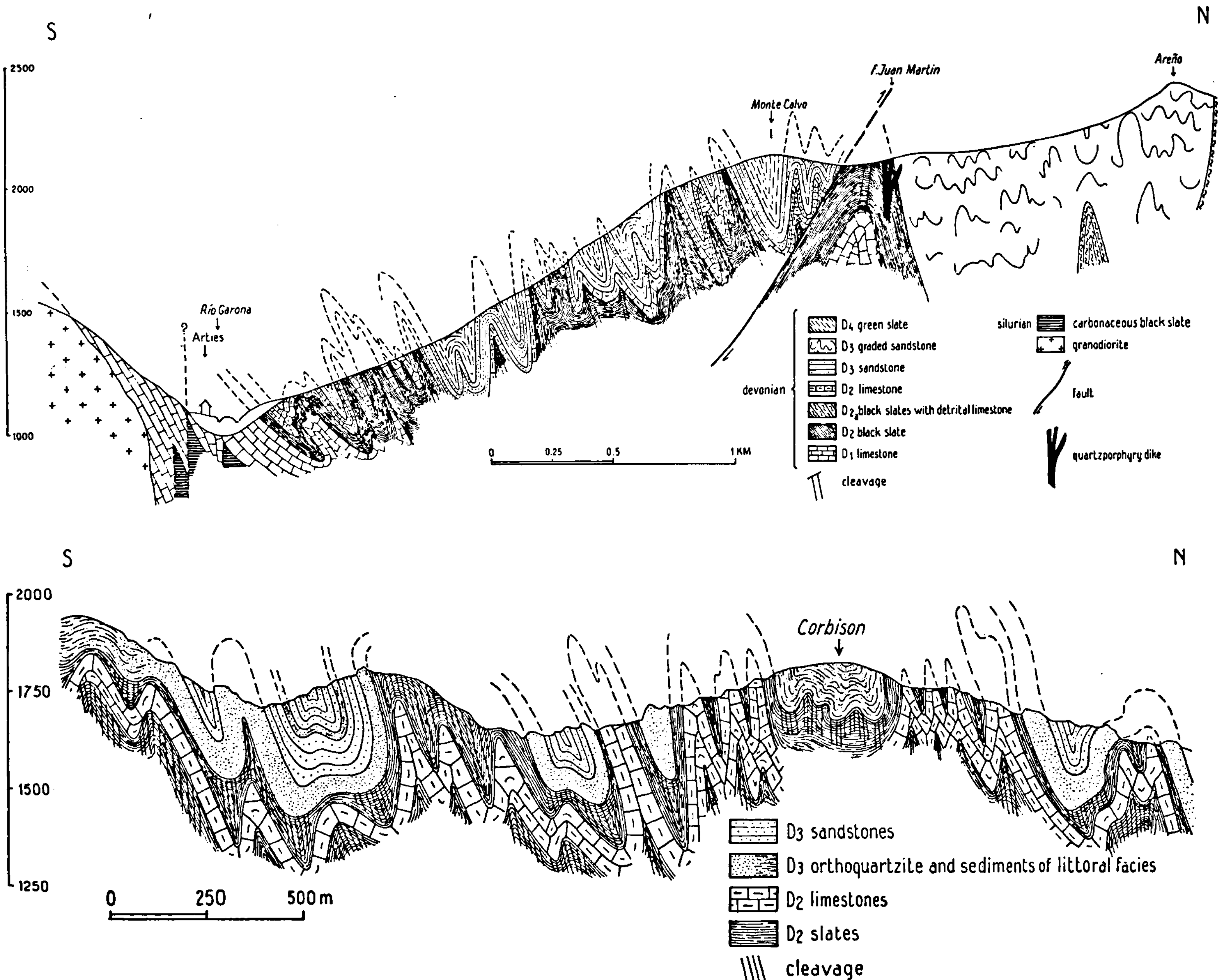


Fig. 23. Two sections through Devonian of the Valle de Arán (sheet 4), upper section north of Arties; lower section through Corbison (after Kleinsmiede, 1960).

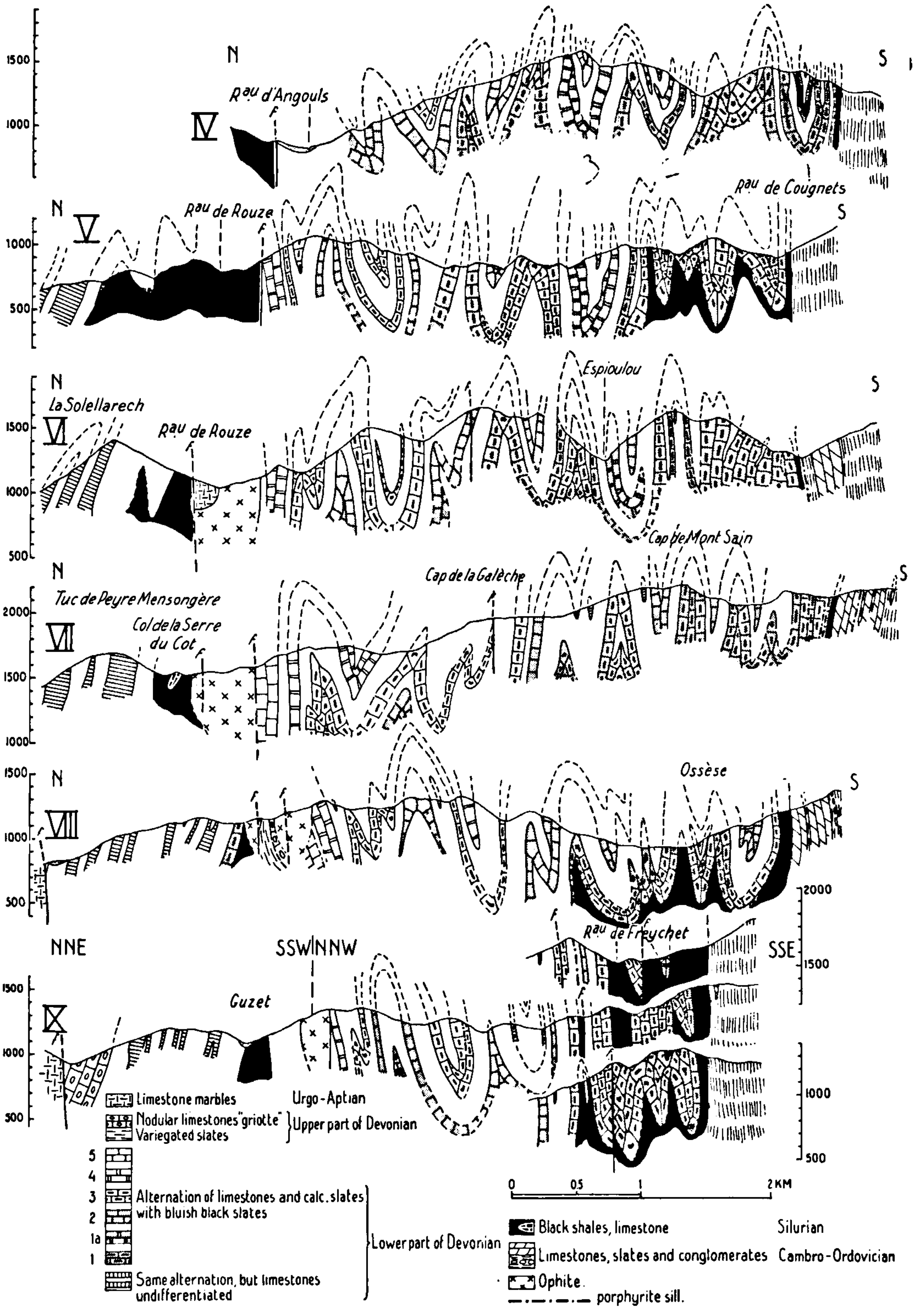


Fig. 24. Cross-sections through Devonian of Salat-Alet area (sheet 5). For location of sections see Plate I (after Zandvliet, 1960).

coincides with a fault along the southern limb of the Llavorsi syncline. This fault, the Llavorsi fault, probably of Alpine age, apparently has cut out a fairly thick rock sequence.

Infrastructure. – Structures within the metamorphic realm are characterized by their flatlying to horizontal attitude. This is clearly the case in the Aston-Hospitalet massif and in the Garonne dome, and outside our map area in the Canigou-Carança massif in the eastern Pyrenees. Later folding phases often have changed these low-dipping structures to a steep orientation. In our map area it has not been possible to locate large scale structures in these rocks, in contrast to the opinion of Guitard (1964), who has described large scale Penninic type nappes in the Canigou massif. Although it is not absolutely excluded that a large nappe or a recumbent fold occurs in the Aston massif, there is very little direct proof for it. In the Hospitalet massif and the Garonne dome certainly no nappes are exposed now.

The most obvious microstructure is a schistosity or slaty cleavage. In the augengneisses of the Hospitalet massif, which is a deformed homogeneous granite, a schistosity is defined by flat feldspar eyes surrounded by a quartz-micafabric curving around the feldspars. In addition an E–W lineation due to elongate feldspar and mica crystals is present. As the original granite had no layering, folds related to this schistosity are absent. Later folds, folding the schistosity are common in the Aston massif. In the micaschists, covering the augengneisses of both massifs, the same fabric elements, a schistosity and a lineation due to the shape of quartz and mica grains are found. In most cases this schistosity is parallel to a layering, which is certainly of sedimentary origin in some outcrops. Isoclinal folds of this layering with the schistosity in the axial plane do occur, but they are quite rare (Fig. 25). The size of the limbs of these folds does not exceed a few decimetres. They are often asymmetric but the presence of a constant sense of asymmetry could not be established.

In the Garonne dome, consisting of Cambro-Ordovician rocks, only the area around Bosost and Bagnères de Luchon is metamorphosed to amphibolite facies grade; in the remainder of this large structure the grade is very near the biotite isograd. In the whole of the Garonne dome the schistosity was originally flatlying, but has been steepened by later folds (Boschma, 1963). Isoclinal recumbent folds also occur in these rocks, but they are even more exceptional than in the Aston-Hospitalet massif. Fig. 25 gives a photograph of a thin section of such an isoclinal fold, with an axial plane slaty cleavage, and a later crosscutting crenulation cleavage.

No signs of folding earlier than this S_1 schistosity have been found in the Aston massif or in the Bosost area. Zwart (1965) thought earlier folds to be present south of the Hospitalet massif.

An interesting problem is the change from the vertical cleavage in low grade rocks to low-dipping schistosity in the high grade rocks. In the Garonne dome there is an abrupt change in attitude, best visible in a section E of Bosost. In the Cambro-Ordovician micaschists the schistosity has an undulating but usually rather gentle dip, the overlying Devonian has a vertical cleavage parallel to the axial plane of large and small folds. The Silurian acts as a decollement horizon, its lower part being parallel to the Cambro-Ordovician but occurring as steep, pinched anticlines in Devonian folds (Fig. 16). Nowhere in the Pyrenees is the particular behaviour of the Silurian so outspoken as in the Garonne dome, although elsewhere a certain degree of disharmony be-

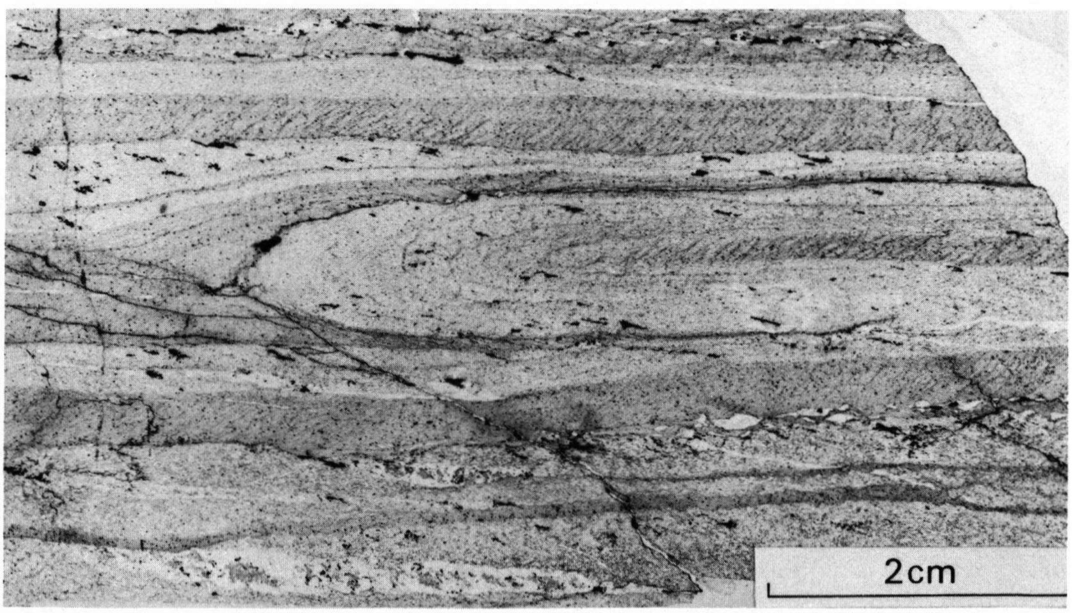


Fig. 25. Photomicrograph of thin section of isoclinal fold in Cambro-Ordovician quartzphyllite near Fos in Garonne dome (sheet 4). Slaty cleavage is parallel to layering, except in fold hinge. Oblique structure in pelitic layers is F_4 crenulation.

tween folds in the Cambro-Ordovician and in the Devono-Carboniferous is due to decollement on the very plastic Silurian slates.

The transition between the two types of structure is much more gradual in the Hospitalet and Aston massifs. This change is shown for the Hospitalet massif in the profiles of sheet 6. In the western end of the Aston massif it has been a subject of a special study by Lapré (1965) and Oele (1966). In that area the low-dipping schistosity gradually changes to a steep cleavage and has the shape of a keel of an inverted ship. For details concerning this problem I refer to Oele (1966). Further problems connected with these two structural domains, as microstructures in schists, superimposed folding, and an interpretation will be treated in a later section.

Post F_1 folding (Table 6)

In the Palaeozoic of the Central Pyrenees several deformation phases have been active after the main phase. The size of these later folds is, with few exceptions, small, and they have not changed the pattern of the geological map. Usually it concerns folds on a millimetre scale up to a size of several metres or tens of metres. Only in the northern border of the Lys-Caillaouas granite, major folds belonging to a late phase have been found. In addition the large structure, forming half a fan of the main phase cleavage, is a late feature.

As has been stated already, not everywhere in the map area these late folds have been studied. Our study was mainly concentrated on the Aston-Hospitalet massif, parts of the Garonne dome around Bagnères de Luchon and Bosost, the Lys-Caillaouas granite, parts of the Pallaresa anticlinorium, and the southern part of the axial zone. Elsewhere, like in all North-Pyrenean massifs, very little work has been done, and consequently there is still a large field open for further investigations. It will also be clear that the conclusions drawn in this section may have to be modified, as more results become available.

From our survey it became clear that locally at least four deformation phases have occurred after the main phase. In the past we have correlated these phases over the whole of our map area. This correlation was mainly based on overprinting criteria; furthermore it relies on orientation and style, and on relationships to metamorphism. It is realized that some of the last named criteria can lead to misinterpretations, especially if large areas are compared. However, the orientations and relation to metamorphism seem to be so constant, that we still adhere to interpretations made by Zwart (1962, 1963a, 1965), Boschma (1963), Lapré (1965), Oele (1966) and Hartevelt (1970).

The discussion can best be split in two sections according to locality. The first, northerly section, comprises the Aston-Hospitalet massif, the eastern part of the Pallaresa antiform, the Bosost area and the Lys-Caillaouas massif. The second section deals with the southern synclinorium.

The northern region. – In the region from the Aston-Hospitalet to the Lys-Caillaouas massif, the presence of three fold generations post-dating the main phase cleavage is well established on the basis of overprinting criteria. These are called F_2, F_3 and F_4.

F_2 folds. Folds postdating the main phase slaty cleavage or schistosity, and having low-dipping to horizontal axial planes and N–S directed foldaxes, have been described from the western part of the Aston massif by Lapré (1965) and Oele (1966), and from the Bosost area by Zwart (1963a). In the Aston massif the folds vary from open to isoclinal and have an S-asymmetry when looking north. The amplitude varies from a few cm to some tens of metres. There is an axial plane crenulation cleavage often grading into a new schistosity. They are overprinted by F_3 folds.

Folds with similar properties occur in the Bosost area in the Garonne dome, but they are not frequent. They fold S_1, but interference with other late fold has not been observed, and one cannot be sure that it concerns F_2 folds. In the Bosost area, however, there is ample evidence for a deformation phase causing rotation of porphyroblasts about N–S axes. This rotation postdates S_1 and predates F_3. It is assumed that the N–S folds in the Bosost area and in the Aston massif are related to these rotational movements, although there is no direct proof. Rotation with similar axes and sense of movement occur in the micaschists of the Lys-Caillaouas massif. N–S directed folds have not been found in that area. These microstructures and the rotation of minerals will be dealt with in a later section. It should be added that N–S folds have only been found in or close to higher grade rocks in the infrastructure. Finally it should be remarked that in our present interpretation these N–S folds have no relation to the folds in the granulite facies rocks of the North-Pyrenean massifs. These are now thought to be Precambrian.

F_3 folds. In the Aston-Hospitalet massif and the eastern part of the Pallaresa anticlinorium F_3 folds occur abundantly. They have been described by Lapré (1965), Oele (1966) and Zwart (1962, 1963a, 1965). According to these authors there are two sets of F_3 folds which together form a conjugate system. One set, F_{3a}, has NW–SE striking, steep to vertical axial planes, the other, F_{3b}, NE–SW striking, steep to vertical axial planes. The first set predominates over the second. These structures fold the compositional layering and S_1, and in case there is an S_2 schistosity or there are F_2 folds, there is a clear overprinting relationship, indicating the younger age of F_3 folds. Interference between the two sets (F_{3a} and F_{3b}) has been found, but no clear succession could be established. Therefore it is not certain that it concerns a conjugate system, as this hypothesis was solely based on the symmetric relationship of the two sets with the overall large scale structure.

As the F_3 folds occur both in the infrastructure and suprastructure, without change of the attitude of the axial plane, it is certain that the formation of these two domains predates F_3. The foldaxes vary within the axial plane from horizontal to vertical, depending on their occurrence in the infrastructure, in the suprastructure or in the transition zone (Fig. 26, and Oele, 1966). Also the shape of the folds depends on their locality. In the flatlying rocks of the Aston-Hospitalet massif, they are open to close and symmetric; in the steep slates tight to isoclinal and strongly asymmetric, the NW–SE set with an S-shape, and the NE–SW set with a Z-shape. The size of the folds varies from microfolds in crenulated schists and slates to mesoscopic folds with amplitudes of a few metres, and occasionally some tens of metres. In the steep phyllitic rocks F_3 folds are always accompanied by a well-developed crenulation cleavage.

Folds belonging to F_2 and F_3 are especially abundant in the transition zone between the infra- and suprastructure, that is in the western plunge of the Aston and Hospitalet massifs (Oele, 1966).

Some evidence that the axial planes of these folds are shearplanes is brought forward by Zwart (1963a) who found rotated cordierite and staurolite crystals in micaschists of the Hospitalet massif with a vertical rotation axis and a sense of rotation in accordance with the supposed shear direction.

Folds with similar orientations as F_3 have been found in the Bosost area (Boschma, 1963; Zwart, 1963). They postdate F_2 movements and predate F_4. They are usually of minor size, but occasionally they may reach dimensions of a few tens of metres.

F_4 folds. Folds with E–W striking steep to vertical axial planes occur frequently in the Aston-Hospitalet massif and in the Pallaresa anticlinorium (Fig. 27). They are less closely associated with higher grade rocks than the F_2 and F_3 folds, and occur at many places in the Cambro-Ordovician phyllites of the Pallaresa anticlinorium. The first description dates from 1954 by de Sitter who described some minor folds and cleavages from the Cardos valley, however without recognizing that these folds are due to a late folding phase. In this outcrop good overprinting relationships occur between F_1 and F_4 folds.

Overprinting criteria with F_3 folds are not rare in the Aston massif where they have been described by Lapré (1965) and Oele (1966). In the flatlying metamorphic rocks the F_4 folds are mostly symmetrical, have limbs of a size up to a few tens of metres, and not uncommonly they are chevron folds. Their degree of closure varies from close to tight.

In the steep, low grade rocks the folds are usually asymmetric and have an S-asymmetry looking west.

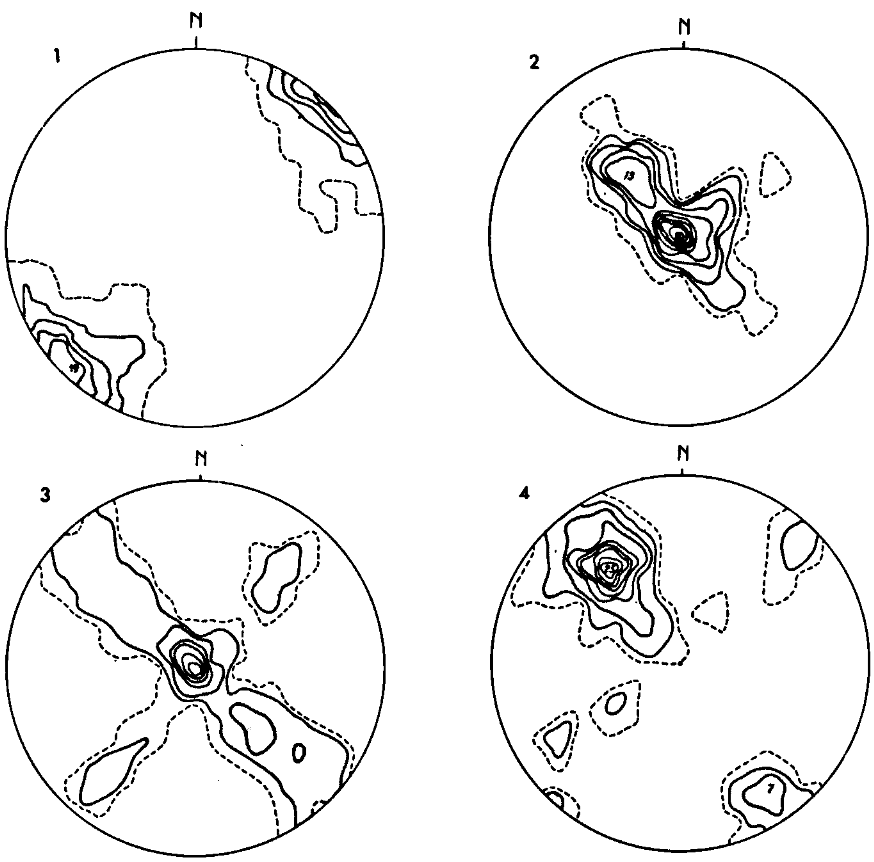


Fig. 26. Stereograms of F_3 folds in western plunge of Aston massif; 1: poles to F_{3a} crenulation cleavage; 2: F_3 foldaxes and lineations in steep phyllites of Hospitalet massif (south of Mérens fault); 3: F_3 foldaxes and lineations in transition zone of steep phyllites to gently dipping schists of Aston massif; F_{3a} is NW–SE; F_{3b} is NE–SW; 4: F_3 foldaxes and lineations in gently dipping schists of Aston massif; contours of diagram 1 at 1, 5, 9, 13 and 17% per 1% area; contours of diagrams 2, 3, 4 at 1, 2 and 5% per 1% area; each diagram has 100 measurements (after Oele, 1966).

This is related to the attitude of the S_4 plane which dips steeper to the north than S_1. S_4 folds are usually accompanied by a crenulation cleavage in the slaty layers of the rocks; it does not penetrate the quartzite layers (Fig. 28).

Due to their size F_4 folds have no effect on the large scale structure.

In the Bosost area in the Garonne dome a similar succession of fold generations has been described by Zwart (1963a) and Boschma (1963).

At least two sets of folds occur, which fold the S_1 ($+S_2$) schistosity. They have NW–SE and E–W striking steep axial planes. Based on correlation with the Aston-Hospitalet massif and on relationships with the metamorphism, it is supposed that the NW–SE set is F_{3a} and the E–W set F_4. The latter set is much more abundant than the NW–SE set, and occurs also frequently outside the Bosost area proper in the lower grade Cambro-Ordovician of the Garonne dome.

A profile north of the Bosost area has been described by Boschma (1963). F_4 folds are open to close, often of the chevron type, and usually have a crenulation cleavage in the pelitic layers. Some of these folds are distinctly larger than in the Aston massif, and may have amplitudes up to a few hundred metres. Parasitic folds of micro- to minor size occur abundantly on these larger folds.

The situation in the Lys-Caillaouas massif is insofar different that late folds of major size, attributed to F_4, are present. The massif occurs in, and forms part of the central anticline. On Profile 6a and on the profiles of sheet 7 and Fig. 29, a major Z-shaped fold, the Lac d'Oo antiform, is visible, folding the lithologic layering, a schistosity in the micaschists, and the sheetlike granite-micaschist complex.

There are obviously at least two generations of folds. An early generation consists of tight to isoclinal folds up to a size of a few hundreds of metres, folding the sedimentary bedding and thick layers of marble and conglomerate. There is a well-developed axial plane foliation formed by a parallel micafabric. The later generation

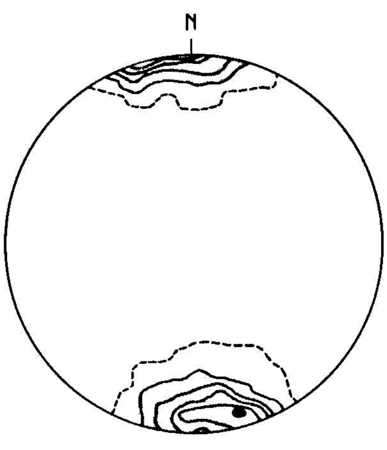


Fig. 27. Stereogram showing poles to S_4 cleavage, Val Ferrera (sheet 5) west of Aston massif; 100 measurements; contours at 1, 2, 5, 8 and 11% per 1% area (after Oele, 1966).

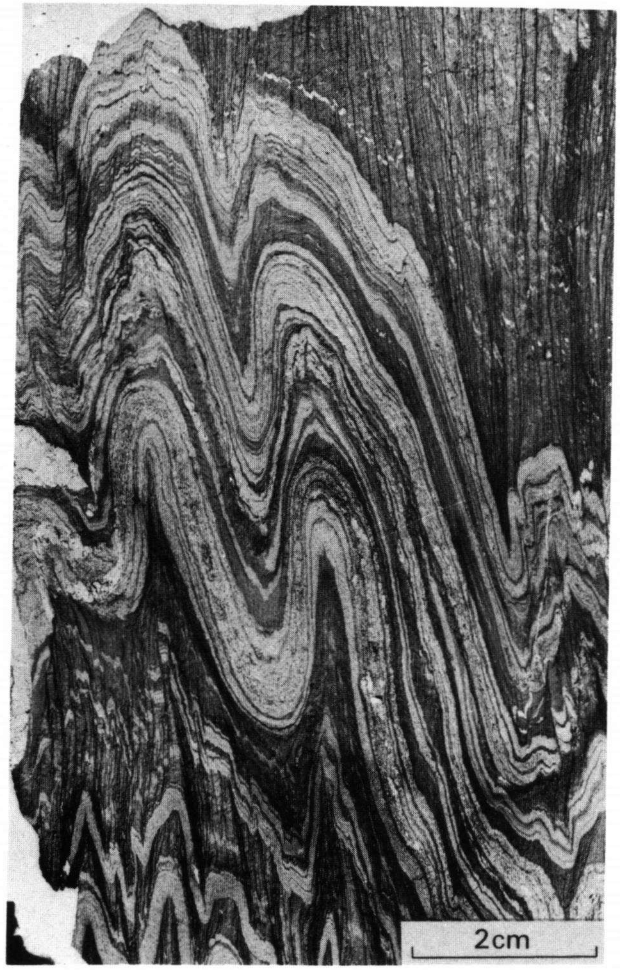


Fig. 28. Photomicrograph of thin section of Cambro-Ordovician quartzphyllite near Lago Areste (sheet 5, 6) west of Aston massif, showing an F_4 fold; S_4 cleavage only in pelitic layers.

forms the large Z-shaped Lac d'Oo antiform, folding bedding and foliation (Fig. 29). The granite with its sheetlike inclusions of micaschist is also folded by these late folds. The absence of a well-developed foliation in the granite, and the fact that the granite has crosscutting contacts with the micaschists, especially in the inclusions, demonstrates that the granite postdates the early set of folds. It is probable that this early set can be correlated with F_1 elsewhere. The late set can be correlated with F_4 based on the attitude of the F_4 axial plane.

Further to the west, SW of the granite body of Lys-Caillaouas another F_4 antiform, the Fredancon antiform (outside sheet 7) has been described by Clin (1964) and Trouiller (1976). According to the last named author, three fold generations occur in that area, of which the first one develops a schistosity and is correlated with our F_1. Of the following two, the last one can be correlated with the Lac d'Oo antiform and our F_4 and is responsible for the formation of the Fredancon antiform. This structure occurs south of the Lac d'Oo antiform, so that two major antiforms are present in the Lys-Cail-

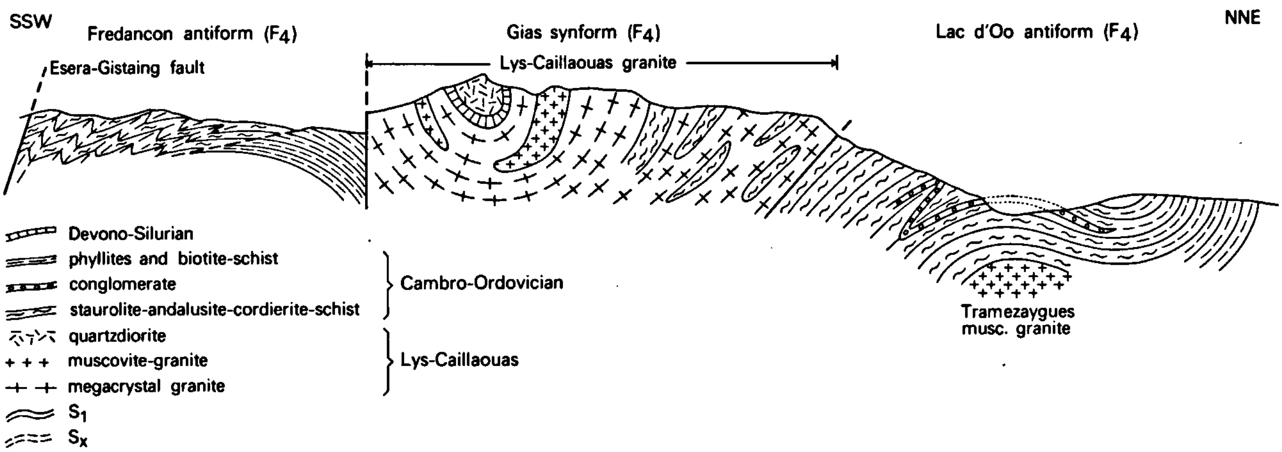


Fig. 29. Diagrammatic section through Lys-Caillaouas massif (Fredancon antiform is in fact lying further west).

laouas massif and surrounding micaschists. Microstructures confirm the existence of at least two fold generations, but in reality three or four deformation phases can be recognized (see p. 55, Tables 2 and 6).

According to our interpretation, S_1 had originally a steep attitude and is locally folded to a horizontal position by F_4. This means that this region belongs to the suprastructure. Some more evidence about the structural sequence can be gained from microstructures which will be dealt with in a later section.

Southern synclinorium. – From this region the F_0 and F_1 folds have been described already. The well-developed cleavage belongs to the F_1 folds. Locally S_1 has been folded in close to tight chevron folds accompanied by a crenulation cleavage which is especially important between the Mañanet and Tor Rivers (sheets 9, 10; Flamisell-Mañanet; Mey, 1967a). This foldset was designated by him as F_2. The axial planes dip moderately to the north, and the foldaxes have an E–W to NW–SE direction. As S_2 dips always steeper to the north than S_1, the folds are asymmetric and have an S-asymmetry when viewed in westerly direction. Locally, where F_2 is very intense the position of S_1, which generally has a gentle N dip in this region, has been completely changed, even to steep dips to the south (Profile 4). It is possible that F_2 in the south can be correlated with F_4 in the north if the attitude of the axial planes and other properties are taken into account. However, a direct correlation has not been made, although observations in the Tor valley by Boschma (1963) seem to confirm the correlation.

In the eastern part of the map area, on sheet 10, Hartevelt (1970) has described two sets of late folds, which as far as directions are concerned, may be correlated with F_3 and F_4 in the north. They are mainly developed as kinkbands.

It is evident that the southern part of the axial zone is quite strongly influenced by Alpine folding. Alpine cleavages in Palaeozoic material have been described by Seguret (1970), Muller & Roger (1977), Majesté-Menjoulas (1979). It is therefore not excluded that F_4 ($=F_2$ of Mey) in the southern synclinorium is of Alpine age. We have no direct evidence that it concerns a Variscan cleavage, although in earlier papers of our group this has been assumed. There are two arguments that late crenulation cleavages have a Variscan age. One is that in hornfelses south of the Maladeta granodiorite, Mey (1968) has found a folded slaty cleavage with a new crenulation cleavage (that is at least F_2) overgrown by cordierite crystals formed by contact metamorphism of the granodiorite, which has certainly a Carboniferous age. Secondly in Stephanian conglomerates near Malpas, pebbles of Palaeozoic material have been found, containing two cleavages. Consequently it is certain that two cleavages of Variscan age do exist. This does not prove, however, that all cleavages in the southern part of the axial zone or in the Nogueras zone are of the same age. There may be Alpine cleavages and folds as well.

Kinkbands. – At many places in the Palaeozoic of the axial zone, kinkbands occur in well-cleaved rocks. They usually are developed as one set and have a constant relation to S_1. The main cleavage is vertical in the north, and has less and less steep dips towards the south. The kinkbands have a shallow dip in the steep slates and a southerly steep dip in low-dipping slates. Their sense of asymmetry is always the same, S-shaped when looking west. Intersections of kinkbands with the main cleavage are usually close to E–W. When F_4 folds are present, as in the Pallaresa anticlinorium, the kinkbands appear to be later. In the south where S_1 is close to horizontal, conjugate systems of kinkbands may be found (see also Zandvliet, 1960 and Zwart, 1963a).

THE CLEAVAGE-FAN IN THE AXIAL ZONE

In the suprastructure of the axial zone one more major structure can be discerned. When only the attitude of the main phase cleavage is considered, irrespective of lithology or stratigraphy, it appears that S_1 is arranged in a fan-like structure.

In the northern part of the axial zone the cleavage has a steep dip to the south, then a zone with vertical cleavage occurs, whereas further to the south the cleavage gradually gets a shallower dip to the north, until it finally is close to horizontal (Profiles 2, 3, 5, and Fig. 30). Thus the main phase cleavage forms a half of a large fan structure. The origin of this fan has been a controversial issue within the Leiden group. Zandvliet (1960) has done some detailed work on it. He has shown that in fact the cleavage does not change its dip gradually, but there are compartments with the same dip, separated from other compartments with a somewhat different dip (Fig. 31). According to Zandvliet there occur faults between the different compartments. In his interpretation the formation of the fan is a late Variscan structure due to uparching of the northern part of the axial zone, resulting in a gravity-induced tilting of the cleavage. This feature probably predates the intrusion of the Maladeta granodiorite, as it is in contact with low-dipping slates. Zandvliet also suggests that the formation of kinkbands is related to the fan formation. This interpretation was accepted by the present author (Zwart, 1963).

Mey (1967a) did not agree with this conclusion. He assumes that in the southern part of the axial zone the cleavage was originally formed with a gentle dip. He puts forward some arguments for this hypothesis (Mey, 1968, p. 200), one of his main arguments being his inability to find the faults along which the cleavage should have been tilted. The same fan structure continues to the west, where it has been described by Muller & Roger (1977). They interprete it also as a feature postdating the formation of an originally steep cleavage. According to these authors it is partly of late Variscan, partly of Alpine age.

The present author is of the opinion that it concerns a late Variscan feature, formed in the way Zandvliet envisaged. Due either to N–S tension, or to vertical uplift in the northern part of the axial zone, the cleavage became rotated to a less inclined position. This could take place along many normal faults parallel to the cleavage with a downthrown northern block. These faults stand out as zone of stronger weathering and thus can be identified in the field.

As can be deduced from the attitude of the unconformable Permo-Triassic, it is quite probable that the southern part of the fan structure has been influenced by Alpine movement, producing an additional tilting.

In addition kinkbands are especially frequent near these faults and indicate the same stress field with a

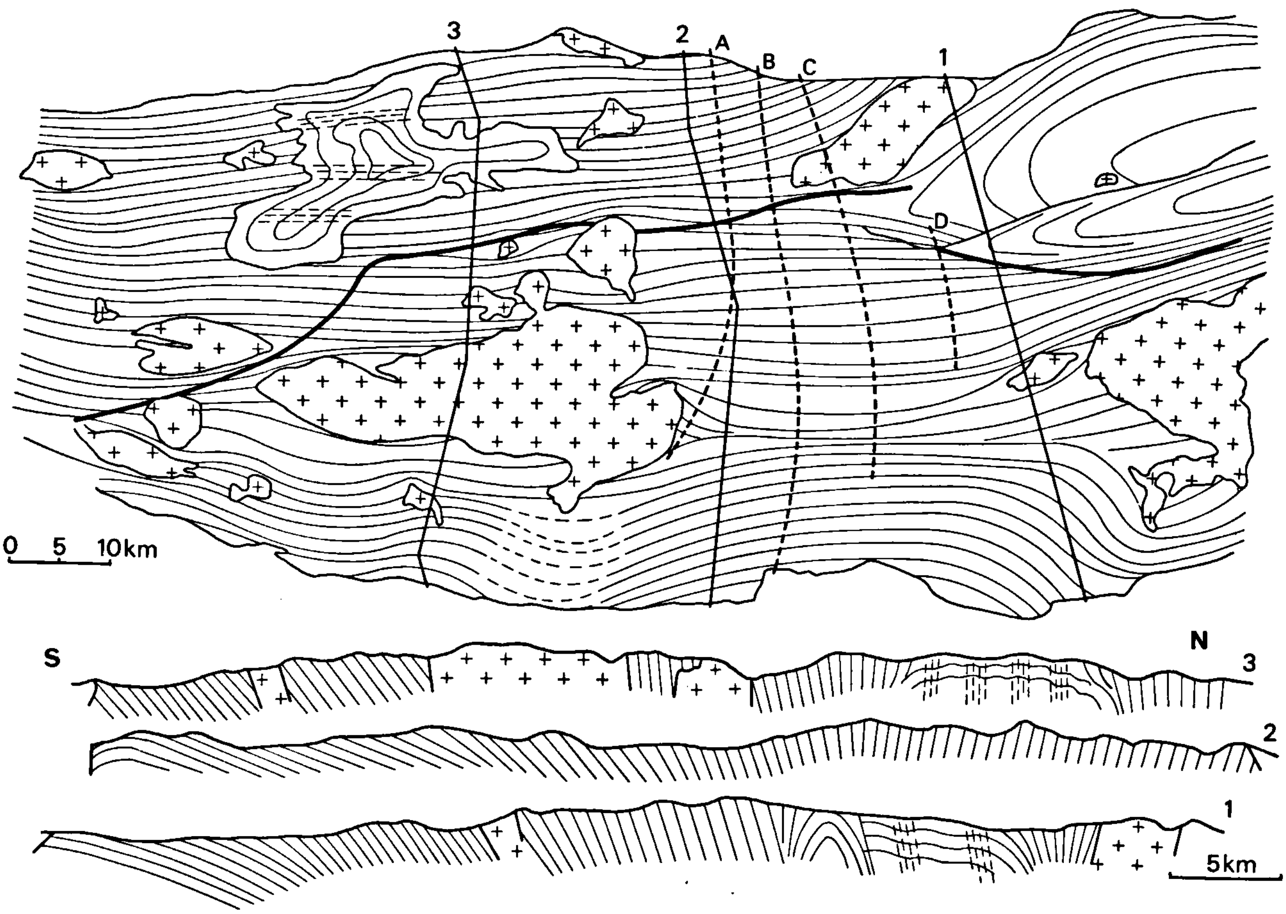


Fig. 30. Map of axial zone of central Pyrenees with strike lines of S_1 cleavage (full drawn lines) and S_4 cleavage (striped lines), where dominant; three cleavage profiles (1, 2, 3) and location of profiles of Fig. 31 (A–C). Thick line separates regions with S-dipping cleavage from regions with N-dipping cleavage.

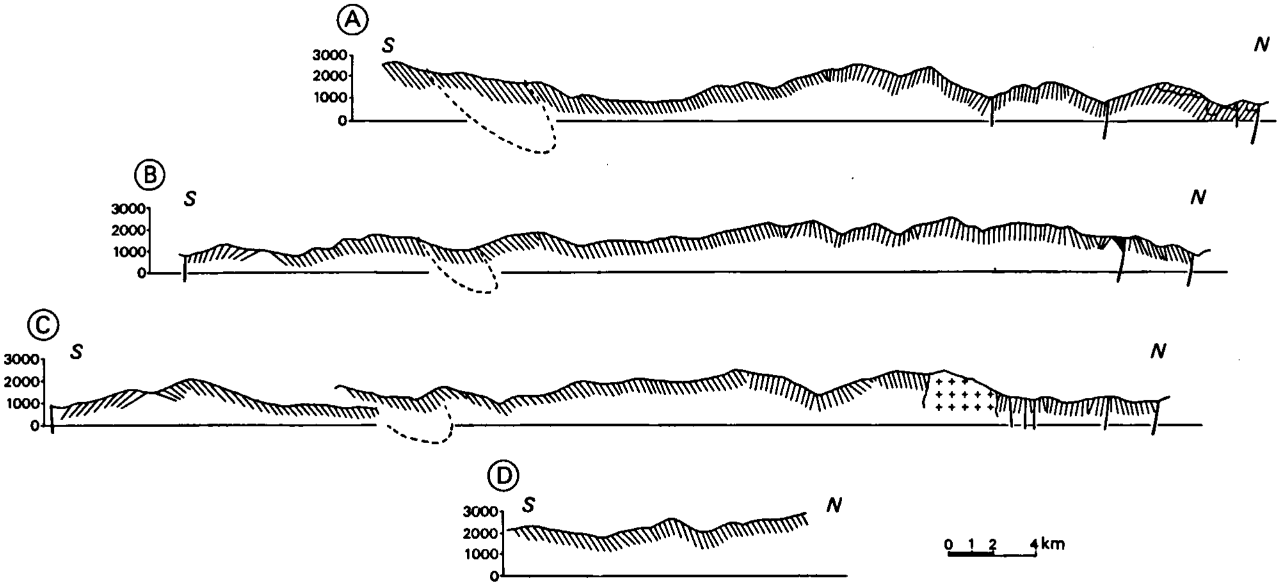


Fig. 31. Cleavage profiles; for location see Fig. 30 (after Zandvliet, 1960).

vertically directed maximum stress and a N–S tension. The mechanism is similar to that described by Hoeppener (1955) for the Rhenic slate belt.

THE GRANODIORITE BATHOLITHS

Several large bodies of granodiorite occur in the axial zone, as well as in the North-Pyrenean massifs. They usually have an elongate shape parallel to the trend of the major structures. They are, however, clearly cross-cutting these structures and postdate them. This is also shown in the contact metamorphic aureoles around the granodiorites, as the contact metamorphic minerals like andalusite and cordierite always overprint the slaty cleavage. In some cases they also overprint late crenulation cleavages like around the Maladeta batholith (Mey, 1968). In other cases as near the Bassiès-Auzat body, crenulations belonging to F_4 deform cordierite and andalusite crystals. This indicates that some granites intruded during or even before the last folding phases. On the other hand the Maladeta granodiorite intrudes already tilted S_1 cleavage, and therefore probably predates the formation of the large fan structure.

On Fig. 30 which is a cleavage strikeline map, it can be seen that the cleavage abuts against the granodiorite bodies, but at the same time the cleavage wraps more or less around the small and large batholiths. This bowing-out effect is interpreted as being due to the shouldering aside by the intruding body, making its way towards shallower levels by forceful intrusion. This accounts at least partly for the space problem involved in the emplacement of these granodiorites. It is remarkable, however, that only a large scale effect has been observed, and that no small scale structures were produced in the slates surrounding the granodiorites. Late crenulations are present, but these occur also elsewhere far away from any granodiorite body. Emplacement and making room for the granodiorite batholiths can only partly be explained by shouldering aside the country rock. There must be at least one more mechanism by which these bodies create room for themselves. It seems likely that a part of the rocks which occurred above the large batholiths has been uplifted during intrusion, and has subsequently been eroded away.

Many large joints occur in the granodiorite bodies. Those shown on the maps have mainly been taken from aerial photographs. In some cases field observations in these joints indicated the presence of mylonitic or even phyllonitic rocks, although displacements on these joints cannot be proved. The age of jointing is difficult to establish, it could be late Variscan or Alpine.

DISCUSSION OF DIFFERENT INTERPRETATIONS

In a number of publications by Mattauer & Seguret (1966), Matte (1969), Seguret & Proust (1968a, b), Déramond (1971) and Déramond et al. (1971) the interpretations as given above have been challenged.

According to their interpretation the schistosity or cleavage in the whole Palaeozoic, from the Cambro-Ordovician to the Carboniferous was born in a horizontal position. In the Devono-Carboniferous it is folded by a third phase to a steep position with the development of a new cleavage. In the infrastructure it gave rise to the large dome or antiformal structures as the Canigou, Aston and Hospitalet massifs.

The main objection against their interpretation is that in the steep Devono-Carboniferous in the northern part of the axial zone, all structures, that is bedding and cleavage, are steep to vertical so that there do not occur any recumbent folds, as is claimed. As has been shown already, earlier folds do exist, but they are not recumbent, but have rather steep axial planes. The second objection is that the cleavage in the steep folds is a well-developed slaty cleavage, which never shows signs of an earlier cleavage despite the examinations of hundreds of thin sections. All crenulations belong to a later set, cutting obliquely through the first folds.

To the present author's opinion there is a confusion with our F_0 folds, which, however, were not recumbent but inclined to vertical, and did not, or only very occasionally, possess an axial plane cleavage. It is possible that further west, outside our map area early folds are recumbent and have a cleavage (Majesté-Menjoulas, 1979), but this is not the case in the region we have mapped. In the area referred to by Matte and Déramond, we have found good examples of F_0 folds. Our interpretation is confirmed by Muller & Roger (1977), who found early, pre-cleavage folds west of the Garonne dome, and especially in the westernmost part of the axial zone, with the same properties as our F_0 folds.

From the French publications it appears that their S_3 must correspond to our F_4. Our F_3 has not been found by these authors despite its very frequent occurrence (Table 6).

Another problem is the division in supra- and infrastructure. In earlier publications the present author has put forward the view that steep and horizontal structures in both domains were formed simultaneously. This view has been challenged by Matte (1969). Our interpretation was mainly based on the reasoning that both domains originated before a certain metamorphic event, and result from the first visible deformational event. However, subsequent observations have brought forward some contradictory evidence, and the simultaneity of both structures is not certain. Unfortunately it is difficult to propose an alternative solution, if our original and the French hypothesis are rejected. I plan to discuss this matter in the section on microstructures.

Yet another problem is the antiformal shape of the large metamorphic massifs. Although the division in supra- and infrastructure must represent an early event in the Variscan structural history, we are convinced that late folding, i.c. F_4 folding has modified these large structures, and it is likely that their present shape is at least in part due to late doming, related to F_4.

POST-VARISCAN STRUCTURES

Although this paper deals mainly with Variscan structures, a few words have to be said about later, Alpine structures. During our survey little attention was paid to these structures, and they have not been fully appreciated during the mapping period. However, good reviews of the Alpine structures have been given by Seguret (1970) and Choukroune (1974) to whom I refer for more details. This short description will deal with the northern border of the axial zone and the North-Pyrenean massifs, with the southern border and the Nogueras zone, and with Alpine deformation within the axial zone.

The northern border

The axial zone is on its north side bounded by a major fault, the North-Pyrenean fault, which cuts obliquely through the Variscan structures (Plate 1). Immediately north of the fault occurs a zone of strongly folded, and in part metamorphosed Mesozoic rocks, which has been the subject of study of Ravier (1957), Avé Lallement (1969) and Choukroune (1974). Within this zone occur the North-Pyrenean massifs as islands of Palaeozoic rocks. In our map area the deepest exposed Precambrian rocks are always found on the southern side of these massifs, indicating an uplift of the northern block along the fault of up to 5000 m or more. In several publications it has been mentioned that strike-slip movements up to hundreds of kilometres have occurred along the North-Pyrenean fault (Le Pichon et al., 1971). In the area we studied, there is no direct evidence for such large movements. In fact the similarity of the Palaeozoic north and south of the fault is quite great and does not favour important transcurrent movements, although this evidence is not so compelling that they are excluded. Strike-slip movement of the Trois Seigneurs massif along this fault to an amount of about 10 km has been postulated by Zwart (1954) but the correlation based on granite stocks is not very certain.

As rocks of Mesozoic age not only rest unconformably on the Upper Palaeozoic sediments, but also on the deeply eroded metamorphic rocks, they must have been uncovered shortly after the Variscan folding. It seems therefore likely that these massifs have been uplifted already at the end of the Variscan orogeny, and that consequently the North-Pyrenean fault dates already from this time.

Movements during the Triassic and related to the intrusion of ophites have been suggested by Muller & Roger (1977). There can be no doubt that important movements during the Alpine orogeny in late Cretaceous or early Tertiary time also took place, probably related to the opening of the Bay of Biscay, the rotation of Spain, and the compression between the Iberian peninsula and stable Europe.

Within the Palaeozoic rocks of the North-Pyrenean massifs there is some effect of the Alpine movements. Cataclasis in the gneisses near the fault is common. The Palaeozoic sediments of the northern border of the massifs must have been folded together with the Permo-Triassic unconformable cover, as these rocks have a steep to vertical position (Profiles 1, 2).

The southern border

The southern border of the axial zone is much more complex than the northern border. In the south the Palaeozoic is unconformably overlain by folded Stephanian, Permian and Triassic, beyond which appear a number of blocks of Palaeozoic rocks, forming the Nogueras zone. The Palaeozoic of the Nogueras zone consists solely of Devono-Carboniferous. There are no older rocks, nor metamorphics. The structure of these blocks is complicated and has been worked out for some of these areas (Mey, 1968; Hartevelt, 1970).

There has been a controversy between Seguret (1964, 1966) and our group (Mey, 1968; Zwart & Mey, 1965) about the allochthonous vs. autochthonous position of the Palaeozoic blocks of the Nogueras zone. It seems now, that seen in the light of new evidence, like drill holes, the interpretation of Seguret is basically correct, and that the Nogueras zone has been derived from the axial zone in the north by thrusting and/or gliding.

The stratigraphy of the Devonian gives some constraints on the origin of the different blocks of the Nogueras zone, as some of the blocks consist of Devonian in Sierra Negra subfacies, like the Gotarta-Malpas, Las Iglesias, Sta. Coloma, Castels and Montsech de Tost block, whereas the Feixa and Erdo blocks contain a Devonian of the Compte subfacies (Fig. 6, Plate 1). From the same figure it is clear that the Compte subfacies in the axial zone occurs SE of the Sierra Negra subfacies. Therefore the Gotarta-Malpas, Las Iglesias and Sta. Coloma blocks were derived somewhere from the axial zone in the north, and root probably in one of the Alpine thrusts south of the Maladeta granodiorite. The Erdo and Feixa blocks probably originate from the Orri dome, whereas the Castels and Montsech de Tost blocks must have their roots in the Llavorsi syncline, most probably in the large fault along the southern side of this syncline. For these reasons the suggestion of Muller & Roger (1977) that the rocks of the Nogueras zone are derived from the Mérens fault in the Aston-Hospitalet massif must be discarded. Also Seguret's profile G (1969) in which the Feixa block roots in the Llavorsi syncline is in our opinion not correct. The Castels block is originating from this syncline, and the Feixa block must have a somewhat more southerly origin.

Alpine structures and faults in the axial zone

Alpine structures, like folds and cleavages in the axial zone have been described by Mattauer (1964), Mattauer & Seguret (1966), Matte (1969), Mey (1968), Wennekers (1968), Muller & Roger (1977).

From our work and the literature it appears that the influence of the Alpine folding in the axial zone gradually decreases in intensity from west to east. This is probably related to the greater depth of erosion in this direction, through which deeper levels, further away from the Alpine influence are reached. In our map area Alpine movements have mainly taken place along faults, and to a limited extent only by folding and cleavage formation. Numerous faults occur in the axial zone, among which a number of large ones, like the Mérens fault (Aston-Hospitalet massif), Bosost fault (Garonne dome), the Esera-Gistain fault (Lys-Caillaouas), the Llavorsi fault, and several faults in the southern synclinorium, as the Sahun, Castanesa, Senet and Bono faults or thrusts (Plate 1). In the northern part of the axial zone the faults have a steep to vertical dip; towards the south they have a northerly dip and are reverse faults, whereas those in the southerly synclinorium are gently dipping to the north and can be called thrusts. The movements on all these faults, including the North-Pyrenean fault, have always the same sense, that is an upthrow of the northern block. Many of these faults have a dip close to the S_1 cleavage, but there are exceptions like the Mérens fault, occurring mainly in flatlying crystalline rocks. The throw on many of these faults is considerable. For the North-Pyrenean fault, which may have an unknown lateral displacement, the vertical movement may reach 5 km, but for the other faults in the northern part of the axial zone it is considerably less, but amounts at least from several hundreds of metres up to more than thousand metres. The displacement on the reverse faults and the thrusts in the south is more difficult to estimate. Also dating of the movements on all these faults is in many cases not possible. In certain cases, like the Mérens and Bosost faults we have assumed Variscan movements based on the occurrence of recrystallized mylonites (under greenschist facies conditions), but this evidence is not conclusive. During the late Cretaceous-early Tertiary there was a thermal event along the northern side of the axial zone which may have been responsible for this recrystallization. This is confirmed by mica ages of Variscan rocks which are rejuvenated and now have Cretaceous or mixed ages. In the author's opinion, fault movements probably date back to Variscan times but they may have occurred again in Alpine time. For a number of other faults, Alpine movements are beyond doubt, as Mesozoic rocks occur in the fault zones, like in the Valle de Arán (sheet 4) between the Arties and Aguamoix valleys, and in the Sahun, Senet and Bono thrusts (sheets 7, 8). Also the Llavorsi fault, being the root of a part of the Nogueras zone, is probably largely an Alpine feature. From the different nature of the S_1 cleavage, slaty cleavage to the north, crenulation cleavage on sedimentary bedding to the south, it can be deduced that a thick succession of the Palaeozoic is cut out by this fault. The same probably applies to the thrust faults occurring south of the Maladeta batholith. It is impossible to estimate how much the displacement on these thrusts is, but it is certain that important shortening of the axial zone has taken place during Alpine movements.

The question arises whether there are also penetrative structures as folds and cleavages of Alpine age in the Palaeozoic. Good proof of such Alpine structures has been given by Muller & Roger (1977) and Majesté-Menjoulas (1979) to the west of our map area. Their occurrence in our area has also been suggested by a.o. Mattauer (1964), Choukroune & Seguret (1973) and Matte (1969). They proposed that part of our F_4 could be Alpine. Variscan age of F_4 can with certainty only be established in metamorphic regions like the Aston, Bosost and Lys-Caillaouas massifs, where andalusite and cordierite overprint F_4 structures or where F_4 crenulations have been recrystallized to polygonal arcs. Around the Maladeta granodiorite where F_4 folds are contact metamorphosed, there is also no doubt about its Variscan age.

In the southern part of the axial zone, where no regional metamorphism of higher grade is present, it is much more difficult to establish the age of these late folds. As has been remarked already it is not excluded that some of these structures are indeed of Alpine age. According to Mey (1967a) the Variscan cleavage has been reactivated by Alpine movements close to the unconformity (Fig. 32). Also the presence of a cleavage in the isolated outcrop of Triassic in the Valle de Arán near Baños de Tredos has been used as an argument, as these rocks show a reasonably well-developed cleavage. However, the Palaeozoic rocks nearby, like the Maladeta granodiorite and Devonian limestones do not show any sign of a cleavage, so that this argument is not very helpful either. It is very well possible that the Triassic was folded in a decollement style, or that the cleavage was acquired by fault movements, as close to the Triassic outcrop several faults occur.

The conclusion is, that to the author's opinion no proof exists for important penetrative movements and small scale folding with cleavage development in the Palaeozoic rocks during the Alpine orogeny, except locally near the unconformity with late Palaeozoic or Mesozoic rocks.

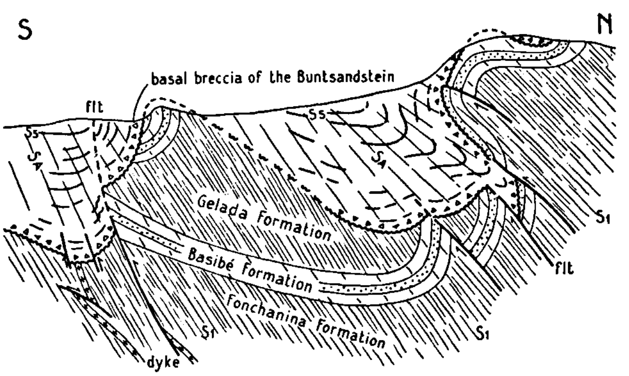


Fig. 32. Relation between Alpine structures in Triassic and reactivated Variscan S_1 cleavage (after Mey, 1967).

MICROSTRUCTURES

Introduction

During our survey in the Central Pyrenees, a great deal of interest in microstructures was developed. These microstructures have been studied in low grade slates and phyllites, as well as in higher grade metamorphics. One of the first publications on this subject is from de Sitter (1954) in which an outcrop of Cambro-Ordovician phyllites in the Pallaresa anticlinorium was described. It is from these rocks that the term 'microlithons' was coined. Photomicrographs of this outcrop were also published in de Sitter's book 'Structural Geology'. More work on slaty cleavages and crenulation cleavages in general was done by Zandvliet (1960), Kleinsmiede (1960), Lapré (1965), Oele (1966), Mey (1967a) and Hartevelt (1970), although little about microstructural details was published by these geologists. The present author has especially studied the microstructures of micaschists (Zwart, 1962, 1963). Porphyroblasts of various minerals in these schists turned out to be very suitable for examining the relationships between metamorphism and deformation.

Slates and phyllites

S_1-cleavage. – In slates and phyllites of Cambro-Ordovician, Devonian and Carboniferous age, various cleavages are usually present. They can be classified as slaty cleavage and crenulation cleavage. Moreover differentiated layering is often present. The late folding phases usually produce foliations of the crenulation cleavage type, the main phase foliation may be a slaty or a crenulation cleavage. In this section only cleavages in pelitic and quartzitic rocks are discussed. Cleavage development in calcareous rocks is a separate problem, on which only few observations have been done.

The nature of the cleavage depends, at least partly, on the locality and on the age of the rocks. In general the main phase cleavage S_1 is of the slaty type in the northern part of the axial zone, and of the crenulation type in the southern part. The same change takes place in rocks from old to young. However, this is only a tendency and there are many exceptions.

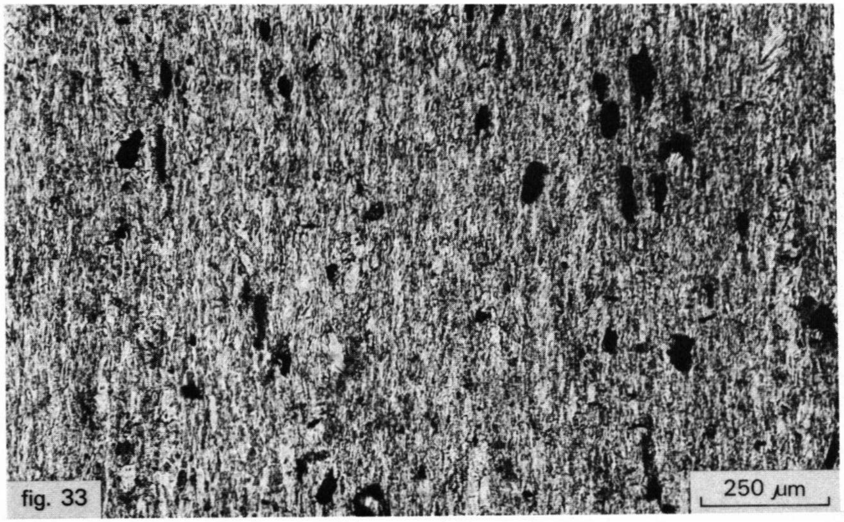
The best slaty cleavages are developed in the Cambro-Ordovician north of the Llavorsi syncline. In these rocks, consisting of white mica, chlorite and quartz with accessory opaques, there is a good parallel orientation of most minerals (Fig. 33*). The phyllosilicates have a lattice orientation, the quartz and the opaques only a shape orientation. There is, however, usually a certain amount of unoriented material, consisting of elongate, eyeshaped chlorite- or micastacks which have their basal cleavage planes at a large angle to the rock cleavage, and of small micacrystals cutting across the cleavage. As a result of the presence of the chlorite- and micastacks the cleavage has a slightly anastomosing pattern. From a comparison with other thin sections it appears that these stacks are probably remnants of an early stage of cleavage development.

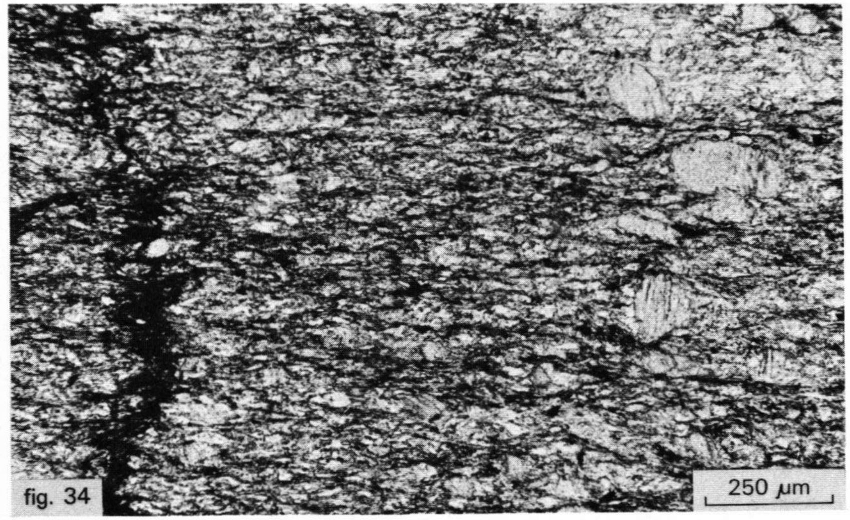
In slates of the Pallaresa anticlinorium, even several km west of the metamorphics of the Aston massif, small, stubby grains of biotite may occur in these slates. The same applies to slates of the Cambro-Ordovician of the Garonne dome. Due to the small size of these crystals it is difficult to determine the relationship to the slaty cleavage, but it looks as if they are crosscutting and formed late.

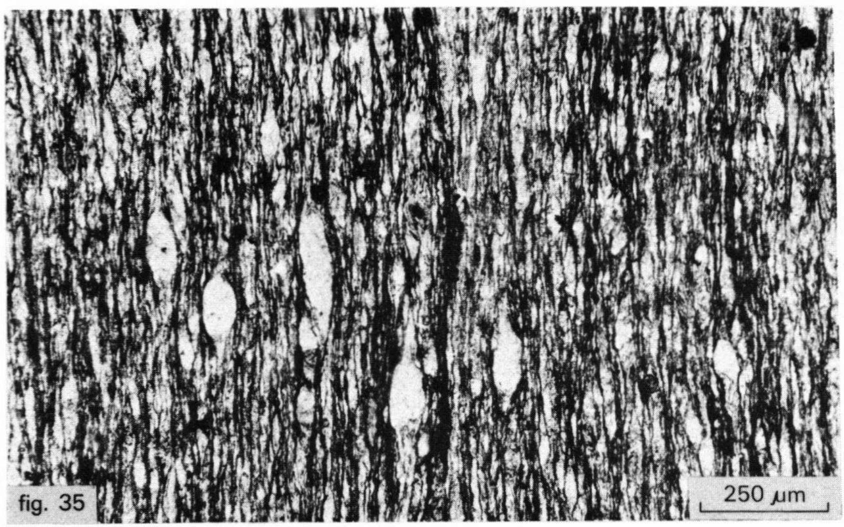
In Fig. 34 a section of a Cambro-Ordovician slate from the Massana anticline in Andorra is shown. The cleavage makes a large angle with the bedding. The black, graphite layer is crenulated. The cleavage consists of an anastomosing orientated mica and chlorite fabric, but there is a large amount of unoriented material between the cleavage planes, mainly made of large ovoid deformed chlorite stacks. Although in handspecimen there is a perfect cleavage, the thin sections show a far from perfect slaty cleavage.

Fig. 35 is a thin section from a Devonian slate from the Ribagorzana valley. There is a well-developed cleavage, especially marked by thin dark layers, which could not be identified very well, but which presumably contain oriented mica. Between the cleavage planes most of the material is unoriented. Lens-shaped chlorite-mica stacks are quite numerous, and are similar to those described from the previous sections. Furthermore there

* Most figures from Figure 33 onwards are photomicrographs of thin sections.







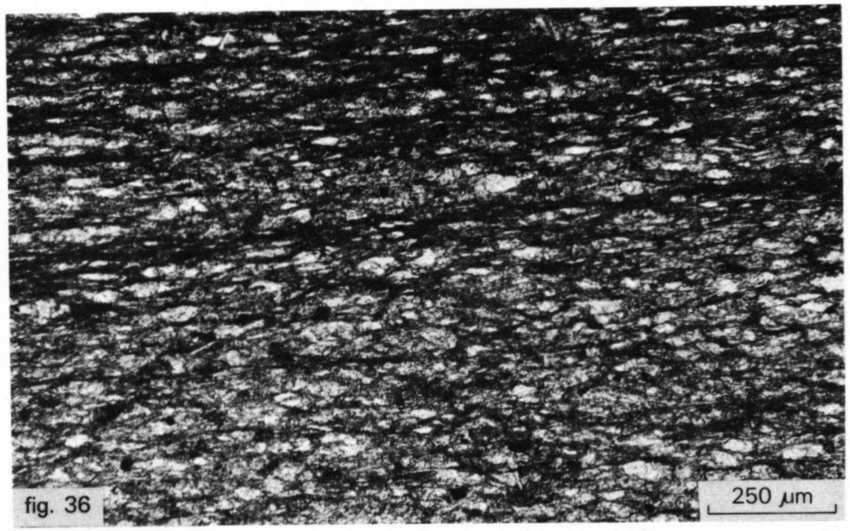


Fig. 33. Cambro-Ordovician slate from Cardos valley (sheet 5).

Fig. 34. Cambro-Ordovician slate from Massana anticline, Andorra, sheet 6.

Fig. 35. Devonian slate, Ribagorzana valley, sheet 8.

Fig. 36. Devonian slate near Gessa, Valle de Arán, sheet 4.

is much unoriented micaceous material between the cleavage planes. There is no bedding, nor any crenulation visible in this section.

A thin section from a Devonian slate near Gessa in the Valle de Aran is shown in Fig. 36. There is a well-developed slaty cleavage with oriented phyllosilicates anastomosing around flat quartz grains. The cleavage is also visible as black films with unidentified material. The angle with the bedding is a few degrees only. With crossed polarizers the presence of much badly oriented micaceous material is visible, and the cleavage is less perfect than visible in Fig. 36.

The Carboniferous of the Civis Formation in the Llavorsi syncline consisting of homogeneous dark slates shows some interesting microstructures as shown in Figs. 37 and 38. In Fig. 37 it can be seen that there is a fairly well developed micafabric anastomosing around large chlorite-muscovite stacks and around small, lath-shaped chloritoid crystals. The chlorite-muscovite stacks are clearly deformed and like in the previous sections must have formed relatively early. In most handspecimens of these Carboniferous slates, muscovite flakes are visible on the cleavage. They are shown on Fig. 38 where many of the muscovites are eyeshaped and more or less oriented in the bedding which makes an angle of about 20° with the cleavage. These muscovites are not intergrown with chlorite as in Fig. 37. The orientation in the bedding makes it likely that they are of sedimentary origin. This is certainly not the case with the chlorite-muscovite stacks, as large chlorites do not survive as a detrital mineral. Therefore these stacks must have grown in an early stage of cleavage formation, perhaps even earlier, during diagenesis. In Fig. 39 from the same thin section as Fig. 37, many chloritoid laths are visible, besides chlorite-muscovite stacks. Some chloritoids penetrate the stacks and are clearly of a later origin. The relationship of the chloritoid with the slaty cleavage can be observed in Fig. 40. The slaty cleavage, in this case consisting of a well-developed orientation of phyllosilicates, is continuous in the chloritoid as an internal fabric. The chloritoid crystals are, however, deformed, and the cleavage curves around the crystals. Therefore the chloritoids postdate the main cleavage development, but in a late phase of flattening on this cleavage the crystals are deformed. As later crenulations are rare in the Llavorsi syncline, it is unlikely that the flattening is related to these late movements. The ubiquitous occurrence of chloritoid in the Carboniferous slates of the Llavorsi syncline is somewhat enigmatic, but they have been found in many specimens from Andorra to the Pallaresa River. Presumably the rock chemistry is important. On the other hand chloritoid is clearly indicative of the greenschist facies, but there is no relation with any of the regional metamorphic areas nor with an intrusive granodiorite.

In some slates a differentiated layering is developed making an angle with the sedimentary bedding. In Fig. 41 of a thin section of a Cambro-Ordovician feldsparrich slate from the Bonaigua pass near the western end of the Pallaresa anticlinorium, such a case is illustrated. The bedding makes an angle of about 20° with the tectonic layering, which consists of alternating thin layers of 0.1 to 0.3 mm thickness. The micaceous layers have a well-developed oriented fabric; the quartz-plagioclase layers also show a cleavage, but the orientation depends on thin dark-coloured films between the quartz grains. It is probable that the quartzo-feldspathic material is dissolved in certain layers and thus produces the micaceous bands. Examples of differentiated layering occur commonly in slightly higher grade rocks (Fig. 42).

In a thin section of a Devonian slate from Las Bordas in the Valle de Arán (Fig. 43), it can be seen that it concerns in fact a very fine-grained crenulation cleavage. As in these rocks there are no signs of the presence of more than one fold generation, it is quite certain that the oriented fabric is of sedimentary origin, and that a first cleavage developed as a crenulation cleavage.

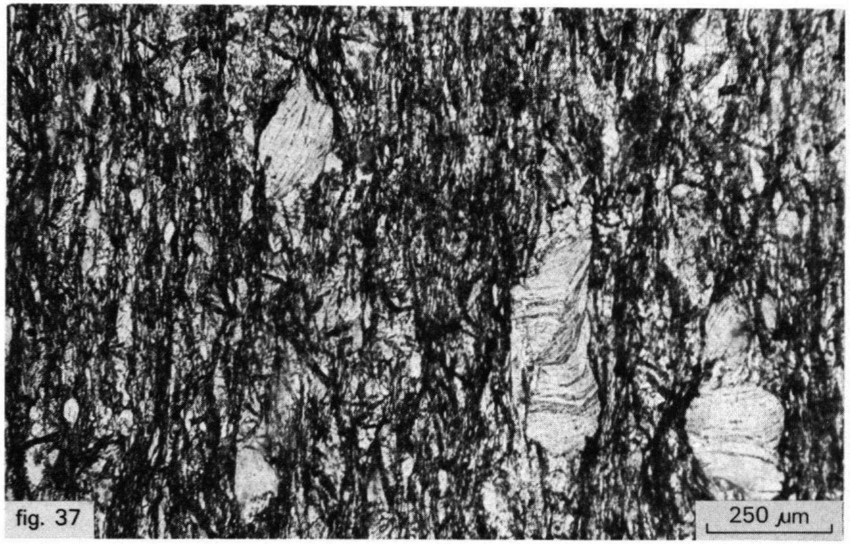
Another section of a Devonian slate is shown in Fig. 44. It concerns a slate from the upper part of the Las Bordas sandstones. The rock is strongly folded and in handspecimen there is a good cleavage. In thin section it appears that there is a large amount of chlorite stacks both in pelitic and quartzitic layers. Again it seems that the large chlorite crystals have developed from sedimentary phyllosilicates in an early phase of cleavage formation. The cleavage is present as a system of anastomosing dark lines around the chlorites. Oriented micas in the cleavage are scarce.

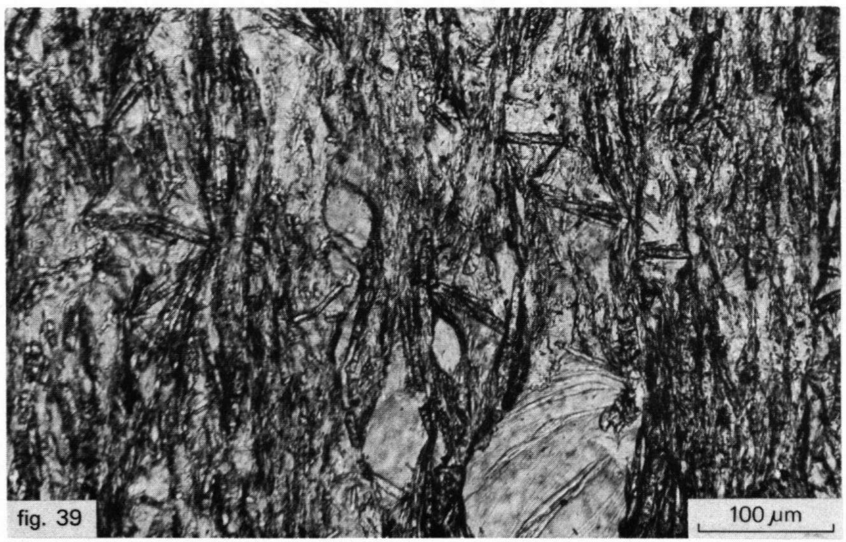
The examples mentioned above are all from the pelitic rocks. Most Cambro-Ordovician rocks are layered quartzphyllites, of which the quartzrich layers show always a poorly developed cleavage. In thin sections they contain equant or weakly flattened quartz grains with unoriented micas in between.

South of the Llavorsi syncline the cleavage, even in the deepest exposed Cambro-Ordovician rocks, is usually badly developed and looks more or less like the thin section just described. There are always many large, deformed chlorite stacks, a crenulated finer grained matrix, and a more or less irregular cleavage, visible as dark lines (Figs. 45 and 46). In the author's opinion the crenulated fabric is a sedimentary bedding, which agrees with the fact that the first set of folds apart from F_0 folds is always the one which has the crenulation cleavage as axial plane. Hartevelt (1970) thought the first fabric to be of tectonic origin and called it S_1, and the crenulation cleavage S_2. It seems quite certain now that Hartevelt's S_2 is our S_1, and belongs to the main phase.

The difference in the development of the S_1 cleavage north and south of the southern limb of the Llavorsi syncline indicates that on the fault bordering this major syncline a large part of the Ordovician sequence is cut out, which then must be due to later, probably Alpine faulting.

Post S_1-cleavages. – Cleavages postdating S_1 are usually developed as crenulation cleavages. One of the first descriptions is from de Sitter (1954) who described an outcrop in the Cardos valley in the Pallaresa antiform. De Sitter did not recognize that this cleavage is folding an existing slaty cleavage, but later work showed that in





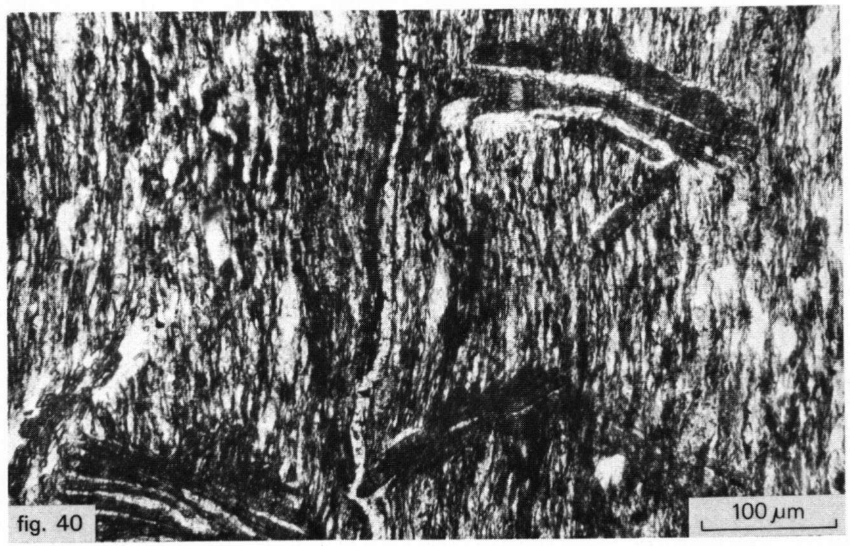


Fig. 37. Chloritoid-bearing Carboniferous slate from Estaro-Llavorsi syncline, sheet 9.

Fig. 38. Carboniferous slate from Llavorsi syncline, sheet 10.

Fig. 39. Same thin section as Fig. 37.

Fig. 40. Chloritoid-bearing Carboniferous slate from Llavorsi syncline, sheet 10.

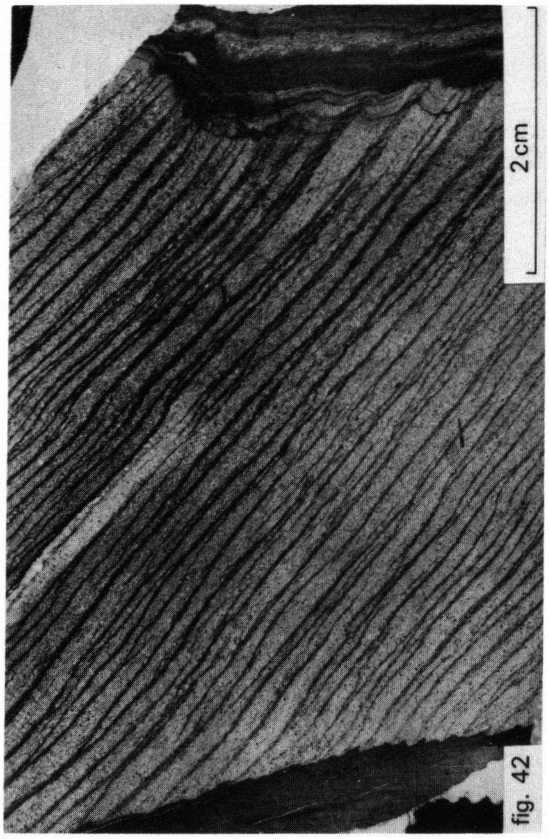


Fig. 42. Cambro-Ordovician schist of Aston massif (sheet 6), showing differentiated layering (oblique) making an angle with the bedding (vertical).

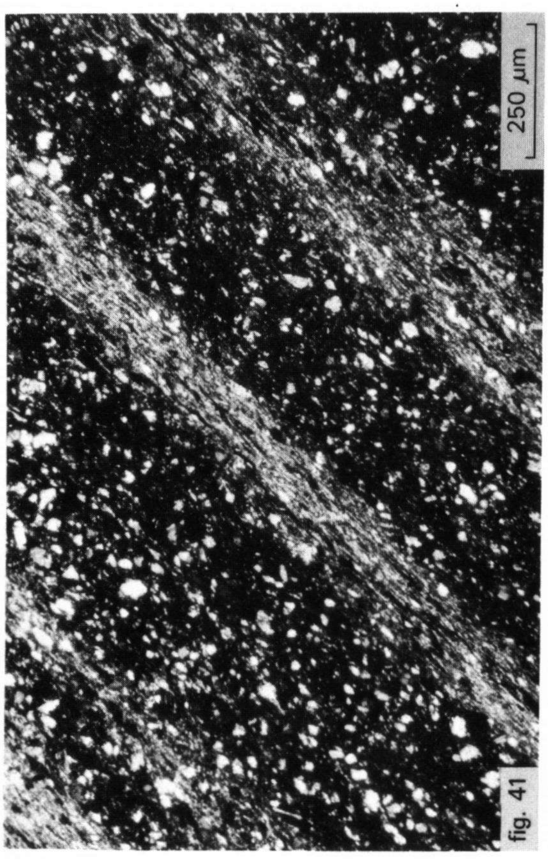


Fig. 41. Cambro-Ordovician slate from Bonaigua pass, sheet 4, showing differentiated layering.

this outcrop very good overprinting relationships exist between two fold generations (Figs. 47, 48, 49). Crenulation cleavages are very common in all rocks in which F_3 or F_4 is present, but they develop only in the well-laminated pelitic parts of the rock. More quartzitic layers are folded by these later fold generations, but do not develop any cleavage (Fig. 28). In general the appearance of crenulation cleavages formed by F_3 and F_4 is remarkably similar, whereever they have been observed.

In Figs. 50 and 51 other examples of crenulation cleavages are shown. They often develop a differentiated layering with alternating sigmoidally folded and straight but compressed S_1 cleavage. In these layers the micas are always strained, indicating that in the low grade slates and phyllites, F_3 and F_4 are postcrystalline.

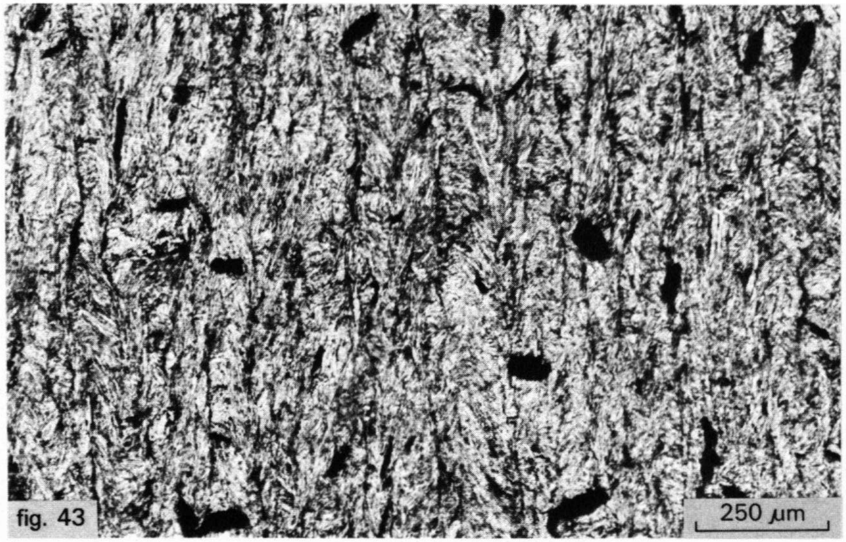
Micaschists

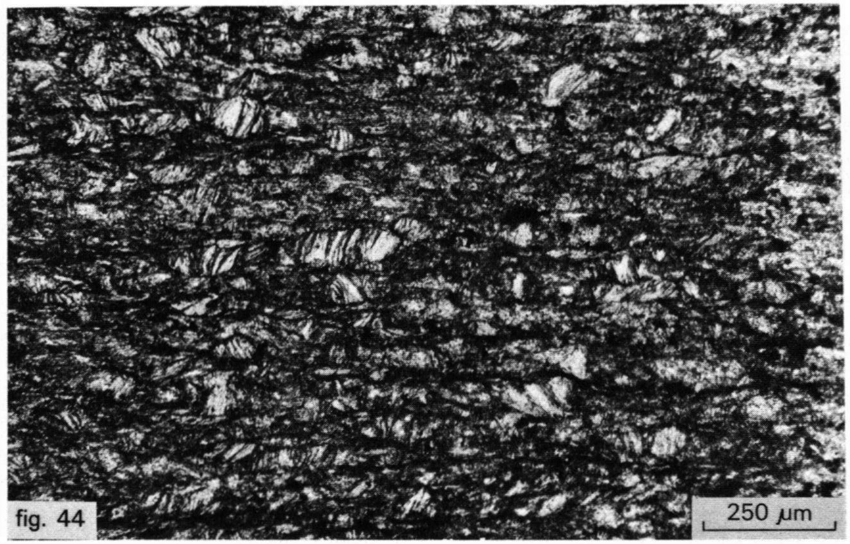
The phyllites described above grade into micaschists in the metamorphic regions like the Aston-Hospitalet massif, the Bosost area in the Garonne dome, and the Lys-Caillaouas massif. In the transition zone where biotite becomes a permanent member of the mineral assemblage, the degree of crystallization of the schists is much higher. The chlorite-muscovite stacks, described in the previous section have disappeared completely, and there is a perfect orientation of muscovite, biotite and flat quartz grains. In the more quartzose parts of the schists a much better orientation of the minerals than in the phyllites is present (Fig. 70). The microstructure in these micaschists is therefore relatively simple, even when they are folded by one of the late folding phases. The micaschists get, however, a much more complicated microstructure when porphyroblasts as biotite, andalusite, staurolite, cordierite, and occasionally chloritoid and garnet occur. These porphyroblasts always overgrow the S_1 schistosity, and usually contain trails of inclusions (S_i) inherited from this schistosity.

They have been extensively treated by the present author (Zwart, 1962, 1963) and for this reason the principles of the usage of these relations will not be discussed here, but only some results from various regions in the map area. As the growth of the porphyroblasts indicates a higher degree of metamorphism, in all massifs this degree is clearly increasing after F_1.

In all three metamorphic areas, the Aston-Hospitalet massif (Zwart, 1965; Lapré, 1965; Verspyck, 1965), the Bosost area (Zwart, 1962), and the Lys-Caillaouas massif, the same feature has been observed, although there exist differences between these areas.

Aston-Hospitalet massif. – In the Aston-Hospitalet massif the main foliation is interpreted as S_1, although F_2 folds may have a new axial plane schistosity. In a number of cases porphyroblasts of biotite, andalusite, staurolite and cordierite overgrow the undeformed S_1, as shown by planar S_i trails in the crystals, which consequently postdate F_1. As most F_2 folds occur outside the area with porphyroblasts, relationships between these two phenomena could not be established. In the Hospitalet massif the internal fabric of biotite, andalusite,





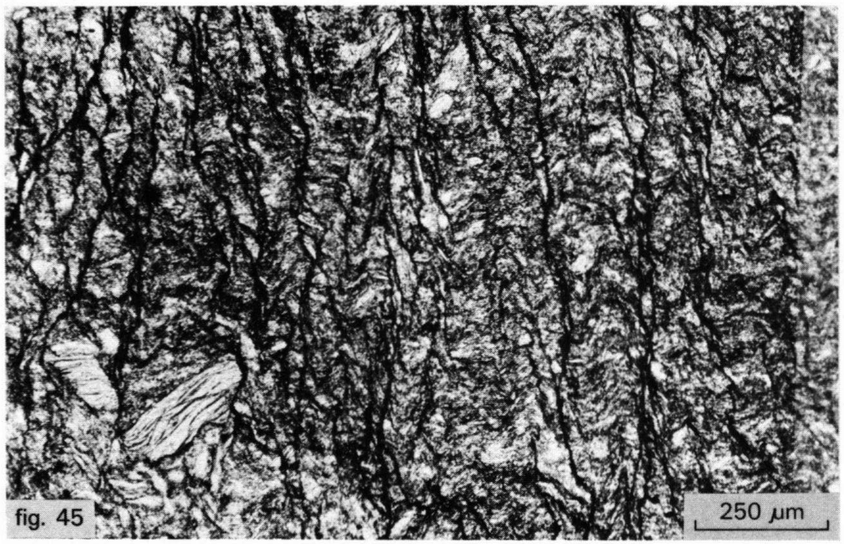
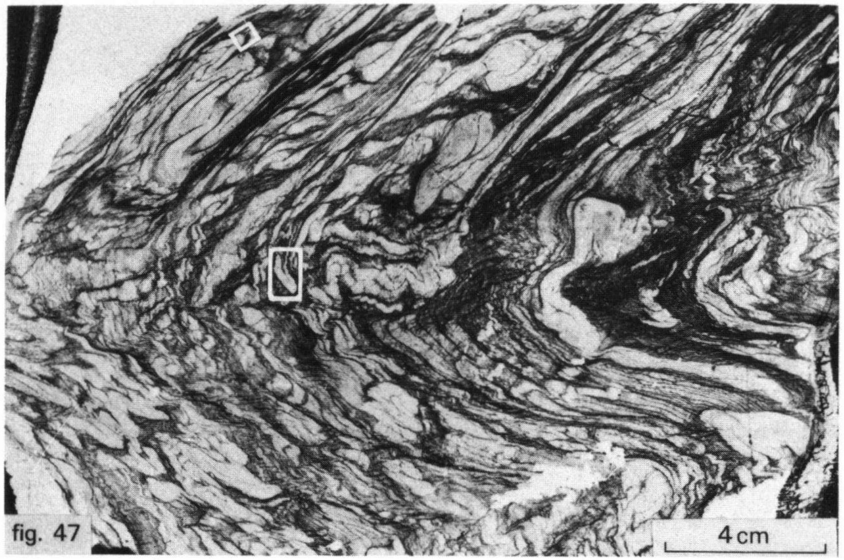


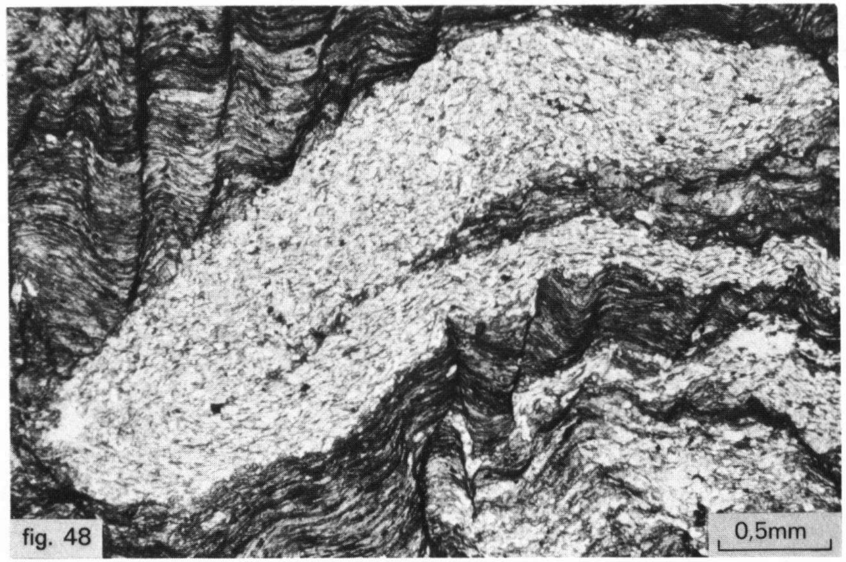
Fig. 43. Devonian slate near Las Bordas, sheet 4, with S_1 crenulation cleavage.

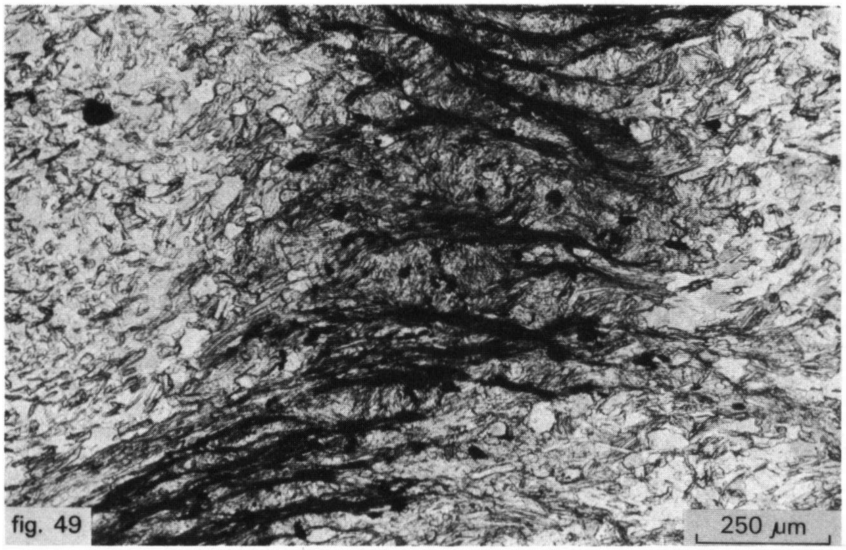
Fig. 44. Upper Devonian slate near Viella, Valle de Arán, sheet 4.

Fig. 45. Cambro-Ordovician slate near Puigcerda (east of sheet 10).

Fig. 46. Cambro-Ordovician slate near St. Julia, Andorra, sheet 10.







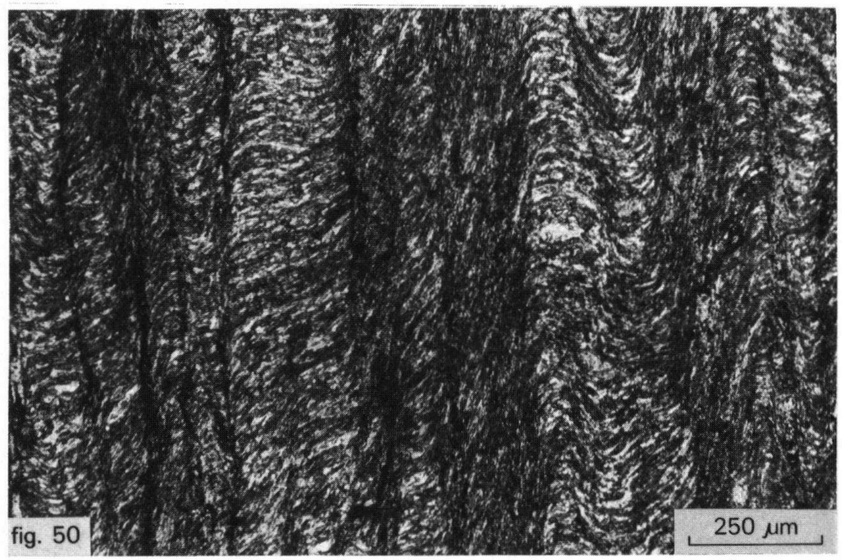


Fig. 47. Cambro-Ordovician slate, Cardos valley, sheet 5, showing F_4 fold with crenulation cleavage, overprinting F_1 folds.

Fig. 49. Detail of Fig. 47 (white rectangle); with F_1 hinge.

Fig. 48. Detail of Fig. 47 (white rectangle); F_1 hinge overprinted by S_4 crenulation cleavage.

Fig. 50. Cambro-Ordovician slate near Esterri de Aneu, sheet 5, with F_4 microfolds and differentiated crenulation cleavage.

staurolite and cordierite is usually planar, but there is a strong influence of the F_3 (NW–SE) folds, which fold the S_1 schistosity, often producing a crenulation cleavage. Most staurolites withstand this folding and are undeformed, whereas S_3 curves around the crystals. Andalusite behaves much more ductile and is folded together with the matrix. The extinction of the crystals rotates to the same degree as the fold, so that originally the inclusion pattern was planar (Fig. 52). Staurolite crystals may exhibit postcrystalline rotation due to F_3 movements. Also biotite is usually deformed, often with kinkbands. Small biotites with planar S_i have an orientation in S_3 caused by mechanical rotation into these planes (Fig. 53). These relations are well developed in Cambro-Ordovician schists of the El Serrat valley in Andorra where S_1 dips to the south. In the central part of the Hospitalet massif like east of the Incles valley, where S_1 has a subhorizontal position, the schists have an E–W lineation, due to cleavage–bedding intersections and fabric habit of quartz and mica crystals. Large, elongate porphyroblasts of staurolite, andalusite and cordierite have a NW–SE orientation parallel to open crenulations of S_1. Thin sections parallel to the S_1 foliation show the L_1 lineation with E–W direction. The internal lineation in the porphyroblasts makes an angle with this direction, but is continuous with the external lineation. From these relationships it can be deduced that these crystals were rotated with a rotation axis approximately perpendicular to the foliation. The long axes of the crystals were thus rotated in the NW–SE direction of the F_3 folds where they attained an equilibrium position (Zwart, 1963, and Fig. 54). In some cases the internal lineation has an S-shape, indicating contemporaneity of growth and rotation. In this case porphyroblastesis continued and took place during F_3. The same conclusion can be drawn from some andalusite crystals with helicitic folds of F_3 in the central part of the Hospitalet massif.

In the Silurian schists in the Valira del Norte valley in Andorra near Llorts, large chiastolite crystals occur in black schists with F_3 crenulations. Again the porphyroblasts overgrow S_1, but are earlier than F_3 (Fig. 55).

In conclusion it can be stated that in the western part of the Hospitalet massif porphyroblastesis took place between F_1 and F_3. In the more central part this process continued into the F_3 folding phase.

In the Aston massif the relations are rather similar. Here also porphyroblastesis starts after S_1, and the influence of F_3 is clearly visible. In Fig. 56 a biotite porphyroblast is shown with a planar S_i. The S_1 matrix is strongly folded by F_3 and produces a crenulation cleavage curving around the biotite crystal. In Fig. 57 the biotite contains, besides a planar S_1, many kinkbands due to F_3 folding. The biotite of Fig. 57 has a folded internal fabric, but the external fabric is more strongly folded and shows an F_3 crenulation cleavage.

Figs. 58 and 59 are from a thin section with staurolite and cordierite porphyroblasts. Both minerals contain

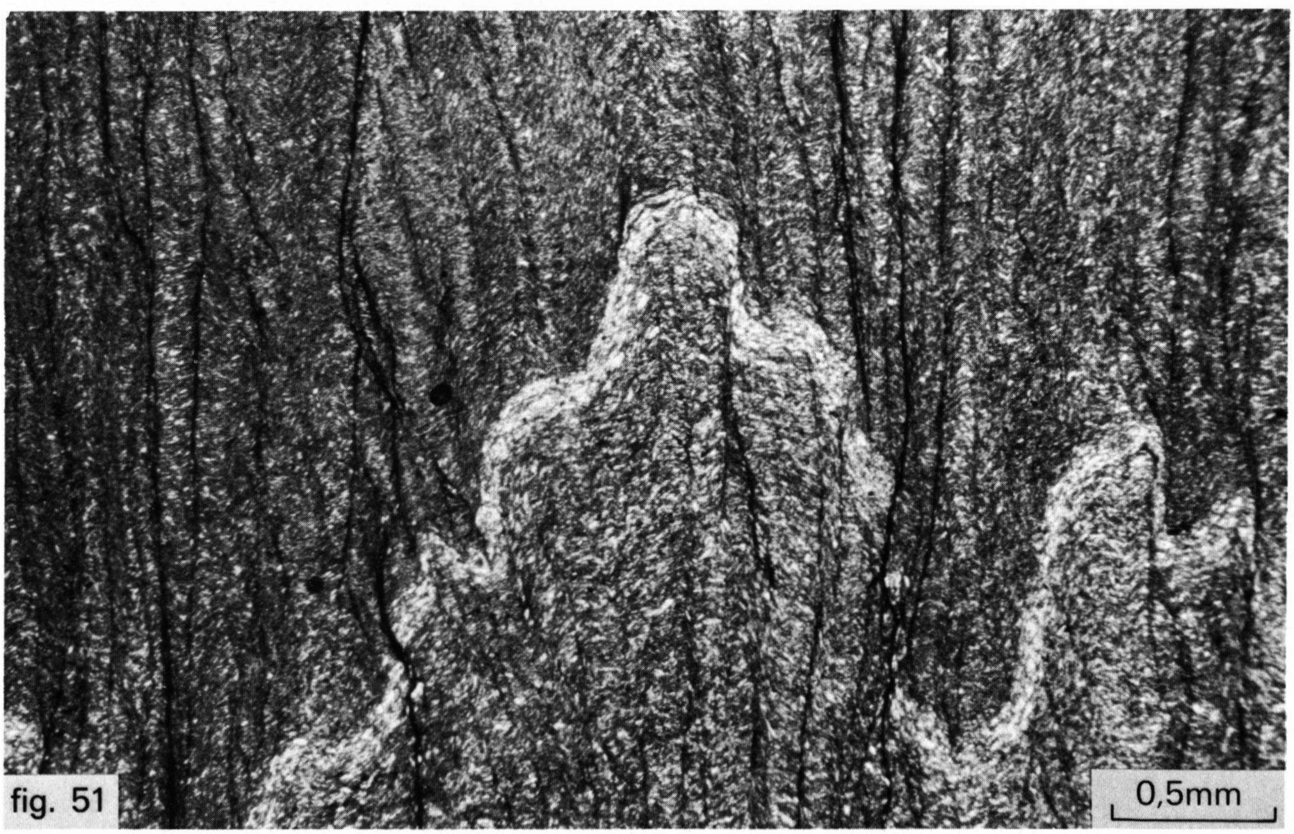
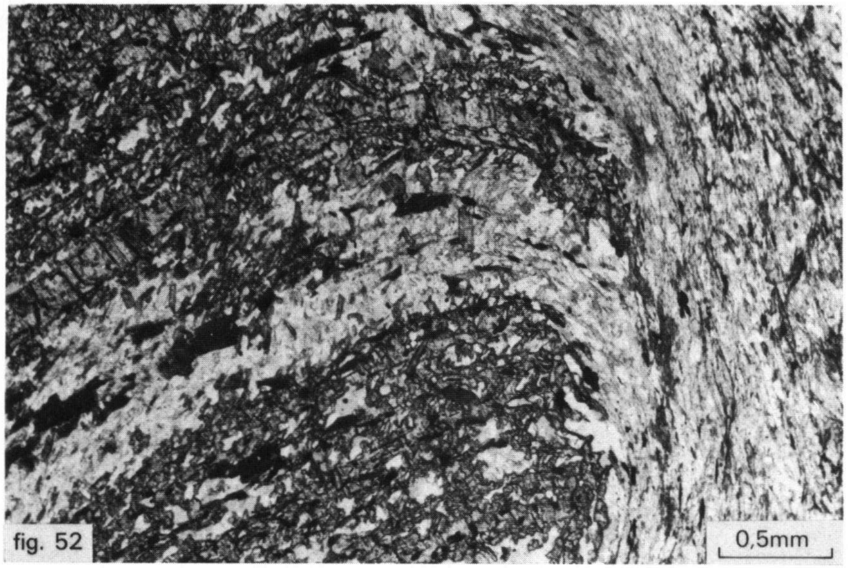
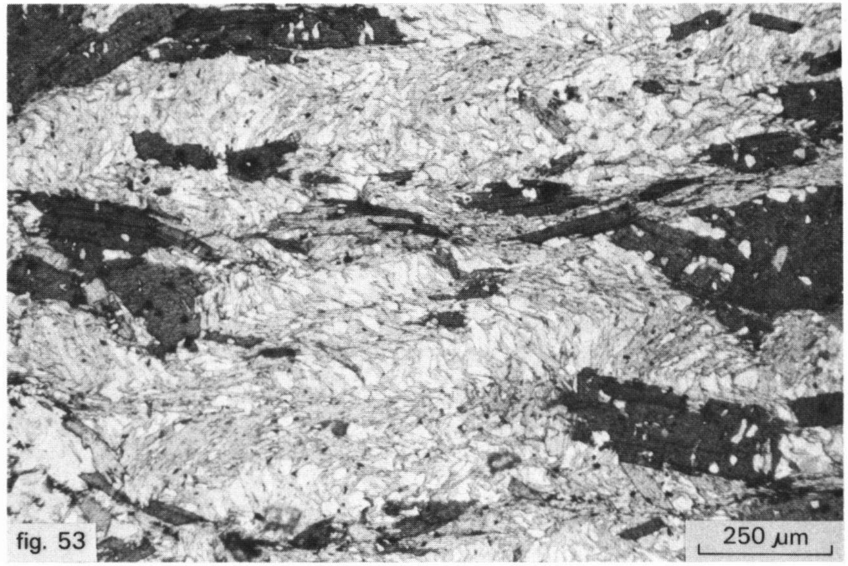
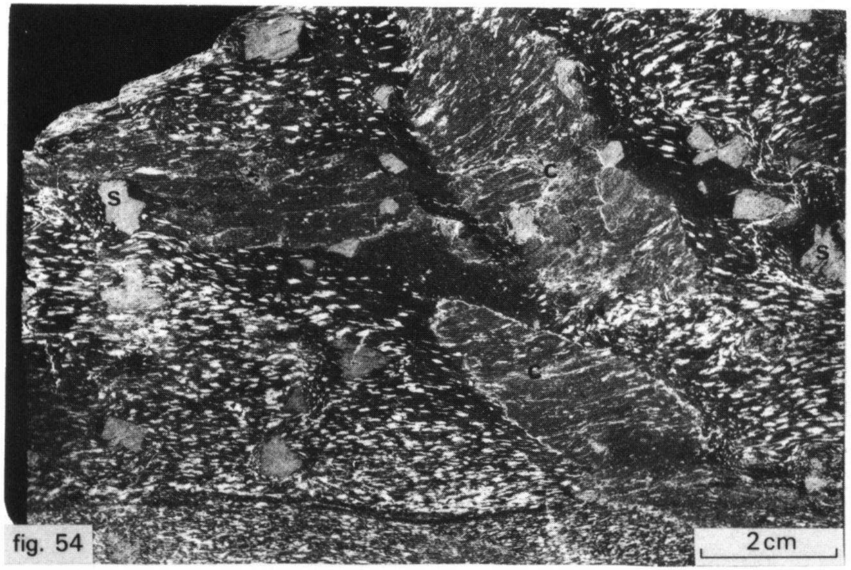


Fig. 51. Cambro-Ordovician slate near Esterri de Aneu, sheet 5, with F_4 folds and crenulation cleavage.







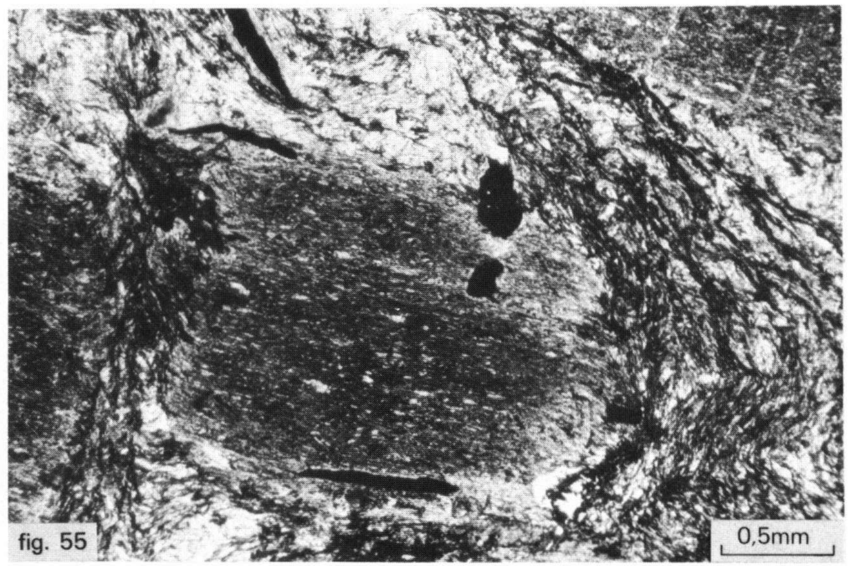


Fig. 52. Andalusite-schist with F_3 folds, folding S_1 schistosity and andalusite, El Serrat, Andorra, sheet 6.

Fig. 53. Phyllite with small biotite porphyroblasts; S_1 is crenulated by F_3; biotite overgrows S_1 and is oriented in S_3; El Serrat, Andorra, sheet 6.

Fig. 54. Staurolite-cordierite-schist, cut parallel to S_1; cordierite (C) rotated into NW–SE position by F_3; on the crest of the Hospitalet massif, W of Port de Fontargent, sheet 6 (negative print).

Fig. 55. Silurian graphite schist; andalusite with planar S_i in matrix folded by F_3; Llorts, Andorra, sheet 6.

Fig. 57. Kinked biotite porphyroblast; kinking due to F_3, same locality as Fig. 56.

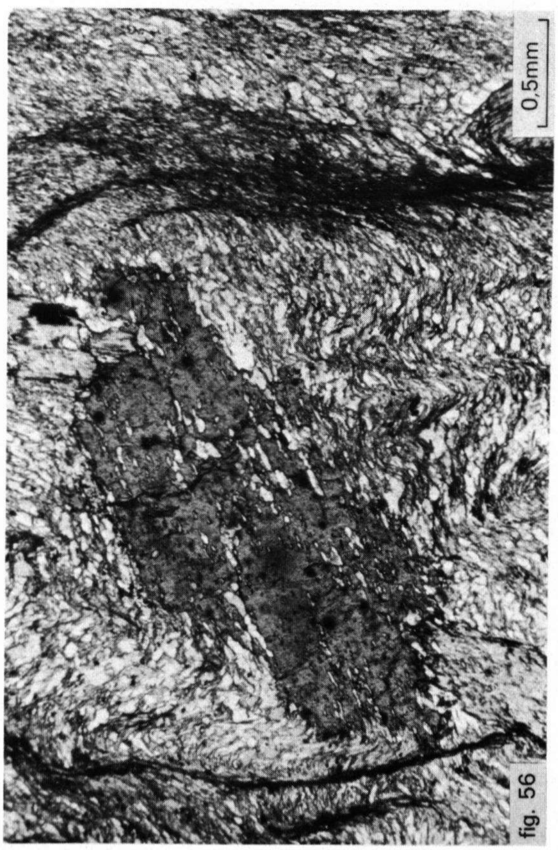


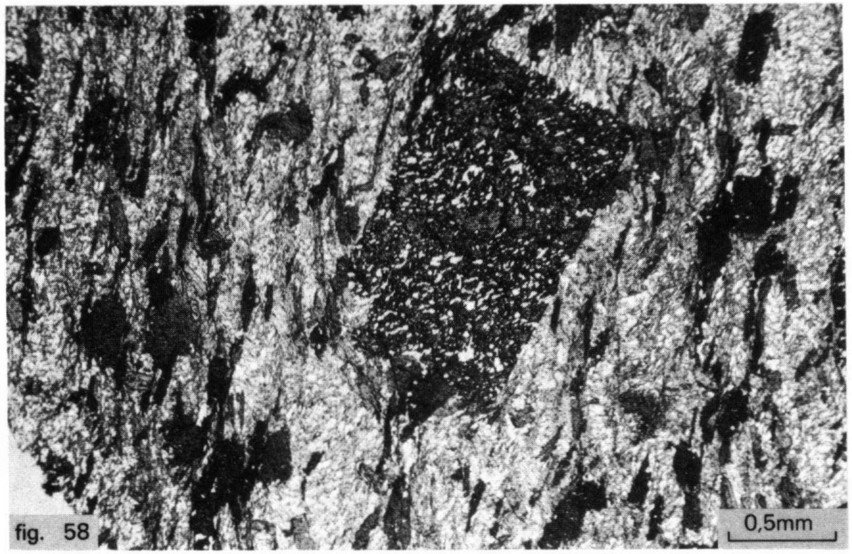
Fig. 56. Phyllite with biotite porphyroblast with planar S_i and F_3 crenulations; Mounicou valley, W part Aston massif, sheet 6.

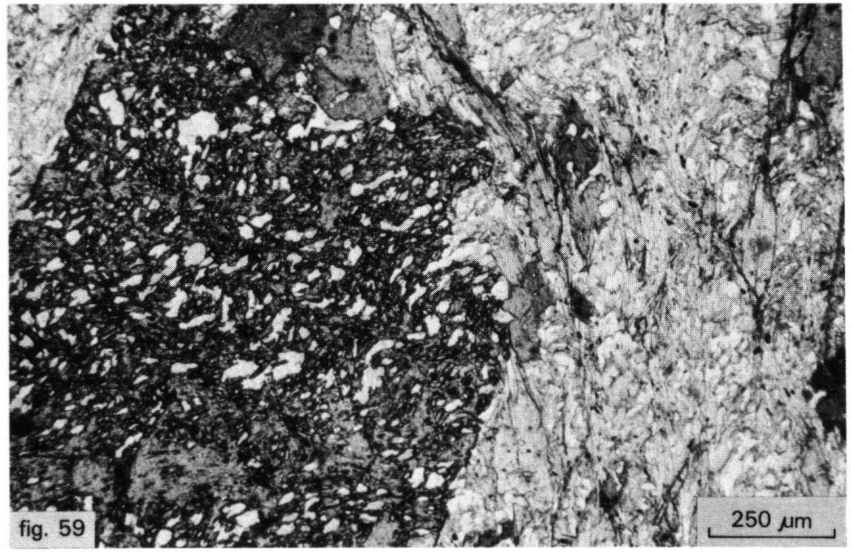
helicitic folds, but the matrix is more strongly folded and has an S_3 crenulation cleavage, in which small biotite porphyroblasts are oriented, probably by mechanical orientation. The following succession of events is indicated: F_1 folding under greenschist facies conditions and development of slaty cleavage, formation of small biotites inheriting planar S_1, folding of S_1, in a certain stage of development of these folds relatively rapid growth of staurolite and cordierite as shown by inclusion pattern, continuation of folding and development of S_3 crenulation cleavages, and orientation of biotite in these planes. In other cases porphyroblasts are later than F_3 folding, as they contain helicitic folds which have the same shape as matrix folds (Fig. 60). In some cases the F_3 differentiated layering is clearly reflected in the crystals like in Fig. 61, where the sigmoidally folded bands are rich in quartz inclusions, and the adjoining bands with planar fabric have almost no inclusions.

In the Aston massif F_4 folds postdate the porphyroblastesis. However, mica's in F_4 fold hinges are recrystallized and contain polygonal arcs. From these data it appears that in the Aston massif, metamorphism lasted from F_1, until beyond F_4, but the peak of metamorphism, as shown by the porphyroblasts, occurred between F_1 and F_3, and during F_3.

Bosost area. – In the Bosost area the succession of structural and metamorphic events is somewhat similar to that in the Aston-Hospitalet massif, but it is more complex and lasting longer. The S_1 schistosity is again formed under low grade conditions and consists mainly of quartz and white mica, possibly with some biotite. The presence of the latter mineral in S_1 is not certain, as the original fabric has changed considerably due to later, higher grade metamorphism, which eventually obliterates all signs of this schistosity. Porphyroblastesis of biotite, andalusite, staurolite, cordierite and to a lesser extent garnet took place after F_1, as they overgrow S_1 and inherit a planar inclusion pattern. These porphyroblasts, especially biotite, grow more or less at random and seem to lack a preferred orientation. In addition there is evidence that the biotites were rather stubby crystals. This can be seen from biotite crystals included in andalusite or cordierite porphyroblasts where they have been shielded for later deformation (Fig. 63). The next event is a deformation phase with a simple shear character which causes all these porphyroblasts to rotate, with a N–S directed rotation axis, and a consequent movement from west to east. That means, looking north all crystals have a clockwise rotation (Fig. 62).

Besides staurolite with planar S_i, other, less abundant crystals of this mineral contain an s-shaped S_i and obviously formed during the second phase of deformation. These crystals still show a certain amount of postcrystalline rotation, indicating that the formation of staurolite has ceased before the end of that phase. From the abundance of crystals with planar S_i it can be concluded that the culminating point of its crystallization falls before the second phase. It is not possible to tell exactly when staurolite started to form as during the interkine-





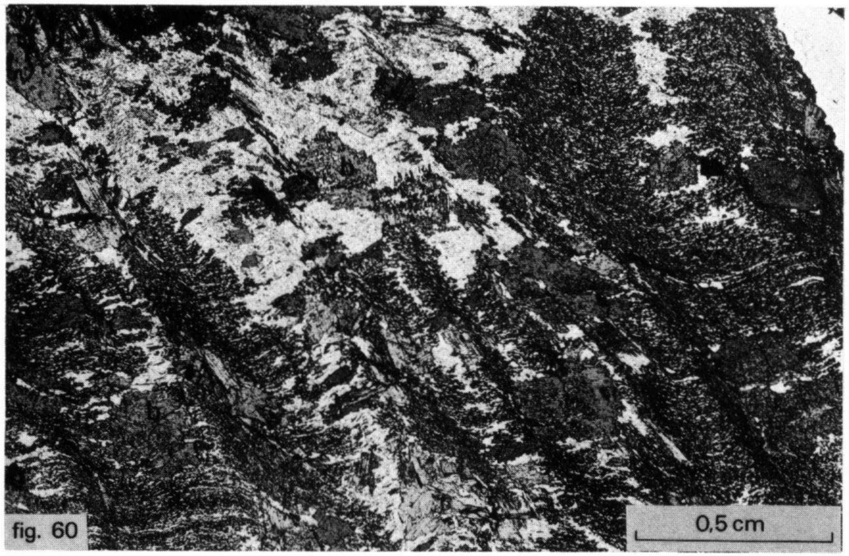


Fig. 58. Staurolite with helicitic folds in phyllite with S_3 crenulation cleavage and oriented biotite; Mounicou valley, Aston massif, sheet 6.

Fig. 60. Biotite and andalusite with helicitic F_3 folds, location as Fig. 58.

Fig. 59. Cordierite with helicitic folds in phyllite with S_3 crenulation cleavage; Mounicou valley, Aston massif, sheet 6.

Fig. 61. Detail of Fig. 60, location as Fig. 58.

matic period no further determination of time is possible.

A similar reasoning can be made for andalusite. Most crystals of this mineral in this zone have been rotated about N–S axes, and crystals with planar and with s-shaped S_i have been found in large quantities. Moreover crystals with planar S_i in the core, and s-shaped S_i in the rim are common in Devonian schists. So in this case also, andalusite crystallization occurred around the beginning of the second phase. The end of andalusite formation falls later than that of staurolite, since post-crystalline rotation of the first mineral has often not occurred, and S_i is continuous without break into S_e. This shows that andalusite crystallization went on until the end of the second phase. A few crystals are even indicative of an age later than this phase. These crystals are usually less well bounded; they are not rotated and invade the matrix without mechanically disturbing it. Such crystals are not very common, but they demonstrate that andalusite crystallization ended shortly after the second phase (Table 2).

As far as the beginning of andalusite formation is concerned, it can be deduced that it falls later than that of staurolite. In several thin sections it has been observed that staurolite is included in andalusite, whereas the reverse relation does not exist. It should be added that it does not concern a replacement process since the included staurolite still exhibits its idioblastic shape. Included staurolite and host mineral are often rotated together. For this reason both minerals are certainly stable together, crystallized simultaneously for a certain time, but staurolite began and ceased crystallization before andalusite.

The same reasoning can be applied to unravel the history of cordierite crystallization. Numerous crystals with planar and s-shaped S_i have been observed, again with N–S directed rotation axis. Consequently staurolite, andalusite and cordierite form a contemporaneous and stable assemblage. It should be added that samples containing all three minerals are by no means exceptional; especially in the Devonian they are of frequent occurrence.

More details about these three minerals can be deduced from their mutual relationships. Not uncommonly staurolite or andalusite are enclosed in cordierite, but again the reverse relation is absent. This signifies that the cordierite formation started after that of the other two minerals, although still before the start of the second phase. Besides rotated crystals one finds commonly other cordierites usually lying with their long dimensions in the schistosity but showing no rotation; they replace the groundmass without deforming it. Like some of the andalusites these cordierite crystals must be considered as having grown postkinematically with regard to the second phase. They may be deformed by the third or fourth phases, so that these phases are post-crystalline. Since such cordierites are definitely more abundant than comparable andalusites, it seems safe to conclude that the crystallization of cordierite outlasted that of andalusite. Therefore the whole period of formation of cordierite, although overlapping that of staurolite and andalusite, falls a little later than the formation of these two minerals (Table 2).

As far as the staurolite included in cordierite is concerned, it is to be noted that in many cases the staurolite is idioblastic especially when included in rotated cordierite. When, however, staurolite is enclosed in post F_2 cordierite, unstable relationships often are involved, and staurolite is actually partially or even wholly replaced by cordierite (Fig. 87). From this evidence it may be concluded that at first staurolite, andalusite and cordierite form a stable paragenesis, but that after staurolite had ceased crystallization, it became unstable, resulting in its replacement by cordierite.

Although garnet is a rare component of the Bosost schists, its relations with these rocks are strongly similar to those of staurolite, and its duration of crystallization is about the same.

Rather equant crystals like biotite and staurolite all show rotation; elongate crystals like andalusite and cordierite lying with their long axis parallel to the rotation axes also rotate (Fig. 63), but crystals with an orientation perpendicular to this axis have an unfavourable position and do not show this feature (see fig. 17, Zwart, 1963). The biotite crystals show some special features. As can be seen from their internal fabric they are rotated, often over an angle of about 60°. When rotated back to their original position, they have a preferred shape orientation making an angle of 60° with the foliation. An

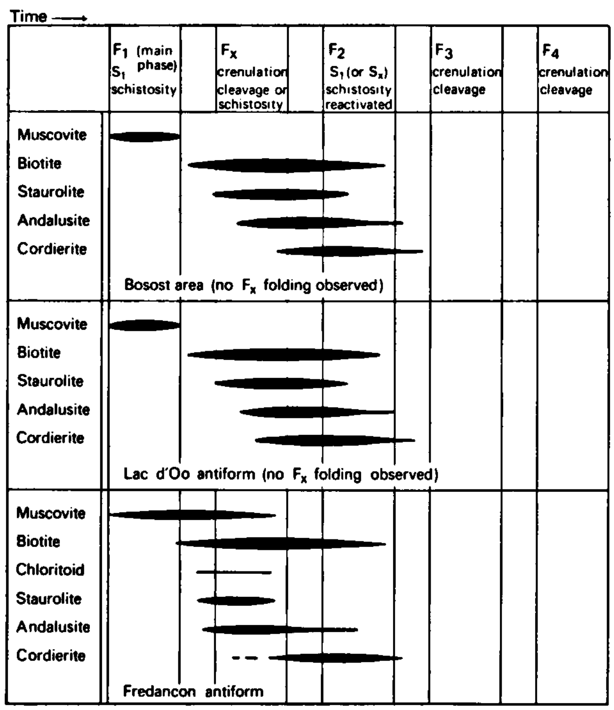


Table 2. Relations between metamorphism in muscovite- and staurolite-andalusite-cordierite-zone and folding phases in Bosost area, Lac d'Oo and Fredancon antiforms.

explanation for such preferred orientation is difficult to suggest. As a result of the rotation the biotites now have a shape orientation in the schistosity but no lattice orientation (Fig. 64). In addition they acquire an elongate habit, the long axis lying in E–W direction, and the short axis parallel to the rotation axis in N–S direction (Fig. 65). For this reason the E–W biotite lineation in the micaschists is not due to F_1 folding, but to the later rotational movements, referred to as F_2. With increasing amounts of strain, the biotites become flatter and more elongate (Fig. 66). They also tend then to get a lattice orientation in or close to the S_1 foliation.

The amount of rotation of any mineral rarely exceeds 90°, but the cumulative effects in a schist unit of at least a few km thickness must be considerable. It has been estimated by the author as a shear of 4–5 (Zwart, 1962).

Metamorphism did not cease after F_2, but increasing temperature first results in the instability of staurolite, and then of andalusite in favour of cordierite, sillimanite and muscovite, which occur in the deepest exposed part of the Bosost area. Staurolite alters to an unoriented aggregate of muscovite, sometimes with some biotite (Figs. 88, 89), to andalusite or cordierite. Andalusite may transform into cordierite, fibrolitic sillimanite or occasionally to muscovite. All these alterations postdate F_2, as can be deduced from the microstructure (Table 3). Cordierite overgrows, for example the bowed out schist around rotated staurolites, without showing undulatory extinction.

In the highest grade zone therefore the mineral assemblage consists of quartz, biotite, sillimanite, cordierite, muscovite and some plagioclase. This zone is also structurally the deepest one (Zwart, 1962, and chapter on metamorphism).

Although most rocks in the Bosost area have a similar metamorphic history within one metamorphic zone, there are some rocks which are more complex. In an outcrop near the village of Lés the schists have a foliation consisting of alternating biotite-rich and poor bands. This foliation is folded in minor, tight, chevron folds with a NNE directed axis and an axial plane schistosity making an angle of 20–40° with the foliation (Figs. 67, 68, 69). In this rock cordierite and andalusite crystals occur with their long axis in the plane of schistosity.

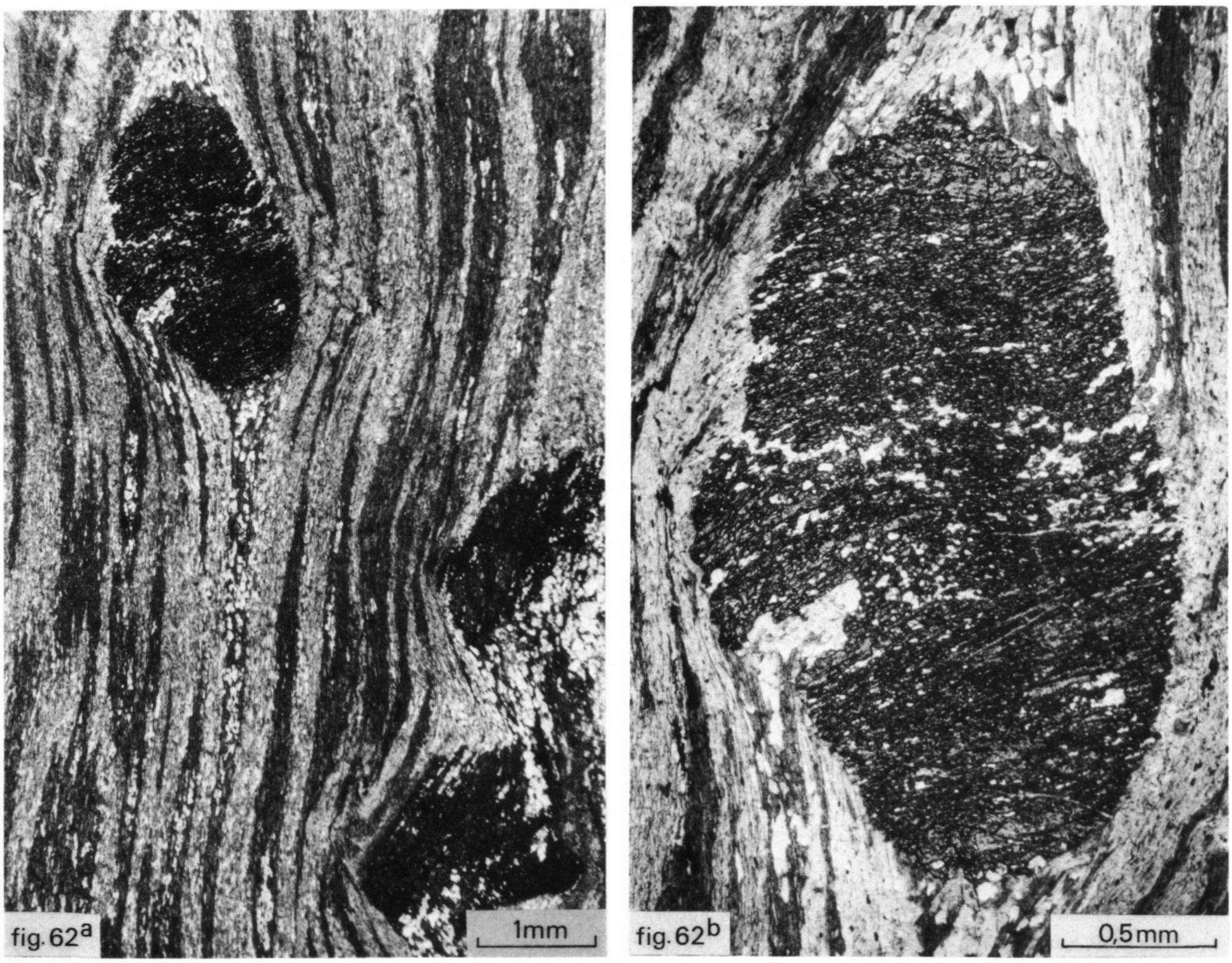
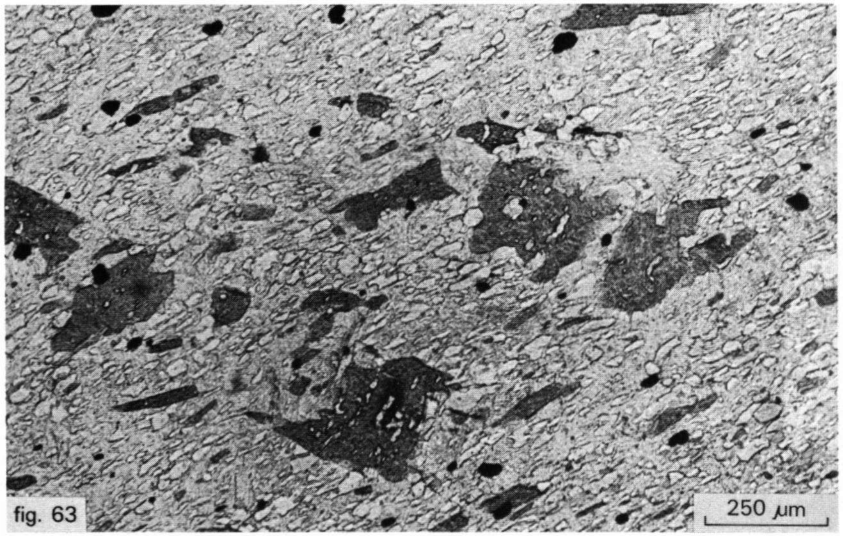
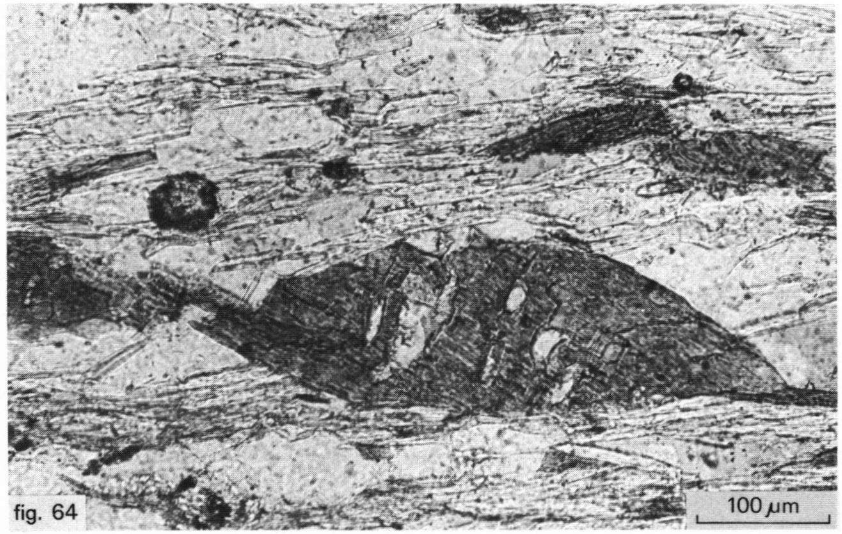
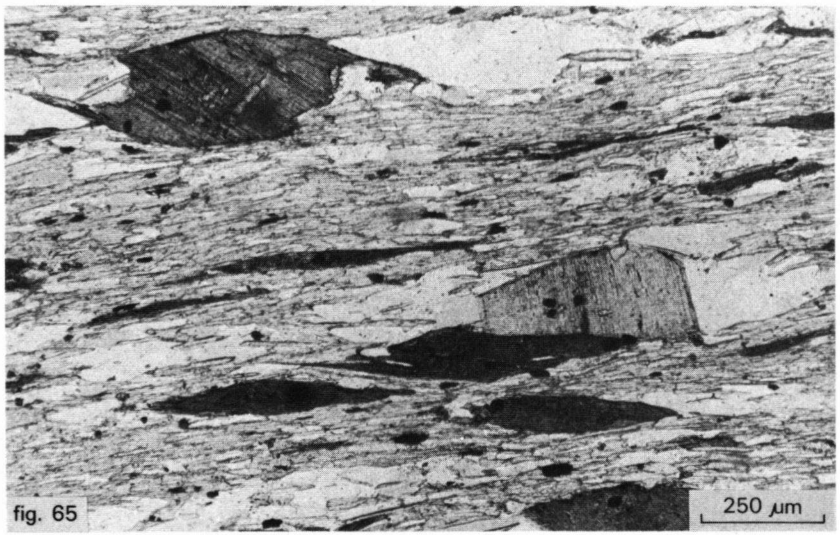


Fig. 62. Rotated staurolites in biotite-schist with differentiated layering; section parallel to E–W lineation, perpendicular to N–S rotation axis; east is to the bottom of photograph; Bosost area, sheet 4.







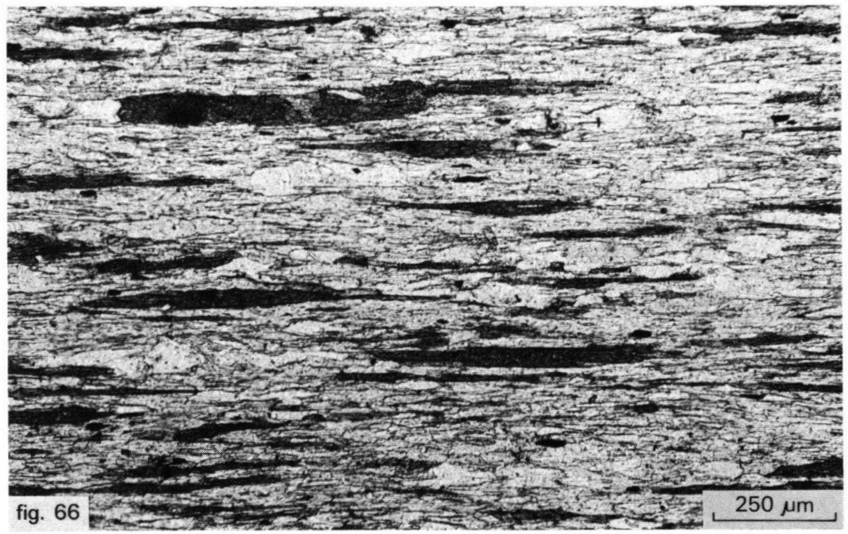
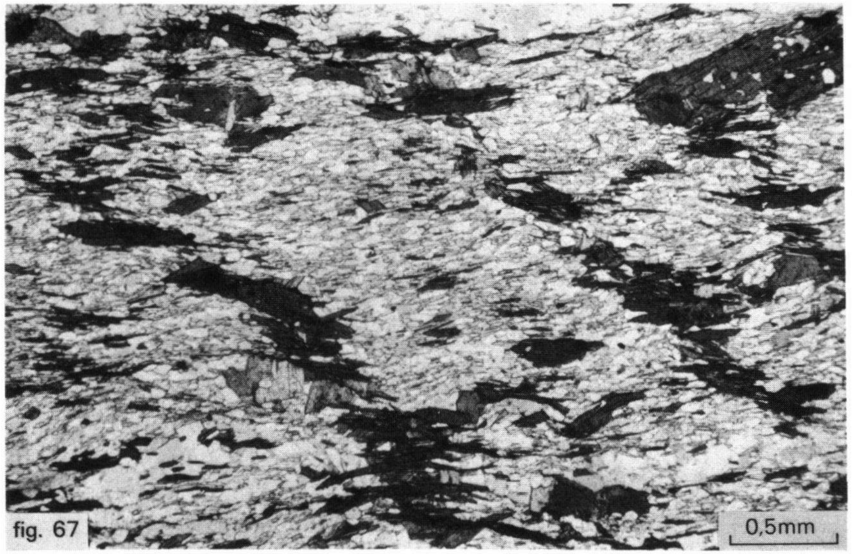


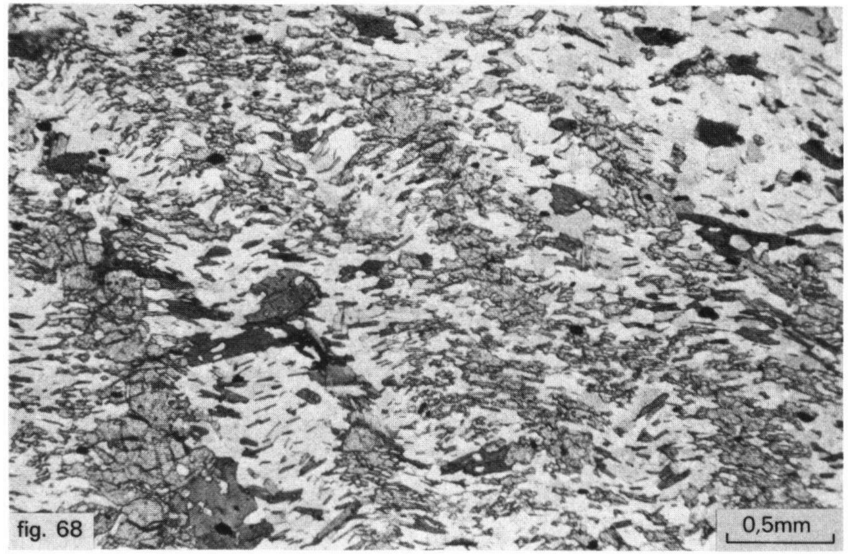
Fig. 63. Unoriented, equant biotite crystals in large cordierite porphyroblast.

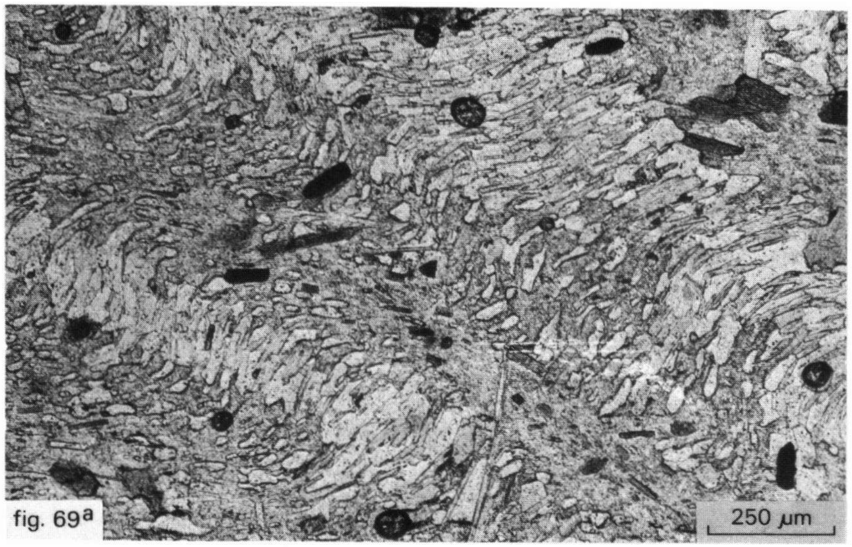
Fig. 64. Biotite porphyroblast rotated into S_1; Bosost area, sheet 4.

Fig. 65. Biotite crystal with shape orientation in S_1 due to rotation by F_2; Bosost area, sheet 4.

Fig. 66. Thin, elongate biotite crystals in S_1, Bosost area, sheet 4.







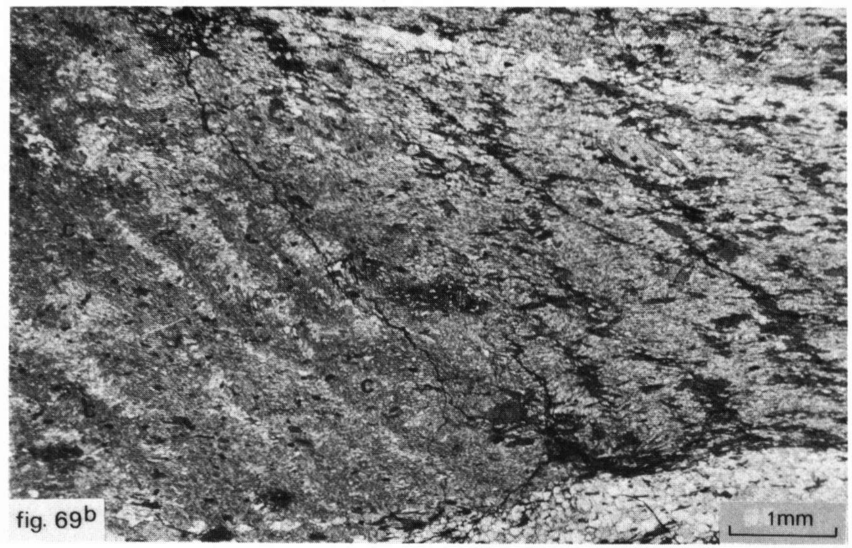


Fig. 67. Biotite layering making an angle with the schistosity; near Lés, Bosost area, sheet 4.

Fig. 68. Helicitic folds in andalusite; axial plane of included crenulations parallel to biotite layering of Fig. 67; location as Fig. 67.

Fig. 69a. Helicitic folds in cordierite; detail of Fig. 69b.

Fig. 69b. Cordierite porphyroblast (on left side of photograph) with helicitic folds, and matrix with biotite layering; same relationships and location as Fig. 67.

These crystals contain helicitic crenulations with an axial plane parallel to the biotite foliation. The succession of events for these particular rocks is probably as follows. The crenulations in the porphyroblasts are at least an F_2 feature and the crenulated fabric is S_1. The biotite foliation, parallel to F_2 axial planes, should also be of F_2 age. The schistosity, determined by a parallel micafabric postdates the foliation and is of a younger age, and can be called S_3. Correlation with surrounding micaschists is not entirely clear, but I suggest that S_1 in these schists and those of Lés is the same. S_2 in the Lés rocks is absent elsewhere, and S_3 in these rocks can be correlated with F_2 or S_2 (N–S folds, rotation with N–S axes) in the other schists of the Bosost area.

In the andalusite-cordierite and especially in the cordierite-sillimanite zone the schistose fabric is undergoing drastic changes. The shape and orientation of biotite crystals in the staurolite-andalusite-cordierite zone has been discussed. In the higher grade zones the biotite porphyroblasts can no longer be recognized, and instead there is now a microstructure of small muscovite and biotite crystals with a shape and lattice orientation. The size of both micas is more or less the same, and the fine-grained muscovite of the lower grade zones has now disappeared. Between the micas occur quartz crystals which are less flat than in lower grade rocks. In Fig. 70 such a micaschist is shown. This rock still has a lineation due to fabric habit of the minerals. Some fibrolite is present in this thin section.

In the cordierite-sillimanite zone the schistosity tends to disappear completely. In Fig. 71 such a rock is shown. Biotite and muscovite have a random orientation, and the quartz grains form a kind of foam texture with triple junctions, where grain boundaries meet at angles close to 120°.

Another feature in these zones is the development of polygonal arcs of micas in F_3 and F_4 folds. Figs. 72 and 73 are from a thin section of micaschist layers in a folded pegmatite sill in the Barrados valley (sheet 4). This pegmatite was emplaced in the schists after F_2, but is folded by F_3 and F_4 (Zwart, 1963, p. 364 and fig. 10). In the micaceous layers a crenulation cleavage is formed, but in Fig. 73 it is clearly visible that the micas are undeformed, and recrystallized after F_4 folding. This indicates that metamorphism still continued during and after this folding phase.

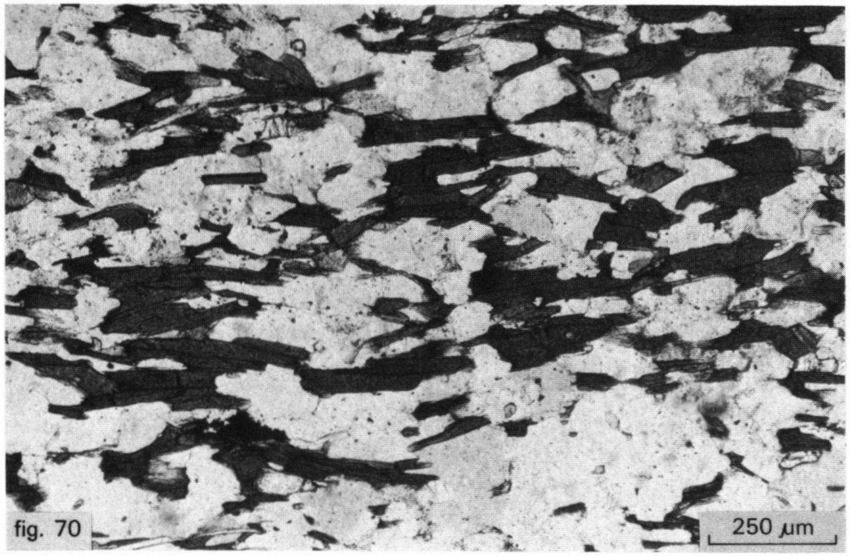
Lys-Caillaouas massif. – In the Lac d'Oo antiform two fold generations can be distinguished, of which the first one with tight to isoclinal folds is accompanied by a schistosity defined by a parallel arrangement of muscovite and quartz crystals. The antiform itself is due to a second fold generation (but F_4 in our scheme) and may be accompanied by a crenulation of the schistosity. The antiformal hinge contains aluminium-silicate-bearing biotite-schists, but to the north the grade drops to biotite-schists and then to phyllites without biotite.

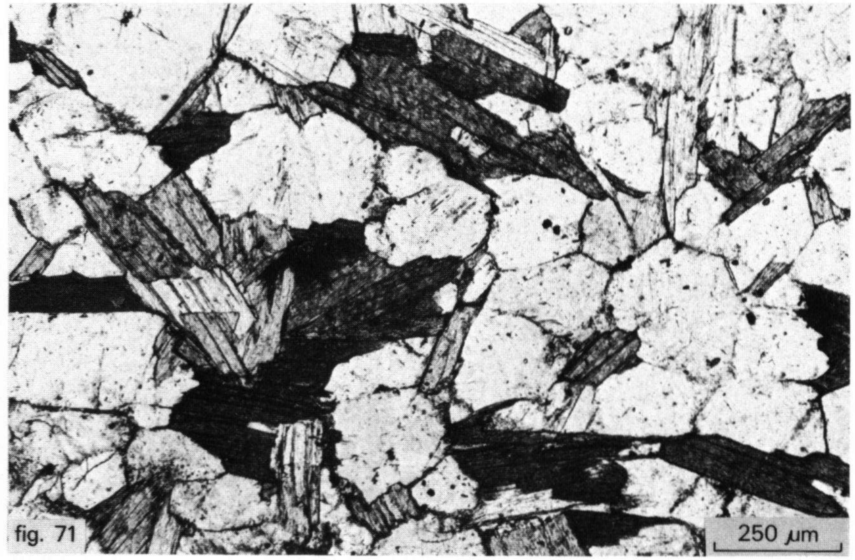
Micaschists not too close to the granite contact contain porphyroblasts of staurolite, andalusite, cordierite, biotite and occasionally garnet and rarely chloritoid. There is an E–W lineation determined by fabric habit of biotite crystals, which have a preferred shape orientation, but no lattice orientation. From included trails of inclusions it is evident that the biotites have rotated into the schistosity. This produces a microstructure similar to the Bosost micaschists (Fig. 74). In addition many crystals of staurolite, andalusite, cordierite and garnet also show rotation with a N–S directed rotation axis, and a clockwise rotation looking north (Fig. 75). This is also similar to the Bosost area. Some andalusite and cordierite crystals have weakly s-shaped inclusion trails, and have grown during these simple shear movements which can be correlated to F_2 in the Bosost area. From included biotites in andalusite crystals, it follows that biotite grew at random over a schistosity, with a rather stubby shape. Due to the rotational movements they got oriented in the schistosity and are stretched resulting in flat elongate crystals.

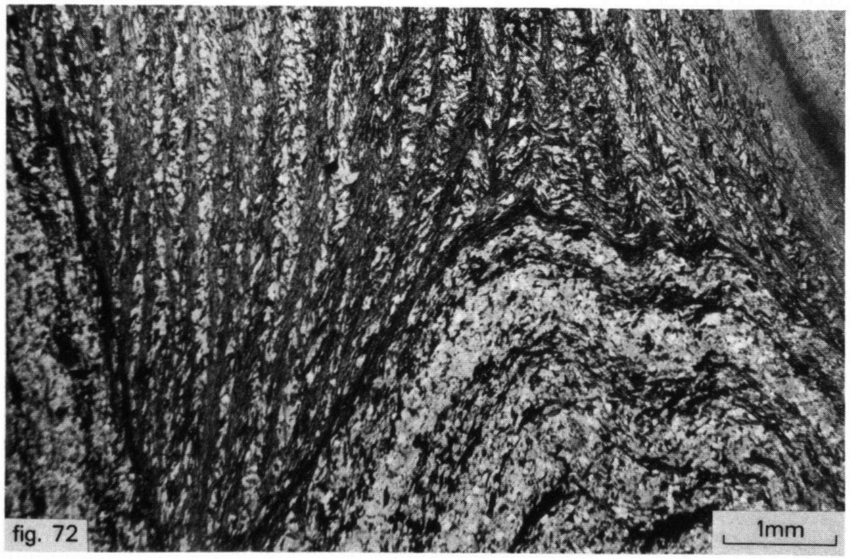
Crenulations belonging to the Lac d'Oo antiform, fold the micafabric and the andalusite crystals, and clearly postdate the peak of metamorphism. Also in this respect the evolution is similar to the Bosost area. Therefore S_1 in Lys-Caillaouas can be correlated with S_1 in Bosost as well as the growth of the aluminium silicates after S_1. The rotational movements with N–S axes and the resulting microstructures are the same in both areas. Crenulations, probably belonging to F_4 are postcrystalline. Closer to the Lys granite the rocks are higher grade, staurolite and andalusite disappear whereas sillimanite (fibrolite) is now a common constituent. The microstructure changes considerably, as the micafabric recrystallizes, and the preferred orientation tends to disappear. The late folds, however, deform mica and fibrolite (Fig. 76). The same features occur in the Bosost area in the deeper exposed part (Table 2).

In the Fredancon area, west of the Lys-Caillaouas massif, and outside our map area, quite interesting microstructures in micaschists occur, which are briefly described here. From published reports (Trouiller, 1976) it appears that there are three generations of folds. A first generation with E–W foldaxes develops a slaty cleavage, which is usually parallel or sub-parallel to bedding. S_1 is folded and a new cleavage S_x develops. On the southside of the Fredancon antiform this cleavage dips steeply to the south; mesoscopic folds have a vergence to the north (Groen, 1978). On the the crest of the antiform, S_x has quite a shallow dip. Its attitude on the northern side of the antiform is unknown. S_1 and S_x have been folded by a third generation with steep axial planes into the major Fredancon antiform. This generation can be correlated with the Lac d'Oo antiform and with F_4 further east. In the eastern part of the Fredancon antiform, in the Cinqueta valley, S_x is developed as a crenulation cleavage (Fig. 77), and the same applies to part of the Rioumajou valley. On the crest of the antiform in that valley, S_x has often developed into a new schistosity. This is indicated by internal fabrics in porphyroblasts. Lineations are common in the micaschists. They are formed by intersections of S_1 or S_x with the bedding, and due to fabric habit of biotite crystals. Both lineations are E–W to ESE–WNW.

In the central part of the Fredancon anticline many







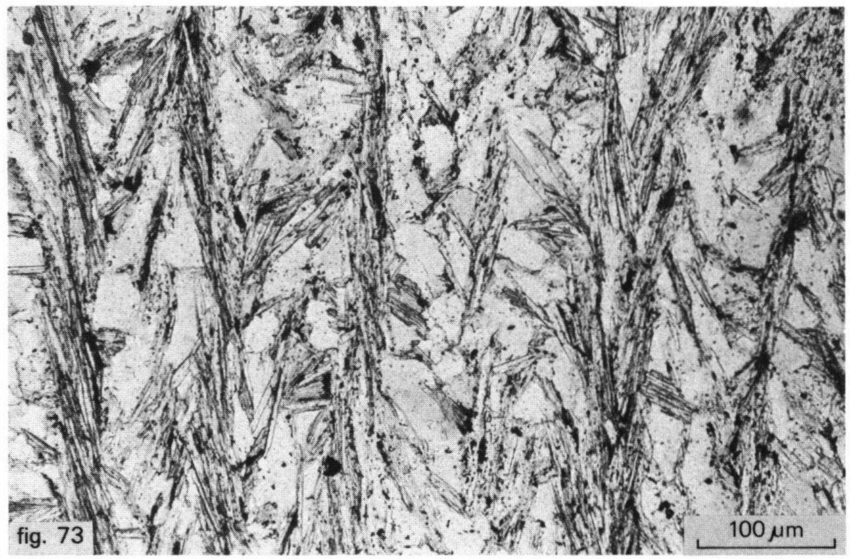
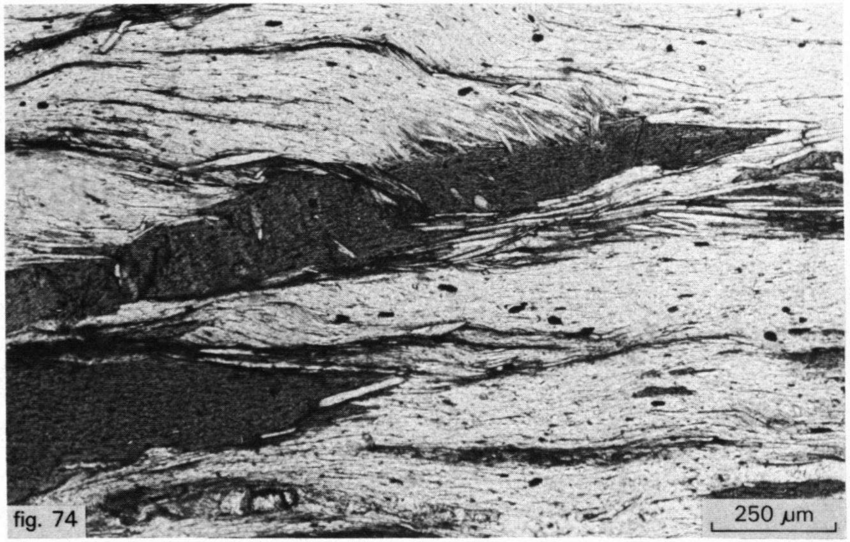


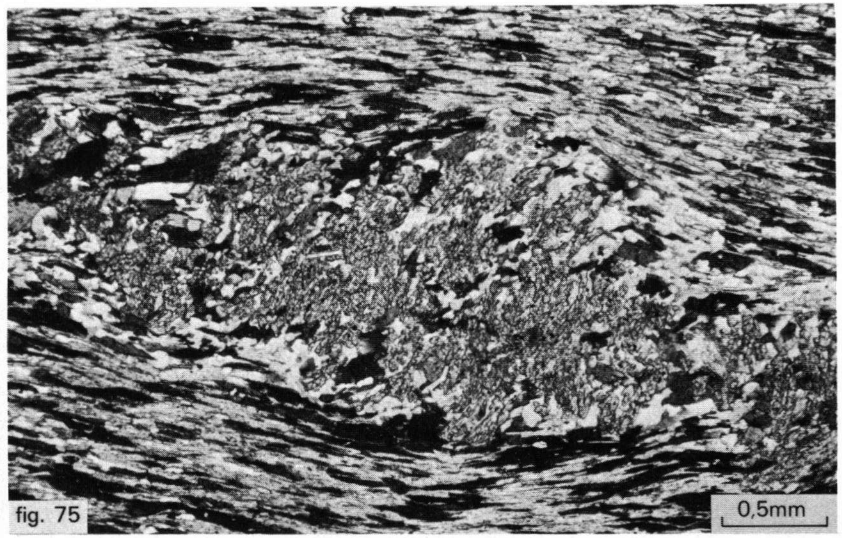
Fig. 70. Micaschists with shape and lattice orientation of biotite and muscovite, Bosost area, sheet 4.

Fig. 72. F_4 fold in layered micaschist, Barrados pegmatite, sheet 4.

Fig. 71. Schist from cordierite-sillimanite zone with unoriented fabric, Bosost area, sheet 4.

Fig. 73. Detail of Fig. 72 showing polygonal arcs of micas.





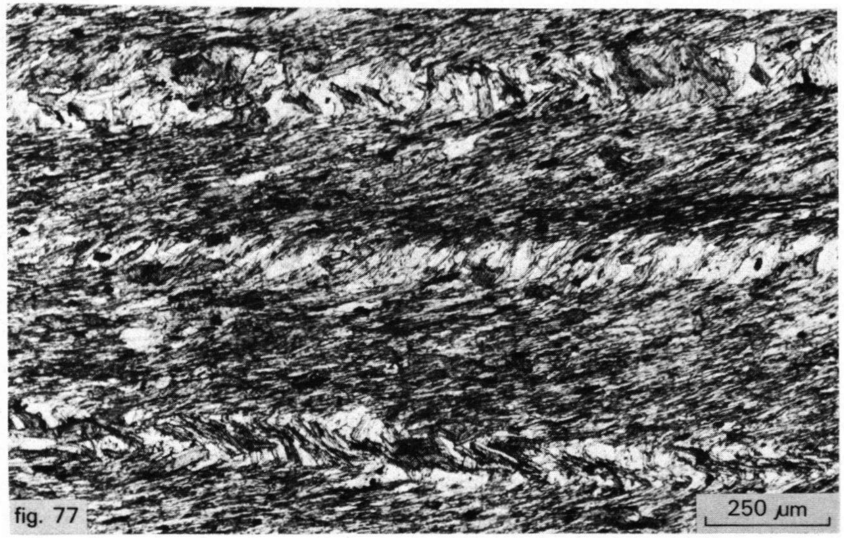
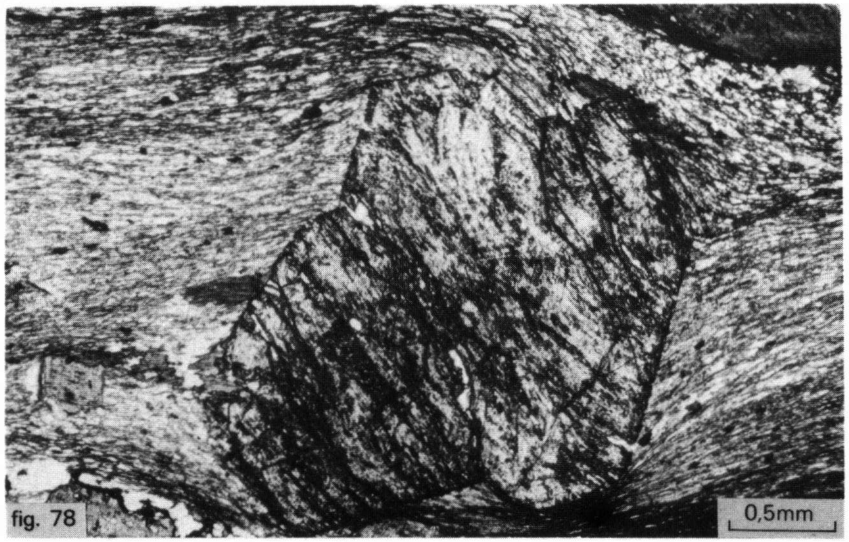


Fig. 74. Biotite porphyroblast rotated in foliation (S_1 or S_x), Fredancon antiform, west of Lys-Caillaouas massif.

Fig. 75. Rotated andalusite with N–S rotation axis, Lac d'Oo antiform, sheet 10.

Fig. 76. Folded fibrolite in micaschist; Lac d'Oo antiform close to granite contact, sheet 10.

Fig. 77. Schistosity (S_2 or S_x) with relics of crenulations southern limb Fredancon antiform, Cinqueta area, SW of Lys granite.



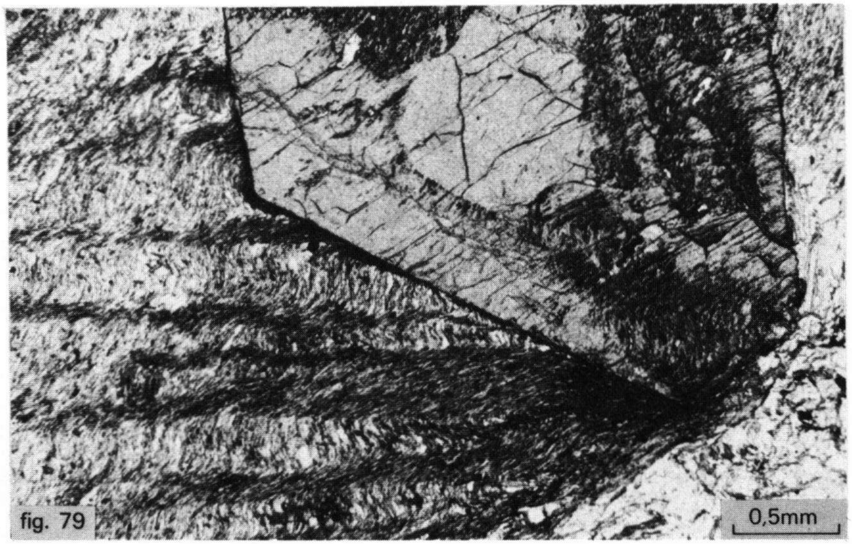


Fig. 78. Rotated staurolite, E–W section, clockwise rotation looking north, Fredancon antiform.

Fig. 79. Micaschist with F_x crenulations in matrix and as helicitic folds in staurolite, Cinqueta area, Fredancon antiform; N–S oriented section.

Fig. 80. Helicitic crenulations in staurolite (right half of photograph) with crenulations preserved in pressure shadow, Fredancon antiform; N–S oriented section.

Fig. 81. Staurolite with helicitic F_x folds in schistose matrix, Fredancon antiform; N–S oriented section.

porphyroblasts of staurolite, andalusite, cordierite, biotite, and to a lesser extent chloritoid and plagioclase occur in a matrix consisting mainly of quartz and muscovite. In all cases these porphyroblasts contain an internal foliated fabric which is inherited from S_1. This internal fabric may be planar, slightly s-shaped or crenulated, depending on the orientation of the thin section. In sections which are cut in E–W direction, parallel to the lineation, the inclusion pattern is planar or somewhat s-shaped, and from the relationship with the external fabric these crystals are clearly rotated (Fig. 78). In N–S sections perpendicular to the lineation many staurolites show an included crenulation cleavage, which in some cases can also be seen in the matrix (Fig. 79), sometimes only in the pressure shadow around the staurolite (Fig. 80), and in other cases the crenulation cleavage in the matrix has made way for a new schistosity, bowing out completely around the crystal (Fig. 81). This crenulation cleavage belongs to a second fold generation. From these relations it is obvious that staurolite started to grow after F_1, as it overgrows S_i, but its main development is during an early stage of a second phase, when a well-developed crenulation cleavage had already formed. Continued deformation obliterated in a number of cases all signs of these crenulations, and a completely new schistosity was produced (Fig. 82). In some examples in sections normal to the lineation a planar S_i is present, indicating growth after F_1 and before the second phase (called here F_x). Similar relations, but less common have been observed in andalusite. However, due to its elongate shape, rotation has not been observed as often, but in several specimens helicitic folds occur. Biotite crystals, usually overgrowing the matrix without any preferred orientation, may also show rotation (Fig. 83).

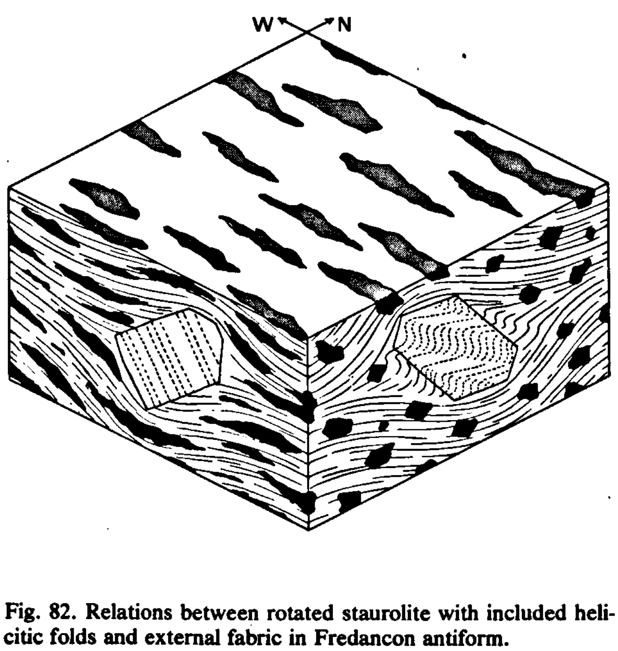


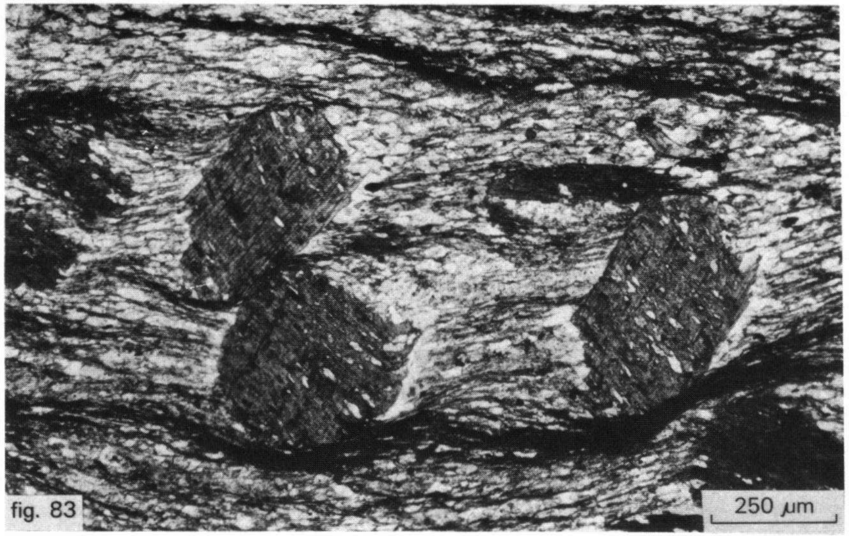
Fig. 82. Relations between rotated staurolite with included helicitic folds and external fabric in Fredancon antiform.

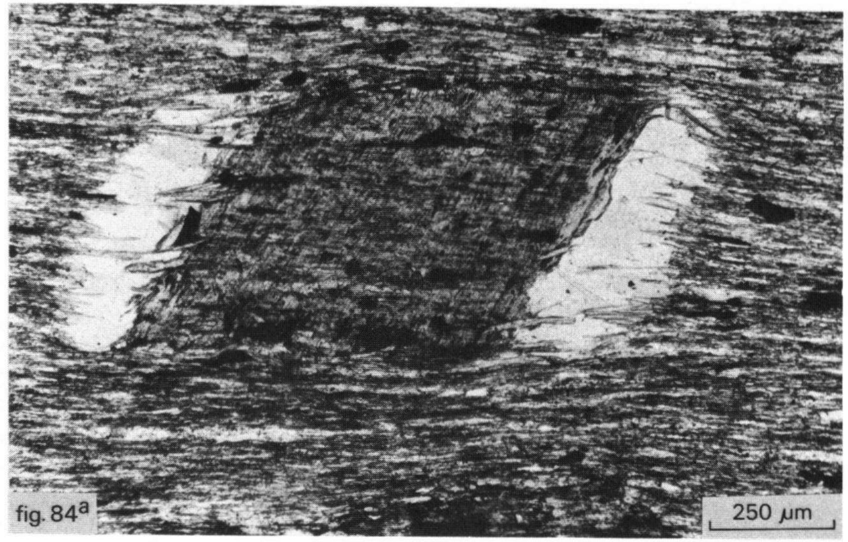
In other cases biotites are flattened and have an eye-shape, they are boudinaged, and sometimes they have pressure fringes, all indicating that F_x deformation was active after these biotites had grown (Fig. 84). In some thin sections biotite crystals are oriented in F_x crenulations (Fig. 85), also indicating growth during that folding phase. Chloritoid, a not very abundant mineral in the Fredancon antiform, also contains included F_x crenulations. As in the same thin section staurolite and andalusite are found, all three minerals can occur together in a stable assemblage. The succession of events in the Fredancon antiform is then as follows: F_1 folding under low grade (quartz + muscovite) conditions producing a slaty cleavage, S_1. Then beginning of growth of some biotite, staurolite and andalusite, followed by a folding phase, called F_x because of the uncertainty with regard to the elsewhere established fold succession, producing mesoscopic tight folds with a crenulation cleavage, S_x, in many rocks. Growth of many biotites, staurolites, andalusites and some cordierite and chloritoid crystals inheriting a crenulated fabric, is visible in N–S sections. Then, perhaps in a late phase of F_x, but otherwise as a new phase correlated with F_2, simple shear causing rotating of all existing porphyroblasts, and producing an E–W mineral lineation, resulting from rotation and extension of biotite in the S_x plane. Some minerals, like biotite and andalusite grew during this rotational stage as shown by s-shaped inclusions. In Table 2 the evolution of the various deformation phases with regard to mineral growth is represented.

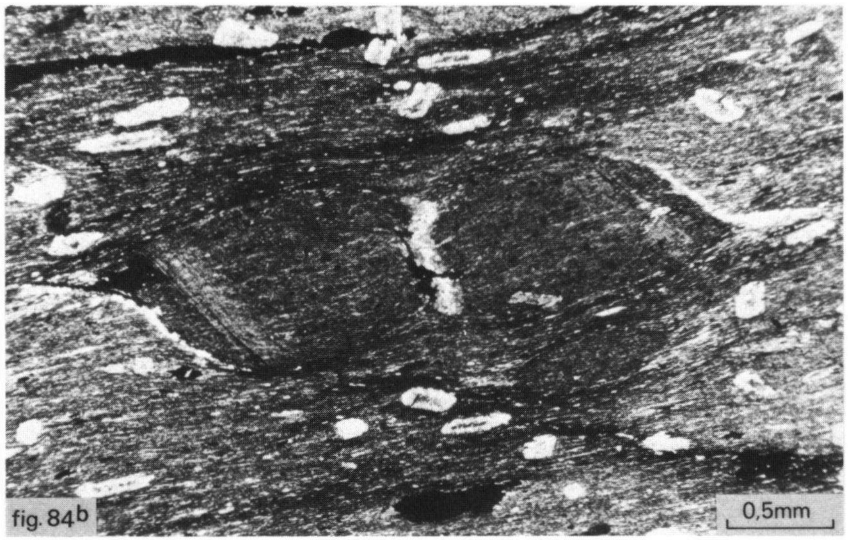
The difference between the Fredancon and Lac d'Oo antiform, is the absence of F_x folds and crenulations, and their inclusion in porphyroblasts in the Lac d'Oo area. The most probable correlation is that the low grade S_1 fabric is the same in both areas, and that the rotational movements with N–S axes are also the same. These movements are referred to as F_2 in our scheme. However, it now appears that in the Fredancon antiform another folding phase occurs between F_1 and F_2, which seems to be absent elsewhere in our map area (Table 2). It is to our opinion not probable that there is any relation with the F_0 folds from other parts of the Pyrenees. It is also difficult to correlate this succession with early N–S folds described from the Gavarnie area (Majesté-Menjoulas, 1979).

Discussion supra- vs. infrastructure

In an earlier part of this chapter the difference between the supra- and infrastructure with respectively steep and flat structures has been mentioned. After the treatment of the microstructures some more remarks can be made. Cleavage and schistosities in the two domains are determined by the parallel arrangement of metamorphic minerals, mainly quartz, chlorite and muscovite. Moreover in the steep slates of the suprastructure, recumbent structures are absent, and vice versa. Steep, late cleavages (S_4) occur in both domains. It also appeared that F_3 folds have the same orientation of axial planes in both domains. From the microstructures it appears that rotation of aluminium silicates with the same sense of movement occurs both in rocks of the infrastructure as







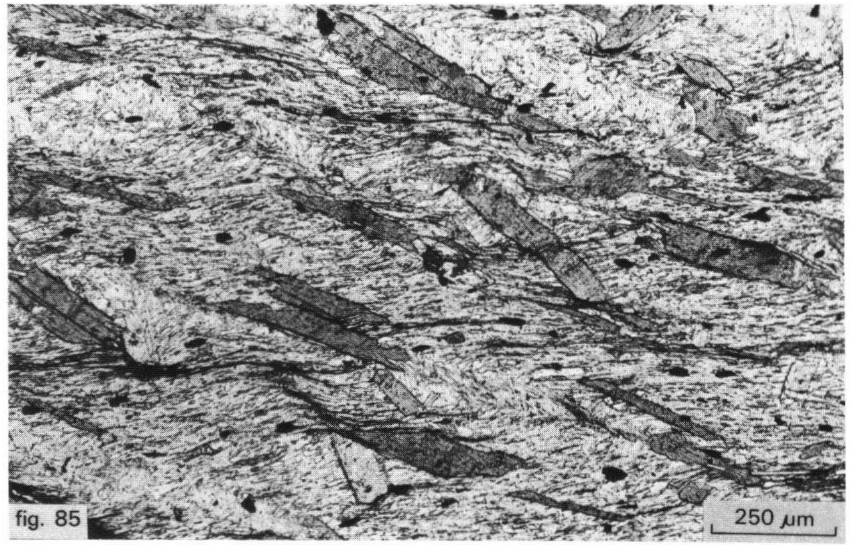


Fig. 83. Rotated biotite porphyroblasts; Cinqueta area, Fredancon antiform.

Fig. 84. Biotite with pressure fringes; Fredancon antiform.

Fig. 85. Biotite oriented in F_x crenulations; Cinqueta area, Fredancon antiform.

the Cambro-Ordovician of Bosost, and in suprastructural rocks as the Cambro-Ordovician rocks of the Lys-Caillaouas massif or the Devonian of the Bosost area. This indicates that the difference between the two domains existed before F_2. Consequently there is very little room in time for the formation of both domains. For this reason I believe that they have formed simultaneously, and to my opinion the arguments brought forward by Matte (1969), stating that the flat structures precede the steep ones, are invalid. Although a mechanical interpretation for the different behaviour of the two domains is difficult to give, it should be remarked that in many regions in the world low grade structures are steep, and high grade structures flatlying.

CHAPTER 4

METAMORPHISM

INTRODUCTION

In the Pyrenees metamorphism has taken place in three periods, a late Precambrian one, a Variscan one and an Alpine one.

Precambrian metamorphic rocks occur in the St. Barthélemy, Castillon, Agly, Arize and Trois Seigneurs massifs, but in the latter two massifs it concerns only small areas. A short description can be found in the chapter on stratigraphy. For more details I refer to Zwart (1954, 1959), Fonteilles (1970) and Roux (1977).

Alpine metamorphism is restricted to a narrow zone in Mesozoic rocks, north of the North-Pyrenean faultzone. A treatment is beyond the scope of this publication, and reference can be made to Ravier (1957), Avé Lallemant (1969) and Choukroune (1974). Therefore only the Variscan metamorphism is treated here.

Metamorphism of Variscan age involves three groups of rocks, the Precambrian basement, Palaeozoic sediments and pre-Variscan granites. According to some authors, e.g. Guitard (1963) these granites also are a part of the Precambrian basement, but to others, as the present author, they are of Ordovician age. Geochronology has not yet given a definite answer to this problem.

The effect of Variscan metamorphism on the Precambrian gneisses has not been worked out in detail. Some remarks have been made about this problem on p. 4 and 5, and these rocks will not be discussed further in this chapter. Metamorphism of Palaeozoic rocks is treated in three paragraphs, depending on grade of metamorphism, viz. slates and phyllites, micaschists and migmatites.

PALAEOZOIC METASEDIMENTS

Slates and phyllites

A large part of the Cambro-Ordovician, most of the Silurian and Devonian, and all of the Lower Carboniferous show a low grade of metamorphism, and are present as slates and phyllites. These rocks all have a foliation parallel to the axial planes of F_1 folds. As has been mentioned in the section on microstructures, the main minerals in these rocks are quartz, white mica, chlorite and opaques. In the Cambro-Ordovician of the Pallaresa anticlinorium and the Garonne dome fine-grained biotite has been found in a number of thin sections. In the Carboniferous of the Llavorsi syncline chloritoid is a ubiquitous constituent. The same mineral has been found in some Devonian and Cambro-Ordovician phyllites as well. So far this mineral, nor biotite has been found south of the southern limb of the Llavorsi syncline. North of this line all phyllites obviously are metamorphosed under lower greenschist facies conditions. The grade south of this line is uncertain. In view of the badly developed cleavage (see section of microstructures) it is not unlikely that the grade is lower, and that one is dealing with anchizonal metamorphism. There are, however, not yet any mineralogical data to support this hypothesis. From the microstructure it is clear that the greenschist facies assemblages, including the development of chloritoid took place during F_1, and there is a close connection between this metamorphism and the main folding phase. Crenulations belonging to F_3 or F_4 are always postcrystalline.

Micaschists

In the metamorphic massifs (St. Barthélemy, Arize, Trois Seigneurs, Aston-Hospitalet, Bosost and Lys-Caillaouas) the phyllites grade into micaschists. These schists are different from the phyllites by their grainsize and the presence of macroscopic visible biotite. With some more increase in grade, large porphyroblasts of aluminium silicates are generally present. These porphyroblasts are staurolite, andalusite, cordierite and sometimes garnet, and occasionally chloritoid. Furthermore sillimanite (often as fibrolite) may occur. The following zones can be distinguished in all metamorphic massifs: 1) muscovite-chlorite (-chloritoid) in phyllites, 2) muscovite-biotite, 3) staurolite-andalusite-cordierite, 4) andalusite-cordierite, 5) cordierite-sillimanite (Fig. 86, Table 3). Zone 1 has been dealt with in the previous section. Zone 2 consists of quartz, muscovite, chlorite and biotite. Zone 3 contains: quartz, muscovite, biotite, staurolite, andalusite, cordierite and occasionally garnet. In zone 4 occur quartz, muscovite, biotite, andalusite, cordierite, and in zone 5 quartz, muscovite, biotite, cordierite, sillimanite and a second generation muscovite. In all rocks opaques and other accessories are present, and usually some feldspar, mostly plagioclase. Zone 5 grades into a sixth zone which consists of migmatitic rocks with quartz, biotite, cordierite, sillimanite, plagioclase and potassium feldspar.

These progressive metamorphic zones are characterized by their remarkable thinness; from the first to the last zone a sequence not exceeding 1500 m in thickness may be present, whereas the sixth zone itself may have a considerable thickness. As has been observed before

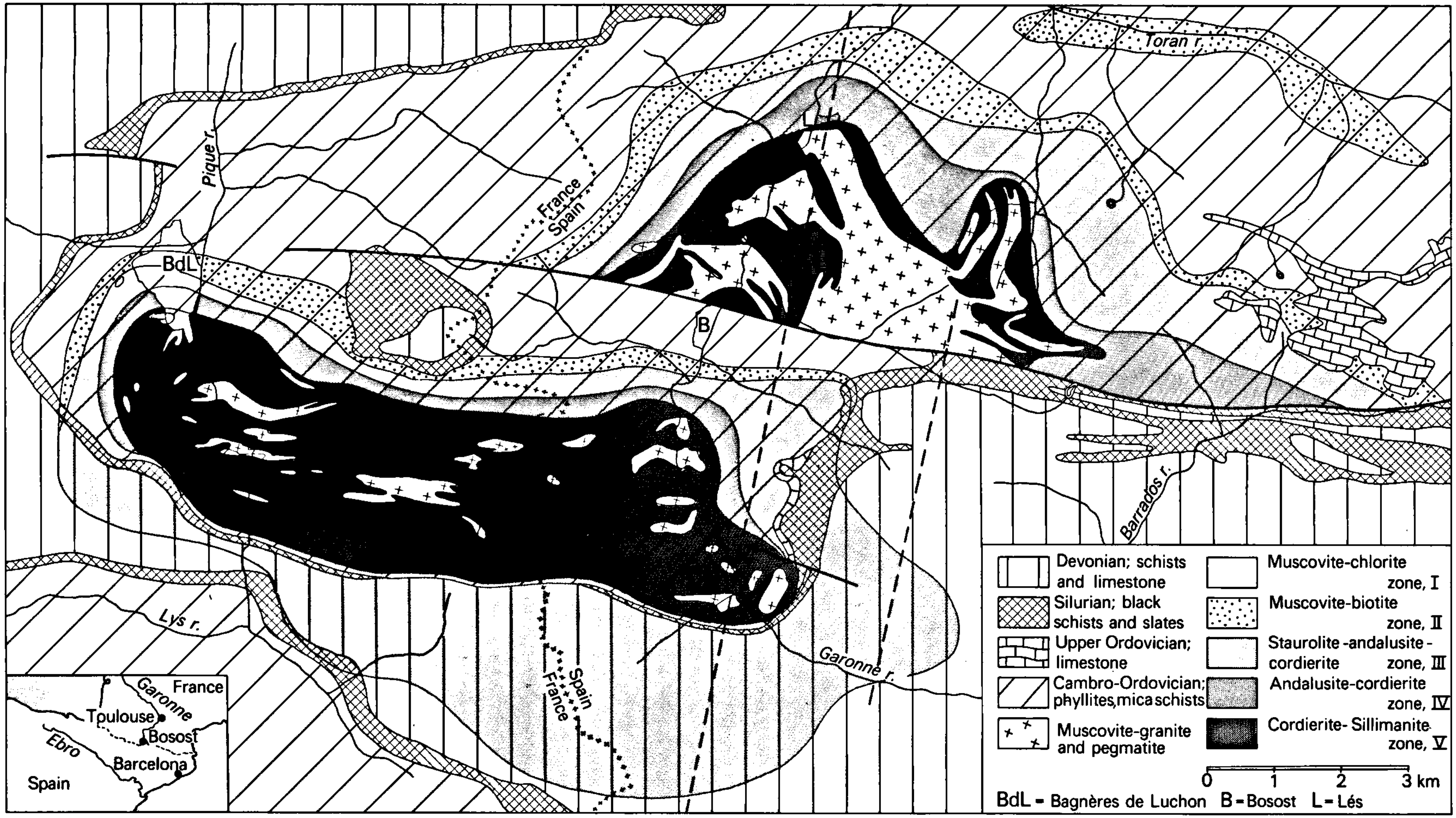


Fig. 86. Map of Bosost area, showing geology and metamorphic zoning.

(Zwart, 1962) this feature is due to the high geothermal gradients and the shallow depth of burial during metamorphism. Another important fact is that metamorphism was not only progressive in space, but also in time. This means that the higher grade zones were formed later than the lower grade ones (Zwart, 1962, 1968). This has been worked out with the aid of microstructures (see also Table 3). The muscovite-chlorite and muscovite-biotite zones for example always date from the main folding phase, indicating that nowhere during F_1 the grade reached beyond the greenschist facies. In some cases biotite was not even present during F_1, and higher grade metamorphism was overprinted on lower greenschist facies schists. This is the case in the Bosost and Lys-Caillaouas areas and in the outer part of the Aston massif. However, in the gneisses of the Hospitalet massif the assemblage quartz, plagioclase, potassium feldspar, biotite, muscovite dates from the main phase and contains no overprinting effects.

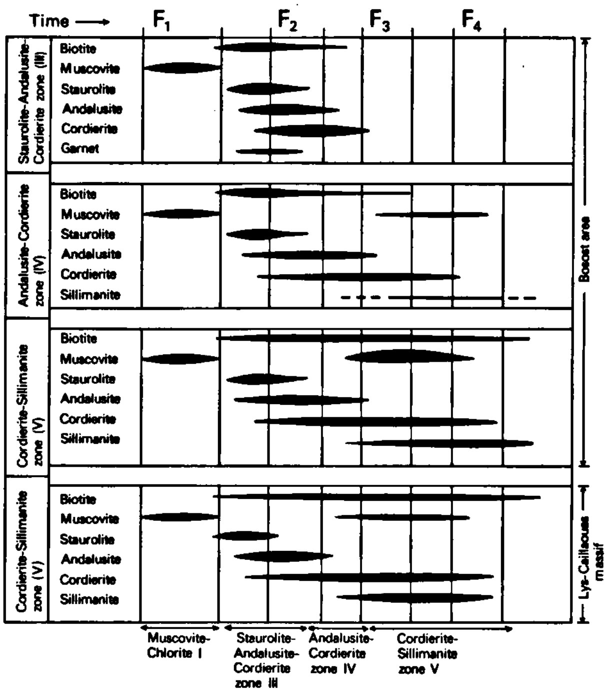


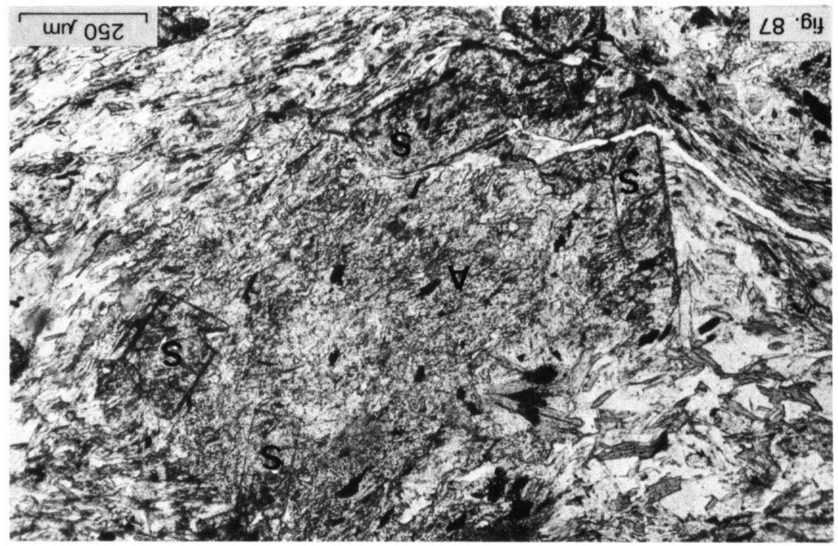
Table 3. Relations between successive metamorphic zones and folding phases in Bosost area and Lys-Caillaouas massif.

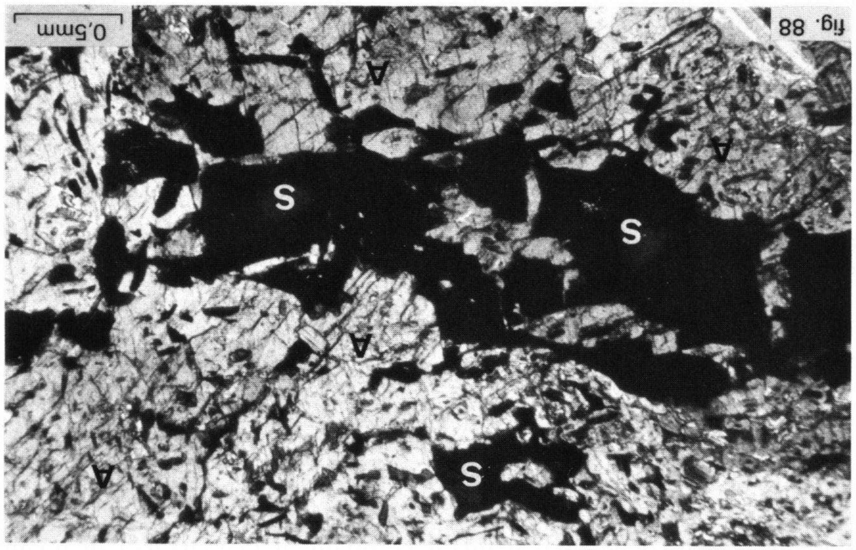
The increase in temperature is best recorded in the micaschists, which consequently have a plurifacial history. From microstructural criteria, the inclusion of one mineral in another and the replacement of minerals, it can be concluded that in general staurolite is the first aluminium silicate to appear. Chloritoid is too scarce in micaschists to make a definite statement about its age. There are, however, schists in which staurolite, chloritoid and andalusite seem to be a stable assemblage. As idioblastic staurolite may be included in andalusite, but never vice versa, andalusite starts to grow after staurolite on increasing temperature (Fig. 87). Somewhat later cordierite appears in the assemblage, so that in fact the rocks pass first through a staurolite zone, then a staurolite-andalusite zone, to finally enter a staurolite-andalusite-cordierite zone. However, there is so little temperature difference between the devélopment of these three minerals, that most schists within the aluminium-silicate-bearing rocks contain all three minerals, and a separate staurolite or staurolite-andalusite zone cannot be mapped in the field. Generally the growth of staurolite is preceded by porphyroblastic biotite (Table 2), as unoriented biotite is commonly present within any of the three aluminium silicates (although rarely in staurolite, Fig. 63). With increasing grade staurolite becomes unstable. It may be replaced by andalusite or cordierite (Figs. 88, 89), so that a separate andalusite-cordierite zone can be distinguished. In this zone there are no reactions between andalusite and cordierite. Moreover this zone is usually very thin, and does not exceed a few hundreds of metres (Fig. 86).

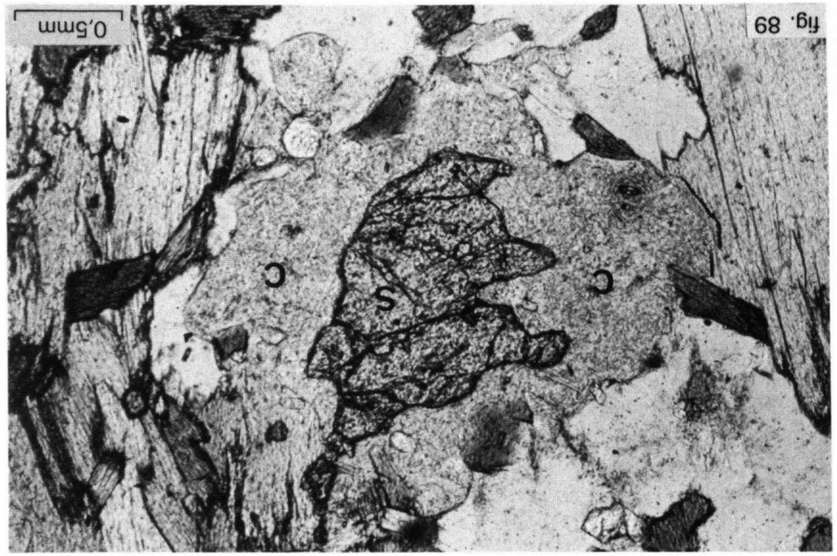
In the next higher grade zone several changes take place. Firstly pegmatite sills and dykes are always present. Secondly staurolite, which still may be present as relics, is transformed to an aggregate of unoriented muscovite or muscovite-biotite (Figs. 90, 91). In a previous paper (Zwart, 1962) I have suggested that this reaction was a retrograde one, and postdates the highest grade of metamorphism. However, the same feature has been observed by Guidotti (1968) from schists in Maine (USA), and was interpreted by him as a prograde succession. This interpretation fits the rocks in the Pyrenees much better, as in the same rocks also fibrolite is present. Thirdly andalusite becomes unstable and is altered to cordierite, fibrolite or muscovite (Fig. 92), so that an assemblage quartz, muscovite II, cordierite, sillimanite appears. This is also the highest grade reached in micaschists. With further increase in temperature the rocks become migmatized and a new assemblage is formed. In the highest grade micaschists there is also a dramatic change in microstructure; as the schistose matrix recrystallizes, quartz, muscovite and biotite form a more or less unoriented aggregate (Fig. 71) indicating that this metamorphism is of a static type and postdates F_1 and F_2 folding. F_3 and F_4 folds may occur in these schists, but seem to have no influence on the schists, other than folding the foliation and producing crenulations in which the micas form polygonal arcs and are clearly recrystallized (Fig. 73).

Migmatites

In the North-Pyrenean massifs (St. Barthélemy, Arize, Trois Seigneurs), the Aston massif, and in a few outcrops near Bosost, the micaschists grade into migmatitic rocks (Zwart, 1954, 1959; Allaart, 1958; Verspyck, 1965). This change is the result of the presence of thin layers, lenses and pods of quartzo-feldspathic material in the schists, giving them a typical migmatitic appearance. At the same time the rocks become coarser grained. As sillimanite is a common constituent of these rocks, they have also been called sillimanite-gneisses. The micaceous layers are foliated, but the quartzo-feldspathic bands are generally unoriented. The thickness of the two components does not usually exceed a few centi-







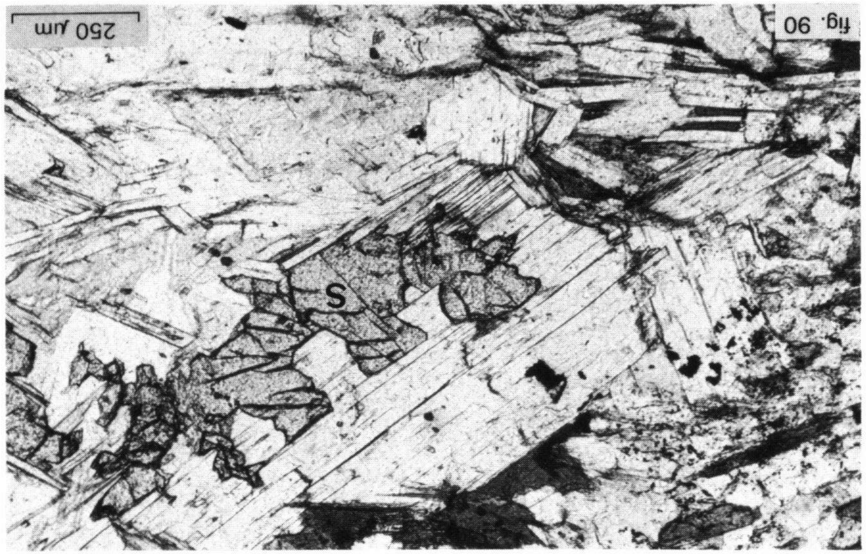


Fig. 87. Idioblastic staurolite (S) within andalusite (A), Devonian schist, Bosost area, sheet 4.

Fig. 89. Staurolite (S) relic in cordierite (C), Bosost area, sheet 4.

Fig. 88. Staurolite (S) relics (in extinction position) in andalusite (A), Etang Fourcat, Aston massif, sheet 6.

Fig. 90. Shimmer aggregate of muscovite, replacing staurolite (S), Bosost area, sheet 4.

Fig. 92. Andalusite (A) replaced by fibrolite (F), Bosost area, sheet 4.

Fig. 91. Staurolite relic in shimmer aggregate of muscovite, Bosost area, sheet 4.

metres. The leucocratic layers are in general conformable with the foliation of the micabands, but crosscutting contacts occur as well. Pegmatites, either as concordant or discordant bodies, occur frequently. With increasing intensity of migmatization, which occurs deeper in the sequence, the rocks tend to lose their migmatitic character and acquire a more granitic structure, although inhomogeneities always remain present, as well as inclusions of quartzite, marble and calcsilicate rocks. These more homogeneous rocks are quartz-diorites as they contain little potassium feldspar. The best outcrops of such quartzdiorites can be found in the Trois Seigneurs massif (Allaart, 1959).

The sillimanite-gneisses are often strongly folded. In these folds the layering and the foliation are deformed, and there is no new axial plane foliation. However, the micas are recrystallized in the foldhinges. Obviously the folds belong to one of the late folding phases, and migmatization clearly postdates F_1 and F_2 deformation. The axial planes and axes of the folds in the migmatites are, however, rather irregular, and it is difficult to assign them to F_3 or F_4. However, in view of the fact that in the Aston massif, metamorphism in micaschists reached its peak before and during F_3, there are good reasons to believe that migmatization took place already during F_3, but continued in F_4 time.

In the quartzdiorites, mobilization is evident from the disruption and rotation of quartzite and limesilicate layers (Allaart, 1959), indicating that these rocks have been very plastic indeed, and must have approached melting conditions (Fig. 93).

The mineralogy of the sillimanite-gneisses is as follows. The micaceous layers consist of biotite, sillimanite, muscovite and often some cordierite, the quartzofeldspathic layers of quartz, oligoclase, minor potassium feldspar and cordierite. From relationships between muscovite and the other minerals, it is evident that muscovite does not belong to the primary assemblage and was formed in a late stage as a retrogressive mineral. Sillimanite occurs as fibrolite formed from biotite, and as prismatic crystals. The quartzdiorites consist of quartz, oligoclase, biotite, sillimanite, cordierite, minor potassium feldspar, and muscovite as a late retrogressive mineral.

It is evident that in these rocks the highest grade reached is the upper amphibolite facies, and under these conditions the rocks start to melt. The presence of much water apparently enhances this process and prevents that granulite facies conditions are reached. The upper amphibolite facies is confirmed by assemblages in carbonate rocks. These contain: diopside, An-rich plagioclase (up to bytownite), grossularite, wollastonite, green hornblende, forsterite, spinel, clinohumite and pargasite. For more details, reference is made to the publications by Zwart (1954, 1959), Allaart (1959) and Verspyck (1965).

Pressure-temperature conditions of metamorphism

The mineral assemblages in pelitic rocks are clearly indicative of the low pressure character of the Variscan metamorphism. The parageneses staurolite-andalusite-

Fig. 93. Rotated inclusion in migmatitic quartzdiorite, Trois Seigneurs massif, sheet 3.

cordierite, and cordierite-sillimanite-muscovite in micaschists, and cordierite-sillimanite-potassium feldspar are critical. Especially the rapid transgression of cordierite-sillimanite-quartz-muscovite rocks to muscovite free assemblages indicates that metamorphism occurred very close to the intersection of the phase boundaries andalusite $\rightleftharpoons$ sillimanite, and muscovite + quartz $\rightleftharpoons$ K-feldspar + sillimanite. According to Winkler this intersection lies at 2.5 Kb and 640°C, and at these conditions anatexis and migmatization should start. The rocks would enter the staurolite-andalusite-cordierite zone (also beginning of the amphibolite facies) at 540 °C and 2 Kb. However, these figures are not in agreement with field evidence. As the stratigraphic column is well known, as well as the structure, the total overburden during metamorphism can be calculated. It has been estimated at 2500 m for the muscovite-biotite zone, 3500 m for the staurolite-andalusite-cordierite zone, and 4500–5000 m for the cordierite-sillimanite-potassium feldspar zone. This would mean pressures of 1 Kb for boundary greenschist-amphibolite facies, and 1.5 Kb for the beginning of migmatization. These are minimum pressures valid for the Bosost area. Further east the isograds lie deeper with regard to the stratigraphic sequence, and pressures have been higher there. This is confirmed by assemblages in the eastern Pyrenees, where the isograds are lying even deeper, and almandine is a common mineral of the micaschists (Guitard, 1965).

Based on our observations in the Bosost area, geothermal gradients must have exceeded 100 °C/km.

The low pressure character of the Bosost area has been confirmed by Sassi & Scolari (1974) using the b_0 values in potassic white micas of low grade rocks. According to their measurements the Bosost metamorphism should be the lowest pressure regional metamorphism found so far. It is in any case corroborated by field evidence.

Geochemistry

A great number of chemical analyses of the metasediments have been executed by the late Mrs. C. M. de Sitter-Koomans. They have been published in de Sitter & Zwart (1959) and Zwart (1959, 1963, 1965). The separate analyses are not reproduced here, but the averages of the various rock groups, that is phyllites, micaschists, and migmatites, the latter divided in sillimanite-gneisses and quartzdiorites, are represented in Table 4.

Fig. 94. Contact metamorphic Cambro-Ordovician slate near Andorra granodiorite (sheet 5), showing unoriented biotite crystals overgrowing slaty cleavage.

Fig. 95. Spotted slate of contact of Bassiès-Auzat granodiorite, with cordierite porphyroblast, predating F_4 crenulations.

From this table it appears that there is no significant difference between phyllites, micaschists and sillimanite-gneisses. They show a high aluminium excess, low sodium and fairly high potassium contents. The composition of the quartzdiorites is different. SiO_2 is higher, Al_2O_3 lower, whereas sodium and calcium are also increased. These differences seem to be significant. Although variation in original chemistry is not altogether excluded, it should be stated, that some quartzdiorites are from rocks fairly high in the Cambro-Ordovician sequence, whereas elsewhere phyllites are taken from the deeper parts. This is possible because the migmatite front lies at different levels, generally deep in the east, and more shallow in the west. Therefore it is probable that differences in the parent rock were not very large and that the variation in chemical composition of the various rock groups is significant. These differences are probably due to introduction of Si and Na in those rocks which were intensely migmatized, mobilized, and perhaps partly molten. The higher CaO content may be the result of the digestion of calcareous layers occurring in the Cambro-Ordovician. For the sake of comparison also an average of 9 analyses of intrusive granodiorites is added in Table 4. It appears that there is very little difference with the quartzdiorites, except that the calcium content is significantly higher. Assuming that this is due to the incorporation of calcareous rocks in the magma, it seems quite probable that these granodiorites are produced by melting of Cambro-Ordovician rocks at a somewhat deeper level. In this way the quartzdiorites can be considered as autochthonous, and the granodiorites as allochthonous members of a granite suite.

	50 phyllites	46 micaschists	9 sillimanite gneisses	20 migmatitic quartz-diorites	9 intrusive grano-diorites
SiO_2	60.2	60.6	60.1	65.6	64.1
Al_2O_3	19.5	19.1	19.3	16.5	16.4
Fe_2O_3	2.3	2.7	1.9	1.5	1.1
FeO	4.3	4.4	4.6	3.1	3.5
MgO	2.3	2.4	2.8	2.1	2.0
CaO	1.1	1.6	2.2	2.2	4.5
Na_2O	2.0	1.9	1.9	3.0	3.2
K_2O	3.5	3.9	4.1	3.4	2.9

Table 4. Average chemical analyses of phyllites, micaschists, sillimanite-gneisses, migmatitic quartzdiorites and intrusive granodiorites.

PRE-VARISCAN GRANITES

In the Aston-Hospitalet massif large bodies of a felsic orthogneiss are found. They have also been mentioned from the St. Barthélemy massif (Zwart, 1954). In the Aston massif this gneiss occurs as a thick sheet, as it is underlain by migmatites and quartzdiorites formed from Cambro-Ordovician pelitic rocks. In the Hospitalet massif the orthogneiss forms the core of a large antiformal structure and its base is not exposed. In both massifs the gneisses are overlain by micaschists of Cambro-Ordovician sedimentary age (Profile 1). In the last-named massif the gneiss occurs as a linear augengneiss with predominant low-dipping schistosity but following the antiformal shape, and E–W lineations. These structures are formed during the main phase of the Variscan orogeny. Consequently the original material predates this major event. The derivation of these gneisses from original granites follows from their very homogeneous character and their granitic chemistry. The contacts with the metasedimentary rocks are always sharp.

The mineralogical composition is quartz, albite/oligoclase, potassium feldspar, biotite and muscovite. Both feldspars contribute to the eyed character of the rocks, whereas quartz and micas form a foliated fabric between these augen.

In the Aston massif, augengneisses occur also, but in this region the gneisses have been migmatized and granitized, resulting in flasergneisses, granitic gneisses and granites. The mineralogical composition is similar to the rocks of the Hospitalet massif. The migmatization is accompanied by mobilization, which results in irregular folding and disorientation of gneiss remnants in a granitic matrix.

In the Aston massif, gneisses with quartz-sillimanite nodules occur at various places. They have been described by Zwart (1965) and Verspyck (1965). They are formed by removal of alkalis during the migmatization, resulting in the formation of shearzones covered with quartz and sillimanite. Eventually these recrystallize into nodules, similar to those described from many other regions with high grade metamorphism (Losert, 1968). The chemical composition of these gneisses is discussed by Zwart (1965). An average of 23 analyses is shown in Table 5.

SiO_2	73.0
Al_2O_3	13.7
Fe_2O_3	0.8
FeO	2.0
MgO	0.8
CaO	1.5
Na_2O	3.1
K_2O	4.6

Table 5. Average chemical analysis of pre-Variscan granites of Aston-Hospitalet massif.

Some controversy about the age of the original granite exists. According to Jäger & Zwart (1968) the granites were emplaced 475 m.y. ago, based on a Rb-Sr isochron.

In the Canigou massif where similar orthogneisses are exposed, Guitard (1970) believes that they belong to a Precambrian basement upon which the Cambro-Ordovician was deposited unconformably. Geochronological measurements have given dates of 535 m.y., that is Lower Cambrian. If Guitard's interpretation is correct, the granites were formed during the Cadomian orogeny. Although conclusive evidence cannot be given, the present author thinks it to be more likely that these rocks belong to the ubiquitously occurring suite of granites within the Variscan belt, and intruded mainly during Ordovician times.

CHAPTER 5

LATE VARISCAN INTRUSIVE ROCKS

INTRODUCTION

In the Pyrenees intrusive rocks occur at many places. They can be subdivided into three groups: 1) granodiorites, 2) muscovite-granites, and 3) dykes. A special place is taken by the Lys-Caillaouas granite, which has a composite structure.

GRANODIORITES

Bodies and batholiths of granodiorite occur commonly in the Pyrenees. They are intrusive in Cambro-Ordovician, Silurian, Devonian and Carboniferous rocks, and in most cases in a low grade environment. Granodiorites in high grade micaschists or migmatites are rare; the only example in our map area is the small stock on the southern side of the Trois Seigneurs massif. The bodies are generally elongate in the direction of the general trend of the Variscan structures, like the Bassiès-Auzat and the Maladeta massifs, but they are discordant with regard to the country rock, and clearly postdate the S_1 cleavage. Their relations to the folds have been described on p. 37.

The granodiorites in the various bodies are quite similar. They are unoriented rocks of medium grainsize and quite homogeneous throughout. The mineralogical composition is quartz, oligoclase/andesine, potassium feldspar and biotite. Hornblende bearing varieties do occur occasionally. For more details reference can be made to de Sitter & Zwart (1959), Zandvliet (1960), Zwart (1965), Mey (1968), and Hartevelt (1970).

The chemical composition is that of a typical granodiorite. An average of nine analyses is represented in Table 4. Their derivation by melting of Cambro-Ordovician sediments seems probable and is discussed on p. 67.

The granodiorites have a well-defined aureole of contact metamorphic rocks, consisting of hornfelses and spotted slates. Common contact metamorphic minerals are biotite, muscovite, andalusite and cordierite (Figs. 94, 95). Staurolite has not been observed as a contact metamorphic mineral, despite its frequent occurrence in regional metamorphic rocks. An explanation for this behaviour of staurolite is difficult to give.

In pelitic rocks close to the contact of the Bassiès-Auzat batholith, corundum has been found in association with andalusite. In the graphitic slates of Silurian age, chiastolite is the main contact metamorphic mineral.

In calcareous rocks the following new minerals have been observed: grossularite, vesuvianite, clinopyroxene, epidote-clinozoisite, green hornblende, bytownite, wollastonite, sphene, zoisite and prehnite.

In the hornfelses, occurring closest to the granodiorite, the original slaty cleavage has more or less disappeared, and the new minerals form an unoriented microstructure. In the spotted slates, forming the outer parts of the aureole, the slaty cleavage is still well recognizable, and relationships between the contact metamorphic minerals, the cleavage and later crenulations can be established. It appears that in some bodies, like the Maladeta, contact metamorphism overprints the F_4 folds. Elsewhere as in the Bassiès-Auzat and Andorra granodiorites F_4 crenulations postdate contact metamorphism. For these reasons it seems probable that some granodiorites intruded before F_4, others after F_4 (Fig. 95). One has to keep in mind, however, that F_4 is not necessarily of the same age throughout the Pyrenees, so that there is no guarantee that there is a time difference in the intrusion of the various bodies. There is no doubt, however, that they are late kinematic, and must be of pre-Stephanian, and probably of pre-Westphalian D age. On the other hand they are post-Namurian, and perhaps even post-Westphalian A.

MUSCOVITE-GRANITE

Muscovite-granites occur as small bodies, stocks and patches within the regional metamorphic terrains of the Pyrenees, like in the North-Pyrenean massifs, Aston massif, Lys-Caillaouas massif, and Bosost area. The country rock is always of amphibolite facies grade, either as aluminium-silicate-bearing micaschists or as migmatites. Sometimes they occur as well-defined, sharply bounded bodies, as in the Trois Seigneurs massif and in the Bosost area, but elsewhere as small patches, sills, dykes and irregular masses in the augengneisses of the Aston massif. The first group must be considered as intrusive, but rather parautochthonous than allochthonous, as they probably were produced not far from the place where they occur. The second group is thought to be produced in place by melting of the felsic orthogneisses. The muscovite-granites are always accompanied by pegmatite sills and dykes. The mineralogical compositon is quartz, albite/oligoclase, microcline, muscovite and biotite. Occurring in high grade rocks, they have no contact aureole. Chemically they resemble the pre-Variscan granites, and it is suggested here that they are the product of melting of these rocks, a process which occurred undoubtedly in the Aston massif.

LYS-CAILLAOUAS MASSIF

The large body of Lys-Caillaouas (Clin, 1959; Clin et al., 1963) is of a particular nature. It is a composite intrusion consisting of several types of granitic rocks. In addition it contains many large inclusions of Cambro-Ordovician, Silurian and Devonian metasediments, all strongly metamorphosed. The whole body occurs in micaschists with amphibolite facies assemblages. It is folded by F_4 folds, and the granite clearly predates this folding phase (Fig. 29).

The lowest part of the massif consists of a megacrystal granite with large potassium feldspar crystals, more or less oriented in an E–W direction. Besides this mineral the rock contains quartz, oligoclase, biotite and muscovite. Large and small inclusions in this granite are mainly Cambro-Ordovician micaschists. The megacrystal granite is overlain by a thin and discontinuous layer of fine-grained muscovite-granite, consisting of quartz, microcline, albite, muscovite and some biotite. Structurally overlying this granite occurs a quartzdiorite forming the core of a synform (synform of Gias, Fig. 29). The quartzdiorite is a medium-grained rock containing the following minerals: quartz, andesine, biotite and hornblende. It has usually a foliated structure. The quartzdiorite may grade into hornblende-gabbro. Large inclusions of Silurian and Devonian metasediments occur in the quartzdiorite, or between the quartzdiorite and an underlying sheet of porphyritic granite. One of these inclusions forms a part of the Gias synform, which obviously is an F_4 structure. Isoclinal or tight F_1 folds occur also in the inclusions. South of this synform the granite massif is cut off by the Esera-Gistain fault (=faille de Consaterre of Clin, 1962). The interpretation of the intrusion of the granite and associated rocks and the structure is as follows: a sequence of Cambro-Ordovician-Silurian and Devonian rocks was folded by F_1 under greenschist facies conditions, and was associated by the development of an axial plane foliation. This foliation had probably a steep to moderate dip to the south. The Siluro-Devonian must have formed the core of a large F_1 syncline, which no longer can be recognized. After F_1 the temperature rose, and aluminium-silicates started to grow in the staurolite-andalusite-cordierite zone. These minerals were deformed or rotated by F_2. After this event the megacrystal granite intruded as a sheet in the Cambro-Ordovician, and the quartzdiorite and muscovite-granite in the Siluro-Devonian rocks, but including large rafts of country rock. Near the contact and in the inclusions the rocks were metamorphosed in the cordierite-sillimanite zone, undoubtedly by the heat of the granite. Corundum has also been found in these inclusions. Finally the granite and the quartzdiorite with inclusions, and the surrounding micaschists were folded by F_4 in two large antiforms (Lac d'Oo and Fredancon) and one large synform (Gias) in between. There is a clear connection between the granite intrusion and the metamorphism. However, part of the metamorphism preceded the emplacement of the granite, but the highest grade reached, clearly coincides with the intrusion. Furthermore the Lys-Caillaouas granite is earlier than the intrusive granodiorites elsewhere (Table 6).

DYKES

Dykes in rocks of Palaeozoic age, older than the Westphalian have been described from most of the map sheets by Keizer (1954), Zwart (1954, 1965), de Sitter &

Zwart (1959, 1962), Kleinsmiede (1960), Zandvliet (1960), Mey (1967) and Hartevelt (1970). They may occur as dyke swarms like on sheets 1 and 4, or more isolated. They are often intruded parallel to the S_1 cleavage, and obviously postdate F_1 folding. Crosscutting contacts have also been observed. Some dykes possess a cleavage, which supposedly was acquired during one of the late folding phases. Their thickness varies from a few tens of centimetres to 20–30 metres, and occasionally to 250 m. Most common are quartzdiorite-porphyrites, containing phenocrysts of quartz, sodic plagioclase, biotite and occasionally potassium feldspar in a strongly altered fine-grained matrix. Less widespread are hornblende-diorite porphyrites in which green or brown amphibole occurs as phenocrysts, and have less quartz. Aplite dykes consisting of quartz, feldspar and muscovite have also been found at various places. Furthermore granodiorite and lamprophyre dykes occur in or near the granodiorite bodies.

Like the granodiorites most dykes are of late-kinematic Variscan age.

This paper (also Zwart, 1963; Oele, 1966)			
Symbol of fold phase	Structures	Metamorphism	
	Faulting? Shouldering aside of S_1 cleavage by intruding granodiorite bodies		
	Tilting and fanning of S_1 cleavage; faulting kinkbands		
F_4	Minor and locally major (Lys-Caillaouas) folds Axial plane: steep to vertical, E–W Foldaxis: E–W, gently plunging to horizontal Foliation: crenulation cleavage, S_4		
F_3	Minor folds; mainly in metamorphic areas Axial plane: steep, NW-SE (F_{3a}) and NE–SW (F_{3b}) Foldaxis: horizontal NW or NE to steeply plunging Foliation: crenulation cleavage, S_3	Upper Amphibolite facies	Cordierite-sillimanite-K-feldspar-zone, VI Cordierite-sillimanite-zone, V.
F_2	Minor folds in metamorphics of Aston massif and Bosost area Axial plane: recumbent Foldaxis: N–S Foliation: schistosity, or reactivation of S_1 with W to E simple shear, causing rotation of minerals in Bosost and Lys-Caillaouas areas	Lower Amphibolite	Andalusite-cordierite-zone, IV
F_x	Minor folds Axial plane: steep to recumbent Foldaxis: E–W Foliation: crenulation cleavage or schistosity (S_x) only observed in Fredancon antiform	Facies	Staurolite-andalusite-cordierite-zone, III
F_1	Major and minor folds in whole Palaeozoic sequence, often called Main phase Axial plane: vertical to inclined in low-grade areas; recumbent in higher grade areas Foldaxis: close to E–W Foliation: slaty cleavage and crenulation cleavage in low-grade areas; schistosity in high-grade areas	Greenschist Facies	Muscovite-biotite-zone, II Muscovite-chlorite-(chloritoid)-zone, I
F_0	Major folds in Devono-Carboniferous; minor folds in Cambro-Ordovician; folds in metamorphics?? Axial plane: vertical to steeply inclined Foldaxis: close to N–S Foliation: usually absent This phase has also been called pre-cleavage or pre-main phase	No, or only local	metamorphism

Table 6. Structural and metamorphic evolution and intrusive activity of Variscan events in the Pyrenees, and comparison with earlier publications.

ACKNOWLEDGEMENTS

It is a pleasant task to express some words of thanks to all those who have been instrumental in producing this memoir.

First of all I should mention the great number of students who took care of the major share of the fieldwork, resulting in the large quantity of geological maps accompanying this final publication. They are too numerous to be listed here, but their names appear on the 1:50,000 sheets. With most of them I spent many hours or days in the field, of which I treasure up many unforgettable memories.

A special word of thanks should go to my secretary Mrs. T. W. Terpstra, earlier secretary to professor de Sitter. Not only has she patiently typed and retyped the manuscript of this publication, but almost all our Pyrenean papers and map descriptions of the past 24 years have first gone through her hands and typewriter.

I am very grateful to Mrs. C. J. P. Roest, who has

	Boschma, 1963	Mey, 1967	Hartevelt, 1970	Muller & Roger, 1977	Déramond, Soula, Majesté-Menjoulas, 1971, 1979	Matte, 1969
Intrusive Activity						
Maladeta and other granodiorite bodies Dykes				Phase 4 S_{v4} cleavage		
	Arching Knickzones			Basculement		
Muscovite-granite and pegmatities; Bassiès-Auzat and Andorra granodiorite	E–W refolding	F_2	Fourth phase			
Lys-Caillaouas granite; dykes	Diagonal refolding		Third phase	Phase 3, P_{v3} S_{v3} cleavage		
					Phase H_3	
	Main phase	Main phase (Formation of cleavage fan)	Second phase S_2	Major phase P_{v2} S_{v2} cleavage	Major phase H_2 SH_2 cleavage	S_3 S_{1-2}
	Pre-main phase	Pre-cleavage folding	First phase S_0	Early phase P_{v1} S_{v1} cleavage	Phase H_1 SH_1 cleavage	

done most of the graphic artwork; all coloured maps, except one sheet, have been done by her, besides hundreds of diagrams and figures.

I also wish to thank Mr. M. L. Brittijn, who executed sheet 9, the coloured profile sheet, and many illustrations of this paper, Mr. T. van Hinte for his share in a number of drawings, and Mr. J. Bult for his contributions and much useful advice.

The photographic work has been done very skilfully by Mr. W. C. Laurijssen and to him also I extend my sincere thanks.

Finally I want to mention Mr. C. J. van Leeuwen, our thin section specialist, who produced thousands of thin sections, large and small, of our Pyrenean rocks, and thus made it possible that Pyrenean metamorphics became well known the world over. Also to him I am very grateful.

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ERRATA

Sheet 9, Pallaresa-Flamisell
in legend: colour of Gelada Formation, D_G, ninth lithology symbol from bottom in left column, should be the same as of Rueda Formation, D_R, in second lithology symbol from bottom in right column.

Geological map of the Pyrenees, 1:200 000
on the map: the Carboniferous around the granite massif of Cauterets, and the syncline north of the Lys-Caillaouas massif belongs, at least in part, to the Devonian (Série de Sia)
in the legend: in 19th lithology symbol from top (phyllite with staurolite-andalusite-cordierite porphyroblasts) add small red dots.